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Geodynamic and Geomorphic Evolution of the Permo–Triassic Appalachian Mountains

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Abstract

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The basic topography of a steady state, accretionary mountain range is a function of the critical taper of the accretionary wedge, which in turn depends upon the convergence flux at the toe, internal rock strength, basal slope, and erosion rate off the top. It is possible to obtain a first-order estimate of this topography for ancient mountain ranges by constraining these parameters using sediment thicknesses in adjacent, flexurally-produced sedimentary basins. Here we apply the technique to the Early Permian central Appalachians, by first obtaining the steady-state mass of the wedge from a flexural loading model of Beaumont and others. Then, the equations of a critically tapered wedge are solved using a Monte Carlo method of parameter selection. Those solutions that match the observed mass of the wedge suggest that the Early Permian Appalachians possessed a central Andean topography with average relief of from 3.5-4.5 km and width of 250-300 km.

Introduction

Elevation and width are two primary geomorphical attributes of a mountain range, and it is of some interest to speculate on their controlling factors. Recent advances in our understanding of crustal convergence now make it possible to offer a first order explanation as to why modern mountain ranges, at the same stage in their evolution, exhibit a wide variation in these basic attributes. This understanding raises the possibility that ancient ranges, now only vestiges of their former selves, can be reconstructed to their once lofty heights by an application of the same principles. Of all the mountain ranges that have existed none would seem more suited to bring back to life than the ancestral Appalachians. They are one of the best studied ranges in the world and the literature is rich in speculation on their geomorphic evolution (e.g., Judson, 1975; Sevon, 1985; Morisawa, this volume). Maybe most importantly, the Appalachian sedimentary basin preserves important, albeit circumstantial, information.

Our intention here is to use geodynamic principles and the results of recent basin modelling to better rationalize the evolution of Appalachian topography during the Permo–Triassic. We have attempted to answer two questions: (1) what variables and feedback loops control the topographies of orogenic mountain belts in general, and the Appalachians in particular? and (2) how high and wide were the Appalachian Mountains in the central Atlantic region at the end of the Alleghanian Orogeny?

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Controls of the topography of mountain ranges

To first order, the topography of an orogenic mountain range – that is, its mean elevation, relief, and width – is controlled by the processes of crustal thickening and denudation (e.g., Molnar, 1988; Molnar and Lyon-Caen, 1988). Crustal thickening can be produced by a variety of mechanisms including thrusting-related shortening and magmatic addition. In convergent orogenic belts such as the Appalachian or Alpine chains, crustal thickening via folding and thrusting predominates. Such crustal thickening can occur on length scales ranging from short (0–5 km) imbricate thrusts and buckle folds seen in accretionary wedges, to the megathrusts of continental convergence (sometimes referred to as continental subduction) where 'thrust' sheets of greater than 10 km in thickness may be displaced tens or hundreds of km. The resultant mountain belt geometry is additionally "filtered" by the response of the lithosphere to this added mass of crust which acts to deflect the underlying lithosphere. This deflection will normally be flexural in nature, that is, it will be a broad regional downwarp rather than a mirror image of the surface topography that would be expected if local isostatic compensation occurs. The magnitude and wavelength of this lithospheric deflection is controlled by the strength or flexural rigidity of the lithosphere and thus the net topographic shape of the mountain range and adjacent sedimentary basins will depend on both the size and shape of the load (i.e., the region of crustal thickening), and on the strength of the underlying plate.

Several approaches use these basic ideas to evaluate the evolving and steady-state geometry of a mountain belt. For thickening via folding and thrusting, the width and height of the region of thickened crust is limited by the magnitude of thickening (Molnar and Lyon-Caen, 1988), because the horizontal driving forces must work against gravity and the gravitational potential energy of a range grows as the square of its height. When this work is greater than the work necessary to break through the crust in an adjacent lowland area, thickening commences there and the range grows wider. Davis et al. (1983) and Dahlen and Suppe (1988) have combined this energy view with a failure criteria for the crust to calculate the wedge geometries of many convergent orogens. Alternatively, one may evaluate the lithospheric response to mountain loads. Often, the amplitude and shape of the peripheral sedimentary basin can be used via a flexural model to place constraints on the excess mass present in the thickened crust of the mountain range (e.g. Beaumont et al., 1987, 1988).

Our contribution here is to combine the approaches mentioned above in order to develop a model for the evolution of the Appalachian system that has its mass distribution at any time controlled by the response of the lithosphere to the load of the mountain range, and its geometry controlled by the combination of energy considerations, an assumed failure criterion for the region, and erosion. Below we describe the models used, starting first with the wedge model which provides us with the geometry (including effects of erosion), and then the flexural loading model which, through an analysis of the basin geometry, constrains the excess mass in the range. Secondly, we describe the application of this method to the geometry of the Appalachians. Given a lack of hard data to constrain many of the parameters, we use a Monte Carlo approach to evaluate the most probable width and heigth of the Permian Appalachians in the region of Pennsylvania.

Theory

Wedge model

Dahlen (1984), Dahlen et al. (1984), and Dahlen and Suppe (1988) consider the crustal thickening in active fold and thrust belts to be analogous to the wedge of snow or soil in front of a bulldozer blade (Fig. 1). As material is

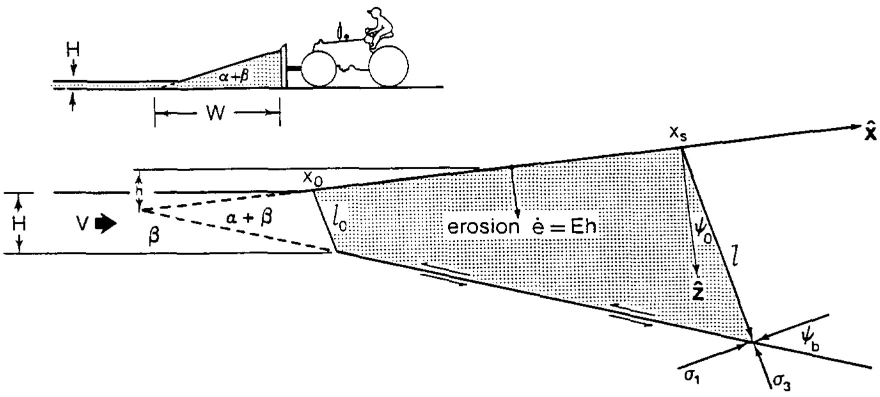

Fig. 1. Schematic of the Dahlen and Suppe (1988) critically tapered wedge. Material enters the toe (x_0) at the rate of HV km^2 Myr^{-1}, where H is the thickness of the incoming material and V is the convergence velocity. The material is stretched along the axis of least principal stress, σ_3, and narrowed along σ_1 as it passes back through the wedge. The basal slope of the wedge is β and the topographic slope is α. Ψ_0 and Ψ_b are the acute angles between the axis of principal compressive stress and the topographic and basal surfaces, respectively. Erosion proceeds perpendicular to x at a rate of $\dot{e}$ km Myr^{-1} which is a function of h, the height of the wedge surface above the vertex. A steady state occurs for some width W (equal to $x_s - x_0$) when $\dot{e}$ equals HV (modified from Dahlen and Suppe, 1988).

scraped up, the wedge grows in height until it reaches a critical surface slope or taper governed by the relative magnitudes of the sliding resistance along the base and the strength of the wedge material. After reaching critical taper, the wedge grows self-similarly as new material is added at the toe, becoming wider and thicker but maintaining the same surface slope. This equilibrium taper is called critical because in thinner wedges, compressive stress will exceed the rock strength, as given say, by the Coulomb failure law, and the wedge will deform by thickening until the critical taper is achieved. Finally, the wedge reaches a steady state when accretionary influx is balanced by erosive efflux. The Coulomb wedge model has been successfully applied to several active fold and thrust belts and accretionary wedges (Dahlen and Suppe, 1988) with good agreement between observed wedge geometries and theory for reasonable values of rock strengths.

The wedge width and height can be calculated from a mass balance of the material entering and leaving the wedge (Fig. 1) (Dahlen and Suppe, 1988). In Figure 1 a column of rock aligned with the principal stresses moves back through the wedge, elongating in the σ_3 direction from l_0 to l while conserving its cross-sectional area. The flux into the wedge due to convergence is:

$$\dot{A} = (1 - \Delta\rho/\rho)HV \qquad (1)$$

and the flux out the back of the wedge at position x_s is:

$$l\dot{x}_s \cos\psi_0 = x_s \dot{x}_s \frac{\sin(\alpha+\beta)\cos\psi_0}{\cos\psi_b} \qquad (2)$$

where the dot denotes differentiation with respect to time, ρ is the density of the material, $\Delta\rho$ is the change in density as material enters the wedge, and the remaining variables are defined in Fig. 1.

The flux off the surface of the wedge due to erosion is rationalized as follows. Regression equations of modern erosion rate as a function of relief (Ruxton and McDougall, 1967; Ahnert, 1970; William Hay, pers. commun., 1988) seem to show a linear dependence, and thus as a first approximation Dahlen and Suppe (1988) let:

$$\dot{e} = Eh \qquad (3)$$

where $\dot{e}$ is the erosion rate perpendicular to the surface of the wedge, E is a constant with units of Myr^{-1}, and h is height (km) of the surface above the vertex of the wedge. Note that because of self-similarity, the surface slope of the wedge doesn't change, and therefore in the model the increase in erosion rate must be assumed to arise from complex changes in vegetation, elevation-dependent processes such as glaciation, and from increasing slopes of streams entrenched into the wedge. The total erosive flux off the surface of the wedge from x_o to x_s is:

$$\int_{x_0}^{x_s} \dot{e}dx = \frac{1}{2}E \sin \alpha (x_s^2 - x_0^2) \tag{4}$$

By conservation of mass eqn. (1) must equal 2 plus 4. Solving for $x_s(t)$, the location of a rock column in the wedge, yields:

$$x_s(t) = (W + x_0)[1 - \{1 - (x_0/(W + x_0))^2\} \exp(-\eta t)]^{1/2} \tag{5}$$

where $x_s(0) = x_0$ when $t = 0$, and:

$$w = \left[\frac{2\dot{A}}{E \sin\alpha} + x_0^2\right]^{1/2} - x_0 \tag{6}$$

$$\eta = \frac{E \sin\alpha \cos \psi_b}{\sin(\alpha + \beta) \cos \psi_0} \tag{7}$$

$$x_0 = \frac{H \cos \psi_b}{\sin(\alpha + \beta) \cos(\psi_0 + \alpha)} \tag{8}$$

As t approaches infinity, x_s approaches $x_0 + W$, and thus W is the steady-state width of an eroding, critically tapered wedge. By construction in Fig. 1:

$$h = x \sin \alpha \tag{9}$$

Dahlen and Suppe (1988) have applied eqn. (5) to the accretionary wedge of which Taiwan is a part, using the parameters in Table 1. The observed steady state width, W, is 87 km which is predicted from eqn. (6) only if $E = 1.27$ Myr^{-1}. This yields an erosion rate $\dot{e}$, averaged over the length of the wedge, of 5.7 km Myr^{-1}. Taiwan streamload data indicate a mean erosion rate of 5–6 km Myr^{-1} over the last 30 years (Li, 1976) and fission track data indicate a rate of 3–10 km Myr^{-1} over the last million years (Liu, 1982), values that are surprisingly close to the predicted. The steady-state height above the point x_0 at a distance $x_0 + W/2$ from the origin is:

$$h = \frac{W}{2} \sin \alpha \tag{10}$$

or 2.2 km, also in reasonable agreement with

TABLE 1

Values of independent parameters in modern wedges (from Westbrook et al., 1982; Davis et al., 1983; Dahlen, 1984; Dahlen and Suppe, 1988)

Wedge	α (°)	β (°)	$(\alpha+\beta)$ (°)	ψ_0 (°)	ψ_b (°)
Taiwan	2.9∓0.3	6∓1	-	2.6	11.5
Barabdos	0 to 3	2	3	-	-
Makran (Gulf of Oman)	-	-	4	-	-
Eastern Aleutians	-	-	7	-	-
off Oregon Coast	-	-	8	-	-

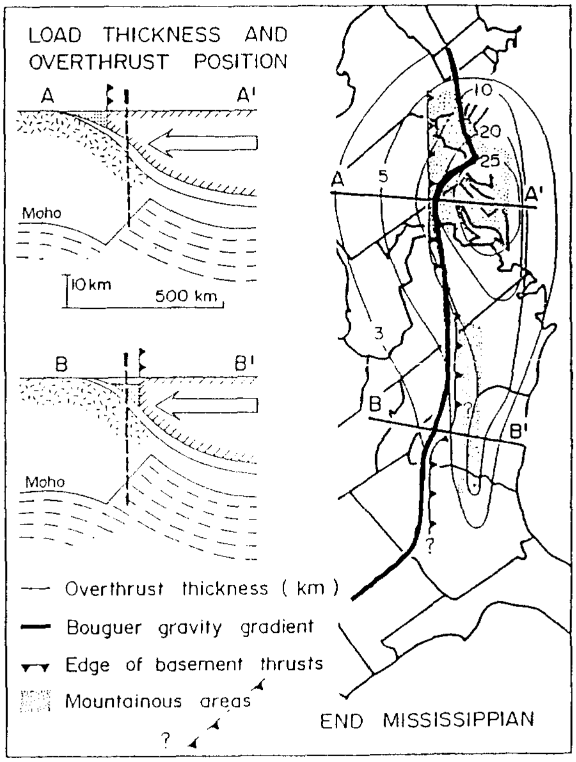

Fig. 2. Results of the Beaumont et al. model at the end of the Mississippian, showing overthrust thickness (km) in relation to the Bouguer gravity gradient. See text for interpretation (from Beaumont et al., 1987).

the geomorphology of Taiwan, where the central ranges average 2–3 km in elevation with a few peaks reaching 4 km.

Implicit in this theory is the assumption that fold and thrust belt mountains grow in height and width until their erosion, given by $W \langle \dot{e} \rangle$, balances $\dot{A}$, at which time they are in steady-state. Thus high erosion rates (on the order of 10^1 km Myr^{-1}) should produce narrower (50–200 km) and lower (1–5 km) mountain ranges than low erosion rates (on the order of 10^0 km Myr^{-1}) which should produce wider (>500 km) and higher (ca. 9 km) ranges (Dahlen and Suppe, 1988). As Dahlen and Suppe (1988) point out, this is substantiated by the Andes, whose width is greater in the arid zone between 10° and 30°S latitude than to the north or south.

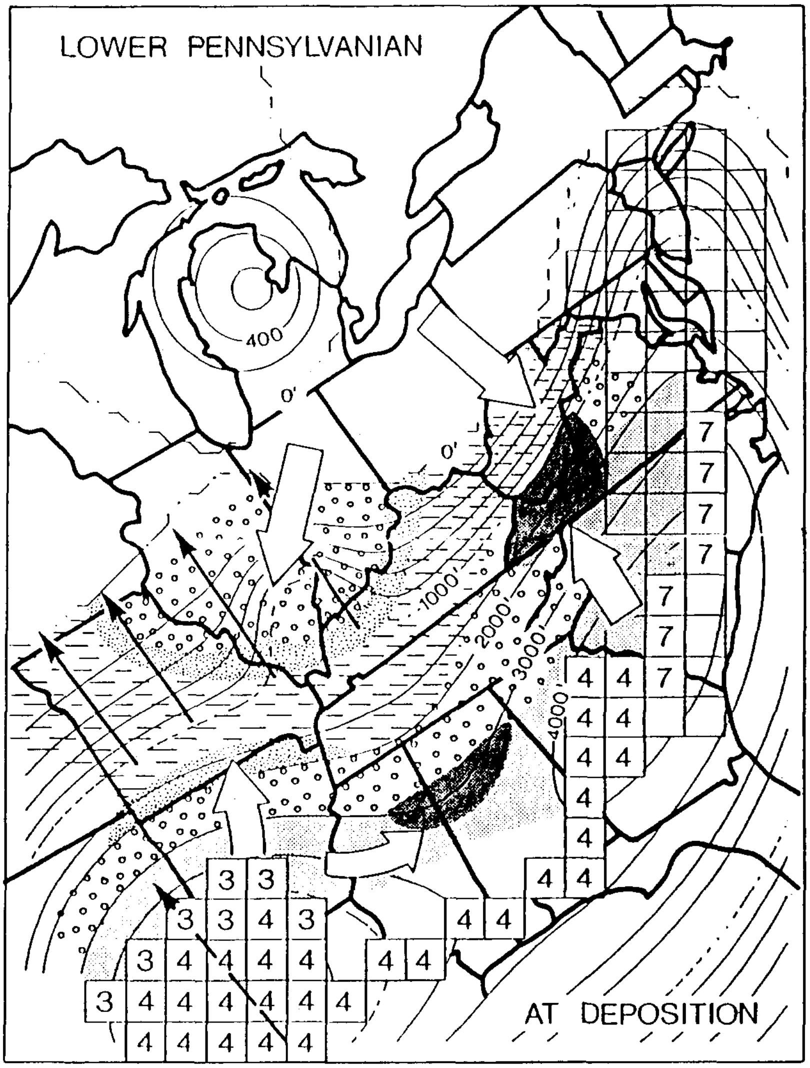

Fig. 3. Predicted sedimentary isopachs (in ft for comparison with original data) from the Beaumont et al. model for the Early Pennsylvanian. Numbers in squares are the load thicknesses in km that generate the necessary subsidence in the foreland basin. Dash and dot patterns denote observed shales and sandstones, respectively; large arrows show the major dispersal directions (from Beaumont et al., 1987).

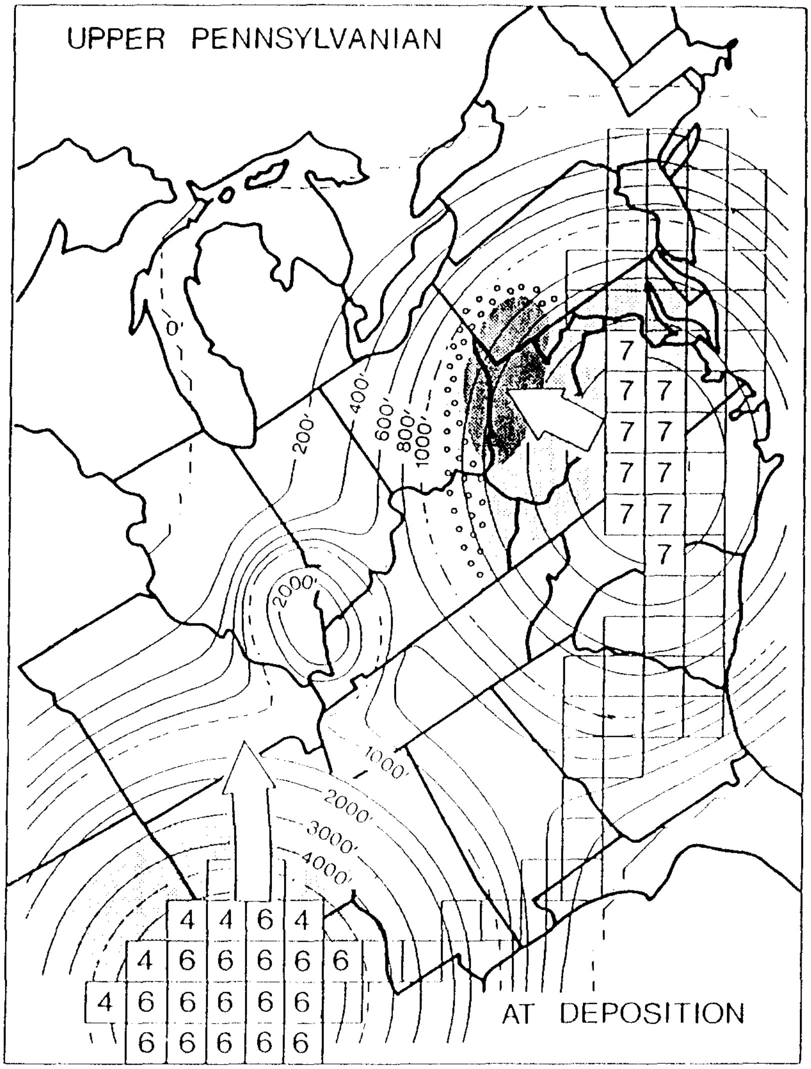

Fig. 4. Predicted sedimentary isopachs (in ft for comparison with original data) from the Beaumont et al. model for the Late Pennsylvanian. Numbers in squares are the load thicknesses in km that generate the necessary subsidence in the foreland basin. Dot pattern denotes sandstones; large arrows show the major dispersal directions (from Beaumont et al., 1987).

Flexural loading model

When the lithosphere is subjected to supracrustal loading it flexes downward, producing a sedimentary basin adjacent to the load. The width and depth of the basin are functions of the magnitude and rate of application of the load, and the rigidity of the crust. This simple relationship between basin geometry and load suggests that one could use the history of basin development as recorded in the stratigraphic record to back-calculate the magnitude and timing of the applied loads. This method has been elegantly pursued by Quinlan and Beaumont (1984), Stockmal et al. (1986), Beaumont et al. (1987), Beaumont et al. (1988), and Jamieson and Beaumont (1988), and summarized for the Appalachians in Slingerland and Beaumont (1989). Insofar as the loads represent a thickening of the crust, their magnitudes and rate of application should be some measure

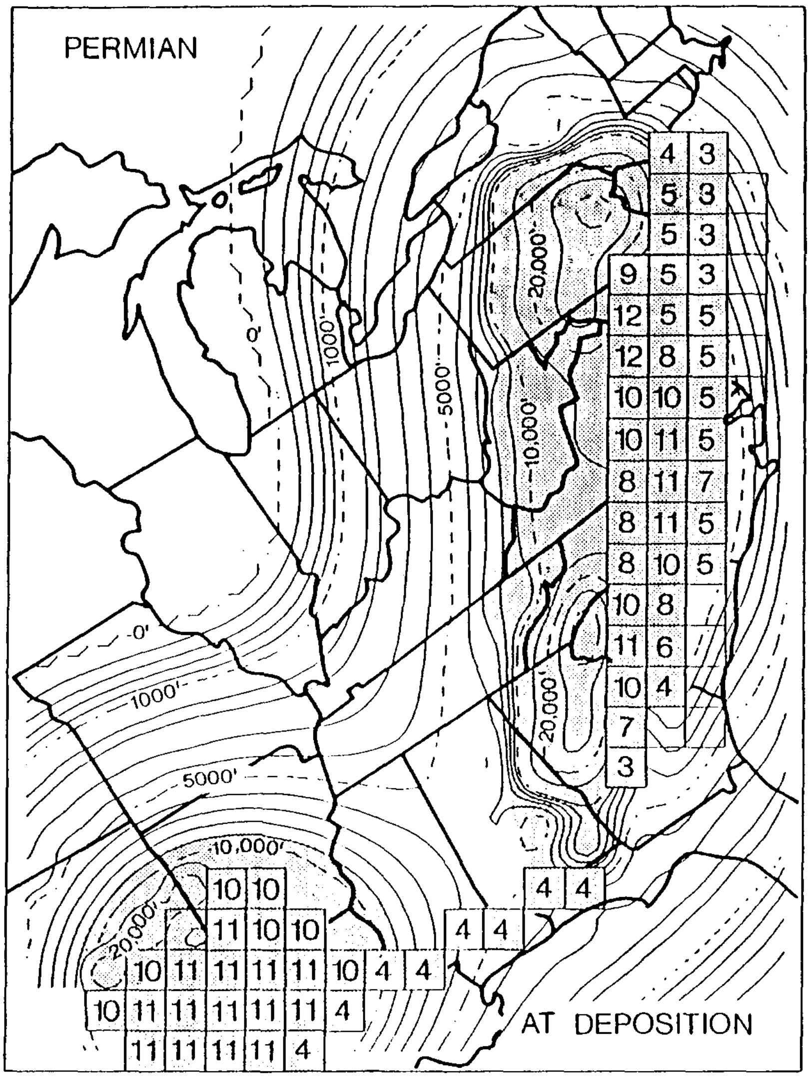

Fig. 5. Predicted sedimentary isopachs (in feet for comparison with original data) from the Beaumont et al. model for the Permian. Numbers in squares are the load thicknesses in km that generate the necessary subsidence in the foreland basin. Sediment thicknesses are inferred, for the most part from coal moisture, vitrinite reflectance, fluid inclusion data, and thermal maturation of shales (from Beaumont et al., 1987).

of the width and height of any mountain range they create.

To back-calculate the magnitude and timing of the crustal thickening, Beaumont and coworkers model the lithosphere as a linearly viscoelastic body in which the viscosity is temperature dependent. The lithosphere is supported by an inviscid or low viscosity fluid half-space (for details, see Quinlan and Beaumont, 1984). The area of the Appalachians is subdivided into a 20×31 grid, each a square 88.9 km on a side. There are twenty timesteps of non-uniform length during which loads are applied along the eastern margin of the Paleozoic Appalachians and the crust allowed to flex. Overthrust loads and sedimentary fill have a density of 2400 kg m^{-3} except for one timestep when the effect of higher density sediment is included. Load thickness scales linearly with density; therefore loads of 2800 kg m^{-3} would be 14% thinner. The

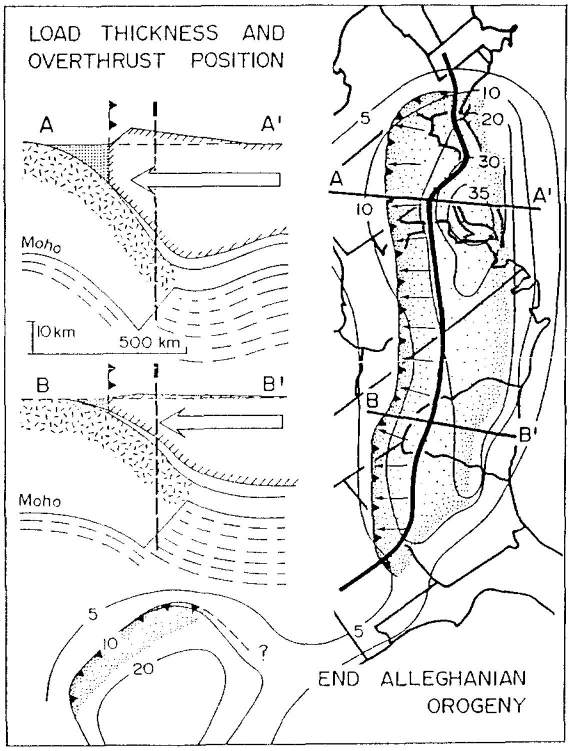

Fig. 6. Results of the Beaumont et al. model at the end of the Alleghanian orogeny, showing overthrust thickness (km) in relation to the Bouguer gravity gradient. See text for interpretation (from Beaumont et al., 1987).

thicknesses of overthrust loads are constrained in two ways - results from the mathematical model of lithospheric flexure must match both the Paleozoic and present sedimentary isopachs. In the case of the Permian strata, now largely eroded, the thickness of the eroded strata is estimated from the moisture content and rank of the coals (Beaumont et al., 1988).

The results (Figs. 2–6) show the loads that best fit the sedimentary isopachs and depths of burial. At the end of the Mississippian, the cumulative thicknesses of overthrusts and basin fill reached a maximum of over 25 km in the Delmarva area, outboard of the present Bouguer gravity gradient (Fig. 2). Insofar as this gradient represents the transition from continental crust to stretched and oceanic crust, these loads on thinned crust can be accommodated without significant topography (Beaumont et al., 1988). During the Early Pennsylvanian (Fig. 3), little if any thrusting occurred in the central Appalachian region. This is inconsistent with previous interpretations of the Pottsville gravels, but one must remember that although those gravels are coarse, they also are quite thin. Levine and Slingerland (1987) suggest they are more directly a result of increased precipitation than of increased elevation in the source terrain.

By the Late Pennsylvanian, thrusting had moved northward to the central Appalachian region (Fig. 4), although the major loads remained south of New Jersey. At the end of the Permian, thrusting had moved into southern New England and also cratonward across eastern Pennsylvania and Maryland (Fig. 5). These results are corroborated by vitrinite reflectance and fluid inclusion studies of the eastern anthracite fields of Pennsylvania (MacLachlan, 1985; Orkan and Voight, 1985; Levine, 1986), suggesting tectonic emplacement of up to 10 km of overburden there during the Permian. The cumulative thicknesses of model loads and foreland basin deposits during the Alleghanian orogeny (Fig. 6) suggest that overthrusts extended significantly westward onto the unthinned craton, and therefore may have had significant topography (Beaumont et al., 1988). Unfortunately these results are not unique – thicker loads further outboard (that is, wider mountains) – could match the observed sediment thicknesses equally well). Also, the effects of erosion are not taken into account.

Application to the ancestral Appalachians

By combining the critical wedge theory with the loading history proposed by Beaumont and coworkers, we hope to develop a self-consistent model of topography and load which uniquely satisfies the available data from the central Appalachian region. We use a Monte Carlo simulation because few of the parameters in eqns. (1)–(10) are well constrained. The results of this modelling are distributions of solutions, each solution simultaneously satisfying the physical constraints of the wedge theory and the observational constraints of the basin geometry and mountain load. The physical constraints of the wedge theory are that the steady state width and average height should be given by eqns. (6) and (10), respectively. These equations along with (1), (3), and (8), are solved using various combinations of the parameters described below. The observational constraint is that the steady-state load of the wedge per unit length of mountain range (stippled area in Fig. 1), given by:

$$L=\tfrac{1}{2}(W^2+2Wx_0)\frac{tan(\alpha+\beta)}{1-tan\ (\alpha+\beta)tan\ \Psi_0} \tag{11}$$

must be equal to the Permian loads calculated by Beaumont and coworkers (Fig. 5). These amount to 5800 ± 580 km^3 per unit length of mountain range in the central Atlantic region. Thus, wedge model solutions are obtained for a very large number of random combinations of the independent parameters, V, H, and so on, but only those that produce a load of 5800 ± 580 km^2 are retained.

Parameter estimation

Table 2 lists the ranges of parameters used in the simulations. The convergence velocity, V, is obtained from the plate convergence velocities at compressive margins today, estimated by Dahlen and Suppe (1988) to be $10<V<100$ km Myr^{-1}. Paleogeographic reconstructions (Ziegler et al., 1979) yield a value of roughly 66 km Myr^{-1} during the Late Pennsylvanian, consistent with past and present velocities in the region of the Alps (Le Pichon et al., 1988), and within the range of modelled values. Sediment thicknesses entering present-day mountain belts are in the range $1<H<10$ km (Dahlen and Suppe, 1988). It is difficult to estimate H for the late Paleozoic Appalachian wedge, but the convergence flux $\dot{A}$, equal to VH, can be constrained.

The minimum possible value of $\dot{A}$ is obtained from the thicknesses of the loads predicted by Beaumont and others (Table 3) by dividing the km^3 loaded (per km along the thrust belt) by the interval over which the loading occurred. In table 3 it is assumed that the Permian loading occurred over the first 28 my of that period as suggested by the timing of Alleghanian remagnetization (Miller and Kent, 1988). The remagnetization age of Appalachian redbeds and limestones youngs northwards. The assumption that the remagnetization reflects the influence of fluids driven ahead of thrust sheets allows Miller and Kent (1988) to infer that major thrusting and large-scale fluid migration ended near the Permo–Carboniferous boundary in the southern Appalachians but continued on through a part of the Permian in the central Appalachians. Note that this corroborates the northward progression of loading inferred by

TABLE 2

Ranges of values used in Monte Carlo simulations of the wedge model

$\dot{A}$	10–1000	km^2 Myr^{-1}
V	10–100	km Myr^{-1}
H	1–10	km
E	0.1–10	Myr^{-1}
$(1-\Delta\rho/\rho)$	0.9	-
ψ_b	11.5	degrees
ψ_0	2.6	degrees
α	1–3	degrees
β	2–6 or 2–12	degrees

TABLE 3

Cross-sectional compressive fluxes, A, into the toe of the Alleghanian wedge (km^2 Myr^{-1}), estimated from the loads in Figs. 2–6

Interval	Northern Virginia	Southern Pennsylvania
Early Pennsylvanian	none	none
Late Pennsylvanian	57	none
Early Permian	80	54

Beaumont and co-workers (Figs. 2–6) and is consistent with the observation that this secondary natural remnant magnetization is independent of bedding attitude, indicating it is post-tectonic (Van der Voo and French, 1977; Scotese et al., 1982). This timing also is consistent with a Rb–Sr whole-rock age of 273 my for Brevard zone ultramylonite from near Rosman, North Carolina (Hatcher et al., 1988). The values obtained in this manner (Table 3) are comparable to the Barbados wedge away from the Orinoco delta which has a compressive influx of 20 km^2 Myr^{-1}. Taiwan, by contrast has a compressive influx of 500 km^2 Myr^{-1}. The values in Table 3 are minimum fluxes however, because they do not account for the material entering the wedge that is eroded during the interval. Thus, the range of $\dot{A}$ from 10 to 1000 km^2 Myr^{-1}, obtained from the product of HV, appears to cover the expected values for the late Paleozoic Appalachians.

Determining the best estimates of E, the erosion constant in eqn. (3) requires considering its range of values today as a function of climate and estimating the paleoclimate of the central Appalachians in Late Pennsylvanian–Early Permian time. Data of Whitehouse (1988) show that regional erosion rates across the Central Southern Alps of New Zealand are proportional to mean elevation. Given the quality of the data, when $\langle \dot{e} \rangle$ is plotted versus h as in eqn. (3), the proportionality constant E, lies between 3 and 6 Myr^{-1} for this wet region where precipitation values range between 4 and 12 m yr^{-1}. Dahlen and Suppe (1988) derive $E = 1.3$ Myr^{-1} for the mountain belt of central Taiwan, another wet climate. Dry regions such as the central Andean Altiplano, experience average erosion rates of 0.5 km Myr^{-1} (Dahlen and Suppe, 1988), yielding an $E = \mp 0.15$ Myr^{-1}, an order of magnitude lower (although see Flemings and Jordan, 1989, for higher values based on volumes of foreland basin fill). This contradicts the results of Ahnert's study (1970) where there was no obvious variation of E with precipitation. However, we feel this arises because Ahnert's database consists only of 20 mid-latitude small drainage basins. More recently, Pinet and Souriau (1988) determined E using a worldwide data set from continent-scale drainage basins. Denudation rates were determined from the total sediment load of rivers (dissolved and suspended) and h was taken as the mean elevation within the basin. As expected, they found that the dissolved load was mainly correlated with precipitation and not with relief. In the case of the suspended sediment load, two correlations with mean elevation arise, one for basins associated with recently active orogenies and one not. For the former, the linear regression equation is:

$$\dot{e} = 0.419h - 0.245 \qquad (12)$$

where $\dot{e}$ is in km Myr^{-1} and h is in km. The non-zero intercept can be interpreted to mean that a threshold elevation of about 0.58 km (at which value $\dot{e}$ equals 0) separates an erosive from a depositional terrain. Above the threshold elevation, E in eqn. (3) is approximately 0.42. They found no significant correlation with mean annual precipitation, which ranged from 0.76 to 2.03 m yr^{-1} among the basins. Based on these studies, we conclude that E varies from 0.1 to 1.0 for regions with less than about 2 m yr^{-1} annual precipitation, and from 1 to 10.0 for wetter regions.

The Permian climate of the central Appalachians can be estimated by two methods: (1) mapping paleoclimatic indicators such as evaporites, and (2) paleoclimatic modelling using atmospheric general circulation models. Most researchers (Ziegler et al., 1979; Crowley et al., 1987; Parrish et al., 1986; Barron, this volume) who have summarized the paleoclimatic indicators agree that the Pangean climate generally was one of widespread warmth and aridity. Numerical climate models corroborate this (Kutzbach and Gallimore, in press), calculating year-round aridity for all regions of Pangea but the eastern coastal regions, the tropical west coast, and the regions poleward of 40° latitude. This is true for various CO_2 concentrations, solar luminosities, and two different continental mean elevations (0 and 1 km). The model does not include the ancestral Appalachians, and the possibility arises that they created a local orographic climate wetter than surrounding areas (e.g., Hay et al., 1982). Computer simulations similar to Kutzbach and Gallimore's, but including crude estimates of mountains, did not indicate this however (E. Barron, pers. commun., 1989). Thus, we conclude that the appropriate erosion constant, E, is most probably in the range of 0.1–1 Myr^{-1}, but we include simulations with $1.0<E<10.0$ Myr^{-1} for comparison.

Ψ_0 and Ψ_b the acute angles between the axes of principal compressive stress and the top and bottom of wedge, must be consistent with Byerlee's law. Here we take these as constant at the Taiwan values of 2.6 and 11.5°.

The values of α and β vary today from 0–3° to 2–6°, respectively (Table 1). But, as Jamieson and Beaumont (1988) point out, the geometry of the wedge on the edge of a rifted continental margin is likely to be more complex than a simple triangle. The basement ramp is likely to have slopes of up to 12° in the region above the transition from attenuated to unattenuated crust, leveling off cratonward to around 2°. It should be expected that this steeply dipping buttress would produce a region of high topography not accounted for in the simple wedge model. While we are not prepared to construct a wedge model with multiple basal slopes, we can examine the influence of varying β through shallow and steep slopes. Thus, β will take on the ranges of either 2–6° or 2–12° in the various numerical simulations.

Results of the Monte Carlo simulation

The results of the Monte Carlo simulation are presented in Figs. 7–9 as frequency by number versus mean height and width of the steady-state mountain range. It is important to note that in all simulations, only a few percent of the many thousand trials came within 10% of the loads required by eqn. (11). This indicates to us a certain robustness of the method in that there are relatively few combinations of parameter values that match the observed load. Also, the average convergence velocities of these acceptable simulations are all near 65–70 km Myr^{-1}, values remarkably close to the 66 km Myr^{-1} estimated from paleogeographic reconstructions as mentioned earlier.

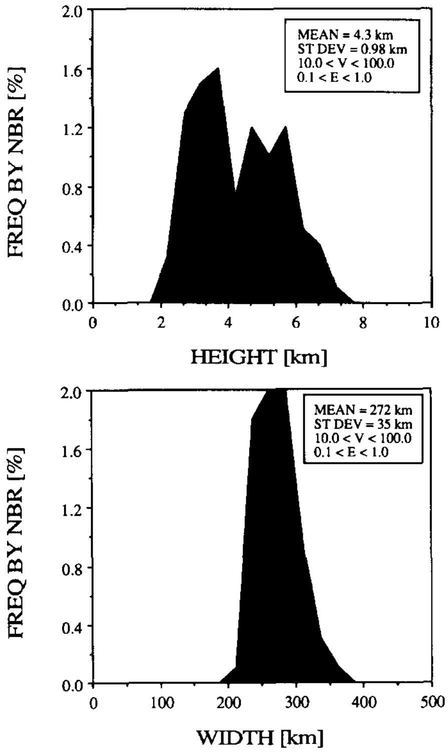

Fig. 7. Results of the Monte Carlo simulation for a dry climate and wedge basal slope of 2–6°. (a) Predicted mean height of the wedge (with respect to the ground surface at its toe) for the Permian central Appalachians; (b) predicted steady state width. The histogram gives the frequency of occurrence of height or width computed for various combinations of the parameters described in the text. V is convergence velocity (km Myr^{-1}) and E is the erosion constant (Myr^{-1}),.

Assuming a dry climate and shallow basal slope of the accretionary wedge (2–6°), the most probable width and cross-sectional average height of the central Appalachians are 272 km and 4.3 km, respectively (Fig. 7). Note that this height is measured not from sea level or the

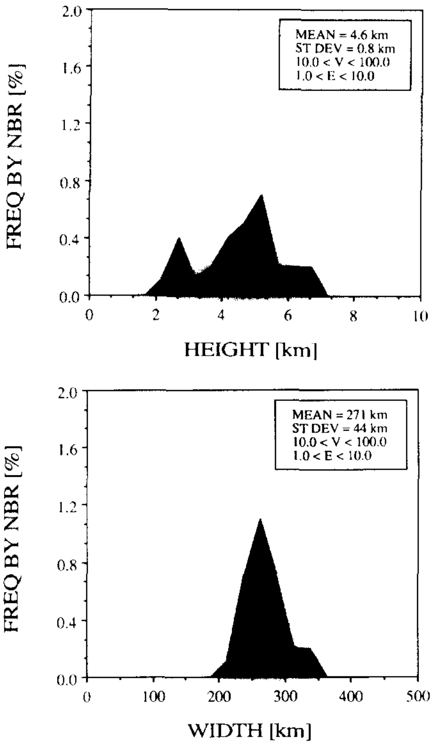

Fig. 8. Results of the Monte Carlo simulation for a wetter climate and wedge basal slope of 2-6°. (a) Predicted mean height of the wedge (with respect to the ground surface at its toe) for the Permian central Appalachians; (b) predicted steady-state width. The histogram gives the frequency of occurrence of height or width computed for various combinations of the parameters described in the text. V is convergence velocity (km Myr^{-1}) and E is the erosion constant (Myr^{-1}).

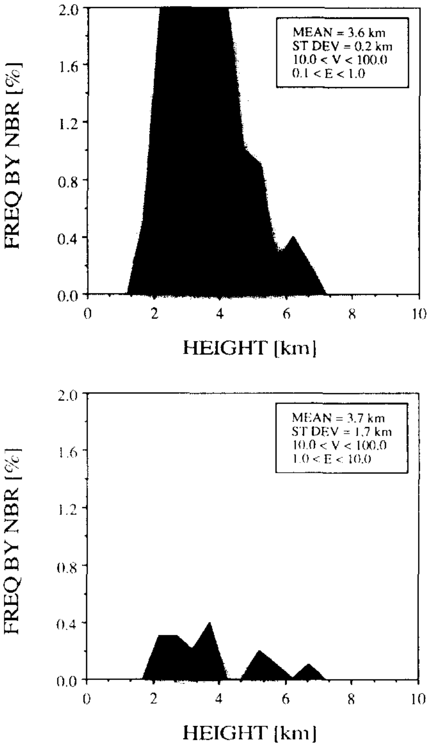

Fig. 9. Predicted mean height of the wedge (with respect to the ground surface at its toe) for the Permian central Appalachians for (a) a dry and (b) a wetter climate and wedge basal slope of 2-12°. The histogram gives the frequency of occurrence of height computed for various combinations of the parameters described in the text. V is convergence velocity (km Myr^{-1}) and E is the erosion constant (Myr^{-1}).

vertex of the wedge (Fig. 1) but from the elevation of point x_0 (Fig. 1). In the Permian the seas had withdrawn from the Appalachian basin, and therefore x_0 almost certainly was at or above sea level. Thus the heights are crude estimates of elevation or more precisely, estimates of relief. For comparison, the central Andes from 10° to 30°S comprise about a 300 km wide zone, all above 3 km and much above 4 km, with peaks rising to 5 km. Assuming a wetter climate (Fig. 8), does not vary the topography substantially because the wedge must still produce a load sufficient to flex the foreland basin. Allowing a steeper basal slope to the wedge (Fig. 9), only reduces the mean height from 4.3 to 3.6 km for a dry climate. Thus, with some confidence we can say that the Early Permian central Appalachians were similar to, or slightly less majestic than, the central Andes.

Although it confirms our well entrenched intuition (Rodgers, 1987; Slingerland and Beaumont, 1989; Cleaves, this volume), this conclusion is dependent upon so many premises that independent tests still seem warranted. The least demanding test is that eqn. (3) and the erosion constant used to calculate the steady state height should provide a subsequent rate of denudation consistent with observations. Fortunately, the Mesozoic basins of the central Atlantic states allow us to test this requirement. All available evidence indicates that the rocks immediately beneath the *unconformity* of the Newark basin in eastern Pennsylvania are identical to those adjacent to the basin margins (Manspeizer and Cousminer, 1988), an observation first made by Davis (1889). For example, the Furlong fault near Doylestown, Pa., fully 25 km into the basin, brings Cambro-Ordovician formations to the surface identical to those exposed on the north side of the basin. This indicates to us that the ancestral Appalachians must have been worn down to roughly their present stratigraphic level of erosion by earliest Carnian time (230 Myr ago using the DNAG time scale). Forty kilometers to the north in the Southern Anthracite Field, vitrinite reflectance and fluid inclusion data (Orkan and Voight, 1985; Levine, 1986) indicate that an additional 9–10 km of overburden lay atop Middle Pennsylvanian rocks presently exposed at the surface. Thus, if we assume that this was the overburden thickness at the end of thrusting, roughly 9–10 km of the Appalachians must have been eroded over the interval from 258 to 230 Myr.

Although it is not the point of this paper, this observation that the Appalachians experienced substantial pre-Triassic erosion, bears further discussion. Two problems are noteworthy: (1) where were the temporary and ultimate repositories of the erosional debris, and (2) if the central Appalachians reached their present degree of exhumation in Early Triassic time, what is the source of the siliciclastics on the present eastern United States passive margin? In answer to the first question, the paleo-lithological maps of Ronov et al. (1984) show shales of Late Permian age in the west-central United States. It is possible that a vast alluvial wedge of debris spread westward from the Appalachians, much as the present apron of the Colorado Rocky Mountains spreads eastward, only to be eroded at a later time by the early Mississippi River. In answer to the second question, the isopach maps of Poag and Sevon (this volume) indicate that 4 km of denudation of the Appalachians is necessary to account for the post-Early Jurassic sediment of the Atlantic passive margin. Roughly one-third to one-half of this comes from the Adirondack and New England terrains, leaving 2–3 km of denudation of the central Appalachians. Additionally, apatite fission-track thermochronology across Pennsylvania (Roden and Miller, this volume) indicates removal of only 5–6 km of overburden from Late Permian through Early Triassic. Thus, we can refine our estimate of pre-Carnian erosion to conclude that of the 9–10 km of overburden present in southeastern Pennsylvania in Early Permian time, roughly 6 km was removed prior to the Carnian, and the remaining 3 km removed subsequently. The pre-Carnian denudation yields an erosion rate of about 0.2 km Myr^{-1}. This can be compared to the erosion rate predicted from eqn. (3) for mountains about 4 km in average height (at the start of the interval) and an erosion constant of 01–1 Myr^{-1}. The erosion rates are similar and also in line with the higher rates for the post-Permian Appalachians presented in Table 1 of Sevon (this volume). Thus the model is at the least, internally consistent.

Another test of the model arises from the *P–T–t* histories of metamorphic and plutonic rocks in the orogen (Jamieson and Beaumont, 1988). Based on the closure temperatures of various mineralogie systems, they conclude that the southeastern Pennsylvania region experienced minimal uplift and exhumation during the Pennsylvanian. This changed during the Permian into significant net uplift relative to sea level with little exhumation, consistent with our postulate of mountain growth at that time.

Conclusions

Determining the ancient geometry of a mountain range requires knowledge of the mechanisms which formed the crustal thickening, the rate at which mass was eroded from the system, and the response of the lithosphere to the loading event itself. In the case of the ancestral Appalachians, the loading effect is reasonably constrained by the resulting size and shape of the Appalachian basin; the distribution of the load in space is less well constrained. By coupling a model of crustal response to thickening (given by the critical wedge theory) with a model of the basin evolution, we can determine not only the mass of the Appalachians but their shape as well. The technique indicates a central Andean topography with average relief on the order of 3.5–4.5 km and width of 250–300 km. Denudation rates generated by the high relief, and possibly Mesozoic extension, were sufficient to remove approximately 6 km of the orogen prior to Carnian time, suggesting that over half of the 10 km total of post orogenic unloading in southeastern Pennsylvania (Jamieson and Beaumont, 1988, Fig. 25) was accomplished early.

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