Geology

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Geology 1990;18;430-434
doi: 10.1130/0091-7613(1990)018<0430:SMOFBI>2.3.CO;2

Stratigraphic modeling of foreland basins: Interpreting thrust deformation and lithosphere rheology

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ABSTRACT

We incorporate the processes of erosion and deposition in a numerical model to predict the stratal geometries and facies patterns produced during episodic thrusting in a nonmarine foreland basin. The resultant stratigraphic record is characterized by a stairstepped facies package in which each retrogradation of facies (toward the thrust) marks the onset of a thrusting event. The retrogradation of facies coincides with the migration of the forebulge toward the thrust and the generation of an erosional unconformity. In the past, changes in basin wavelength during basin evolution have been interpreted to record viscoelastic relaxation of the lithosphere. This model suggests that changes in basin wavelength are a natural consequence of the interplay between thrust and sediment loading on an *elastic* lithosphere.

INTRODUCTION

Trends in sedimentary facies through space and time are often used to infer the structural history of a bounding mountain belt. For example, a traditional interpretation is that deformation results in uplift and erosion of the source area and the deposition of a rapidly prograding clastic wedge in the foreland basin. Thus, Armstrong and Oriel (1965) and Wiltschko and Dorr (1983) dated the times of movement of the Sevier thrust belt from the ages of conglomerates that were shed into the adjacent foreland basin. The opposite interpretation has also been proposed: during uplift, subsidence is assumed to be so rapid that coarse sediment is trapped adjacent to the thrust belt while fine-grained sediment is deposited in the rest of the basin; during quiescence, subsidence is slow and coarse strata prograde into the distal basin (Beck and Vondra, 1985; Blair and Bilodeau, 1988; Heller et al., 1988). Unfortunately, it is extremely difficult to test these interpretations rigorously because it is difficult to constrain the age of thrust movement independently.

Although this problem is difficult to study empirically, advances in our understanding of the mechanisms by which basins form and fill allow us to examine the problem theoretically. Here, we have integrated the processes of structural deformation, isostasy, erosion, and deposition in a quantitative basin model. In general this approach can be used to examine how particular variables (e.g., deformation, climate, rheology) control the character of strata. Herein we apply these techniques to study the evolution of a foreland basin during episodic thrusting.

MODEL

Foreland basins are asymmetric subsiding troughs that form adjacent to active fold and thrust belts. Price (1973) suggested a causal relation between the emplacement of thrusts and the subsidence of adjacent foreland basins. Since that time, a variety of studies have quantified the relation among thrust loading, lithospheric flexure, and foreland basin formation (e.g., Beaumont, 1981; Jordan, 1981). These models could predict the overall geometry, but not the complex stratigraphy, of the basin (e.g., unconformities, facies migrations, and lithology).

To overcome this limitation, we have taken the original modeling approach and incorporated a third component: the time-dependent erosion of mass from the uplifted thrust belt and subsequent deposition in the adjacent foreland basin. As in previous models, a crustal load, which is defined by a thrust geometry and a magnitude of shortening, is compensated by the lithosphere (Fig. 1A). In addition, we erode mass from the thrust belt and deposit it in the bounding sedimentary basin (Fig. 1B). Though we illustrate how these components are linked over one large-scale shortening event (Fig. 1), the model is actually iterated over short (10 000 yr) time intervals. Below we briefly review these components; see Flemings and Jordan (1989) for a more complete discussion of the modeling approach.

We model the structural development of a mountain belt as a fault-bend fold deforming over a crustal-scale ramp (Fig. 1A). The High Andes, the Himalayas, and the western United States are each inferred to have a major crustal-scale ramp along which shortening occurs (Molnar and Lyon-Caen, 1988). Similarly, zones of thick-skinned deformation (e.g., the Laramide of the western United States, the Tien Shen of China, and the Sierras Pampeanas of Argentina) are inferred to result from shortening on a crustal-scale ramp (Jordan and Allmendinger, 1986; Molnar and Lyon-Caen, 1988). Thus, this simple model has direct analogy to thick- and thin-skinned mountain belts.

We use an elastic rheology to describe the deformation of the lithosphere in response to loading. Both elastic (Jordan, 1981) and viscoelastic models (Beaumont, 1981) have been proposed, yet the degree of lithosphere relaxation that occurs on basin-forming time scales is controversial. Because this model proposes a new mechanism for changes in foreland basin geometry, the results can later be compared to the predictions of other rheological models.

We simulate erosion and deposition by assuming that mass redistribution follows a diffusive process in two dimensions. Diffusion is valid if the flux of sediment is proportional to the slope of the topography. In erosional regimes, the diffusion equation is used to describe the denudation of fault and shoreline scarps (Nash, 1980; Hanks et al., 1984), whereas in depositional settings it describes both fluvial and deltaic processes (Begin et al., 1981; Kenyon and Turcotte, 1985).

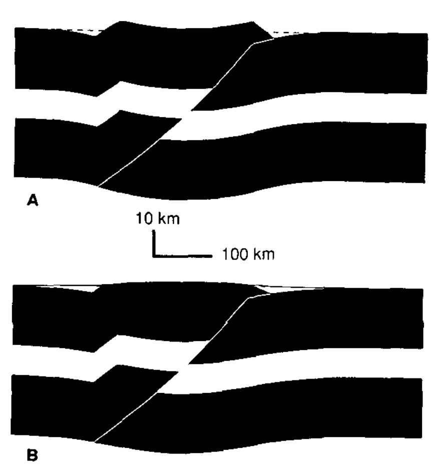

Figure 1. Linking three components of model. A: Crustal scale fault-bend fold after 40 km of shortening (ramp angle of 12° and total crustal thickness of 50 km) (modeled after Suppe, 1983), compensated by elastic lithosphere (effective elastic thickness = 25 km). B: Topography is eroded and mass redistributed from thrust belt to basin. Basin at right edge of fault-bend fold is also shown in Figure 3D.

PARAMETERS USED IN THE MODEL

The assumption of episodic deformation in thrust belts has been based largely on observed progradations and retrogradations of facies in bounding sedimentary basins (e.g., King, 1955). Because we wish to address the problem of episodicity it would be circular to assume it from the strata alone. Instead, information about the rate and duration of thrust movement must be based on crosscutting relations in the thrust belt. For example, in the Sevier belt of the western United States the average shortening rate was 1.4 mm/yr as measured over the 100 m.y. history of the thrust belt (Royse et al., 1975). In contrast, two particular thrusts within the overall Sevier belt shortened much more rapidly for short time intervals: the Meade thrust shortened at a rate of ~6.0 mm/yr over 12 m.y. (Royse et al., 1975), and the Prospect fault shortened at a rate between 5 and 10 mm/yr during an interval of 1 to 2 m.y. (Wiltschko and Dorr, 1983). Similarly, the Precordillera fold and thrust belt of Argentina has shortened at between 3 and 7 mm/yr for the past 10.5 m.y., yet it shortened at between 6 and 16 mm/yr for 2.7 m.y. within this period (Sarewitz, 1988).

On the basis of the summary above, our model has thrusting at 10 mm/yr between 0 and 2.0 m.y. and between 4.0 and 6.0 m.y., with quiescence from 2.0 to 4.0 m.y. and 6.0 to 8.0 m.y. For these simulations the lithosphere is assumed to be unbroken and to have an elastic thickness of 25 km, a value determined by previous modeling studies of retro-arc foreland basins (Jordan, 1981).

In the transport model, the efficiency with which mass is transported over a given topographic slope is proportional to the transport coefficient used in the solution of the diffusion equation. One transport coefficient must be constrained for the erosional regime of the mountain belt and another for the depositional regime of the bounding basin. The denudation rates of mountain belts are inferred to be proportional to relief on the basis of measurements of sediment load in rivers draining mountain belts (Fig. 2). Pinet and Souriau (1988) suggested that the denudation rate in actively deforming mountain belts is much higher than in old inactive mountain belts (Fig. 2). The modeled denudation rate for the crustal-scale fault-bend fold in Figure 1 also predicts that denudation rate is proportional to relief and that when no uplift is occurring, the denudation rate drops rapidly (Fig. 2). The transport coefficient of 1000 m^2/yr produces an erosional history in the range of the empirical results (Fig. 2). The resultant flux of sediment into the basin (10 to 100 m^2/yr) is comparable to observed fill rates of foreland basins (Flemings and Jordan, 1989).

We simulate fluvial transport using a transport coefficient of 10 000 m^2/yr, which is of the order of diffusivities used to describe fluvial and deltaic mass transport (Begin et al., 1981; Kenyon and Turcotte, 1985). These values predict depositional gradients within the range of those found in depositional settings (less than one degree).

RESULTS

The 8 m.y. basin simulation is presented at 2 m.y. intervals, illustrating both time lines and facies (Fig. 3, A–D); the time lines record isochronous surfaces at 1 m.y. intervals. Two facies are distinguished on the basis

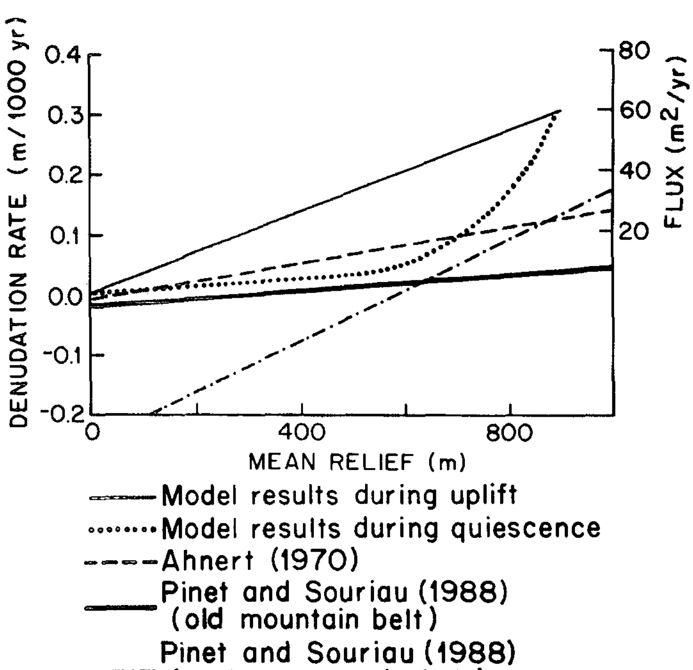

Figure 2. Relation between denudation rate and relief (mean elevation) in mountain belt modeled (see Fig. 1). For comparison, empirical relations between denudation rate and relief (mean elevation) are shown for two previous studies. Sediment flux to basin (shown on right) is denudation rate multiplied by width of mountain belt (~200 km).

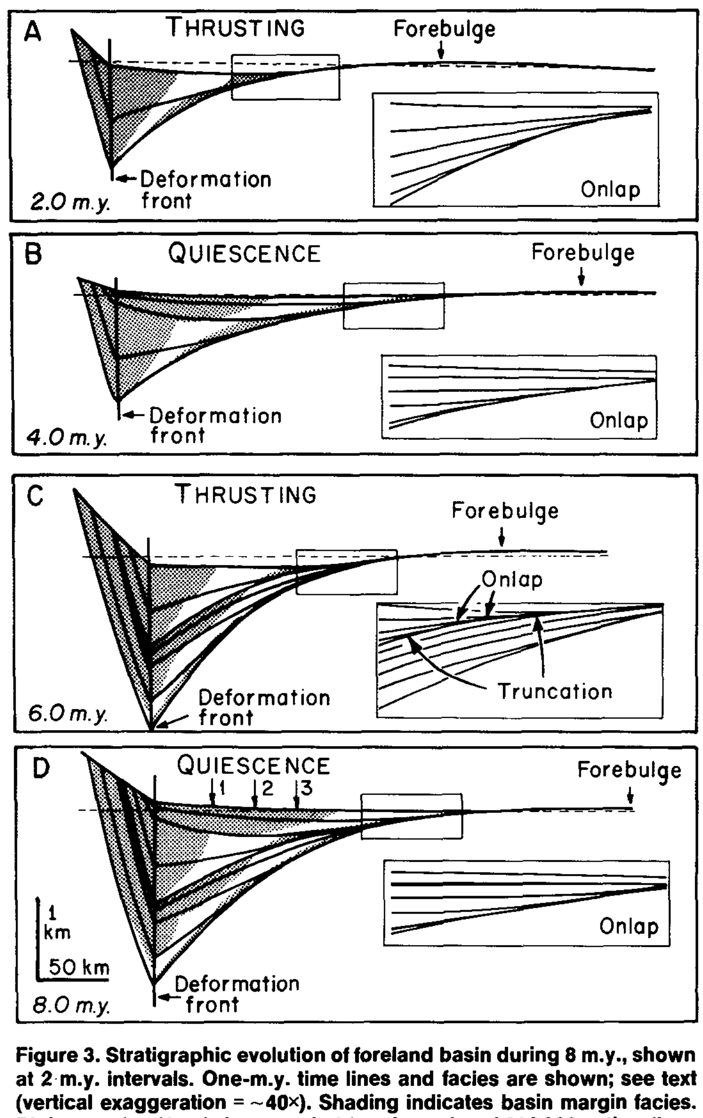

Figure 3. Stratigraphic evolution of foreland basin during 8 m.y., shown at 2 m.y. intervals. One-m.y. time lines and facies are shown; see text (vertical exaggeration = ~40×). Shading indicates basin margin facies. Right margin of basin is expanded (see boxes) and 200 000 yr time lines are used to illustrate onlap and truncation (vertical exaggeration = ~100×). Bedding on left has been uplifted and deformed by leading edge of fault-bend fold (deformation front migrates by 20 km during second phase of thrusting). Numbers in D locate accumulation curves plotted in Figure 5.

of depositional slope: the basin-margin facies was deposited at a gradient greater than 0.0005, whereas the basin-axis facies accumulated at a smaller gradient. The transport law describes only the transport of a single grain size, and we do not predict, quantitatively, textural changes in the depositional system (Gessler, 1971). Nevertheless, shear stress is proportional to gradient in the fluvial system that we model (Begin et al., 1981), and we expect it to be a primary control on the downstream sorting, by size, of sedimentary particles. Thus, the facies definitions are meant to differentiate qualitatively, but consistently, relative changes of depositional energy and sedimentary texture.

During the initial 2 m.y. of deformation, the time lines diverge toward the thrust, recording asymmetric subsidence in wedge-shaped stratal geometries that thicken in the direction of the thrust (Fig. 3A). This asymmetric subsidence results from lithospheric flexure due to crustal thickening in the thrust belt. As deformation proceeds, the relief in the thrust belt increases, which results in greater erosion and greater sediment supply to the foreland basin. As the sediment supply increases, and as the thrust propagates, successive wedges of strata onlap the distal basin margin (Fig. 3A). While the thrust propagates 20 km, the facies migrate approximately 50 km into the basin (Fig. 3A), migrating at more than twice the rate of thrust movement. The basin margin facies is largest volumetrically in the mountainward side of the basin, where it is derived from the thrust belt, yet it is also present in a narrow band on the right side of the basin, where it is derived from the uplifted forebulge.

After deformation ceases (2–4 m.y.), lens-shaped stratal packages are deposited (Fig. 3B): time lines are closely spaced proximal to the mountain belt, diverge in the basin center, and then onlap the forebulge on the right margin of the basin. These stratal packages are smaller in volume than those deposited in the previous thrusting phase, because the erosion rate drops off rapidly during quiescence (Fig. 2). Although the supply of sediment to the basin is lower, the basin continues to widen rapidly (Fig. 3B); the facies also continue to prograde (a total of almost 60 km). The topography of the forebulge is more subdued during quiescence and a smaller volume of basin-margin facies is derived from it (Fig. 3B).

During quiescence, the mountain belt continues to erode; this results in uplift as the lithosphere isostatically compensates for the removal of mass. Because there is no thrust-induced subsidence adjacent to the thrust belt, there is little accumulation in the basin proximal to the thrust (Fig. 3B); in a more extreme case of erosion in the mountain belt, an unconformity would be developed in the strata proximal to the belt. In this example, the sediment fills the topographic low between the thrust and forebulge that was created in the previous thrusting phase. As sediment fills this space, it further loads the lithosphere, amplifies the space that is present, and shifts the forebulge farther away from the thrust.

The transition from quiescence to renewed thrusting is marked by the abrupt shift from broad lenticular stratal packages (Fig. 3B) to wedge-shaped stratal packages that build outward from the thrust front (Fig. 3C). Rapid subsidence adjacent to the thrust traps the available sediment, and the concentration of loading at the thrust forces the forebulge to step toward the thrust. While wedge-shaped stratal packages are deposited adjacent to the thrust, older strata are uplifted and truncated by erosion of the migrating forebulge (Fig. 3C). As these wedge-shaped packages continue to fill the basin, the forebulge is forced outward again and strata onlap the underlying eroded horizon with angular unconformity (Fig. 3C).

When the basin narrows in response to renewed thrust loading, the facies step approximately 60 km toward the thrust (Fig. 3C). As a result, low-energy facies in the basin center overlie the high-energy facies, and in distal areas there is a drainage reversal (where previously the sediment was derived from the mountain belt, it is now derived from the forebulge). In addition, the forebulge is more sharply defined, and the volume of basin-margin facies derived from the bulge is consequently larger. As thrusting continues, facies migrate approximately 40 km in the direction of thrust movement.

The final phase, from 6 to 8 m.y., is structurally quiescent. No thrust-induced subsidence occurs, but as the mountain belt continues to erode, the basin widens and lens-shaped packages onlap the distal margin of the basin (Fig. 3D); the forebulge is progressively forced farther from the thrust, widening the basin. Facies prograde almost 100 km over this phase (Fig. 3D).

Viewed chronostratigraphically, the stratigraphic record of episodic thrusting is a stairstepped stratigraphic package bounded by unconformities (Fig. 4). Such a chronologic signature, where the majority of time is recorded in progradation and little time in retrogradation, is similar to observed chronostratigraphic signatures within clastic wedges in foreland basins (e.g., Swift et al., 1985). The stepping of facies toward the thrust occurs across a time-correlative surface and coincides with the onset of the formation of the unconformity in the distal part of the basin (Fig. 4).

These results may be understood intuitively to result from two assumptions: the lithosphere compensates elastically to loading, and erosion is a time-dependent function of topography. Because of these assumptions, when rejuvenated thrusting occurs there is initially more space (created by thrust-induced subsidence) than can be filled by sediment; as a result, the basin narrows, and facies step toward the thrust at the beginning of a thrust cycle. During continued thrusting and during quiescence, the supply of sediment outpaces the subsidence adjacent to the thrust, and facies prograde as the basin widens. The general stratigraphic signature produced is not a result of the particular erosional model used; it would result from any erosional model in which the erosion rate increases with relief.

STRATIGRAPHIC RECORD OF THRUST EVENTS

This model provides a quantitative interpretation for the origin of retrograding facies in foreland basins which has important implications, both for dating thrust movement and for interpreting lithosphere rheology. These results suggest that while, in general, sedimentary facies prograde in the direction of thrust movement, these progradations are punctuated by retrogradations (toward the thrust) at the onset of thrust cycles. Thus, in zones proximal to the thrust, rapidly fining-upward lithologies may record the onset of a thrust event, whereas in the more distal parts of the basin, an erosional unconformity will mark the onset of a thrust event. In contrast,

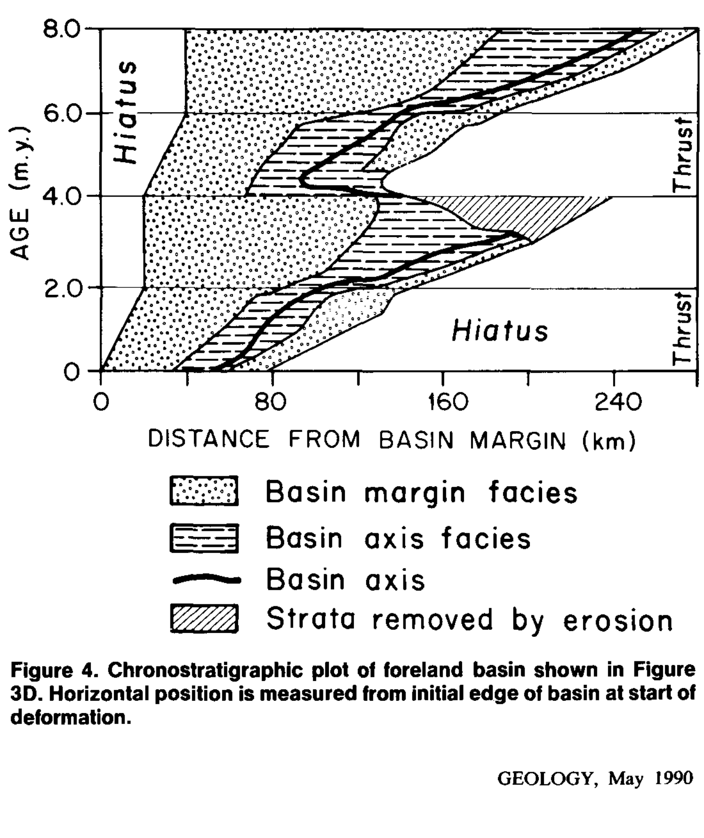

Figure 4. Chronostratigraphic plot of foreland basin shown in Figure 3D. Horizontal position is measured from initial edge of basin at start of deformation.

coarsening-upward lithologic sections may result from either continued thrusting or quiescence.

In general, these interpretations support recent work that has stressed the importance of thrust loading as a mechanism for generating retrograding facies (Beck and Vondra, 1985; Blair and Bilodeau, 1988; Heller et al., 1988). In particular, Heller et al. (1988) also predicted wedge-shaped stratal geometries and near-source trapping of coarse facies during deformation and the progradation of coarse facies in the form of lens-shaped stratal geometries after thrusting ceases. Heller et al. (1988) emphasized that during quiescence an erosional unconformity will form proximal to the thrust due to erosional unloading. In contrast, because we find that the proximal unconformity is not always formed and does not extend far into the basin, we stress the larger and more consistent unconformity that is formed in the distal basin.

It should be emphasized that some recent work supports the traditional interpretation that thrust deformation results in the progradation of facies and that retrograding facies record times of quiescence. For example, Burbank et al. (1988) found that conglomerates are deposited in a basin throughout a thrust cycle. This suggests that in certain basins the rate at which coarse sediment is supplied may be so high that it overwhelms thrust-induced subsidence; as a result, gravels prograde throughout a thrust cycle. Even so, we would predict preferential extraction of some of the coarse-grained strata proximal to the thrust during thrust-induced subsidence (e.g., Paola, 1988). In some situations we would not expect a retrogradation of facies toward the thrust. For example, if the rate of thrust propagation were extremely rapid relative to the amount of thrust induced subsidence (due to a very low angle thrust ramp), there might be insufficient subsidence to shift the facies toward the thrust. Perhaps most important, if renewed deformation results in a new thrust that steps basinward, this step may overwhelm any shift of facies due to thrust loading.

In light of the general disagreement over how to interpret facies migrations, many workers concentrated instead on the accumulation history of the basin. These workers inferred the onset of thrusting from an increase in the rate of accumulation of strata, which they interpret to record thrust-induced subsidence (e.g., Heller et al., 1986). However, the modeling presented here suggests that the accumulation history varies strongly as a function of the distance from the thrust front (Fig. 5). An increase in the accumulation rate will coincide with a thrusting event only very proximal to the thrust (Fig. 5); in fact, as the accumulation rate is measured at successively more distal positions in the basin, the increases in accumulation rate are progressively more out of phase with the time of a thrust event. In the most distal section, the times of thrusting coincide with times of lowest accumulation rate (Fig. 5).

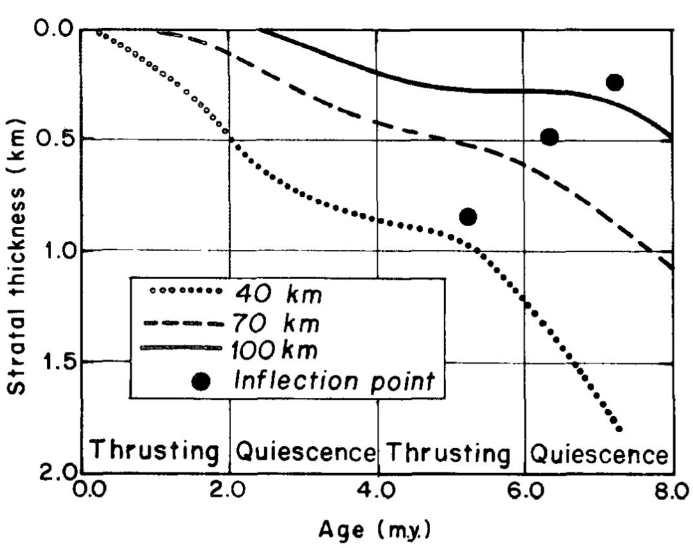

Figure 5. Stratal thickness vs. time for modeled foreland basin. Positions of section are located in Figure 3D; position is measured from initial edge of basin at start of deformation.

INTERPRETING LITHOSPHERE RHEOLOGY

In the past, changes in basin wavelength have been one of the primary pieces of evidence that the lithosphere relaxes over basin-forming time scales. Quinlan and Beaumont (1984) and Beaumont et al. (1988) argued that lithospheric stress relaxation was the dominant control on migration of facies and forebulge over orogenic time scales (>50 m.y.); they inferred changes in basin wavelength to record the interplay between periodic crustal deformation and lithospheric relaxation. Figure 6 is an attempt to compare this model with that of Quinlan and Beaumont (1984); however, in Quinlan and Beaumont's (1984) model loading is instantaneous, whereas in our model loading occurs over a period of time. In their model, during thrust loading the lithosphere is assumed to be stiff, the basin is broad and shallow, and the distance between thrust and forebulge is large (Fig. 6A). During quiescence, relaxation of the lithosphere causes subsidence proximal to the thrust, migration of the forebulge toward the thrust, and narrowing of the basin (Fig. 6B). During a subsequent thrust event, the lithosphere is again stiff and the forebulge migrates away from the thrust (Fig. 6C).

Quinlan and Beaumont (1984) and Beaumont et al. (1988) used this model to infer the time and magnitude of the loads necessary to result in the Appalachian basin stratigraphy. Subsequently, Tankard (1986) found detailed evidence of forebulge migration and used this as evidence of a viscoelastic lithosphere. Similarly, Schedl and Wiltschko (1983) used a

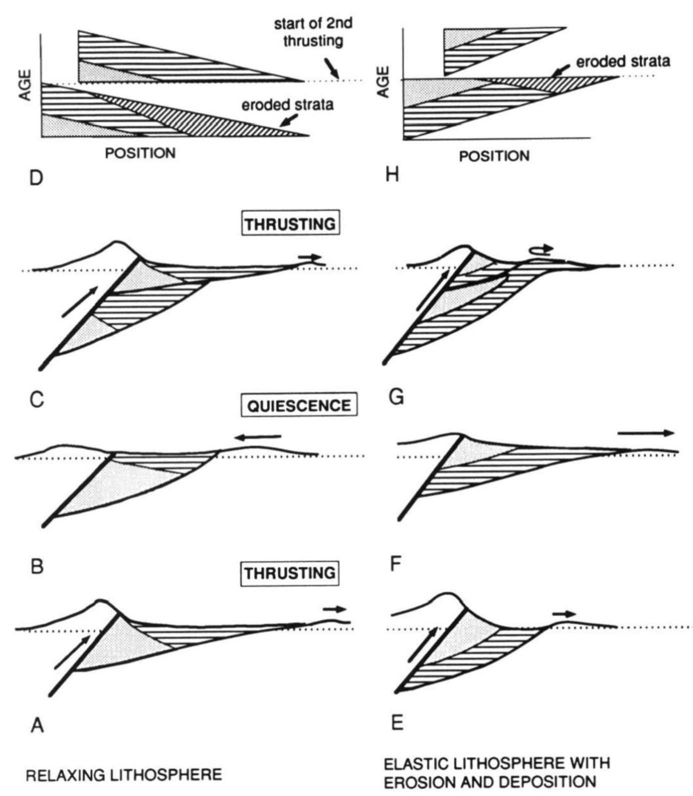

Figure 6. Stratigraphic response in sequential cross sections (A, B, C) predicted by Quinlan and Beaumont (1984, Fig. 18). D is chronostratigraphic plot of C (Quinlan and Beaumont, 1984, Fig. 19). Stratigraphic response predicted by this study in sequential cross sections (E, F, G); H is chronostratigraphic plot of G. Dotted pattern = coarse-grained, proximal deposits; horizontal lines = fine-grained distal deposits; diagonal bars (D and H) = strata that have been eroded.

viscoelastic lithosphere to model facies migrations and applied the model at intraorogenic time scales (<10 m.y.).

In contrast to this interpretation, our model predicts sweeping changes in basin wavelength during episodic thrusting on an elastic lithosphere. Furthermore, we present an interpretation of the timing of thrust deformation opposite to the relaxing lithosphere: the basin narrows and the forebulge and facies migrate toward the thrust at the onset of deformation (Fig. 6G); the basin widens and the forebulge and facies migrate away from the thrust both during continued deformation and after deformation ceases (Fig. 6, E and F). Our result is presented at a shorter time scale than that of Quinlan and Beaumont (1984); however, on an elastic lithosphere, the general response will occur regardless of the time scale (migration of forebulge and facies toward the thrust at the onset of thrusting). The central question becomes which of these processes (lithospheric relaxation or erosion and redistribution of sediment on an elastic lithosphere) dominates foreland basin evolution.

Figure 6 provides hypotheses that can be tested against field observations. The ideal test would be to examine a basin in which the timing of thrust migration is constrained independently (e.g., by crosscutting relations). Even without such direct evidence, the different models predict distinct stratigraphic architectures. In the relaxation model, the erosional unconformity is cut during quiescence (Fig. 6D) and retrogradation of facies occurs gradually while the unconformity is being cut. In contrast, in our model the unconformity is cut and facies migrate rapidly toward the thrust at the onset of a thrust cycle, while facies migrate gradually outward during both continued thrusting and quiescence (Fig. 6H). However, until time-dependent loading and sedimentation are incorporated into the Quinlan and Beaumont (1984) approach, these direct comparisons must be viewed with caution. The solution, if it can be resolved, will come from both continued stratigraphic modeling and detailed chronostratigraphic and lithostratigraphic correlation in a variety of foreland basins.

CONCLUSIONS

By linking time-dependent erosion and deposition to foreland flexural models, we have explored the stratigraphic response to episodic thrust deformation in a nonmarine foreland basin. The model predicts that thrusting is recorded by an overall progradation of lithofacies, yet this progradation is punctuated by retrogradation of facies (toward the thrust) at the onset of a thrust cycle. These facies shifts coincide with the migration of the forebulge toward the thrust and the generation of an erosional unconformity. Because the model predicts changes in basin wavelength on a nonrelaxing lithosphere, the question of whether or not the lithosphere relaxes significantly on basin-forming time scales must be reexamined.

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ACKNOWLEDGMENTS

This project is supported by the Shell Oil Company Foundation and the Donors to the Petroleum Research Fund of the American Chemical Society (18887-AC2). R. Allmendinger, C. Beaumont, J. Beer, D. Sarewitz, and an anonymous reviewer critically reviewed the manuscript. We thank C. Beaumont for constructive discussions concerning future tests of the stratigraphic models. Institute for the Study of Continents (INSTOC) Contribution 118.

Manuscript received May 25, 1989
Revised manuscript received November 6, 1989
Manuscript accepted November 20, 1989