Late Cenozoic flexural deformation of the middle
U. S. Atlantic passive margin

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Abstract. Despite the century-long recognition of regional epeirogeny along the middle Atlantic passive margin, relatively few studies have focused on understanding postorogenic uplift mechanisms. Here, we demonstrate that epeirogenic uplift of the central Appalachian Piedmont and subsidence of the Salisbury Embayment represent first-order, flexural isostatic processes driven by continental denudation and offshore deposition. Our results show that regional epeirogenic processes, present on all Atlantic-type passive margins, are best resolved by specific stratigraphic and geomorphic relationships, rather than topography. A simple one-dimensional geodynamic model, constrained by well-dated Baltimore Canyon trough, Coastal Plain, and lower Susquehanna River (Piedmont) stratigraphy, simulates flexural deformation of the U.S. Atlantic margin. The model represents the passive margin lithosphere as a uniformly thick elastic plate, without horizontal compressive stresses, that deforms flexurally under the stress of strike-averaged, vertically applied line loads. Model results illustrate a complex interaction among margin stratigraphy and geomorphology, the isostatic response to denudational and depositional processes, and the modulating influence of exogenic forces such as eustasy. The current elevation, with respect to modern sea level, of fluvial terraces and correlative Coastal Plain deposits or unconformities is successfully predicted through the synthesis of paleotopography, eustatic change, and margin flexure. Results suggest that the middle U.S. Atlantic margin landward of East Coast Magnetic Anomaly is underlain by lithosphere with an average elastic thickness of 40 km (flexural rigidity, $D = 4 \times 10^{23}$ N m), the margin experiences an average, long-term denudation rate of approximately 10 m/My, and the Piedmont has been flexurally warped between 35 and 130 meters in the last 15 m.y. Long-term isostatic continental uplift resulting from denudation and basin subsidence resulting from sediment loading are accommodated primarily by a convex-up flexural hinge, physiographically represented by the Fall Zone. Our results elucidate an inherent danger in using topography alone to constrain late-stage passive margin deformation mechanisms. Only through careful synthesis of field stratigraphic and geomorphic elements such as fluvial terraces, Coastal Plain deposits, and offshore stratigraphy can age control be extended from the offshore depositional setting to the erosional dominated continent. This study demonstrates that despite a relatively subdued topography, the middle U.S. Atlantic margin experiences progressive flexural isostatic deformation similar to that proposed for high-relief margins characterized by great escarpments. Thus margin topographic diversity remains a function of other factors, such as lithospheric composition and/or structure, supracrustal stratigraphy and structure, degree of drainage integration, drainage divide migration, and climate.

Introduction

The geomorphology and stratigraphy of the U.S. Atlantic margin (Figure 1) provide an opportunity to investigate the interaction between regional-scale surficial and crustal processes in a passive margin setting. U.S. Atlantic passive margin studies have traditionally focused on the evolution of sedimentary basins during the synrift and postrift stages [e.g., Kligord et al., 1988] These studies, constrained by stratigraphic [e.g., Scholle, 1977, 1980; Poag, 1985] and geophysical [e.g., Grow et al., 1979; Kligord et al., 1988] data elucidate a complex tectonic and isostatic response attributed to crustal extension [e.g., McKenzie, 1978; Weissel and Kanem, 1989], thermal evolution [e.g., Sawyer et al., 1982], and sediment loading [e.g., Démont et al., 1982].

In contrast, few studies have focused on understanding postorogenic uplift mechanisms on the subaerially exposed continental portions of passive margins despite the century-long recognition of broad-scale, the regional epeirogenic movements [Davis, 1985; King, 1955; Hack, 1982], and the constraints that topography can place on rifting [Summerville, 1991] and epeirogenic mechanisms. This lack of study reflects the paucity of datable post orogenic sediments in the erosional landscape and consequent inability to establish the timing and rates of deformation through geomorphic and stratigraphic methods. As a result, uplift mechanisms arise from genetic models developed to interpret visually striking topographic features of passive margins such as the South African Great Escarpment [King,
Figure 1. Location map and composite cross section of the middle U.S. Atlantic passive margin (modified from Grow et al. [1988]). Susquehanna River terrace profiles, Coastal Plain stratigraphic sections, geodynamic model cross-sections, and offshore load volumes [Peng and Sevon, 1989] are defined along A-A'. Figure 3 is oriented along cross section D-D'. Syn-rift basins are shown with the diagonal-line ruled pattern. DR, Delaware River; JR, James River; PR, Potomac River; RR, Rappahannock River; SKR, Schuylkill River; SR, Susquehanna River; YR, York River; BAL, Baltimore; HAR, Harrisburg; NYC, New York City; PHIL, Philadelphia; WASH, Washington D.C.

1955; Partridge and Maud, 1987), or the "peneplain" summits of Appalachian ridges [Davis, 1889, 1902]. Diverse topographic expressions of opposing Atlantic-type margins of similar age and origin as exhibited by the middle Atlantic United States and South Africa [Ollier, 1985] suggest, but do not necessitate, significantly different post rift tectonic/isostatic evolution.

The middle U.S. Atlantic passive margin, encompassing the Appalachian Piedmont, Coastal Plain (Salisbury Embayment),
PAZZAGLIA AND GARDNER: LATE CENOZOIC FLEXURAL DEFORMATION

and Baltimore Canyon Trough (Figure 1), contains a wide array of geomorphic and stratigraphic data that can be used to investigate late stage passive margin evolution in terms of the nature, rates, and timing of isostatic deformation. This margin is particularly well-suited for flexural isostatic analyses given that the Baltimore Canyon Trough is one of the most thoroughly studied passive margin sedimentary basins [e.g., Grow et al., 1988] with a well-known depositional history and lithostratigraphic and chronostratigraphic reference sections [Poag, 1985], the stratigraphy and age of Coastal Plain marine [Owens and Gohn, 1985, Ward and Strickland, 1985] and fluvial deposits [Owens and Minard, 1979; Pazzaglia, 1993] is well-known, and fluvial terraces of the Susquehanna River have been mapped and dated by longitudinal and petrographic correlation to the Coastal Plain stratigraphy [Pazzaglia and Gardner, 1993].

Given this unique set of geologic constraints and our present knowledge of passive margin isostatic evolution, we hypothesize that (1) post-Oligocene epeirogenic uplift of the Appalachian Piedmont [Hack, 1982] and subsidence of the Salisbury Embayment [Owens, 1970; Owens and Gohn, 1985] are first-order responses to flexural isostatic processes driven by offshore deposition [Cronin, 1981] and continental denudation; and (2) the late stage margin geologic evolution, characterized by flexural isostatic processes, is generally representative of other Atlantic-type passive margins. A corollary of the latter is that diverse topographic expressions of opposing Atlantic-type margins are a function of other factors.

A simple geodynamic model constrained by Piedmont fluvial terraces, Coastal Plain stratigraphy, and known sediment loads in the Baltimore Canyon trough, is constructed to evaluate these hypotheses. Specifically, this paper elucidates the late Cenozoic isostatic and geologic evolution of the middle U.S. Atlantic margin, offers insight into the genesis and significance of the Fall Zone, provides an independent analysis of the middle U.S. Atlantic margin flexural rigidity, and speculates on the geomorphic implications presented by diverse topographic expressions of isostatically similar passive margins.

Geologic Setting

Baltimore Canyon Trough

The largest and deepest depositional basin along the Atlantic margin, the Baltimore Canyon trough, extends for 400 km roughly subparallel to the coast from Long Island to Virginia (Figure 1). The trough spans 60 to 100 km between the hinge zone and the East Coast Magnetic Anomaly (ECMA) and is filled with upward of 18 km of middle Jurassic to Quaternary sediments [Klitgord et al., 1988] (Figure 1). The first 40-60 m.y. of post rift development were dominated by rapid subsidence resulting from crustal thinning and thermotectonic cooling. The remaining 120-100 m.y. have been dominated by regional flexural isostatic subsidence driven by sediment loading [e.g., Steckler et al., 1988].

Early Cenozoic sediments in the Baltimore Canyon trough consist of muddy, calcareous deposits reflecting generally low sedimentation rates, not exceeding 5000 km²/m.y. [Poag and Sevon, 1989]. Deposition was more or less continuous until the Oligocene, when it was interrupted by one or more intervals of subaerial and submarine erosion resulting in a regional unconformity [Poag, 1985]. However, following the Oligocene hiatus, late Cenozoic sedimentation exceeded 30,000 km²/m.y. A thick wedge of middle Miocene siltstone was deposited as pragra-

tional late highstand and lowstand deltas over 1000 m thick across the shelf [Poag, 1985; Greenlee et al., 1988, 1992; Poag and Sevon, 1989]. Subsequent Pliocene and Pleistocene deposition continued as prograding silicilastic deltas 400-800 m thick on the outer shelf and slope [Poag, 1985; Poag and Sevon, 1989].

Coastal Plain

The middle Atlantic Coastal Plain is the subaerially exposed portion of the Salisbury Embayment, a large arcuate-shaped basin approximately 300 km in diameter flanked on the north by the South Jersey Arch, on the south by the Norfolk Arch, and on the west by the Fall Zone [Owens, 1970] (Figure 1). Detailed stratigraphic reconstructions obtained from field mapping and borhole analyses have demonstrated that the Salisbury Embayment has a complex Cenozoic depositional history attributed to isostatic, tectonic, and eustatic processes [e.g. Brown et al., 1972; Newell and Rader, 1982; Newell, 1985; Owens and Gohn, 1985; Ward and Strickland, 1985].

Appalachian Piedmont rocks continue as Salisbury Embayment basement seaward of the Fall Zone, dipping seaward more steeply than anywhere else along the Atlantic Coastal Plain [Hack, 1982]. Cretaceous to Eocene predominantly marine deposits unconformably overlie the basement, filling the basin in a seaward thickening wedge [Owens and Gohn, 1985; Ward and Strickland, 1985]. A basin-wide unconformity of Oligocene age [Ward and Strickland, 1985] represents a sustained period of subaerial erosion which separates the early Cenozoic and Cretaceous deposits from the overlying Chesapeake Group (Figures 2 and 3). Chesapeake Group sediments (Old Church, Calvert, Choptank, St. Marys, Eastover, Yorktown, and Chawan River Formations) are late Oligocene to late Miocene, unconformity-bound, quartz-rich, shelly sands interbedded with micaceous clays, marls, and diatomaceous clays that have well-established macrofauna (molluscan) and microfauna ages [Ward and Strickland, 1985].

Several upper Coastal Plain and Fall Zone fluvial deposits, like those present at the mouth of the Susquehanna River, represent the proximal, updip equivalents to the marine Chesapeake Group facies. Petrographic and lithostratigraphic correlations along the Fall Zone and downdip into the Salisbury Embayment allow for a regional chronostratigraphy of these fluvial deposits [Pazzaglia, 1993] (Figure 2). The petrology of Salisbury Embayment fluvial-deltaic and marine deposits begin as late Oligocene to middle Miocene, mature, quartz sediments and progress toward more heterolithic and more immature submarine facies and subarkoses in the late Miocene to early Pliocene and finally to arkoses, litharenites, and greywackes in the Pliocene and Quaternary (Figure 2). These petrographic trends are mirrored by the major fluvial units at the mouth of the Susquehanna River and are used, with additional, limited biostratigraphic data, to assign a late Oligocene-Miocene age to the Bryn Mawr Formation quartz arenite, a Pliocene age to the informally designated Perryville formation feldspathic sandstone, and a Pleistocene age to the Pennsauken Formation arkose and subarkose (Figure 2).

At least three distinct unconformities, separating three depositional packages or phases are recognized within the Bryn Mawr Formation [Pazzaglia, 1993]. If the unconformities represent major hiatuses, correlative downlap to sequence boundaries, the depositional phases would signify fluvial aggradation in response to three, third-order marine
transgressive events (Figure 2, Salisbury Embayment relative base level). Conceptually, Bryn Mawr Formation phase 1 represents late Oligocene upper braid plain deposition chronostratigraphically correlatable downdip to the Oligocene marine Old Church Formation; Bryn Mawr Formation phase 2 represents a major period of extensive distal braid plain deposition chronostratigraphically equivalent downdip to the middle Miocene, marine Calvert, Choptank, and St. Marys Formations; and Bryn Mawr Formation phase 3 represents upper braid plain deposition inset into phases 1 and 2.

Figure 2. Correlation and genetic relationship between upper Coastal Plain and Fall Zone fluvial deposits, Salisbury Embayment marine deposits, Piedmont fluvial terraces, and eustatic fluctuations [modified from Pazzaglia, 1993].

Figure 3. Cross section B-B' showing four timelines constructed by correlation of Piedmont terrace profiles to Coastal Plain deposits and unconformities. The Bethany and Manokin formations are informal lithostratigraphic units here assigned to the Chesapeake Group.
chronostatigraphically correlative downdpip to the late Miocene, fluvial deltaic Bethany and Manokin formations [Pazzaglia, 1993] (Figure 2).

**Piedmont Terraces**

Fluvial terraces, mapped and correlated on the basis of petrography and elevation, flank the lower Susquehanna River [Pazzaglia and Gardner, 1993] (Figure 3). Upland terraces (Tg1, Tg2, and Tg3) are degraded strath terraces exhibiting cover bed deposits of pebbles and cobbles incorporated within colluvial roundstone diamictics. The petrography of these deposits begins almost exclusively as massive vein quartz for the highest and oldest terrace Tg1 and becomes progressively more heterolithic with the introduction of quartzite clasts in Tg2 and more labile sandstones and siltstones in Tg3 (Figure 2). These petrographic trends mirror those of the upland gravels and Coastal Plain deposits and allow for downstream correlation to dated deposits. Terraces Tg1 and Tg2 are hypothesized to reflect strath genesis during Bryn Mawr Formation phases 1 and 2 and Calvert, Choptank, and St. Marys Formation deposition (Figure 2) [Pazzaglia and Gardner, 1993], whereas terrace Tg3 reflects strath genesis during the Bryn Mawr Formation phase 3 and Eastover, Manokin, and Bethany Formation deposition [Pazzaglia and Gardner, 1993] (Figure 2).

Lower strath terrace QTg exhibits a more heterolithic suite of texturally and compositionally immature cover bed deposits also preserved within roundstone diamictics. QTg reflects strath genesis during the Pliocene to early Pleistocene based on downstream correlation to the petrographically similar Perryville and Pensauken Formations (Figures 2 and 3) and the infrequent occurrence of exotic, glacially-derived clasts.

**Terrace Genesis**

Our model for terrace genesis along the lower Susquehanna River requires a graded river with a fixed base level to attain and maintain a characteristic longitudinal profile on an isostatically dominated margin [Pazzaglia and Gardner, 1993]. In terms of the movement of mass through the fluvial system, a graded river is defined as one that will transport all of the mass moving vertically up through the bed of the stream, as well as that supplied by hillslopes in the drainage basin, while maintaining a relatively fixed position in space [Leopold and Bull, 1979; Knox, 1975]. Given the dynamic feedback between uplift of mass and fluvial erosion on an isostatically dominated passive margin with a crustal root, a river can attain and maintain a graded fluvial profile if base level remains stable at the river mouth over graded time periods [Schumm and Lively, 1965]. As the crustal root is consumed and isostatic uplift diminishes, the profile will flatten, more quickly for uniform erosion and Aiyi isostasy, less quickly for non uniform erosion and spatially variable and/or flexural isostasy.

External modulating factors such as climate and base level fluctuations cause the stream to adjust its profile resulting in the creation of strath terraces. Arguably, the dominant modulator for the lower Susquehanna River, because of its proximity to the coast, is relative base level change. In this system, relative base level reflects a complex interaction between eustasy and passive margin isostasy. Erosionally driven continental isostasy results in progressive, vertical uplift over graded time periods. Over this same time span, eustatic rise and fall occurs at variable amplitudes and frequencies (Figure 2). A prolonged period of relative base level stability is achieved by relatively low-frequency eustatic rise acting in concert with continental isostatic rise (Figure 2). The Susquehanna River is able to attain and maintain a graded profile, cutting straths as it sweeps laterally during these periods. Periods of relative base level fall ensue at the eustatic maxima and subsequent fall which act counter to the steady continental isostatic rise. Strath terraces are generated downstream of an upstream-propagating fluvial knickpoint as the channel incises attempting to adjust to the new, lower base level. Thus strath genesis spans a time range commensurate with a period of Coastal Plain deposition, whereas the strath terrace is created during hiatuses in Coastal Plain deposition.

**Time Lines**

Integration of Baltimore Canyon Trough, Salisbury Embayment and Piedmont stratigraphy defines four time lines to constrain the geodynamic model. These time lines reflect nearly isochronous surfaces used to bracket known offshore loads and an equivalent period of continental denudation. In our model, maximum age of a given strath terrace would be approximately equal to the unconformity at the top of a Coastal Plain unit deposited during the coincident period of strath genesis.

Time line 1 (TL1) lies at the base of the late Oligocene in the Baltimore Canyon trough and Salisbury Embayment (Figure 3). Landward, it extends as the unconformity at the base of the Bryn Mawr Formation, and then projects, with an unknown gradient, over the present Piedmont (Figure 3). The age assigned to this time line is 20 Ma, which is the approximate age of the Oligocene-early Miocene sequence in the Baltimore Canyon trough and Salisbury Embayment. Time line TL2 (15 Ma) is anchored by the middle Miocene sequence in the Baltimore Canyon Trough, deposited primarily as a late highstand phase [Greenlee et al., 1992] during Choptank Formation prodgradational deposition [Kidwell, 1984]. Further inland, it is defined by the unconformity between the Calvert and Choptank Formations in the Coastal Plain and upland terrace Tg1 in the Piedmont. Time line TL3 (8 Ma) is anchored by the late Miocene sequence in the Baltimore Canyon Trough and continues landward as the contact between the Bethany and Manokin formations (Eastover Formation) in the Coastal Plain, and upland terrace Tg3. Time line TL4 (2.5 Ma) is anchored as the Pliocene-Pleistocene sequence in the Baltimore Canyon Trough, the contact between the Beaverdam and Pensauken Formations in the Coastal Plain, and lower terrace QTg.

**Geodynamic Model**

Our simple geodynamic model simulates flexural deformation based on the assumptions that the U.S. Atlantic passive margin is in isostatic equilibrium [Kerner and Watts, 1982], original time line geometry can be reconstructed from geologic and eustatic sea level data (Figures 2 and 4) and the passive margin lithosphere can be simulated as a uniformly thick, perfectly elastic plate, without horizontal stresses, that will respond flexurally to strike-averaged, vertically applied line loads. Initial geometry of a time line must be known because current time line elevation with respect to modern mean sea level (E0x) is the sum of a paleoelevation or topography (E0x), the change in eustatic sea level (ΔSL), and isostatic deformation (k(x)) since the time of its creation (Figure 4):

\[ E_{0x}(x) = E_{0x}(x) + ΔSL + k(x). \]  

(1)
Paleotopography and change in eustatic sea level can be obtained by regional geomorphic and stratigraphic relationships and from published sources. Values for isostatic deformation will be generated by the geodynamic model.

**Paleotopography and Eustasy**

Analyses of the Susquehanna River fluvial profile [Pazzaglia and Gardner, 1993], eustatic sea level curves [Ilaq et al., 1987; Greenlee and Moore, 1988; Dowsett and Cronin, 1990; Schroeder and Greenlee, 1993] (Figure 2) and known Pleistocene Coastal Plain and shelf paleochannel gradients [e.g., Colman et al., 1990] (Figure 4) provide for the reconstruction of time line paleotopography and paleoeustasy. Projection of the upper, graded portion of the Susquehanna River across its lower convexity intersects the Bryn Mawr Formation on the Fall Zone and upper Coastal Plain approximately 60 m above modern sea level (Figures 3 and 4). This suggests that the paleo-Susquehanna River was graded to a relative base level about 60 m higher than present in the Miocene.

Despite the fact that the Piedmont and Fall Zone are hypothesized to have experienced long term isostatic as well as possible tectonic [Hack, 1982] uplift, late Tertiary marine deposits (Chesapeake Group, Figure 2, column A) onlap the upper Coastal Plain and locally extend even onto the Fall Zone [Mixon et al., 1989]. Apparently, in this geologic setting, relative base level changes along the Fall Zone are dominated by eustatic fluctuations rather than isostatic/tectonic deformation. Thus the elevation of fluvial deposits at the mouth of the Susquehanna River, such as the Bryn Mawr Formation, represents a reasonable, first-order approximation of paleo-sea level. The presence of a major Coastal Plain Oligocene unconformity suggests a paleo-sea level equal to, or lower than, modern sea level for TL1. The elevations of the Bryn Mawr (phases 2 and 3), Perryville, and Pensauken Formations suggest paleo-sea levels of approximately +60 m above mean sea level (msl) for TL2, +50 m msl for TL3 and +35 m msl for TL4. These estimates correspond reasonably well with values depicted on eustatic sea level curves.

This paper primarily presents results where steep, incised fluvial profiles are paired with eustatic lowstands and graded, low-gradient fluvial profiles are paired with eustatic highstands. (Figure 4). Modeling results not presented here strongly suggest that other combinations, such as steep fluvial profiles during times of high sea level, are not tenable (F.J. Pazzaglia, unpublished data, 1993). Piedmont and Coastal Plain depositional gradients assume a fixed shoreline near the Fall Zone which is generally supported by facies distributions.

**Isostatic Deformation**

Isostatic deformation will be calculated by a geodynamic model that simulates the lithosphere as an infinite, unbroken, elastic plate of uniform thickness. The approximation of flexure for a thin, unbroken elastic plate [Turcotte and Schubert, 1982] is:

$$\frac{D\partial^4 W}{\partial x^4} + \frac{P\partial^2 W}{\partial z^2} + (\rho_m - \rho_s) g \frac{W}{r} = q$$  \hspace{1cm} (2)

where $D$ is flexural rigidity, $P$ is horizontal stresses, $W$ is vertical stresses, $\rho_m$ is lithosphere density, and $\rho_s$ is sediment density. Assuming the absence of horizontal stresses ($P = 0$) and a uniform-thickness plate ($D = constant$), this differential equation has a well-known one-dimensional analytic solution:

$$W_b(x) = W_o e^{-x/\alpha} \left( \cos \frac{x}{\alpha} + \sin \frac{x}{\alpha} \right)$$  \hspace{1cm} (3)

where $W_o$ is the deflection at the point of loading or unloading, $x$ is the distance along the plate, $W_b(x)$ is deflection of the plate at point $x$ from $W_o$, and $\alpha$ is the flexural parameter defined by plate flexural rigidity. The flexural parameter and flexural rigidity are related by:

$$D = \frac{E h^3}{12(1-\nu^2)}$$  \hspace{1cm} (4)

$$\alpha^2 = \frac{4D}{\rho_m g}$$  \hspace{1cm} (5)

where $E$ is the plate elasticity (parameterized as 70 X 10^9 Pa), $\nu$ is Poisson's ratio (parameterized as 0.25), and $h$ is the elastic thickness.

Plate deflection at the point of loading (or unloading) ($W_o$) is defined as

$$W_o = \frac{q \Delta x \Delta y}{8D}$$  \hspace{1cm} (6)

where

$$q = \rho_s g \Delta x \Delta y$$  \hspace{1cm} (7)

$g$ is acceleration of gravity (9.81 m/s^2), and $\Delta x \Delta y$ is the cross-sectional area of the load.
The model is composed of 17 equally spaced, 50-km-wide cells (Figure 5), parallel to cross section A-A′ (Figure 1). Sediment loads in the Salisbury Embayment and Baltimore Canyon Trough (q(b)) (Table 1), determined from known cross-sectional areas (A_n) obtained from isopach maps [Poag and Nevon, 1989], and sediment densities (ρ_s) [Scholle, 1977] are applied in cells 1 through 12 (Figure 5). Because the analytic solution does not allow lateral variations in flexural rigidity, if sediment for a given time line is not present seaward of the ECMA, loads equal in magnitude to the most seaward positioned load are applied to minimize the effects of a strong oceanic lithosphere. Erosional unloading on the continent (q(c)) (Table 1) determined by the product of rock density (ρ_r), and a correction area defined by Δx of 50 km, and Δy equal to the product of erosion rate (e) and time line age (t_n) is applied in cells 13 through 17. Geochemical mass balance studies of saprolite production rates suggest average Piedmont denudation rates ranging from about 5 to 50 m/Myr. (Cleaves et al., 1970, 1974, Cleaves, 1989, 1993; Pavich, 1985; Pavich et al., 1989; Pavich, 1989).

### Error Analysis

Model results must be viewed within the context of model and data error introduced by the stratigraphic elevation and age range, the relatively short (~20 Myr) deformation time span, and the model parameterization. The total vertical elevation range afforded by Coastal Plain deposits and Piedmont fluvial terraces is not greater than 600 m (Figure 6). Reasonable estimates for

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Table 1. List of Vertically Applied Line Loads (q)

The error in calculating the palaeotopography and eustatic sea level alone (Figure 4) can account for 20% of this elevation range (Figure 6), but careful consideration of geologically reasonable combinations of palaeotopography and eustatic sea level.

![Figure 5. Model parameterization. COST B-2 core porosity is from Scholle [1977]](image)
Figure 6. Estimated sources of error in the geodynamic model and field databases. All of the field data are contained within in the zone between the upper and lower data limits.

should minimize this largest error source. Absolute error in the constraining data set is very difficult to calculate given the diverse topographic and subsurface databases from which they are obtained. Reasonable estimates are about ±10 m for the terrace data and about ±25 m for the subsurface data (Figure 6).

Parameterization error reflects a combination of uncertainty in the calculation of offshore sediment load and density (Figure 6), erosion rate, number of cells, and lithospheric deviation from a pure elastic rheology. This remains the smallest error source and should not significantly affect results.

Model Results

Model simulations are organized into three groups (I through III) that show the effects of different flexural rigidity and erosion rate on flexural deformation (Table 2 and Figure 7). In all 12 simulations, the offshore load remains fixed, and relative comparisons of time line correspondence to Piedmont terrace and Coastal Plain stratigraphic data are determined by the sum of the squares of their residual (ΣR²) (Table 2). We will use the ΣR² as a semiquantitative, yet objective way of evaluating the relative degree of correspondence between model results and field stratigraphic data.

Group I simulations (Figure 7, simulations A through C, and Table 2), employ a constant erosion rate of 10 m/m yr., and exhibit good correlations to the stratigraphic and terrace data including the observed depth of sediments in the COST-B2 core [Prange, 1985]. The best correspondence to observed Coastal Plain stratigraphy, especially for TL3 and TL4, is for a plate of lower flexural rigidity (elastic thickness is 20 km), whereas better correspondence to terrace data is illustrated by plates of higher flexural rigidity (k is 40 or 60 km). All simulations show TL2 consistently projecting above its Coastal Plain datum right down the middle of the Calvert-Choptank Formations, whereas TL1 projects consistently below its Coastal Plain Oligocene un-

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<th>Table 2. Summary of Model Simulations</th>
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a High eustatic sea level equals 0 m asl for TL1, 161 m asl for TL2, 150 m asl for TL3, and 135 m asl for TL4. Low eustatic sea level equals zero for all time lines.

b Regional stratigraphic and geomorphic data require that TL1 and TL4 always maintain a steep fluvial gradient defined by y = ΔSL + 0.008 x.

c This simulation is shown graphically in Figure 7.
Figure 7. Geodynamic model simulations. $1 = \Sigma R^2$ is the sum of the squares of the residual ($\times 10^3$) between model generated timelines (TL1, TL2, etc.), terrace profiles, and correlative Coastal Plain unconformities. COST-B2 location is shown on Figure 1.
conformity datum. Perhaps the most interesting feature exhibited by group I simulations is the narrow zone between model km 550 and 650 where all of the time lines intersect. The intersection zone is narrower and virtually coincident with the Fall Zone for simulations (group III, Table 2) where all time line sea levels are modeled as modern sea level.

Group II simulations (Figure 7, simulations F and H, and Table 2) use an erosion rate ranging from 5 to 15 m/m.y. These simulations exhibit a reasonable correspondence to the Coastal Plain data, but a much poorer correspondence to the terrace data. These results demonstrate that the Piedmont portion of the model is very sensitive to relatively small variations in the erosion rate and suggests a reasonable long-term average denudation rate of about 10 m/m.y.

Group III simulations (Table 2) also employ a constant erosion rate of 10 m/m.y., but incorporate no eustatic fluctuations and the assumption that Piedmont straths are generated during eustatic lowstands and steep fluvial profiles. These profiles generally exhibit a reasonable correspondence to the Coastal Plain data but much a poorer correspondence to the terrace stratigraphy.

**Discussion**

**Model Implications**

Geodynamic model results show that post-Oligocene subsidence of the Baltimore Canyon Trough and Salisbury Embayment and uplift of the Appalachian Piedmont result from flexural isostatic deformation responding to long-term continental denudation of about 10 m/m.y. and offshore sediment loading. The best correspondence between model-generated time lines and field stratigraphic data for a uniform thickness elastic plate (Figure 7, simulation H, and Table 2) suggests an average elastic thickness of 40 km which corresponds to a flexural rigidity (D) of $4 \times 10^{23}$ N m. These results are in line with several geophysically based studies which estimate the elastic thickness of the lithosphere underlying the U.S. passive margin to be between 20 and 60 km.
Pazzaglia and Gardner: Late Cenozoic Flexural Deformation


Model results have broad implications for deposition and erosion on the Fall Zone and Coastal Plain, successfully predicting zones of erosion, deposition, and complex stratigraphic relationships. For example, TL1, TL2, and TL3 generally pass through the Bryn Mawr Formation (Figure 7, simulations A, H, and C) supporting our field stratigraphic interpretation that the Bryn Mawr Formation is a polygenetic unit deposited in at least three phases from the late Oligocene to late Miocene. Additionally, the intersection of TL2, TL3, and TL4 on the Delmarva Peninsula at Smyrna, Delaware (Figure 7, simulations A and H) supports recently described stratigraphic relations where older Pleistocene fluvial deposits (TL4) have removed late Miocene fluvial deposits (TL3) and lie unconformably on middle Miocene marine deposits (TL2).

The geodynamic model also generally supports our model for terrace genesis [Pazzaglia and Gardner, 1993]. Simulations which pair relative high sea level conditions and graded fluvial profiles (Table 2 and Figure 7, simulations A, B, and C), in accordance with our model, generally exhibit a good correspondence to petrographically similar terrace and Coastal Plain deposits suggesting that the strath terraces were cut during periods of graded fluvial profiles and Coastal Plain deposition. In contrast, simulations which remove the eustatic signal from the time lines and employ steep fluvial profiles (Table 2) clearly exhibit a much poorer correspondence to mapped terrace profiles and suggest terrace correlations which violate petrographic data. These results, combined with Pleistocene paleochannel data which documents episodes of Susquehanna River fluvial incision during glacioeustatic lowstands [Colman et al., 1990], cast serious doubt on the appropriateness of pairing graded, low gradient fluvial profiles with low eustatic sea level conditions. Thus suberial exposure and erosion of the Coastal Plain and Fall Zone, and steeper fluvial gradients, are more consistent with eustatic drawdowns.

While the results support flexural isostasy as the primary process for margin uplift and subsidence, discrepancies between model results and stratigraphic data, such as the overall steeper basinward gradient of Coastal Plain deposits and known high-angle faults along the Fall Zone, may indicate second-order tectonic processes. Many faults, some associated with complex graben structures, occur beneath the Coastal Plain and along the Fall Zone [Mixon and Newell, 1977; Mixon and Powers, 1984; Newell, 1985; Benson, 1990]. The U.S. Atlantic margin is currently in a state of northeast-directed compressional horizontal stress [Zoback and Zoback, 1989]. Given the known location of faults, it may be possible to resolve the local stress field imparted from regional compression and flexural bending to estimate fault offset and better quantify tectonic deformation.

The initial model parameterization assumption of no horizontal stresses on the elastic plate ignores the well-known horizontal compressive stress field. Horizontal compressive stresses with magnitudes up to a few kilobars have demonstrated that flexural deflections of the lithosphere of up to 100 m could occur in a passive margin setting [Cloetingh et al., 1985]. Thus the flexural effects modeled herein may be amplified by the existing horizontal compressive stresses leading to an interpretation of a somewhat more rigid lithosphere. The onset of horizontal compressive stress for the U.S. Atlantic margin, estimated from stratigraphic modeling, occurs between 30 and 40 Ma during the Oligocene [Cloetingh and Koel, 1989]. These results closely correspond to the onset of regional uplift and subaerial exposure of the passive margin, as well as (glacio-) eustatic drawdown, followed by the major pulse of middle Miocene deposition in the Baltimore Canyon Trough. However, several reverse faults in the Salisbury Embayment are known to have been active in the Cretaceous [Mixon and Newell, 1977; Mixon and Powers, 1984] suggesting that margin compression may have initiated significantly earlier.

Fall Zone

Fall Zone relief along the Atlantic margin primarily reflects the contrast between metamorphic, crystalline Piedmont rocks and unconsolidated Coastal Plain sediments. Increased relief and dissection of the middle Atlantic Fall Zone between the James and Hudson Rivers has been used to argue for late Cenozoic uplift [Hake, 1982]. We support this conclusion and suggest that the uplift is attributed to the Piedmont flexural upwarping shown in our model simulations.

Long-term uplift of the Piedmont and subsidence of the Salisbury Embayment are accommodated by flexural bending of the lithosphere across a convex-up fixed hinge. Time line intersections, exhibited by all model simulations, implies that the Fall Zone marks the landward portion of this hinge. The relatively narrow zone of time line intersection (Figure 7) suggests that the spatial relations between the offshore load and subaerial erosion has been rather consistent, resulting in a fixed Fall Zone, at least throughout the late Cenozoic. The Baltimore Canyon Trough and Salisbury Embayment, nestled into the Pennsylvania reentrant of the Appalachian Orogen causes the flexural bend to be steeper and more narrow than anywhere else along the Atlantic margin. This, in part, explains why the middle Atlantic Fall Zone has been uplifted, while the upper Coastal Plain, just 10 km basinward, remains relatively unaffected by uplift or subsidence. Superposition of the steep, narrow flexural bend with the contrast in rock hardness amplifies the middle Atlantic Fall Zone physiographic expression. In contrast, the gentler and wider flexural hinge along the southern Atlantic margin, reflecting a more widely distributed offshore load, is not narrowly superposed with the contrast in rock hardness, resulting in a more subdued physiography.

Piedmont Uplift and Erosion Rates

Epeiric uplift and broad warping of the central Appalachian Piedmont has been suggested by studies of uplifted or warped peneplains [Stose, 1928, Campbell, 1933] terrace correlations along the Susquehanna River [Pazzaglia and Gardner, 1993], and stream disequilibrium profiles [e.g., Hack, 1957, 1973, 1982]. Our model, constrained by terrace stratigraphy, supports and quantifies late Cenozoic epeiric uplift and warping of the Piedmont (Figure 8). Using an age of 15 Ma for TL2 and lithospheric thickness of 40 km, total uplift of the Fall Zone, central Piedmont (Holtwood), and Great Valley, 0, 50 and 100 km from the river mouth is 35, 95, and 130 m corresponding to rates of 2.3, 6.3, and 8.7 m/m.y., respectively (Figure 8).

Continuous flexural uplift of the Piedmont primarily depends on the rate of denudation because the principal effect of the offshore sediment load flexes the Piedmont downward (Figure 8). The peripheral bulge from the offshore load only results in ap-
approximately 20 m of uplift (above modern sea level) 100 km inland from the Fall Zone (Figure 8). Denudation rates of 5 and 15 m/m.y., illustrated by group II simulations, generally produced poor correspondence to terrace data. Thus long-term Piedmont denudation rates must lie very closely to the 10 m/m.y. used in group I simulations. Other, independent Piedmont denudation rate estimates range from 45 to 90 m/m.y. using the volume of offshore sediment [Brown, 1989; Poag and Sevon, 1989], 1.5 to 30 m/m.y. from apatite fission track data [Zimmerman, 1979], 5 to 50 m/m.y. from monitored drainage basin dissolved-load studies [Cleaves et al., 1970; 1974; Cleaves, 1989, 1993], and about 4 m/m.y. from saprolite production rates [Pavich, 1985; 1989; Pavich et al., 1989]. Except for the mass balance calculations, most of these rates are consistent with the 10 m/m.y. long-term average rate determined by our model.

Origin of Diverse Topographic Expressions

Many regional geomorphic studies of passive margins allude to the importance of episodic uplift, punctuated by periods of tectonic quiescence to produce the observed geomorphic, stratigraphic and topographic features such as regional erosion surfaces or fluvial terraces [Davis, 1889; King, 1955; Hack, 1982; Partridge and Maud, 1987; Stone, 1991]. A mechanism proposed to explain cyclic epeirogeny suggests that episodic isostatic uplift can be driven by continuous denudation [King, 1955; Schumm, 1963; Partridge and Maud, 1987]. However, other studies have demonstrated that an isostatic response must proceed progressively with denudation [Gitchrist and Summerfield, 1991], where “cyclic” geomorphic features are best explained as resulting from lithologic, structural, or local base level control. This study demonstrates that for the U.S. middle Atlantic passive margin, flexural isostatic deformation is a continuous process of variable rate, profoundly modulated by eustatic sea level fluctuations. We propose that fluvial and marine deposits, particularly in the flexural hinge zone (Fall Zone), are very sensitive to erosion and deposition accompanying relative base level changes. The interaction between progressive flexural uplift or subsidence and eustatic fluctuations, further modified by lithology and structure, can produce geomorphic and stratigraphic features like strath terraces, fluvial disequilibrium profiles and regional Coastal Plain unconformities, without invoking episodic uplift events.

While the continental portion of the U.S. Atlantic margin appears to respond to late Cenozoic flexural isostatic processes previously demonstrated to be important for the Atlantic margin basins [Steckler et al., 1988] and other Atlantic-type margins such as South Africa [Summerfield, 1985; Ollier, 1985; Partridge and Maud, 1987; Gilchrist and Summerfield, 1990; ten Brink and Stern, 1992], the effects of this flexural response are not equally reflected in the respective topographic profiles. This fact underscores the caution that should be used when using topography alone to constrain passive margin isostatic or tectonic processes.

Topographic disparities may, in part, reflect differences in lithospheric composition and/or elastic thickness. For example, the U.S. Atlantic margin has been subdivided into distinct structural provinces [Brown et al., 1972] which in part correspond to well-known onshore lineaments [Gold, 1994] and offshore fracture zones. The structural provinces may represent regions of distinctly different lithosphere in terms of gross composition and thickness which will elicit different flexural responses.

Alternatively, the development of margin drainage, strongly influenced by the interaction of flexural uplift and marginal supracrustal stratigraphy and structure, may also influence the creation of topographic disparities. For example, the southern Appalachian margin arguably experiences flexurally induced epeirogenic uplift similar to the central Appalachians but topographically more closely resembles the South African margin, at least in terms of having a large escarpment represented by the Blue Ridge. Integration of drainages across the southern Blue Ridge or South African Great Escarpment generally has not occurred; thus an escarpment, continually uplifted by flexural isostasy and maintained by a wide drainage divide hung up on resistant lithologic(ies), persists. The integration of large east flowing drainages in the central Appalachians (such as the Potomac and Susquehanna Rivers, through the Ridge and Valley where the Blue Ridge is particularly thin or missing) and the lack of a regional, flat-lying caprock results in rapid migration of the divide to the Allegheny Plateau and the subsequent disintegration of a single, prominent escarpment. This results in a more subdued topography of the middle U.S. passive margin. To a lesser degree, margin topography disparities may also be attributed to long-term climatic and associated weathering processes. Quantitative results evaluating the differences in topographic expression between the southern and middle U.S. Atlantic margin in terms of gross crustal or lithospheric structural differences remains to be examined.

Summary and Conclusions

This paper demonstrates the dynamic feedback between regional-scale denudational and depositional surficial processes and lithospheric responses in a passive margin setting. Synthesis of geomorphic and stratigraphic elements of the middle U.S. Atlantic passive margin can successfully constrain a simple one-dimensional geodynamic model designed to simulate flexural isostatic deformation. Model results are consistent with the hypotheses that late Cenozoic epeirogenic uplift of the Appalachian Piedmont [Hack, 1982] as well as subsidence of the Salisbury Embayment [Owens, 1970] and Baltimore Canyon Trough [Steckler et al., 1988] are first-order responses to flexural
isostatic deformation driven by offshore deposition and continental denudation.

This study represents a major attempt to constrain model input with field-based geomorphic and stratigraphic data. Results indicate that the middle U.S. passive margin, landward of the ECMA is underlain by lithosphere with an average elastic thickness of about 40 km \( (D = 4 \times 10^8 \text{ N m}) \). Long-term continental denudation rates are tightly constrained at about 10 m/m.y. which is consistent with other denudation estimates [Pavich, 1985, 1989; Pavich et al., 1989; Cleaves, 1989, 1993]. Piedmont terrace \( \text{T}_g \), used to constrain a middle Miocene time line, indicate 35 to 130 m of eperogenic, convex-up, flexural uplift from the Fall Zone to the Great Valley over the last 15 m.y. We propose that the Fall Zone represents a long-lived subaerially exposed portion of the flexural hinge where long-term basin subsidence and continental uplift is accommodated.

There exists an inherent danger in using topographic profiles alone to constrain late stage passive margin deformation mechanisms. Only through careful interpretations of field stratigraphic and geomorphic elements such as fluvial terraces, Coastal Plain deposits and offshore stratigraphy can age control be obtained from the depositional setting to the erosionally dominated continent. This study demonstrates that despite a relatively subdued topography, the middle U.S. Atlantic margin is undergoing progressive flexural deformation nearly identical to that proposed for other passive margins. Topographic disparities among passive margins of similar age and genesis are concluded to be a function of other factors, among which are lithospheric composition and thickness, supracrustal stratigraphy and structure, degree of drainage integration, drainage divide migration, and climate.

Acknowledgements. Acknowledgment is made to the Donors of The Petroleum Research Fund, administered by the American Chemical Society, for the support of this research. Additional support to Pazzaglia was furnished through a Shell Fellowship by the Penn State Department of Geosciences, an Earth System Science Fellowship through EOS NASA grant NAS-5-30556, and a Student Research Grant of the Geological Society of America. Sincere thanks are extended to the Maryland Department of Natural Resources, D. Kline and P. Reichard of York Building Products Inc., G. Coufon, F. Kirk, V. Brinton, M. Abrogaat and the Holwood Land Management Office, and all of the residents of Cecil and Harford Counties Maryland, and York and Lancaster Counties, Pennsylvania, for their unselfish cooperation, hospitality, and support in providing access to private land. Discussions and field support with D. Merritts, J. Owens, and K. Ramsey are gratefully acknowledged. E. Cleaves, K. Fairgong, R. Slingerland, and G. Tucket provided helpful comments on early versions of this manuscript. Thorugh and insightful reviews by H. Mills, M. Steckler, and D. Merritts greatly improved the manuscript; however, we accept full responsibility for the interpretations and conclusions advanced herein.

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(Received March 29, 1993; revised September 10, 1993; accepted November 3, 1993.)