Macrogeomorphic evolution of the post-Triassic Appalachian mountains determined by deconvolution of the offshore basin sedimentary record

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ABSTRACT

A perplexing macrogeomorphic problem is the persistence of topography in mountain ranges that were initially formed by orogenic events hundreds of millions of years old. In this paper, we deconvolve the post-Triassic macrogeomorphic history of a portion of one of these ranges, the central and northern Appalachians, using a well-documented offshore isopach sedimentary record of the US Atlantic margin.

Topography is an important signature of tectonic, eustatic and/or geomorphic processes that produces changes in the erodible thickness of the crust (ETC). We define ETC as the total thickness of crust that would have to be consumed by erosion to reduce the mean elevation of a landscape to sea level. We use the term 'source flux', designated by \( \nu \), to describe the rate of change in ETC attributed to deep-seated geological processes such as crustal thickening, crustal extension, magmatic intrusions or dynamic flow in the mantle. In a mountain belt, the rate of change of mean elevation with respect to a base level, designated by \( \varepsilon' \), can be represented as \( \varepsilon' = (\nu - k_d \varepsilon' - \dot{\varepsilon}) - \dot{l} \), where \( k_d \) is a proportionality constant relating the mean mechanical erosion rate to mean elevation, \( \dot{\varepsilon} \) is the mean chemical erosion rate, \( \varepsilon \) is a compensation ratio (held constant for Airy isostasy at 0.21) and \( \dot{l} \) is the rate of eustatic sea-level change. This equation describes the sum of constructive source terms, two destructive erosion terms and a eustatic sea-level term.

We use this simple linear equation to develop a landscape evolution model based on the concept of a unit response function. The unit response function is analogous to a unit hydrograph which describes the relationship between input (rainfall) and output (discharge) in a hydrological system. In our case, we solve for the general relationship between the source flux into the mountain belt and the erosional flux out of the belt. Offshore sediment volumes are used to determine the erosional flux. Drainage basin area is treated as either a constant (pinned drainage divide) or as increasing through time (migrating drainage divide). We use a third-order polynomial fit to a global sea-level curve to account for long-term eustatically driven changes in ETC and in drainage basin area. Chemical erosion is treated as a constant fixed at 5 m Myr\(^{-1}\).

We consider two end-member models. The first is a 'tectonic' model in which the source flux is allowed to vary while \( k_d \) is assumed to be constant over geological time and equal to its mean Pleistocene value of about 0.07 Myr\(^{-1}\). The second is an 'erodibility' model in which \( k_d \) is allowed to vary, reflecting changes in climatic, climatic variables and rock-type erodibility, while the source flux is held constant at zero. The 'tectonic' model reveals four important increases in the source flux, ranging from 200 to 2000 m Myr\(^{-1}\) that occur over short (<10 Myr) time spans, followed by a protracted period (>25 Myr) where \( \nu \) drops below zero to values of -1000 to -6000 m Myr\(^{-1}\). The 'erodibility' model produces a topography that decays in a step-like fashion from an initial mean elevation that ranges between \( \sim 1800 \) and 2300 m.

These models cannot unequivocally distinguish the relative importance of tectonic vs. climatic processes in the macrogeomorphic evolution of the post-rift Appalachians, but they do
provide some first-order quantitative prediction about these two end-member options. In light of existing stratigraphic, geological and thermochronological data, we favour the tectonic model because most of the events correlate well in time and form with known syn- and post-rift magmatic events. Nevertheless, the most recent episode of increased sediment flux to the offshore basins during the Miocene remains difficult to explain by either model. Limited evidence suggests that this event may reflect asthenospheric flow-driven uplift and the development of dynamically supported topography at a time when mechanical erosion rates were increasing in response to a cooling terrestrial climate.

INTRODUCTION

The Appalachian Mountains hold many intriguing macrogeomorphological paradoxes, not the least of which is how mountainous topography is maintained for hundreds of millions of years after orogenic processes cease. Recent studies (Poag, 1985, 1992; Grow et al., 1988; Kiltzgord et al., 1988; Poag & Sevon, 1989), among others, which have documented the structure and stratigraphy of Atlantic margin basins, further complicate this paradox when one considers that these offshore basins contain the detritus of no less than 7 km of rock, presumably eroded from the central and northern Appalachians, in the past 180 Myr. Poag & Sevon (1989), Poag (1992) and Poag & Ward (1993) have used these data to interpret qualitatively the post-orogenic climatic and tectonic history of the central Appalachians. The sediments preserved in the offshore basins provide an alternative approach to traditional geomorphic or stratigraphic reconstructions of paleo-topography that are hampered by the paucity of datable post-orogenic sediments in the Appalachian landscape.

Our approach is far from the first attempt to grapple with paleo-topographic reconstructions of the Appalachian landscape. Several geomorphological and geological paradigms on long-term landscape evolution were conceived (Davis, 1889, 1899), challenged (Hack, 1960; Schumm, 1963; Judson, 1975) and modified (e.g. Gardner & Sevon, 1989) in the heart of the Appalachian Mountains. At the core of these models has been the unresolved roles that tectonic forcing (Hack, 1982), climatic change (Barron, 1989) and rock type (Hack, 1980) have played in shaping the landscape. As several modern macrogeomorphic landscape studies have demonstrated, Davisian and Hackian views represent end-member ideas that cannot capture all of the complexities of long-term landscape evolution (Young & McDougall, 1993; Kooi & Beaumont, 1996). In this regard, Davisian theory holds that relief in a landscape should decay exponentially as interfluves are rounded and lowered, and valley bottoms widen. Hackian dynamic equilibrium holds that a characteristic relief will be maintained as the landscape becomes well adjusted to tectonic uplift, structure, rock type and climate. Differences in these paradigms arise in part from their inferred temporal and spatial scales of observation. Nevertheless, proponents of the cycle of erosion and dynamic equilibrium agree on one critical point: that the erosional response of a landscape following a land-surface uplift event is both complex and significantly longer than the original orogenic event. This paper will capitalize on that point and consider both Davisian and Hackian views on long-term landscape evolution in reconstructing post-rift Appalachian topography.

Our approach begins with an introduction of the geological and geomorphic setting of the central and northern Appalachians and the Atlantic margin basins. We then introduce our model equation and the general concepts behind it. At the core of the model is the idea that an elevation-dependent erosion law can be used to deconvolve a basin sedimentary record. Lastly, we describe how the model can be inverted to solve for changes in post-Triassic tectonic processes or landscape erodibility. The data we use for both inversions come from the volume of siliciclastic detritus shed from the post-rift Appalachians in the past roughly 175 Myr into the offshore basin of the Atlantic Ocean. In this respect, the Appalachians represent the most appropriate landscape to test a deconvolution model because no other orogen has its record of erosion so well preserved, understood and documented as the offshore Atlantic margin basins.

This approach, though simple, should be broadly applicable to any well-drained erosional landscape in which the eroded sediments have been preserved in adjacent clastic sedimentary basins. We recognize that sophisticated numerical models are available for simulation of regional landscape evolution, but the use of such models would not be appropriate in our case because the erosional record contains very limited information about the spatial distribution of erosion rates. Instead, we focus on a one-dimensional description of the relationship between spatially averaged variables, such as mean elevation, mean source flux (defined below) and mean erosion rate. This approach avoids much of the complexity that we would have to address if we were to model geomorphic processes at a local scale. We do not intend for this paper to resolve the relative roles of

* Note: this paper uses the program DECON developed by the authors which is available as both QuickBASIC source code and in QuickBASIC compiled executable form free from the anonymous ftp site: gilbert.geology.yale.edu, /pub/brandon/outgoing/decon.bas –or– decon.exe.
tectonism vs. climate, nor do our model assumptions represent the only plausible approach to deconvolving a sedimentary basin. What we do hope to gain by this exercise is a first-order, quantitative understanding of the processes that have shaped the post-orogenic Appalachian landscape.

GEOLoGICAL AND GEOMORPHIC SETTING

Our study area lies roughly between lat. 36°00' and 46°00' N, long. 69°30' and 80°00' W (Fig. 1) which captures all of the Appalachian drainage that contributes to the offshore basins of the US middle Atlantic margin (cf. Braun, 1989). This area includes the southern New England and central segments of the Appalachian Mountains, and three offshore sedimentary basins (Klitgord et al., 1988): the Salisbury Embayment which underlies the Coastal Plain, the Baltimore Canyon Trough which underlies the continental shelf and the Hatteras Basin which underlies the continental rise (Fig. 1b). The Appalachian Mountains rise west of the Fall Zone, a region with up to 150 m of relief that separates the resistant metamorphic rocks of the Appalachian Piedmont from the soft, unconsolidated sediments of the Coastal Plain (Fig. 1).

Elevation and local relief vary considerably across the different physiographic provinces of the central Appalachian Mountains (Fig. 1). The Piedmont lies about 250 m above sea level. It is underlain by high-grade metamorphic rocks and exhibits an upland surface of low relief (<20 m) punctuated by river gorges where local relief does not exceed 180 m. Similarly, the Ridge and Valley, underlain by folded Palaeozoic sedimentary rocks, rises to only 650 m above sea level and exhibits about 300 m of local relief. At least 9000 m of Palaeozoic foreland basin sedimentary rocks remain beneath the Ridge and Valley and Allegheny Plateau (Fig. 1b), an observation consistent with the low relief of this area and low rates of exhumation. In contrast, the Adirondack, White and Green Mountain ranges of the New England Appalachians, and the Blue Ridge of Virginia are considerably higher and steeper than the Ridge and Valley. High peaks in the Adirondack and White Mountains rise over 1500 m above sea level. Summit of the Blue Ridge reach greater than 1000 m above sea level and loom 800 m above the Piedmont along the Blue Ridge escarpment. Particularly in New England, erosion has exhumed structurally deep parts of the Appalachian orogen exposing resistant mid-crustal Proterozoic and lower Palaeozoic metamorphic and igneous rocks.

The central and northern Appalachian Mountains were built by several Palaeozoic orogenic events culminating with the Permian Alleghenian orogeny. Size and relief of the Permian Appalachians may have been similar to the modern central Andes, which have a mean elevation of about 3500–4500 m (Slingerland & Furlong, 1989). Erosion during the Permian and Early Triassic presumably removed most of the Alleghenian topography with virtually all of the detritus being shed west into and beyond the Appalachian foreland basin. An increase in topography and relief was reintroduced into the Appalachian landscape in the Late Triassic and Early Jurassic associated with continental rifting which ultimately led to the opening of the Atlantic ocean. A reversal of Appalachian drainage from the west to the east, which continues to the present, began with the formation of late Triassic and Jurassic rift basins (Judson, 1975). The modern offshore sedimentary basins formed during and subsequent to the rift and have long served as an effective trap of detritus shed eastward from the post-rift margin.

Conventional wisdom holds that eastern North America has been tectonically quiescent since it entered the drift stage in the Late Jurassic, evolving as the classic Atlantic-type passive margin. Contemporary deformation of the middle Atlantic margin does exist in the form of a broad, flexural warp, centred across the Fall Zone, between the upwarped central Appalachians and subsided Salisbury Embayment (Pazzaglia & Gardner, 1994). Smaller tectonic features, such as high-angle reverse faults, are superposed on this warp (Mixon & Newell, 1977; Mixon & Powars, 1984; Newell, 1985; Prowell, 1988; Gardner, 1989). It is important to point out that these faults are small, with the known maximum total displacement on the order of tens to maybe 100 m since the Late Cretaceous. Similarly, the total amount of flexural up-warping of the Appalachian Piedmont is less than 20 m in the last 15 Myr.

Herein lies a paradox: studies of offshore sedimentary basins have clearly demonstrated several dramatic increases in sediment accumulation rates (Poag, 1985; 1992; Poag & Ward, 1993). The implication is that these events are related to increases in mechanical erosion rates. There remains considerable disagreement as to whether these increases in mechanical erosion rate were caused by tectonically driven uplift or climatically driven changes in rock erodibility.

DATA
Drainages

The onland portion of our study area is drained by 10 major east-flowing river systems, stretching from the James River in Virginia to the Merrimack River in New Hampshire (Braun, 1989; Fig. 1a). The western divide of these drainages currently lies along the Blue Ridge Mountains in Virginia, on the Allegheny Plateau across Maryland and Pennsylvania, and north of the Adirondack massif and White Mountains in New England (Fig. 1a). The location of the present drainage divide reflects westward extension of Atlantic drainages during and following Mesozoic rifting and is not particularly well adjusted to structure or rock type (Judson, 1975). Numerous examples of barbed tributaries, stream capture and obsequent streams all point to the continued westward march of the divide south of the glacial boundary (Judson, 1975). Clearly, the combined area of east-flowing drainage is larger now than when it first formed. Unfortunately, it is impossible to know precisely the initial size and subsequent growth rate of the drainage over the past 200 Myr. Following the physically based models of escarpment retreat on continental margins, it has been proposed that divide migration occurs rapidly after rift flank uplift (Kooi & Beaumont, 1994; Tucker & Slingerland, 1994).

The combined drainage area that contributes sediment directly to the offshore sedimentary basins is about 274,000 km². The 14,000-km² Finger Lakes and Champlain lowland, which are currently drained by the north-flowing St Lawrence River, were probably drained by southeast-flowing rivers prior to Pleistocene glaciation. And a similar-sized portion of the upper Coastal Plain was subaerially exposed during most of the Cenozoic (Fig. 1a). Thus, we estimate that the total drainage area supplying sediment to the offshore sedimentary basins currently has an area of about 300,000 km².

Digital elevation data were used to construct a hypsography of the drainage area (Fig. 2). The current mean elevation of the central Appalachian drainages is ~340 m above modern sea level. A more useful measure would
be the mean elevation relative to long-term sea level which is probably best represented by the long-term position of eustatic sea level. The long-term eustatic record was determined by fitting a third-order polynomial to the eustatic curve of Haq et al. (1987) which shows that long-term sea level currently sits at about \(-72 \text{ m}\) (Fig. 3). From this we conclude that relative to long-term base level, the central Appalachian drainages have a mean elevation of \(\sim 412 \text{ m}\).

The onshore drainage area has changed with time due to headward advance of the drainage divide and to eustasy. The effect of these two factors can be represented by

\[
A(t) = (1 - a_1 l(t)) A_f e^{-k_s t},
\]

where \(t\) is the age relative to present in Ma; \(A_f\) is the present drainage area at \(t = 0\) Ma; \(A(t)\) is the drainage area as a function of age; and \(l(t)\) is eustatic sea-level height as a function of age. The constants \(a_1\) and \(k_s\) represent the fractional changes in drainage basin area per metre rise in eustatic sea level, and per million year headward advance of the divide, respectively. The hypsography of the central Appalachians (Fig. 2) indicates that \(a_1 \sim 0.0016 \text{ m}^{-1}\); however, \(k_s\) is not known. As a result, we consider two cases. The first is a pinned divide with \(k_s = 0\), which would imply that the present dimensions of the drainage area were achieved soon after Late Triassic - Early Jurassic riftings. The second case is a migrating divide with \(k_s = 0.005 \text{ Myr}^{-1}\) which means that the drainage would increase at a rate of \(0.5\% \text{ Myr}^{-1}\). For this case, the drainage area during the Early Jurassic would have been half that of the modern drainage.

### Offshore basins

The middle Atlantic offshore sedimentary basins contain a stratigraphic record extending back into the Early Jurassic (Aalenian \(\sim 178 \text{ Ma}\)). Most of this sediment lies in the 400-km-long, 18-km-deep Baltimore Canyon Trough, the largest and deepest continental basin along the Atlantic margin (Fig. 4). Poag (1985, 1992) and Poag & Sevon (1989) subdivided the offshore basin stratigraphy into 23 informal time-stratigraphic units. They subsequently collapsed these into 12 formal allostratigraphic formations (Poag & Ward, 1993). Table 1 shows Poag's (1992) estimates of siliciclastic sediment volumes for the original 23 time-stratigraphic units. Poag's inventory accounts for all significant sedimentary accumulations including deep-sea sediments of the continental rise. We have recalculated the siliciclastic fluxes using the timescales of Harland et al. (1990) and Cande & Kent (1992) and the depth–porosity curve for the COST-B2 well (Scholle, 1977). These fluxes, plotted against geological age in Fig. 5(a), are given in terms of the solid-rock volume delivered into the basins per million years.

It is appropriate to review the geological history of the offshore basins, especially as it relates to the record of sediment flux shown in Fig. 5(a). During the Jurassic (\(\sim 175\) to \(150 \text{ Ma}\)), the offshore basins filled rapidly with synrift sediments building siliciclastic wedges up to 5 km thick. Alternating episodes of primarily siliciclastic or carbonate deposition occurred from the Late Jurassic to
Late Cretaceous (157–66 Ma). Two distinct pulses of siliciclastic sedimentation occurred in the Barremian (~130 Ma) and Coniacian–Santonian (~86 Ma) when sediment fluxes increased from 3 to 7 times that of background. Both siliciclastic and carbonate deposition decreased considerably in the Late Cretaceous and Paleocene. Paleocene and Eocene (66–34 Ma) strata in the Baltimore Canyon trough consist of muddy, calcareous deposits reflecting generally low sedimentation fluxes, not exceeding 5000 km$^3$ Myr$^{-1}$. Deposition was interrupted in the Oligocene (34–24 Ma) by one or more intervals of subaerial and submarine erosion, resulting in a regional unconformity. The loss of section associated with this erosion introduces a downward bias in the estimate of early Cenozoic sediment fluxes, although we feel that the magnitude of this bias is probably small. Following the Oligocene hiatus, siliciclastic sediment fluxes dramatically increased to 30 000 km$^3$ Myr$^{-1}$. A wedge of middle Miocene (16–11 Ma) siltstone over 1000 m thick was deposited (Greenlee et al., 1988, 1992). Pliocene and Pleistocene (5.4–0 Ma) deposition continued as a younger progradational sequence now some 400–800 m thick on the outer shelf and slope.

### Erosion rates for the drainage basin

The relationship between $s(t)$, the flux of siliciclastics shed into the offshore basins (far right column in Table 1), and $e_m(t)$, the mean rate of mechanical erosion of Appalachian drainages, is described by $e_m(t) = s(t)/A(t)$,
Evolution of the post-Triassic Appalachians

Table 2. Erosion rate data discretized to 5-Myr intervals for model input.

<table>
<thead>
<tr>
<th>Time (Ma)</th>
<th>Erosion rate (m Myr$^{-1}$) from basin with pinned divide</th>
<th>Erosion rate (m Myr$^{-1}$) from basin with migrating divide</th>
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<td>0–5</td>
<td>28.68</td>
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<td>5–10</td>
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<td>10–15</td>
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</table>

Fig. 5. (a) Flux of mechanically eroded rock from the Appalachian drainages. (b) Erosion rate over the past 175 Myr for a drainage with a pinned divide. Error envelope is ±11%. (c) Erosion rate over the past 175 Myr for a drainage with a migrating divide growing at the rate of 0.5% Myr$^{-1}$.

where $A(t)$ is given by Eq. (1). (The dot notation indicates the time derivative, e.g. $s(t) = \frac{ds}{dt} \Delta t$.) It is important to note that $s(t) \approx \Delta s/\Delta t$ where $\Delta s$ is the measured solid-rock volume determined for a time-stratigraphic interval of duration $\Delta t$. The mean rates of mechanical erosion are given in Table 2 and Fig. 5(b,c) for the cases of pinned and migrating divides. All of the plots in Fig. 5 show the same general pattern, characterized by four pulses of erosion with each pulse marked by a rapid rise and more gradual decline.

The above erosion rates only account for mechanical denudation. Braun (1989), among others, recognized that the offshore sediments would underestimate the amount of siliciclastics shed from the drainage basin because they do not record rock removed by dissolution. The rate of chemical weathering in the central Appalachians is difficult to constrain, nor is it clear that modern rates are in any way representative of past rates. Contemporary rates of landscape lowering attributed to chemical weathering range from about 2 to 10 m Myr$^{-1}$ (Pavich et al., 1989; Cleaves, 1993). We adopt a constant chemical erosion rate of $e_c = 5$ m Myr$^{-1}$. Our analysis is not sensitive to this assumption given that mechanical erosion (Fig. 5a) is usually much faster than chemical erosion.

**Estimates of error**

The observed data used for this study have large and relatively obvious sources of systematic error. For instance, our estimates of $A(t)$ given by Eq. (1) are merely educated guesses. And while much of the debate about eustasy has focused on the frequency and amplitude of short-term fluctuations, there also remains some doubts about the absolute magnitude of the long-term component of the eustatic record. Nonetheless, we contend that for our analysis, the more important sources of error are for the random errors because they could introduce spurious events into our final results. Differences between Poag &
Sevon (1989) and Poag (1992) suggest estimates of sediment volume are probably within \(\sim 20\%\) of their true value. Thus, we adopt a relative standard error RSE(\(\Delta t\)) \(\sim 10\%\). The 95\% confidence interval reported in Table 4 of Cande & Kent (1992) provides a rough measure of the uncertainties in \(\Delta t\), the time durations assigned to the stratigraphic intervals in Table 1. Averaging and conversion of their confidence intervals to units of standard error suggest that the RSE(\(\Delta t\)) \(\sim 5\%\). Because we use a smoothed polynomial to represent long-term sea level, the high-frequency component of the eustatic curve represents the main source of random error in our estimate of base level \(l(t)\) and drainage area \(A(t)\). We estimate a standard error SE(\(l(t)\)) \(\sim 30\) m based on the residuals of the Haq et al. eustatic curve relative to our best-fit long-term curve. This variation contributes a relative standard error for drainage area RSE(\(A(t)\)) \(\sim 5\%\) (see \(a_3\) in Eq. 1). Propagation of errors due to RSE(\(\Delta t\)), RSE(\(\Delta t\)) and RSE(\(A(t)\)) indicates that RSE(\(\varepsilon_e(t)\)) \(\sim 12\%\). Stated another way, we feel that random errors alone could account for as much as \(\pm 36\%\) \((\equiv 3\) RSE(\(\varepsilon_e(t)\))\) of the variation between individual estimates of mechanical erosion rates. The average variations in our estimates are at least twice as large as this random component, indicating that the observed data should have a resolvable signal.

**METHODS AND CONCEPTS**

Landscapes are the geomorphic expression of constructive tectonic processes and destructive erosional processes, but these processes are difficult to parameterize in a useful way. England & Molnar (1990) examined surface uplift and rock uplift as possible parameters for representing the forcing of the geomorphic system; however, neither of these parameters isolates the tectonic source from the erosional response. Process-based models with spatial dimensions (e.g. Willgoose et al., 1991; Beaumont et al., 1992; Howard et al., 1993; Slingerland et al., 1994; Tuch & Slingerland, 1994) are commonly used to represent geomorphic and rock uplift processes. For our problem, such an analysis seems unwarranted given that our input data are mainly limited to bulk sediment yield. A simpler approach, first outlined by Hay et al. (1989), would be to reconstruct the landscape incrementally by restoring denuded sediments back to their drainages. At the core of this simple approach are three important assumptions: (1) that any given column of crustal material contains some portion that can be removed by erosion, a portion we call the erodible thickness of the crust or ETC; (2) that external geological or plate tectonic forces, which we collectively call source \(v\), can change ETC; and (3) that the mean rate of mechanical erosion of a crustal column is proportional to its mean elevation above base level (Stephenson, 1984; Pitner & Souriau, 1988; Hay et al., 1989; Harrison, 1994).

The isopach data provide a spatially averaged record of the erosional and geomorphic evolution of the central Appalachian drainages. Thus, we focused our analysis on resolving how the spatially averaged state of the geomorphic system evolved with time. The mean elevation \(z\) of the drainage area at any particular time \(t\) is described by

\[
z(t) - l(t) = \varepsilon \text{ETC}(t),
\]

where \(l(t)\) accounts for the mean height of the topography above contemporaneous base level, \(\varepsilon\) is the compensation ratio, and \(\text{ETC}(t)\) is the erodible thickness of the crust. For simplicity, we assume local compensation, so that \(\varepsilon = (p_s - p_\infty)/p_s = 0.21\) where \(p_s = 3400\) kg m\(^{-3}\) for the asthenosphere and \(p_\infty = 2700\) kg m\(^{-3}\) for the crust. Note that time is related to age by \(t = \tau - \tau_0\) where \(\tau_0\) indicates a start-up age.

**Erodible thickness of the crust (ETC)**

The variable \(\text{ETC}\) provides a useful description of the state of the tectono-geomorphic system. It indicates the average thickness of crust that would have to be removed to reduce the average height of the topography to zero relative to contemporaneous base level (i.e. long-term sea level). \(\text{ETC}\) is not limited to only that portion of the crust which lies above base level, but also to that portion that would rise above base level as an isostatic response to erosion. For example, if a landscape has a mean elevation of 1 km above sea level, and that topography is locally compensated (Airy isostasy, \(\varepsilon = 0.21\)), we can estimate how much crustal column would ultimately be consumed as the mean elevation was reduced to base level. Proceeding incrementally, if we remove 1 km of material, the column will rebound roughly 80\% or 800 m. Remove that 800 m, and the column rebounds 640 m. Thirty iterations later, you have reduced topography to \(\sim 0\) m of base level and have consumed 4.8 km of crust by erosion.

An important control on \(\text{ETC}\) is base level, or in other words, the level to which geomorphic forces will attempt to erode the landscape. Equation (2) states that \(\text{ETC}\) is a function of sea level. For instance, a 10-m fall in sea level will increase \(\text{ETC}\) by 48 m. An eustatic rise on the other hand reduces \(\text{ETC}\).

**Sources and sinks**

\(\text{ETC}\) will change as a function of tectonic forcing, erosion and eustatic sea level. Its rate of change is described by

\[
\dot{\text{ETC}}(t) = \dot{v}(t) - \varepsilon_\text{e}(t) - \varepsilon_l(t) - l(t)/\varepsilon,
\]

where \(\dot{v}\) indicates the rate of tectonic forcing which we call the source flux, \(\dot{\varepsilon}_\text{e}\) and \(\varepsilon_l\) are rates of mechanical and chemical erosion, respectively, and \(l\) is the rate of eustatic change. Integration of the source flux with time gives the variable \(v\), which we call the source. The variables \(v\) and \(\dot{v}\) represent useful generic descriptors of the role that tectonic forcing has on the tectono-geomorphic system. It is important to recognize that all of the variables in
Eq. (3) can, at least in theory, take on either positive or negative values. For instance, \( v > 0 \) would indicate tectonically driven uplift whereas \( v < 0 \) would indicate tectonically driven subsidence. Eustatic sea level can rise \( (l > 0) \) and fall \( (l < 0) \). Our analysis here is restricted to the case of positive erosion where total erosion \( \varepsilon = \varepsilon_m + \varepsilon_c > 0 \), but negative erosion is perfectly acceptable description of deposition \( \varepsilon_m < 0 \) or chemical precipitation \( \varepsilon_c < 0 \).

The source flux to a mountain belt can be visualized as the input to the system from deep-seated tectonic or epeirogenic processes. Conversely, the erosion flux from a mountain belt can be viewed as the system output which, in this case, is driven by geomorphic processes. ETC and topography tell us about the balance between those two fluxes. So far, we have focused on source flux, but it is actually more useful here to look at source history for a simpler view of the system input. The generalized examples in Fig. 6 (cf. Allmendinger, 1986) illustrate how various tectonic and epeirogenic processes are distinguished by their source histories. Note that the time derivative of the source history plots would produce the source flux. Each scenario in Fig. 6 will produce its own characteristic erosional response (Schumm & Rea, 1995) which we see recorded as variable rates of sediment accumulation in sedimentary basins. The first example (Fig. 6a) shows horizontal contraction and crustal thickening as might be observed in a typical continental collision zone. In this case, the increase in source with time is directly related to the increase in crustal thickness.

The second example (Fig. 6b) shows continental rifting. In this case, the source history is governed by two processes, thermal buoyancy responding to the rise of hot asthenospheric mantle, and thinning of the lithosphere by horizontal stretching. During the early stages of rifting, the combination of thermal buoyancy and lithospheric thinning might result in an increase in source but, with time, the loss of thermal buoyancy would result in source ultimately dropping below zero, reflecting the permanent result, a thinned lithosphere. The transition from positive to negative source is governed by the time constant for the thermal relaxation of the rift zone.

The third example (Fig. 6c) shows emplacement of mantle-derived intrusives. In this case, both heat and mass are added to the crust, resulting in an increase to an initial high and positive source. Once again, thermal relaxation would lead to a loss of thermal buoyancy and a partial decrease in source. But note that in this case, the source always remains positive because of the increased thickness of the crust following emplacement of the intrusion.

The last example (Fig. 6d) illustrates dynamic topography of Gurnis (1992) in which the topography is dynamically supported by forces associated with the asthenospheric flow. The characteristic of this process is that source is completely recovered with the removal of the dynamic support. As a consequence, the source eventually returns to zero.

### Elevation-dependent mechanical erosion

We adopt a relatively simple erosion law for our deconvolution approach in which the mean rate of mechanical erosion is proportional to the mean elevation above a base level:

\[
\varepsilon_m = k_x(z - l).
\]  

(4)

The constant \( k_x \) can be viewed as a crude measure of the erodibility of landscape at a regional scale. Similar versions of Eq. (4) have been used in other studies (e.g. Stephenson, 1984; Pinet & Souriau, 1988; Hay et al., 1989; Harrison, 1994; Summerfield & Hulton, 1999). A practical advantage of Eq. (4) is that \( \varepsilon_m \) is cast as a function of \( z \) which provides an easy solution for Eq. (2) as is shown below. There is some legitimate skepticism concerning the general use of Eq. (4) as a predictor of mechanical erosion rates. Discussion persists on the relative importance of climate (e.g. Langbein & Schumm, 1958; Ohmori, 1983) or relief (Ruxton & McDougall, 1967; Ahnert, 1970; Pinet & Souriau, 1988) on erodibility. At the continental or subcontinental scale, however, modern sediment yield data are unambiguous; relief plays the dominant role in the rate of landscape ero-
sion (Summerfield, 1991; Milliman & Syvitsky, 1992; Summerfield & Hilton, 1994). Harrison (1994) proposes an exponential relationship between mean elevation and mean mechanical erosion rate at a continental scale, but his relationship only slightly deviates from a linear relationship (see Fig. 2 in Harrison, 1994).

There are two obvious restrictions that must be observed for any reasonable application of Eq. (4). First, an elevation-dependent erosion law seems justified only for well-drained landscapes where fluvial transport is not the rate-limiting step and where ultimate base level is well defined. Thus, we would not advocate the use of Eq. (4) for hyperarid regions, internally drained basins or low-relief plateaus (e.g. Tibetan plateau, Altiplano, Basin and Range) which today make up about one-third of the modern surface of the continents (p. 134 in Snead, 1980).

Another restriction is that the elevation-dependent erosion law provides only a continuum-scale description of the erosion process. The term 'continuum-scale' indicates that the minimum scales in length and time are large enough so that erosion rates vary smoothly and continuously in both space and time. This limitation means that our observations must be at length scales greater than about 10–20 km in order to average out natural variations in erosion rates associated with specific rock types and with specific discrete geomorphic processes, such as mass-wasting, hillslope creep and fluvial sediment transport. Likewise, the time-scale must be of sufficient duration, probably at least several thousand years, to average out variations associated with infrequent maximum-intensity storms and with the delayed or complex response of a river to changes in mean discharge and sediment yield (e.g. graded time of Schumm & Lichty, 1965).

These restrictions may be difficult to meet. For instance, it may not be appropriate to extrapolate short-term erosion rates determined from river sediment yield data to long-term rates of denudation (Gardner et al., 1987) which we need to calibrate the elevation-dependent erosion law. But if the errors in either short-term or long-term rates of erosion are random and uncorrelated, the sampling problem can be improved by restricting our investigations to very large basins or by using erosion records averaged over a suitably long time interval.

We feel that Eq. (4) provides a reasonable first-order representation of the long-term spatially averaged rate of mechanical erosion at a regional scale to long as the landscape has a well-integrated drainage. In our study, \( k_d \) is spatially averaged over the entire ~300,000-km² area of the drainages and time-averaged over the ~5 Myr increments used to represent the mechanical erosion record. We recognize, however, that there may be slow changes in \( k_d \) over the time-scale of several tens of million years. In fact, long-term changes in climate and/or exposed rock type might be manifested in long-term changes in regional-scale erodibility. We return to these issues below.

**Model equation**

The combination of Eqs (2) and (4) gives the rate equation we use for our analysis:

\[ z'(t) = c \nu(t) - k_d z(t) - \xi(t) - \hat{l}(t), \]

where the elevation above modern base level, \( z \), is replaced by \( z' = (z - l) \), the elevation above contemporaneous base level. It is useful to represent the combined effects of source flux, chemical erosion rate and rate of sea-level change by the variable \( \hat{b}(t) \), where

\[ \hat{b}(t) = \nu(t) - \xi(t) - \hat{l}(t)/c. \]

Equation (5) becomes

\[ z'(t) = c \nu(t) - k_d z'(t). \]

If \( k_d \) and \( b \) are constant, then the integrated solution for Eq. (7) is

\[ z'(t) = z_0 e^{-k_d t} + \frac{b}{k_d} (1 - e^{-k_d t}), \]

where \( t \) indicates elapsed time and \( z_0 \) is the mean elevation above base level at \( t = 0 \).

Equation (8) contains two competing terms. The term to the left of the plus sign describes the decay of initial topography, whereas the term to the right of the plus sign describes the construction of new topography. Given enough time, the left term will go to zero and the right term will reach a steady-state elevation \( z_\text{ss} = b/k_d \). Characteristic time, defined by \( t_c = (ck_d)^{-1} \), is commonly used to describe the response time of a first-order dynamic system. For our application, \( t_c \) indicates the amount of time needed to accomplish 63% reduction of an initial landscape or 63% construction of a new steady-state landscape. The system will have a shorter response time if either \( k_d \) or \( c \) is increased, which corresponds, respectively, to an increase in erodibility or a decreased ability to store ETC by isostatic compensation.

**Support for an elevation-dependent erosion model**

Ruston & McDougall’s (1967) study of erosional decay of the Hydrographer’s Volcano, a dormant volcano in Papua New Guinea, provides a good example of elevation-dependent erosion. Mean erosion was determined by reconstructing the original shape of the volcano and then calculating the average amount of erosion that had occurred beneath each reconstructed elevation contour. The duration of erosion is suitable long (~650 kyr) and the eroded lithology was relatively uniform. For erosional decay alone, Eq. (8) reduces to

\[ z(t) = z_0 e^{-\lambda t}. \]

Thus, if the elevation-dependent erosion relation held and \( k_d \) remained constant, we would expect to find a constant ratio between \( z(t) \) and \( z_0 \). This prediction corresponds nicely with the data shown in Fig. 7. We
Evolution of the post-Triassic Appalachians

Fig. 7. Plot of the mean elevation profile of the Hydrographer's Volcano, NE Papua following eruption at 0.65 Ma and in its present eroded form (from Ruxton & McDougall, 1967). The straight line fit demonstrates a linear dependence between mean elevation and mean rate of mechanical erosion.

... can use the slope of the best-fit line to estimate the product \( ck_e \sim 0.7 \text{ Myr}^{-1} \), which corresponds to a characteristic response time of 1.4 Myr. Given the relatively small dimensions of the volcano (~30 km across), it seems likely that the topographic load is not locally compensated so that \( c \) is greater than 0.21 and may approach its maximum value of 1, indicating no compensation. Thus, we conclude that for this area, \( k_e \) lies between 0.7 and 3.3 Myr\(^{-1}\), and probably tends towards the low end of this range.

Pinet & Souriau (1988) examined the relationship between mean elevation and mean mechanical erosion rate for some of the largest river drainages in the world. We have revised their results (Fig. 8) in light of new estimates of natural sediment yield by Milliman & Syvitsky (1992) and a density of 2700 kg m\(^{-3}\), as expected for average crustal rocks. Figure 8 shows that there is no unique value for \( k_e \), assuming of course that Eq. (4) provides an adequate description of long-term regional-scale mechanical erosion rate. We suggest that the variability of \( k_e \) shown in Fig. 8 reflects natural variability in the erodibility of different river drainages. Thus, the application of Eq. (4) requires that \( k_e \) be estimated by local mechanical erosion data. Note that the global average \( k_e = 0.069 \text{ Myr}^{-1} \) would indicate \( c = 69 \text{ Myr} \), which is at least an order of magnitude slower than determined for the Hydrographer's Volcano.

\( \varepsilon_w \) and \( k_e \) for the Appalachian drainage basins

We use three methods to estimate the recent mean rate of mechanical erosion for the Appalachian drainage basins (Table 3). The first is based on historic, fluvial sediment yield data. This approach indicates an \( \varepsilon_w \) of about 12 Myr\(^{-1}\), but it is important to remember that these data reflect a short period (~50 years) of time. The second method utilizes the offshore sediment flux over the past 1.9 Myr (Quaternary flux, Table 1) which indicates an \( \varepsilon_w \) of about 24 m Myr\(^{-1}\). This estimate is probably more reliable, given the longer duration of record.

The third method is based on the relationship between local relief and mechanical erosion rate documented by Ahnert (1970). Ahnert found that regional-scale average mechanical erosion rates in mid-latitude drainages, like those of the central Appalachians, are highly correlated with the average roughness or local relief of the topography. In Ahnert's study, local relief was defined as the maximum difference in elevation within a fixed sample area. We have repeated and expanded Ahnert's (1970) study using 30-s digital elevation data and improved sediment yield estimates from Milliman & Syvitsky (1992). The sample area was standardized to \( R_{10 km} \) defined as the average local relief determined by measuring the maximum vertical range within a 10-km-diameter circle sampled at many points over the area of the drainage. Our results show that Ahnert's law works well for large drainages with areas in excess of 10,000 km\(^2\) (Fig. 9; \( R^2 = 0.90 \)). Some have pointed out that the trend in Fig. 9 is strongly influenced by a small number of high-relief drainages from the Swiss Alps; however, we have found that the best-fit relationship is virtually the same when the Swiss drainages are omitted. The central Appalachian drainages have an average \( R_{10 km} \) of 220 m. The relationship in Fig. 9 would predict a mechanical erosion rate of 44 m Myr\(^{-1}\) (Table 3).

Using these three estimates of mechanical erosion
Table 3. Mechanical erosion rates and $k_d$ for the central Appalachian drainages.

<table>
<thead>
<tr>
<th>$k_d$ estimate method</th>
<th>Mean elevation* (m)</th>
<th>Area (km$^2$)</th>
<th>Load ($10^9$ kg yr$^{-1}$)</th>
<th>$\xi_n$ (m Myr$^{-1}$)</th>
<th>$k_d$ (Myr$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) Estimate based on sediment yield</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rappahannock River</td>
<td>324</td>
<td>112,678</td>
<td>3,821,000</td>
<td>12.54</td>
<td>0.338</td>
</tr>
<tr>
<td>Delaware River</td>
<td>442</td>
<td>300,000</td>
<td>44.00</td>
<td>0.107</td>
<td></td>
</tr>
<tr>
<td>Susquehanna River</td>
<td>292</td>
<td>17,000</td>
<td>1,680,000</td>
<td>14.81</td>
<td>0.034</td>
</tr>
<tr>
<td>Potomac River</td>
<td>292</td>
<td>25,000</td>
<td>720,000</td>
<td>10.67</td>
<td>0.037</td>
</tr>
<tr>
<td>Juniata River</td>
<td>292</td>
<td>62,000</td>
<td>1,800,000</td>
<td>10.75</td>
<td>0.037</td>
</tr>
<tr>
<td>Weighted Mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>406</td>
<td>332,000</td>
<td>17.54</td>
<td>0.329</td>
<td></td>
</tr>
<tr>
<td>(2) Estimate based on Quaternary sediment flux†</td>
<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>(3) Relief-based estimate using Ahnert's Law</td>
<td></td>
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<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean $k_d$ from all three methods</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.70</td>
</tr>
</tbody>
</table>

* Mean elevations for the Delaware, Juniata and Potomac Rivers were directly calculated from 3-arcsec digital elevation data. Other drainages use a mean elevation of the Ridge and Valley and Piedmont Appalachians which is 220 m with respect to modern sea level. All mean elevations in the table are reported relative to present long-term sea level (-72 m). † This estimate is based on Quaternary offshore sediment flux and takes into account eustatically driven changes in drainage area.

Fig. 9. Plot of mean mechanical erosion rates vs. mean relief for mid-latitude basins (> 10,000 km$^2$; $r^2 = 0.90$). All relief measurements were scaled to a 10-km diameter circle using the Brownian fractal relationship: $R(d_i) = (d_i/d_s)R(d_s)^{1/2}$, where $d_s$ and $d_i$ indicate the diameters of different sampling circles, and $R(d_s)$ and $R(d_i)$ are the expected change in local relief sampled within the two circles. Data sources: A–C. relief was determined using 3-arc second digital elevation data and denudation data are from Milliman & Syvitski (1992). D–N. relief was determined using 3-arc second digital elevation data and denudation data from Ahnert (1970), O–W relief and denudation data are from Ahnert (1970), with relief scaled to a 10-km diameter circle. X = modern relief and Pleistocene sediment yield of the central Appalachian drainage.

---

rates, ranging from 12 to 44 m Myr$^{-1}$, the recent $k_d$ for the central Appalachian drainages is estimated to be in the range 0.038–0.11 Myr$^{-1}$. Assuming local compensation ($\epsilon = 0.21$), these $k_d$ values indicate a characteristic time of 44–124 Myr. Our analysis below focuses mainly on an average value of $k_d = 0.07$ Myr$^{-1}$ (Table 3) but it includes some examples at the extremes of the estimated range for $k_d$.

**INVERSION FOR VARIATIONS IN SOURCE FLUX OR ERODIBILITY**

For our problem, we are interested in resolving a record of tectonic forcing or changing erodibility. Equation (8) is not appropriate in this regard because it assumes that $b$ and $k_d$ are constant with time. A useful property of first-order linear differential equations, such as Eq. (7),

is that their general solution can be specified by a series of superimposed solutions. We are interested in a finite number of incremental solutions to match our incremental record of changing erosion rate, which is discretized in 5-Myr intervals.

The basic premise is that the tectonomorphic system acts as a filter that relates the system input, represented by changes in source flux or erodibility, to the system output, manifested by changes in erosion rate. In more formal terms, we state that the system output is determined by convolving the filter with the system input (p. 59 in Brigham, 1988; pp. 187–190 in Menke, 1989). The inverse operation is called deconvolution, where the system input is estimated by convolving an inverse of the filter with the system output. In practical terms, we want to deconvolve the erosion-rate record in order to estimate the record of changing source flux or changing erodibility. In our analysis, we consider two end-member cases: (1) a tectonic model in which \( v \) is allowed to change with time and \( k_s \) is assumed to be constant, and (2) an erodibility model in which \( k_s \) is allowed to change and \( v \) is assumed to be zero.

The tectonic model

We need to specify a unit response function to deconvolve a variable tectonic signal. Our unit response function is a ‘geomorphic’ unit response defined as the erosion-rate record that would result from a tectonic event of duration \( \Delta t \) where \( z' = 0 \) at \( t = 0 \), \( h = 1 \) for \( t > 0 \) and \( \Delta t < t \), and \( h = 0 \) for \( t > \Delta t \). The geomorphic unit response function is analogous to a unit hydrograph in that it describes the response of the system to a unit input. To derive the geomorphic unit response function, we use Eq. (4) (where \( z' = (z - f) \)) to recast Eq. (8) into the following form:

\[
\hat{\epsilon}_{w0}(t) = \hat{\epsilon}_{w0} e^{-\alpha t} + \hat{b}(1 - e^{-\alpha t}),
\]

(10)

where \( \hat{\epsilon}_{w0} \) is the erosion rate at \( t = 0 \). From Eq. (10), we can construct the unit response function

\[
\Gamma(t) = \begin{cases} 1 - e^{-\alpha t} & \text{for } 0 < t < \Delta t, \\ (1 - e^{-\alpha \Delta t}) e^{-\alpha (t - \Delta t)} & \text{for } t > \Delta t. \end{cases}
\]

(11)

\( \Gamma(t) \) describes the response of the system, in terms of the erosion rate record, caused by a unit flux \( \hat{b} \) sustained for a time step from \( t = 0 \) to \( \Delta t \). The plots in Fig. 10 show how the erosion rate and elevation grow and decay in time for a unit response function with \( k_s = 0.07 \, \text{Myr}^{-1} \) and \( \alpha = 0.21 \).

The unit response function is formulated in a discrete form in order to match the discrete nature of the observed erosion-rate record, which is represented by an average rate for each time step. Thus, we introduce a set of discrete variables marked by indices \( i = 1, 2, 3, \ldots \) which indicate the mid-point of each time step, where \( t = (i - 1/2) \Delta t \). The unit response function is discretized by calculating average values of \( \Gamma(t) \), from Eq. (11), for each time step. For \( i \leq 1 \),

\[
\Gamma_{i \leq 0} = 0, \quad \text{and} \quad \Gamma = \frac{1}{\Delta t} \int_0^{\Delta t} (1 - e^{-\alpha t}) \, dt = 1 - e^{-\alpha \Delta t}.
\]

(12a)

where

\[
\Gamma_0 = \frac{1}{\alpha k_s \Delta t} (1 - e^{-\alpha \Delta t}).
\]

(12b)
We have used the standard errors for the observed erosion-rate data to estimate standard errors for our estimates of $h_i$ (see p. 58 in Menke, 1989, for details). The estimates of $h_i$ are determined by finding the matrix $M_{i\tau}$ for $i = 1, 2, 3, \ldots, n$ referring to the midpoints of time steps of geological age, and the same manner as $k$. Equations (12a), (12b) and (12d) can be combined and expanded into a series of linear equations that relate $h_i$ to $v_{ni}$:

\[ e_{n,i} = b_i \Gamma_i \]

\[ e_{n,i} = b_{1i} \Gamma_1 + b_{2i} \Gamma_2 + \ldots + b_{ni} \Gamma_n \]

This system of equations can be recast in matrix form using Einstein's summation notation:

\[ v_{ni} = G_{i\alpha} b_{\alpha} \]

where $G_{i\alpha}$ defines an $n \times n$ response matrix whose elements are given by $\Gamma_{i\alpha} = \tau_{\alpha + 1} - \tau_{\alpha}$. Inversion of Eq. (14) provides a solution for $b_i$:

\[ b_i = G_{i\alpha}^{-1} e_{n,i} \]

The standard errors for $e_{n,i}$ are transformed to $h_i$ by

\[ \text{SE}(h_i) = \text{SE}(v_{ni})^{-1} \]

The estimated standard errors for $h_i$ can then be calculated from the diagonal elements of $\text{SE}(h_i)$:

\[ \text{SE}(h_i) = \text{SE}(h_i)^{-1} \]

At this point, $v_i$ can be calculated from $h_i$ using Eq. (6) and known values of $v$ and $h$. The standard error for $v_i$, $\text{SE}(v_i)$, is assumed to be equal to $\text{SE}(h_i)$. The reason is that $\text{SE}(v_i)$ and $\text{SE}(h_i)$ are relatively small such that the $\text{SE}(v_i)$ is dominated by $\text{SE}(h_i)$ as defined by Eqs (16)–(18).

**The erodibility model**

Now we want to formulate a model that attributes all changes in mechanical erosion rate to changes in $k_{\tau}$, representing regional-scale changes in climate and/or rock type. This problem can be solved incrementally by starting at the present, $i = 0$, and working backwards through time. The following algorithm illustrates the calculation of $k_{\tau}$ and $\text{SE}(k_{\tau})$ for each time step, $i$, where $n$ is the total number of steps. The index notation is the similar to that used above: $i$ indicates the middle of the $i$th time step, and $i + 1/2$ and $i - 1/2$ indicate the top and bottom of that time step. The age is given by

\[ \tau = \tau_0 - i \Delta \tau \]

1. $i$ is set to $n$.  
2. $\zeta_{i+1/2}$ is set to the mean elevation above contemporaneous base level at $\tau = 0 \text{ Ma}$.  
3. $\text{SE}(\zeta_{i+1/2})$ is set to the standard error for $\zeta'$ at $\tau = 0 \text{ Ma}$, which is assumed to be zero.  
4. In a loop, do the following while $i > 0$:  
5. Calculate $\hat{h}_i$; remember $v$ in the erodibility model is equal to 0.  
6. Use a numerical method to solve the following equation for the unknown $\hat{e}_{n,i}$ given known values for $\hat{e}_{n,i}$, $\hat{h}_i$, $\zeta_{i+1/2}$, $\zeta'$ and $\Delta \tau$:

\[ \hat{e}_{n,i} = \zeta_{i+1/2} \hat{h}_i e^{i(\Delta \tau/2)} - \hat{h}_i (1 - e^{i(\Delta \tau/2)}). \]

7. Calculate the change in elevation that occurred during
the time step:
\[ z_{i+1/2} = z_{i+1/2} e^{\frac{\Delta t}{k_i}} - \frac{b_i}{k_i} (1 - e^{\frac{\Delta t}{k_i}}). \]  
(21)

8 Calculate standard errors for the estimated \( k_i \) and \( z_{i+1/2} \), using the conventional procedure for the propagation of errors:
\[ \text{SE}(k_i)^2 = \left( \frac{\partial k_i}{\partial z_{i+1/2}} \text{SE}(z_{i+1/2}) \right)^2 \]
\[ + \left( \frac{\partial k_i}{\partial \epsilon_m} \text{SE}(\epsilon_m) \right)^2 \]  
(22)

\[ \text{SE}(z_{i-1/2})^2 = \left( \frac{\partial z_{i-1/2}}{\partial k_i} \text{SE}(k_i) \right)^2 \]
\[ + \left( \frac{\partial z_{i-1/2}}{\partial \epsilon_m} \text{SE}(\epsilon_m) \right)^2. \]  
(23)

9 Decrement counter to indicate moving backward in time by one time step, i.e. \( i = i - 1 \).

10 Return to step (4).

The calculation starts with \( z_{i-1/2} \) elevation above base level at the top of the time step. Given \( \epsilon_m \), Eq. (20) is solved numerically for a constant \( k_i \) operating over the duration of the time step, designated as \( k_i \). This information is used to solve for \( z_{i-1/2} \), which is the elevation above base level at the bottom of the current time step. Equation (20) is derived from Eq. (10) by setting \( \epsilon_m = z_{i+1/2} k_i \) and \( t = -\Delta t/2 \) (the negative sign indicates that the solution is moving backwards in time). Equation (21) is derived from Eq. (8) by setting \( t = -\Delta t \). SE(\( k_i \)) and SE(\( z_{i-1/2} \)) are determined by the propagation of errors, assuming, once again, that the standard errors for these estimated values are dominated by the standard errors for the observed values of \( \epsilon_m \).

Step (9) decrements \( i \) to the next older time step in the sequence, and step (10) returns the calculation to the start of the loop at step (4). Now \( z_{i-1/2} \) and SE(\( z_{i-1/2} \)) are, by definition, equal to \( z_{i-1/2} \), and SE(\( z_{i-1/2} \)) determined in the previous pass through the loop. The loop continues until the oldest time step is finished, designated by \( i = 1 \) and \( t = t_0 + \Delta t/2 \). The interested reader is referred to the source code for the DECON program for a more detailed account of these calculations.

**MODEL RESULTS**

**Tectonic model**

We first present results from our tectonic model considering both a pinned and a migrating divide in the drainage

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**Fig. 11.** Deconvolution of the source flux assuming a tectonic model with an average \( k_i \) of 0.07 Myr \(^{-1}\) and a pinned divide. (a) The observed erosion rate and \( \pm 1 \) standard error envelope based on the estimated relative standard error RSE \( \epsilon_{i}(t) \) ~ 12%. The circles indicate calculated values predicted by the inversion for the source flux (i.e. Eq. 15 solved by the damped least-squares method; see text for details). Note that the difference between calculated and observed are prescribed to have a mean of zero and a standard deviation of about one. (b) The estimated source flux record with the \( \pm 1 \) standard error envelope calculated using Eqs (16)–(18). (c) The integrated mean values for source, elevation and erosion. The integrated erosion is calculated from Fig. 5 and the assumed constant rate of chemical erosion. Consequently, the integrated erosion will vary between the models depending only on the type of divide assumed, pinned or migrating. The integrated source is determined from the estimated source flux given in (b). Elevation is given relative to modern long-term sea level which sits at \(-72\) m with respect to modern short-term sea level (see Fig. 3). This model, and all others that follow, use \( \beta(t) \) based on the long-term eustatic sea level curve in Fig. 3, and \( \beta = 5 \) m Myr \(^{-1}\).
basin, and three different $k_d$ values (0.038, 0.07 and 0.11 Myr$^{-1}$). The graphical format used for displaying the models is the same for every figure. Part (a) of the figure shows the observed mechanical erosion rates (heavy line) with a one standard error envelope (light lines), and the calculated mechanical erosion rates (open circles) determined by the inverse solution. Part (b) shows the estimated source flux, $v$, (heavy line), with one standard error limits (light lines). Part (c) shows integrated results for the mean source, $v$, mean elevation above base level, $z$, and mean total erosion, $e$. In the following discussion, we focus mainly on the predicted source flux, $v$, and the integrated source history, $v$.

All four tectonic models (three with a pinned divide and one with a migrating divide) exhibit rapid, peaked increases followed by protracted periods of slowly decreasing or steady $v$. As $k_d$ increases from 0.038 to 0.11 Myr$^{-1}$ (Figs 11–13) the range in source flux decreases and mean elevation decreases. The reason for these relationships is that the observed sediment flux requires lower topography as $k_d$ increases. Lower topography means a smaller ETC, which in turn means a smaller flux. The migrating divide model must account for higher mechanical erosion rates during the earlier part of the record. As a result, the solution for this set of models show greater ranges for $v$, and elevation during the Early Jurassic (Fig. 14). For the model simulation using a $k_d$ of 0.07 Myr$^{-1}$ and a pinned divide, mean elevation of the Early Jurassic Appalachians is estimated to have been about 800 m (Fig. 11). Between 500 and 900 m of mean elevation are predicted to be added to the landscape during the several, short-lived pulses of $v$ (Fig. 11).

**Erodibility model**

Our erodibility model solves for a variable $k_d$ and a source flux held constant at zero. For the erodibility model utilizing a pinned divide, $k_d$ varies from less than 0.01 to about 0.13 Myr$^{-1}$ (Fig. 15). Mean elevation of the Early Jurassic Appalachians is predicted to begin at about 1800 m and decay to the present 412 m in a stepwise fashion. Likewise, for the model using a migrating divide, $k_d$ varies from less than 0.01 to 0.14 Myr$^{-1}$, but is generally higher for the Early Jurassic Appalachians (Fig. 16). The greater $k_d$ values in the early history of the record are expected given the smaller drainage basin area predicted by the migrating divide model (Fig. 16).
Evolution of the post-Triassic Appalachians

Fig. 14. Deconvolution for the source flux assuming a tectonic model with an average \( k_s \) of 0.07 Myr\(^{-1} \) and a migrating divide. See Fig. 11 for further details.

The migrating divide model predicts a mean elevation for the Early Jurassic Appalachians of about 2300 m, which is about 500 m greater than that for the pinned divide case.

**DISCUSSION**

Our procedure for deconvolving the offshore basin sedimentary record of the middle Atlantic passive margin provides a first-order analysis of the evolution of mean topography for the post-Triassic Appalachian Mountains. Comparisons with regional, corroborative geological and geomorphic observations allows us to interpret our results in the context of the relative roles that tectonics, epeirogeny, climate, or rock type play on long-term landscape evolution. Other studies have already made the connection between pulses of offshore basin siliciclastic accumulation and postulated, coincident periods of tectonism in the drainage basin (Poag & Sevon, 1989; Poag, 1992; Poag & Ward, 1993). Here we evaluate these earlier interpretations in the context of our model results and concentrate on additional geological and geomorphic evidence in the form of marine eustatic transgressive sequences, fission-track thermochronology, palaeobotanical data and regional geomorphic relations. The discussion below reflects our view that the tectonic model, parameterized with a \( k_s \) of 0.07 Myr\(^{-1} \) and based on the pinned divide erosion data, best explains the offshore sedimentary record and the macrogeomorphic evolution of the post-Triassic Appalachian Mountains.

**Tectonic and epeirogenic influences on topography**

If we consider the offshore basin sedimentary history as a record of variable source flux, then we need to ask: What are the processes responsible for these variations?
We pose this question in the context of numerous studies which have debated the role of episodic land-surface uplift on post-rift passive margin landscape evolution (Schumm, 1963; Judson, 1975; Gilchrist & Summerfield, 1991). Perhaps the most striking features of Fig. 5 are the rapid transitions between slow and rapid sediment fluxes. We are often biased towards attributing these results to tectonic processes; however, tectonic features capable of producing such rapid and short-lived pulses of increased sediment flux, such as active thrust faults (Burbank & Verges, 1994), have simply not been identified on the Atlantic margin. The faults that have been described are typically reactivated rift and pre-rift structures with less than 100 m of post-rift offset. Long-term rates of offset for known fault systems all fall below 1 m Myr⁻¹ (Gardner, 1989).

So we are left with interpreting our tectonic model results in terms of various epeirogenic processes. Our interpretations here will be guided by a comparison of the source history to those predicted by the deep-crustal processes illustrated in Fig. 6 and are summarized in Fig. 17. Let us first consider rifting because this is an obvious process for the Jurassic history of the Appalachians. The immediate post-rift Appalachians (Slingerland et al., 1989) may have looked similar to a modern rifted margin such as those in the Red Sea region (Steckler & Omar, 1994). Our results show 500 (Fig. 13) to 1900 m (Fig. 14) of mean elevation during the time of rifting which straddles the mean elevation range for modern rifted margins. From 170 to 130 Ma, the tectonic model predicts an increase in source, followed by a large decrease to far less than the initial source at 170 Ma. This pattern is precisely what we would expect for a
rifted margin that sees an initial increase in source in response to upwelling of hot mantle, but then subsides as an isostatic response to crustal thinning (Fig. 6b).

As the Atlantic margin passed into the drift stage, Appalachian topography was reduced by erosion and thermal subsidence producing an Early Cretaceous topography with, on average, less relief than present (Figs 11–14). Negative source (source reduction of ETC) for most of the Cretaceous (Figs 11–14) is consistent with onlap of the low-elevation portions of the Appalachians, such as the Pennsylvania Piedmont, during eustatic highstands. Times of marine deposition in the upper Coastal Plain and Fall Zone in the late Early Cretaceous (Potomac Group), Late Cretaceous (Matawan and Monmouth Groups) and Early Tertiary (Pamunkey Group) generally correspond with model predicted times of subsidence in response to negative source (Fig. 17). We do not contest the fact that eustasy probably played the major role in allowing the seas to transgress onto the Fall Zone during these times. We suggest only that periods of margin subsidence during a time of high eustasy (Fig. 2) would represent the most likely times for marine deposition at or near the Fall Zone.

Several times during the drift stage, the Appalachians were subjected to peaked, short-lived episodes of increased source (Figs 11–14). We explore the possibility that deep-crustal thermal processes may be responsible for these pulses (Poag & Sevon, 1989; Poag, 1992; Poag & Ward, 1993). Magnetic activity in a passive setting is not unique to the Atlantic margin. Other passive margins, such as eastern Australia, contain post-orogenic, post-rift igneous rocks (Young & McDougall, 1983, 1985, 1993). Several syn- and post-rift magmatic/thermal events are well documented for the Atlantic margin (de Boer et al., 1988; Fig. 17). Many of these events are directly related in the passage of the Atlantic margin across hotspots (de Boer et al., 1988), such as those responsible for the New England and Bermuda seamount chains.

The tectonic model suggests that the introduction of relief, and subsequent increases in rate of erosion, were accomplished by short-lived (~10 Myr or less) increases in source at rates not exceeding 600 Myr \(^{-1}\) (Figs 11–14). Such rates of crustal thickening are more consistent with intrusion of magmas and thermally produced changes in crustal density (Fig. 6c,d) rather than contractional tectonic processes (Fig. 6a). In fact, the model appears to capture three different and distinct post-Triassic epeirogenic/magmatic events (refer to Fig. 11).

From 130 to 100 Ma, the model predicts a small increase in source, followed by a nearly symmetrical decrease to rates similar to the source at 130 Ma. This pattern is precisely what we would expect for a passive margin that was affected by dynamic topography (Fig. 6d). We are careful not to place too much emphasis on the model's capture of this event because its magnitude is fairly small compared with the uncertainties associated with the source flux estimates (see the SE envelope for source flux in Fig. 11b). From 100 to 60 Ma, the model predicts a large increase in source, followed by a symmetrical decrease to a rather constant value greater than the source at 100 Ma. This pattern is consistent with a passive margin initially inflated by magmatic and thermal processes, and followed by diminished subsidence because the density and thickness of the crust has now been fundamentally changed by the intrusions (Fig. 6c).

The pulses of increased source in the Cretaceous may be related to White Mountain and/or New England seamount magmatic activity (WM and NES of Fig. 17). On the basis of sediment dispersal patterns in the offshore basins, we agree with Poag's (1992) interpretation that the initial source event in the Barremian (124–119 Ma) increased mean elevation for the entire central and northern Appalachians, while the second event in the Coniacian–Santonian (88.5–84 Ma) was more localized in New England. Although not well represented in the offshore sedimentary record, a small increase in sedimentation rate at approximately 50 Ma is captured by our model as an increase in source and mean elevation. This event corresponds well in time with the known Shenandoah intrusive suite in Virginia (Fig. 17). Magnetostratigraphic data suggest that intrusive bodies associated with this event were emplaced in probably less than 1 Myr (Lovlie & Opdyke, 1974).

The largest and most poorly understood event in the source history of the post-Triassic Appalachians is the youngest one, spanning 25 Ma to the present. During this interval, sedimentation in the offshore basins reached its highest level since synrift deposition (Fig. 2b). No less than 1.1 km of rock, spread evenly across our drainages, is needed to account for the Miocene to present volume of offshore sediment (Braun, 1989). Thermochronology data for the Appalachian basin support a Miocene to present increase in exhumation rates (reviewed in Boettcher & Milliken, 1994) that may have produced the increased sediment flux. The fission-track data indicate the removal of 1.5 ± 0.5 km of rock in the last 20 Myr (Boettcher & Milliken, 1994). Such a slab of eroded rock should not be expected to have been evenly derived across the drainage basin. Rather, higher standing areas should have been exhumed more deeply and over a longer period of time than the low-standing regions near the coast. The fission-track data do, in fact, support a nonuniform pattern of erosion. Unroofing of the New England Appalachians occurred in the Cretaceous (Zimmerman et al., 1975; Miller & Duddy, 1989) which corresponds reasonably well with intrusive and volcanic activity of the White Mountains at that time. In contrast, most of the Piedmont and Ridge and Valley, which now stand relatively low, were initially exhumed in the Permian immediately following the Alleghanian orogeny (Zimmerman, 1979; Roden & Miller, 1989; Boettcher & Milliken, 1994). These data cast serious restrictions on the importance of rock type in controlling long-term rates of erosion (Hagedorn, 1980) because they demonstrate along-strike variations in erosion histories for areas that are otherwise underlain by similar rock suites.
Two independent lines of evidence point to epeirogenesis in the New England Appalachians as the cause of the increased Miocene erosion. Although no igneous bodies of Miocene age have been described, the 125-Ma Bear seamount is capped with a Miocene fossil coral reef (de Boer et al., 1988). The implication is that top of the seamount was in the photic zone, which is an intriguing observation considering that the middle Miocene sea level was no less than 60 m higher than present (Greenlee et al., 1988), and the top of the seamount is now well below the photic zone. Also during the Miocene, wide braided alluvial plains composed of coarse gravel and sand prograded into the Salisbury Embayment. Much of the material was derived from the north, possibly from a former New England Coastal Plain (Pazzaglia, 1993).

Our tentative interpretation is that the Miocene to present source event is related to asthenospheric flow and the formation of dynamic topography (Fig. 6d). The primary support for this bunch is the (apparent) Miocene uplift of the Bear seamount and the absence of late Cenozoic magmatism. If this interpretation is correct, then the source curve should drop another 1000 m when the dynamic support is completely removed (Fig. 11).

**Topography as a function of changes in rock erodibility**

We consider the alternative hypothesis that the offshore sedimentological record reflects, to a first order, changes in rock erodibility perhaps related to climate change (Poag & Sevon, 1989; Poag, 1992; Poag & Ward, 1993) or to the sequential exhaustion of rock types of variable erodibilities (Hack, 1980; Figs 15 and 16). The erodibility model, which assumes no source flux, suggests that the modern mean elevation of the Appalachians of 412 m would have decayed from a rifted margin with an initial mean elevation between 1800 (Fig. 15) to 2300 m (Fig. 16). This prediction seems unlikely given that the estimated mean elevation is significantly higher than that observed for modern rift margins and indicative of relief of more in line with the Alleghenian Appalachians (Slingerland & Furlong, 1989).

Nevertheless, there is little doubt that climate has changed for the post-Triassic Appalachians (Barron, 1989; Fig. 17). Climatic reconstructions for the central Appalachians show a general trend from a wet-dry seasonal climate in the Jurassic to a wet, warm temperate, mild seasonal climate in the Cretaceous (Barron, 1989). This trend would suggest a Cretaceous landscape with greater overall rates of chemical weathering in a setting already experiencing significant reductions in mean elevation and relief by mechanical erosion. The Cenozoic is marked by a period of long-term climatic cooling, supported by diverse geological and geochemical proxies including palynological data (Tiffany & Traverse, 1994), palaeobotanical data (Wolfe, 1978, 1985) and the marine oxygen isotope record (Miller et al., 1987). The Palaeocene and Eocene Appalachians straddled a well-defined mid- to high-latitude rainy belt, with the floral data suggesting greater warmth and humidity in the Eocene (Barron, 1989). Several important long-term palynoflora trends suggest dramatic terrestrial cooling by the late Oligocene (Tiffany & Traverse, 1994; Traverse, 1994). Long-term trends in pollen support three important late Cenozoic climatic cooling events for the middle Atlantic Coastal Plain (Pazzaglia et al., in press; Fig. 17). The first occurred in the late Oligocene and succeeded in distinguishing Palaeogene from Neogene florals. The second occurred between the middle and late Miocene, in which most thermophilic taxa became extinct and exotic taxa gave way to more cool-temperate forms. The final major change took place in the latest Pliocene or early Pleistocene, when remaining exotic taxa disappear, paving the way for completely modern assemblages. Given these palaeoclimatic observations, late Cenozoic changes in climate may account for a significant portion of the dramatic increase in mechanical erosion (Fig. 5b,c) model predicted changes in k_d (Figs 15 and 16).

Rock types of variable erodibility may play an important role in controlling local erosion rates (Hack, 1980) and concomitant changes in k_d, but we question their importance at larger spatial scales. We do not contest the Hackian view that the relative differences in rock hardnes between a quartz arenite and a carbonate in the humid temperate Appalachians will always dictate that the quartz arenite will stand higher. However, at the regional scale, proponents of rock-type control on landscape erodibility will have to explain the order of magnitude differences in sediment yield, as well as the order of magnitude changes in k_d. If anything, one might expect k_d to decrease systematically throughout the Cenozoic as the soft cover of sedimentary rocks is stripped from the Appalachians exposing the more resistant Grenville basement, but, in fact, the opposite is estimated by the erodibility model for the last 20 Myr (Figs 15 and 16).

Furthermore, if rock type represented the dominant factor in controlling mechanical erosion, then the New England Appalachians should not be an important source of sediment. But, the geological and fission track data presented above contradict this prediction.

**Landscape evolution paradigms**

It is a useful exercise to consider our model results in the context of prevailing paradigms for long-term landscape evolution (Gilbert, 1877; Davis, 1899; Hack, 1960). Some of our results support the Davisian model of an impulsive uplift, followed by exponential decay of mean elevation as ETC is erosionally consumed, resulting in a landscape of low relief. In particular, the 75-Myr period immediately following rifting, and the 65-Myr period through the Late Cretaceous and early Cenozoic appear to follow exponential decay of mean topographic elevation, in close approximation to the characteristic time for decay predicted for the Appalachian landscape using
an average $k_e = 0.07$ (Fig. 10). But does the model support the actual creation and preservation of peneplains in the Appalachian landscape? Very low rates of uplift for most of the Cretaceous are consistent with a low-standing landscape, but terrestrial lignite deposits of that age preserved in sinkholes (Pierce, 1965) suggest that the Appalachians were not completely inundated by Cretaceous seas. Even if a peneplain was produced in the Appalachian landscape at this time, it would be difficult to preserve it given that no less than 1.1 km average thickness of rock has been removed from the Appalachians since the middle Miocene (Braun, 1989). Unless the erosion was highly variable (Hack, 1982), essentially sparing the Ridge and Valley at the expense of New England and the Blue Ridge -- an assumption not supported by fission track data -- any Cretaceous erosion surface would likely have been obliterated (Figs 11–14).

Other aspects of our analysis more closely mimic the concepts of dynamic equilibrium (Gilbert, 1877; Hack, 1960) rather than the Davisian geographical cycle. For instance, the tectonic model (Fig. 11) predicts that the source flux has been negative as well as positive. Thus, tectonic and epeirigenic processes cannot be reduced to a simple set of impulsive positive pulses as envisaged by the Davisian model. Instead, the eroded flux of sediment shed from the mountain range was continually adjusting to a complex input signal, which is more in line with the Hackian view. An interesting feature of our analysis is that the unit response function, which is simply a unitary example of the Davisian model, can be used to represent the tectono-geomorphic history of the Appalachians. In this sense, the complex interplay between tectonics and erosion advocated in the Hackian model can be viewed as consisting of a number of superimposed, Davisian unit response functions.

We recognize that this simple linear model does not represent the full complexities of the tectono-geomorphic system, even when simplified to the regional scale of an entire mountain belt. Nonetheless, our approach does provide a useful conceptual basis for understanding the macrogeomorphic evolution of mountain belts as well as first-order quantitative predictions of changes in source and mean elevation that can be compared to geological evidence.

CONCLUSIONS

We have presented two simple end-member models for deconvolving the sedimentary record of a depositional basin along separate, but parallel, assumptions which allow us to evaluate the roles of tectonic (epeirigenic) processes, climatic processes and rock type on long-term landscape evolution. These models, constrained by the well-documented sediment volumes of the middle US Atlantic margin, serve to clarify the roles that tectonic forcing and variability erodibility had on the highly variable sediment fluxes recognized in previous studies (Poag & Sevon, 1989; Poag, 1992; Poag & Ward, 1993). We recognize the numerous assumptions built into our approach and thus caution against the dogmatic use of our models. Instead, we feel that the models help to focus some long-debated questions about the tectono-geomorphic evolution of the Appalachians. They also provide quantitative predictions about the distribution, timing and magnitude of erosion in the central and northern Appalachians, predictions that might be testable with further work. Of particular interest is the temporal record of erosion rates, which might be accessible by fission-track dating of detrital apatite and zircon from the offshore sedimentary sequences (cf. Brandon & Vance, 1992; Garver & Brandon, 1994).

Our model suggests that three of the four major pulses of sedimentation in the Atlantic margin basins (Fig. 5) were produced by rapid, short-lived increases in source. The pattern of change in source suggests that the Jurassic event was related to rifting and post-rift thermal relaxation; the two Cretaceous events to asthenospheric flow and magmatism, respectively, with subsequent thermal relaxation; and finally the Miocene to present event to asthenospheric flow and development of dynamically supported topography. The present high mean elevation of the New England Appalachians may reflect this youngest source event. Our tectonic model probably overestimates the importance of source during the most recent event (Figs 11–14) because this time is coincident with major climatic cooling, Northern Hemisphere glaciation, and a concomitant increase in $k_e$ and mechanical erosion. Overall, we feel that the role of rock type in controlling landscape erodibility is a secondary consideration at our scale of observation. With refinements, and a better appreciation for the functional relationship between elevation, relief and mechanical erosion rates, this deconvolution approach may help to force critical re-evaluations of the prevailing long-term landscape evolution paradigms as they apply to old orogenic belts such as the Appalachian Mountains.

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Evolution of the post-Triassic Appalachians


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