7. INTRODUCTION TO MODELING BASINAL STRATIGRAPHY: THE INTERPLAY BETWEEN SEDIMENT SUPPLY, SUBSIDENCE, AND SEA LEVEL

A. INTRODUCTION

Karl Popper once wrote, "Out of theories we create a world: not the real world, but our own nets in which we try to catch the real world." One of the more exciting applications of quantitative analysis of basin development is its use as a tool in determining the relative importance of tectonic, eustatic and climatic effects on the development of basinal stratigraphy. The goal in modeling the development of basinal stratigraphy is not to try to explain every detail observed in the real world, but instead serves as a framework from which we can view the real world. As such, simplified basin-filling models can guide us to look for those critical field relations that can be used to distinguish between the fundamental factors governing the formation of the basinal stratigraphy.

Geologists tend to think of certain factors as the primary controls on the development of basin fills, including: source area uplift rate, source area lithology, rain fall, temperature regime, sea level changes, and basin subsidence rates. In contrast, basin models can address only the most basic controls, such as: basin subsidence, eustatics, volume of sediment supply (flux), sorting of the sediment supply, and rates of sediment transport. The problem is that the relationship between the geologists' factors and the modeling parameters are far from straight forward. For example, changes in rainfall might affect sorting, flux and transportability of the sediment supplied to a basin. Changes in source area lithology can affect all of the same modeling factors. Because of these complex interactions it is difficult to go from the model results back to the primary geologic controls. Nonetheless, models serve a useful function in guiding geologic interpretations, pointing out critical relationships that might not otherwise have been considered, and helping to decipher the relative importance of fundamental basinal controls. As such, basin modeling is a powerful tool in the analysis of sedimentary basins.

In this chapter we focus on the importance of basin subsidence as a major control on basinal stratigraphy and the use of subsidence analysis as a tool to separate tectonic effects from regional sea-level controls when interpreting the stratigraphic record. It has been known for years that the primary controls on deposition of stratigraphic sequences are the rates of sediment flux, subsidence and changes of sea level (e.g., Sloss, 1962). However, relatively few studies have successfully separated the effects of these controls when interpreting the rock record. Usually sedimentary sequences are interpreted based on the particular bias of the worker - either emphasizing tectonics or eustatics. This uncertainty frustrates attempts to sort out tectonics from sea-level effects which must be done if records of tectonic timing and global eustasy are ever to be established.

Clearly, techniques must be developed to evaluate the interplay of sedimentation, subsidence and sea level in the rock record. One approach that we advocate here is the integration of regional stratigraphic information with results predicted by theoretical modeling. In this and the remaining chapters we will discuss some examples of how these factors affect the stratigraphic record and how each factor may be distinguished from the others. In this chapter most of our examples come from foreland basins, particularly the U.S. western interior adjacent to the Sevier belt, but the basic concepts can be applied to other basin types.

B. THE INTERACTION OF SEDIMENTATION, SUBSIDENCE AND SEA LEVEL

The western part of the Cretaceous interior seaway represents a foreland basin that formed adjacent to the uplifting Sevier thrust belt (Fig. 7.1; Jordan, 1981). Sedimentation in the basin is controlled by three primary factors: sediment input - primarily resulting from the uplift of thrust belt source areas; subsidence of the basin - mostly driven by emplacement of the thrust belt and associated flexure of the lithosphere; and changes in regional sea level which may affect the entire basin. Note that it is the *rates* at which these factors act, and not their magnitude, that exerts the greatest influence on the development of stratigraphic sequences. For example, at one time emphasis was placed on the magnitude of eustatic highstands and lowstands as causes of sedimentary sequence boundaries (e.g., Vail et al., 1977), but following the work of

Pitman (1978) these boundaries are now recognized to be controlled by the *rate of change* of sea level (Haq et al., 1987).

The influence of these controls on the formation of depositional packages is diagramed in Figure 7.2. To develop a prograding depositional wedge you need to decrease water depth in the vicinity of the shoreline. This can be done by either increasing the rate of sediment flux, decreasing the subsidence rate or increasing the rate of sea-level fall. This relationship is shown by:

$$R_{shoreline} = \frac{R_{sed} - R_{sub} + R_{sl}}{slope} \tag{7.1}$$

where $R_{shoreline}$ is the rate of shoreline progradation, R_{sed} is the sedimentation rate at the shoreline, R_{sub} the local subsidence rate and R_{sl} is the rate of sea-level fall and *slope* is the slope of the basin floor. From this equation you can see that a change in any of these factors will exert an equal influence on the ability of a depositional wedge to prograde out into a basin. Therefore, it may be difficult to know if a transgressive or regressive event is controlled by any one or some combination of these mechanisms.

In most tectonic settings all three major factors may be changing simultaneously, therefore it is surprising that studies of some basins suggest that transgressive-regressive cycles are synchronous throughout the basin. For example, at one time Kauffman (1977) showed 10 transgressive and regressive cycles (labeled as T and R on Fig. 7.3a) spanning Cretaceous time in the Western Interior. If this interpretation is correct it suggests that either eustatic changes dominate the Cretaceous section, even adjacent to the thrust belt, or that the balance between subsidence, sedimentation and sea-level was nearly identical along the entire seaway, both of which seem unlikely in an active tectonic setting. In light of this it is reassuring that other studies suggest that these events are not synchronous. For example, Lillegraven and Ostresh (1990; Fig. 7.3b) show the change in shoreline position during the Bearpaw transgression (T9 of Fig. 7.3a). Although a transgression occurred to the north in Montana and southern Canada during this time interval, a regression took place at apparently the same time farther south in

Colorado and adjacent areas. This nonsynchroneity implies that changes in regional sea level were not the only factors controlling the transgressive and regressive record during this time, but that changes in sediment supply and subsidence had major influences as well, as might be expected during a time of active tectonism in the thrust belt.

In the following sections, we will illustrate equation 7.1 by showing several examples of (1) how similar looking stratigraphy may be produced by varying the three primary controlling factors; (2) how basin filling sequences respond to tectonic events via the interaction of the controlling factors; and (3) how tectonics may be distinguished from sea level by looking at the geometry of preserved stratigraphic sequences. Again, our examples will come from foreland basin settings, but the principles apply to all basin settings. In this discussion we will refer to *regional sea-level* changes, which are changes in sea level seen on the scale of continents or large parts of continents versus *eustasy*, which are changes in sea level that are truly global and synchronous.

C. SUBSIDENCE VERSUS SEA LEVEL AS A CONTROL ON SEQUENCE STRATIGRAPHY

Following equation 7.1, nearly identical stratigraphic sequences can be produced by varying subsidence rate or rate of sea-level change or both. We will look at an example of sequence stratigraphy along the distal margin of a foreland basin. Flexural models for foreland basins show that times of tectonic loading are correlated with times of basin subsidence (Beaumont, 1981; Heller et al., 1986; Jordan, 1981). Hence, the distal margins of foreland basins record flexural subsidence, but can also be influenced by changes in regional sea level (e.g., Klein and Willard, 1989). If thrust loading is intermittent, subsidence in the distal part of the foreland basin will be episodic, as the thrust belt moves, stops and then moves again. This interplay between episodic subsidence and sea level may make sequence stratigraphy very difficult to interpret.

An example from the Permian Basin of west Texas (Sarg and Lehmann, 1986) illustrates this difficulty. In the cross section of the basin (Fig. 7.4a) a thrust belt lays off to the right and so only the distal margin of the foreland basin is shown. Concentrating on deposition during the

Guadalupian time interval, the familiar stratigraphic signatures of onlap and downlap associated with relative sea-level changes can be seen (Fig. 7.4). Typically onlap is interpreted to result from a relative sea-level rise and downlap from a relative sea-level fall (Vail et al., 1977). The cause of this relative rise and fall can be either local, that is the result of local variations in sediment supply and/or subsidence; regional, that is related to vertical continental movements unrelated to local tectonic events; or caused by global changes in eustatic sea level.

Figure 7.5a is a synthetic stratigraphic cross-section that demonstrates how the sequence can be produced by simply varying the subsidence rate associated with episodic thrust movement and keeping the sediment flux constant. This cross section shows the synthetic stratigraphy along the distal margin of a foreland basin with the thrust belt off to the right. In this model the thrust belt is always moving basinward, so that subsidence is continuous and migrates to the left over time. This generates coastal onlapping for all five units shown (Fig. 7.5A). The rate of subsidence increases while units 1,2 and 3 are deposited, causing a shift in depocenters towards the coast, to the left. Subsequently the rate of subsidence decreases slightly during deposition of units 4 and 5 producing a downlap surface (DLS on Fig. 7.5b). The resultant time-space diagram (Fig. 7.5b) are similar in appearance to that seen in the Guadalupian interval of the Permian Basin (Fig. 7.4b) and is similar to those interpreted as sea-level changes elsewhere (Vail et al., 1977), specifically a period of onlapping with an offshore unconformity, followed by a period of downlapping. Although this sequence does indeed represent a relative change in local sea level (ie. the water depth does change), here it is caused by slight changes in subsidence rates and not by changes in eustatic sea level.

D. TWO-PHASE MODEL FOR FORELAND BASIN STRATIGRAPHIC SEQUENCES

Subsidence can, and often does, exert as much a control on basin filling as does sediment supply. If we focus on wholly nonmarine basins for a moment, so that sea level effects are excluded, then many studies (e.g., (Rust and Koster, 1984; Steel et al., 1977) refer to gravels that prograde in tectonic settings as "synorogenic", that is they result from an increase in sediment supply generated by active tectonism. However, if active tectonism produces an increase in basin subsidence rates, then more space is created to trap the sediment (accommodation space) and so the clastic facies tend to *aggrade* instead of *prograde*. Hence, there is a paradox with "synorogenic" conglomerate - there is more conglomerate supplied during active tectonism which promotes progradation, but there is commonly an increase in basin subsidence to trap that sediment which tends to inhibit progradation. Increased subsidence can occur in concert with tectonic activity not only in foreland basins but in any basin type.

This concept is developed in greater detail in Chapter 9. Here we focus on an example of this concept that looks at two end member facies tracts that might develop during basin subsidence and subsequent erosion. Specifically we suggest a two-phase depositional model for foreland basin facies tracts that is rooted in the concept that subsidence exerts a major control on regional depositional systems (Heller et al., 1988).

1. Introduction

As we have discussed earlier in this course, foreland basins are asymmetric structural depressions that form due to flexure of the lithosphere under the bounding thrust belt and associated tectonic loads. The adjacent thrust belt serves as the primary source for basin-filling sediments. The depth and width of the basin depend on the mass and geometry of the adjacent thrust belt, attendant subsurface and deposited sediment loads, and the flexural rigidity of the lithosphere beneath the foreland basin.

A corollary of the flexural model of foreland basin development is that flexural uplift will occur as the load is removed. For example, if the thrust belt is eroded after emplacement, the basin will rebound slightly to reestablish regional isostatic balance. Sediments previously deposited in the foreland basin will also be uplifted and may begin to be eroded. Continued erosion will further reduce the size of the load leading to more isostatic uplift, and so on, thereby reducing the thickness of the foreland basin fill. If the thrust belt is the sole tectonic load, the basin may eventually be eroded away, leaving only a regional unconformity as evidence that it once existed. This example, although extreme, suggests that if movement of a thrust belt is discontinuous, the adjacent foreland will undergo two phases of basin development. One phase of sedimentation (synorogenic) is associated with rapid tectonic subsidence of the basin during load emplacement in the adjacent thrust belt. A succeeding phase of basin evolution (post-orogenic) occurs when erosion of the thrust belt dominates and there is flexural rebound in the proximal part of the foreland basin, leading to redistribution of deposits across the basin. These contrasting phases of foreland-basin development each reflects the interplay between sediment supply and subsidence and are discussed separately.

Coarsening-upward conglomeratic sequences are potentially important sedimentary indicators of tectonic events, especially in fault-bounded basins (Steel et al., 1977; Van Houten 1974). Although grain-size distribution supplied to a basin is controlled by many factors, including type and relative abundance of rock types in the source area, source-area relief, climate, vegetation, and sea level (Folk, 1974), it is the relationship between grain size and source-area relief that is emphasized in the tectonic interpretation of conglomeratic sequences (e.g., Rust and Koster, 1984). Thus, orogenic events in thrust belts have been dated by the age of conglomeratic units in the adjacent foreland basin (e.g., Wiltschko and Dorr, 1983). Nonetheless, times of active thrusting are also times of rapid basin subsidence, which competes with sedimentation rate in the development of foreland basin sequences (Heller et al., 1989).

2. *Basin Response to Tectonic Loading (Synorogenic Phase)*

Flexural models indicate that subsidence of a foreland basin is commensurate with thrust emplacement and that the most rapid subsidence occurs beneath and immediately adjacent to the advancing thrust sheet and diminishes exponentially away from the load (Beaumont, 1981; Jordan, 1981). Deposition of sediment derived from the uplifted thrust sheet tends to follow the subsidence pattern, the thickest accumulation occurring immediately adjacent to the thrust front (Fig. 7.6a). There is considerable evidence that most of the down-stream fining in aggrading alluvial systems is due to selective deposition of the least transportable (usually the coarsest) clasts (cf. Shaw and Kellerhals, 1982). Paola (1988) has used this connection between subsidence-induced sedimentation and down-stream fining to model the cross-sectional geometry of gravel bodies in alluvial basins. According to this model (Chapter 9), rapid proximal subsidence, such as occurs in foreland basins, should produce thick but areally restricted gravel bodies, because most of the gravel supplied will be selectively deposited near the source (Fig. 7.6a). For example, in Death Valley, a rapidly subsiding extensional basin, coarse grains dominate along the mountain front and grade rapidly into very fine-grained deposits a few kilometers away on the valley floor (Denny, 1965). In fact, the fans along the east side of Death Valley are generally more aerially restricted than those along the west side, reflecting the greater subsidence rates along the east valley margin.

In nonmarine foreland basins, the distribution of coarse-grained deposits is affected by the presence of large lakes or trunk streams that flow parallel to the thrust belt (Miall, 1978). This drainage pattern reflects asymmetric basin subsidence. Alluvial progradation from the thrust belt across the basin is inhibited by flexurally induced back-tilting of the lithosphere beneath the basin which deepens toward the thrust belt (Fig. 7.6a). The position of the lakes or longitudinal trunk streams is controlled by the balance between the rate of subsidence, which tends to force these features towards the adjacent thrust load, and the rate of sediment supply which tends to force alluvial progradation out across the basin. Meanwhile, streams derived from beyond the flexural wave length of the basin may flow toward the subsiding foreland region, resulting in fluvial units in the distal part of the basin which prograde toward, and not away from, the thrust belt (Fig. 7.6a).

The foregoing model of nonmarine foreland-basin sedimentation (Fig. 7.6a) suggests that during times of tectonism and uplift in the thrust belt, coarse alluvial deposits accumulate only in the most proximal part of the basin and grade rapidly into fine-grained deposits toward the distal part of the basin. The rate of grain-size diminution will depend primarily on the distribution of subsidence rate in the basin and the grain-size distribution of sediment supplied to the basin. Farther out in the basin, deposition will be primarily fine-grained sediment of fluvial or lacustrine origin; fluvial systems in the distal part of the basin may flow toward the thrust belt down the flexurally-tilted distal basin margin or off the forebulge area, which extends many tens of kilometers beyond the distal edge of the basin.

An example of this sedimentary model associated with times of load emplacement in the adjacent thrust belt (the synorogenic phase) is provided by deposits of Aptian (mid-Cretaceous) age in western Wyoming. The development of the western margin of the Cretaceous seaway in the western interior is the flexural result of thrust-load emplacement in the Sevier orogenic belt to the west (Jordan, 1981). The Aptian units extending from the Idaho-Wyoming sector of the thrust belt to the eastern side of the Bighorn basin are correlated and shown in cross section (Fig. 7.7). In this cross section the Greybull Sandstone interval consists of quartz arenites that were deposited in streams that flowed westward (Kvale and Vondra, 1985; Mackenzie and Ryan, 1962) or northwestward (Gustason and Ryer, 1985) toward the thrust belt. The position of the Greybull Sandstone along the distal part of the foreland basin (Fig. 7.7), its correlation with fine-grained lacustrine rocks closer to the thrust belt, and the rapid thickening of these deposits towards the adjacent thrust belt agree with the predicted stratigraphy associated with rapid foreland-basin subsidence during the orogenic phase of thrust-belt development. Comparison of our reconstructed basin geometry with the results of simple elastic flexural models by using the palinspastically reconstructed thickness of the Paris thrust sheet (Royse et al., 1975) shows that the basin geometry is compatible with a lithospheric flexure model (Fig. 7.7).

3. Basin Response to Thrust Belt Erosion (Post-orogenic Phase)

A second phase of sedimentation takes place after a thrust belt is emplaced and erosional processes dominate over uplift and advance of the thrust sheet. Isostatic rebound associated with erosion of the thrust belt will lead to flexural uplift of the adjacent foreland basin. The amount of uplift will decrease away from the thrust belt, but over time more distal parts of the basin will become involved. As the thrust belt rebounds, sediments shed from the thrust sheets will be deposited farther out across the basin. At the same time sediments previously deposited in the proximal part of the foreland basin are also uplifted and eroded, transported more distally across the foreland region and redeposited (Fig.

7.6b). Hence, the sedimentary record predicted for times when erosion predominates in the thrust belt consists of relatively coarse-grain deposits extending across the more distal finer-grained parts of the foreland basin and beyond. Moreover, these coarse deposits would correlate with erosional surfaces in the proximal part of the basin and in the thrust belt (Fig. 7.6b). As uplift of the foreland basin continues, the coarse proximal deposits will continue to be eroded and redistributed out across the foreland region. As the proximal part of the foreland basin is uplifted during this second phase, the basin shape becomes less asymmetric and the reworked coarse deposits will tend to have a sheetlike geometry (Fig. 7.6b) rather than forming an asymmetric wedge as they do during the orogenic phase of deposition (Fig. 7.6a). Thus, the characteristic signature of the post-orogenic phase of sedimentation is widespread gravel sheets that correlate with unconformities near the eroded thrust belt.

An example of this style of deposition comes from the late Cenozoic record of the western Great Plains. Uplift of the southern Rocky Mountains in late Miocene and Pliocene time resulted in the deposition of fluvial sands and gravels of the Ogallala Formation and related units (Scott, 1975; Stanley and Wayne, 1972). In Colorado and New Mexico, these units has largely been eroded away within 150 km of the Rocky Mountain front, exposing Cretaceous and early Cenozoic sedimentary units (Fig. 7.8). Deposition of sediments with a Precambrian basement provenance (in part, reworked Ogallala Formation) continued into Pleistocene time in Kansas and Nebraska (Scott, 1975; Stanley and Wayne, 1972). The uplift and redistribution of Ogallala deposits has not been caused solely by erosion of the Rocky Mountains, but probably by lithospheric thinning under the Rockies and western Great Plains (Angevine and Flanagan, 1987; Eaton, 1986) as well. Nonetheless, a similar stratigraphic record would be generated by erosion alone, except at a much slower rate.

4. Discussion

These simple models suggest that, in the proximal part of a basin, coarsening-upward sequences consisting primarily of sandstone and conglomerate accurately record the initial uplift and advance of thrust sheets. In contrast, in the more distal parts of the basin, active loading in the thrust belt is recorded by the fine-grained lower parts of the coarsening-upward sequences. The coarse-grained upper parts of the distal sequences, instead, represent periods when thrusting in the source area has ceased and/or has been overwhelmed by erosion. In this context, the proximal sequence records the onset and climax of thrusting, whereas the distal sequence represents the waning and cessation of thrusting and the attendant increasing influence of erosion in the source area (Fig. 7.9).

Our discussion has not included specific durations and areal extent over which these two phases of sedimentation occur, because specific sedimentary response is a function of several factors that vary over time and space within a thrust belt as well as between thrust belts. Factors such as uplift and erosion rates control the rate and style of sedimentary responses modelled here. The two phases of basin formation may occur as distinct phases, even though erosion takes place as the thrust belt is being emplaced. During the time that the thrust belt is primarily advancing and rising, thereby loading the lithosphere, the basin subsides rapidly and the orogenic model of sedimentation dominates, even though some erosion occurs. The post-orogenic phase dominates when erosion rate overwhelms the uplift rate on the advancing thrust, so that the magnitude of the tectonic load decreases. Many factors are involved in these rates - episodic movement on the thrust fault, emplacement rate and geometry of the fault, flexural rigidity, climatic fluctuations, and lithologic and topographic variations. As a result it is likely that alternation between the two phases will occur many times during a single orogenic event as the threshold between uplift and erosion rates is crossed. Hence, the final record may be more complicated than the models shown here, with many small unconformities in the proximal basin and many thin gravel sheets prograding into the distal part of the basin. The specific wavelength over which the different sedimentary responses occur is also controlled by many variables: the size and distribution of the thrust load, erosion rates, the grain-size distribution and transport mechanics of the erosional products, and the flexural rigidity of the lithosphere. These factors may be difficult to determine and can vary along a thrust system as well as over time. In addition, if new thrust sheets are emplaced before older thrust plates are eroded, then the older sedimentary record may be buried by the new foreland depositional sequence before the post-orogenic phase occurs. Our understanding of long-term sediment production and transport is not sufficient to allow us to quantify the terms "proximal" and "distal" basin deposits as used here.

Nonetheless, a basic principle is illustrated by this example. The effects of both sediment supply and basin subsidence rates must be taken into consideration in order to accurately interpret the significance of the stratigraphic record. In this simplified model other potentially important variables including the rate of sediment supply, the sorting of the sediment supply and the rate transport of the sediment across the basin are not allowed to vary. This model for sedimentation can be made much more complex if these variables as well as sea-level changes are considered. In Chapter 9, changes in these other variables are discussed. Next, we will describe some techniques that might be used to differentiate between regional sea level and tectonic subsidence as controls on basin-filling sequences.

E. SEPARATING TECTONIC FROM REGIONAL SEA-LEVEL EVENTS IN FORELAND BASIN SEQUENCES

One technique that might be used to discriminate between tectonics and eustatics as controls on stratigraphic sequences concerns the origin of foreland basin formation and the geometry of basin-filling deposits (Heller et al., 1987). In order to do this we assume that as a thrust belt advances towards the center of a foreland basin, any given spot out in the basin will record increasing subsidence as the tectonic load moves closer. The subsidence rate will decrease once the site is overrun by the thrust sheet or the thrust stops moving. An idealized subsidence curve for foreland basins that illustrates this behavior is shown as Figure 7.10. During each thrust advance, the rate of subsidence at a given spot in the basin increases and then reduces as thrusting ceases. This ideally produces a convex-up profile to the subsidence curve characteristic for foreland basins (cf. Fig. 6.1).

An example from the central Utah sector of the thrust belt (Lawton, 1986) shows that several deltaic sequences prograded into the western interior basin during Late Cretaceous time (Fig. 7.11). Prograding deltaic cycles such as these are fairly common in shallow marine foreland basins

(Dickinson, 1976). Potentially these sedimentary wedges that fill the basin will also preserve the asymmetric geometry of the subsiding foreland basin if they are deposited during a time of major thrust loading. Thus we predict that by examining the geometry of these prograding sequences the relative importance of regional sea-level changes versus tectonic subsidence can be documented (Fig. 7.12, note the thrust belt lies to the west in these diagrams). Tectonically-induced progradational wedges would preserve the asymmetry of the foreland basin (Fig. 7.12a). Deltaic wedges related to regional sea-level fluctuations would tend to be more tabular with only a gentle increase in thickness back towards the sediment source area (Fig. 7.12b) partially controlled by the slope of the basin floor. At a finer level, it is possible that the internal facies of the shelf sequences would reflect this same asymmetry, with tectonic deltas having more stacked facies sequences, and eustatic deltas having more tabular facies distributions.

These expected distributions can be seen in preserved sections, such as from the central Utah foreland (Fig. 7.11). Both tabular deltaic sequences, such as the Ferron sandstone, and more asymmetric progradational events, such as the Funk Valley Formation, are preserved. Following this approach, these events result from primarily regional sea-level changes and tectonic events, respectively.

F. SUPERIMPOSED TECTONIC AND REGIONAL SEA-LEVEL EVENTS

The problem becomes more complex if both regional sea-level and tectonic events occur at the same time (Fig. 7.13). If short-term eustatic events occur as a tectonic delta is prograding (Fig. 7.13a), the asymmetric shape of the deltaic complex might still be preserved, even though small-scale fluctuations are also occurring. In addition, if the eustatically-induced progradations occur at slow enough rates, each of them will also show a dramatic thickening back towards the thrust front. In contrast, if short-term eustatic changes are superimposed on an overall sea-level fall (Fig. 7.13b) each of the small-scale transgressions will tend to be tabular as is the overall deltaic sequence.

A variation of this model involves a comparison of subsidence histories from the proximal and distal parts of a foreland basin to separate the effects of long-term subsidence and short-term sea-level changes that occur simultaneously in the stratigraphic record (Hill and Heller, 1988). Our example of superimposed tectonic and eustatic events comes from the Coniacian to lower Campanian interval of the western interior. Specifically, we compare two paleobathymetric records (Fig. 7.14) from the western interior - one from Boulder, Colorado (Beatty, 1985) and one from closer to the Sevier thrust belt at Hams Fork, Wyoming (Carlisle, 1979). These plots show estimated water depth versus stratigraphic thickness and age for the units shown (Fig. 7.14). The water depth estimates are based on benthic forams, planktic-to-benthic ratios, and megafossils.

Each of these studies interprets there to be seven deepening events where relative increases in water depth are recorded in each section. Of these events, numbers 2 and 7 (Fig. 7.14) are correlated to each other based on planktic foraminifera. Eustatic sea level changes would seem to be a major control on the stratigraphy, since each of the seven events apparently occurred penecontemporaneously across the entire region. However, notice the stratigraphic sequence at Boulder spans only about 80 meters of section, whereas the Hams Fork section occupies closer to 1800 meters of section. Clearly these two sections are subsiding differentially during this time interval. Differential subsidence between the proximal and distal foreland basin is expected if tectonic subsidence was also occurring during this time interval.

The differential subsidence between the two sites can be quantified by a direct comparison of the two sections assuming exact correlation of the seven events (Fig. 7.15a). Relative differences in water depth are included in the analysis, with uncertainties shown with vertical error bars. Notice that there is an increase in relative subsidence over time approaching event 4, followed by a decrease in relative subsidence. If this data is plotted in terms of cumulative subsidence (Fig. 7.15b), the subsidence curve has a convex up shape that, as discussed above, is characteristic of foreland basins. Therefore, this curve may be recording tectonic subsidence associated with movement on the Sevier thrust belt to the west on which shorter term sea-level events are superimposed.

Thus it seems that, although equation 7.1 is correct, there may be ways of sorting out tectonic effects from sea-level events that utilizes the geometry of basin subsidence. In this example the long term tectonic subsidence could be isolated from the stratigraphic record by assuming a flexural basin model for the Sevier foreland basin. Superimposed on this long-term response are shorter term sea-level events. It is not clear, as yet, if the sea-level events shown here represent local sea-level changes (due to local variations in sediment supply and subsidence rates), regional sea-level changes (due to episodic thrust loading and/or epeirogeny) or global eustatic changes. Regardless, the approach used at least discriminates between the first-order effects caused by tectonic loading and the small effects of uncertain origin. This interpretation is facilitated by use of subsidence modeling.

G. CONCLUSIONS

Thus we propose several techniques that might be used to separate out eustatic from tectonic events, at least in foreland basin settings. Many of these techniques are, as yet, not well established. However, at present data is being collected with which to test these concepts. Regardless, the underlying concepts of comparing the results of field studies with those predicted from theoretical studies such as those discussed in this short course, can be a very powerful tool when used in conjunction with other basin analysis techniques.

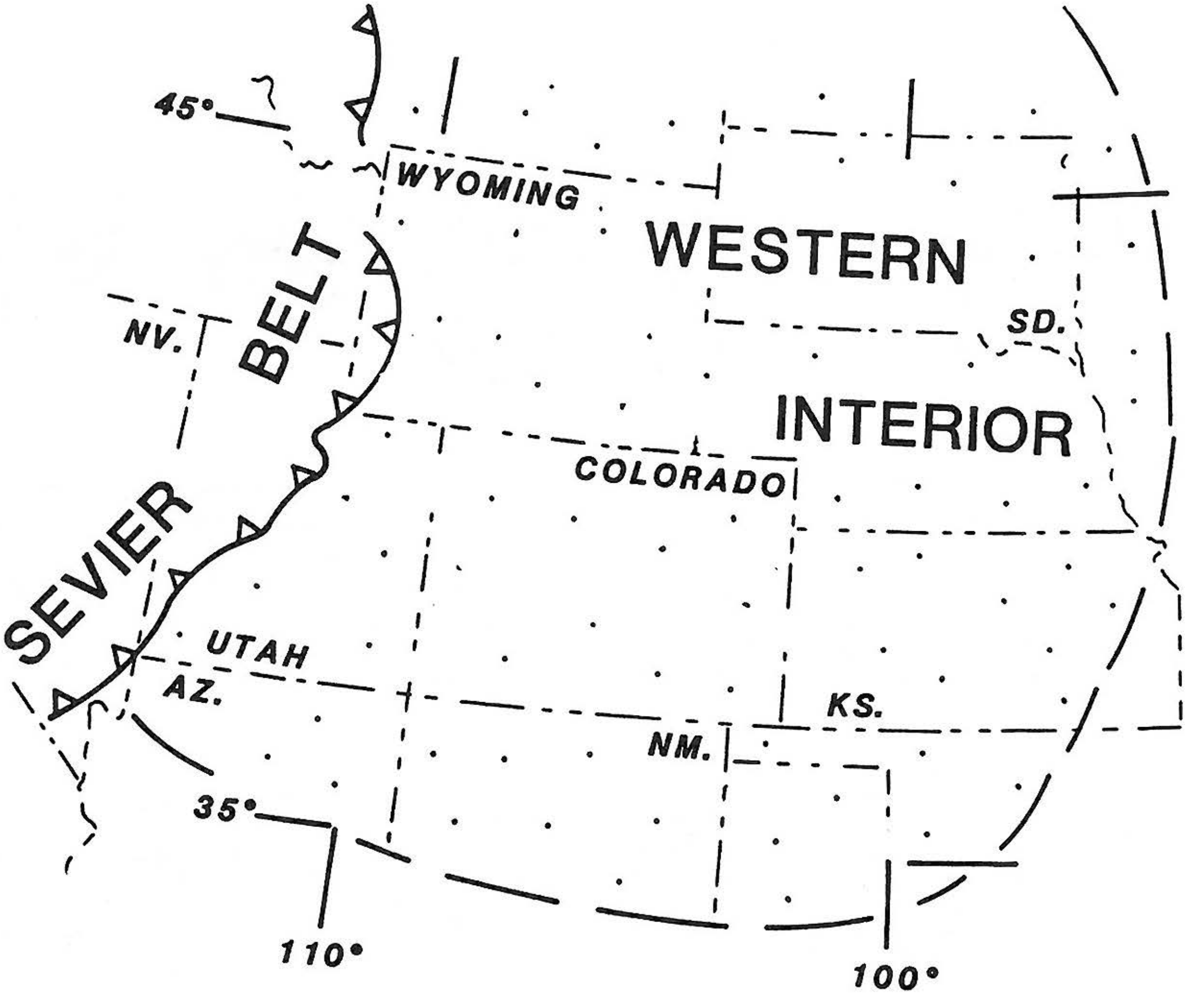


Figure 7.1 Map showing limits of foreland basin adjacent to Sevier orogenic belt in west-central U.S.A.

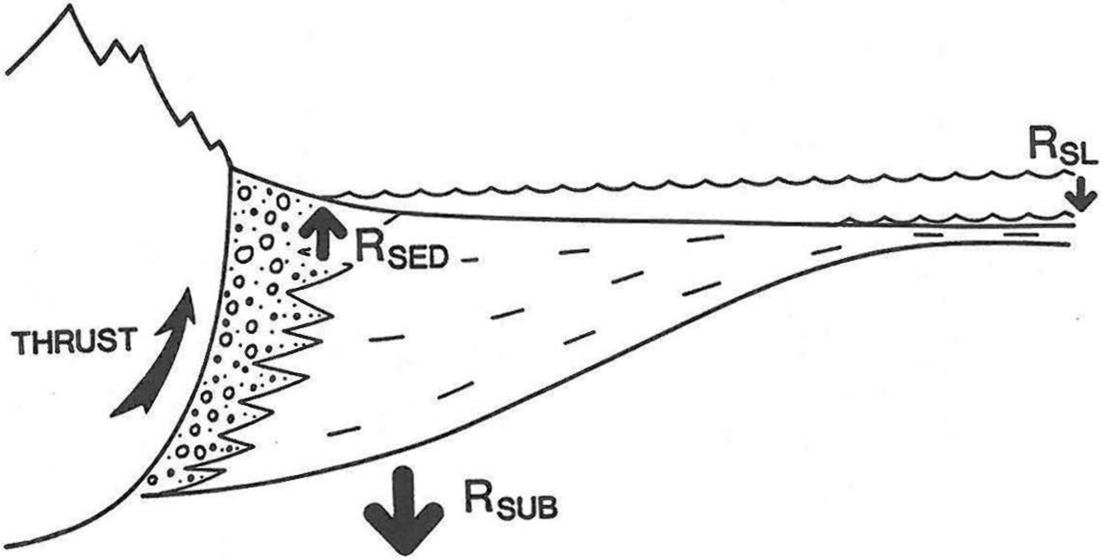


Figure 7.2 Diagrammatic representation of foreland basin deposition showing directions in which sedimentation rate (Rsed), subsidence rate (Rsub), and rate of relative sea-level fall (Rsl) would generate a regressive shoreline sequence.

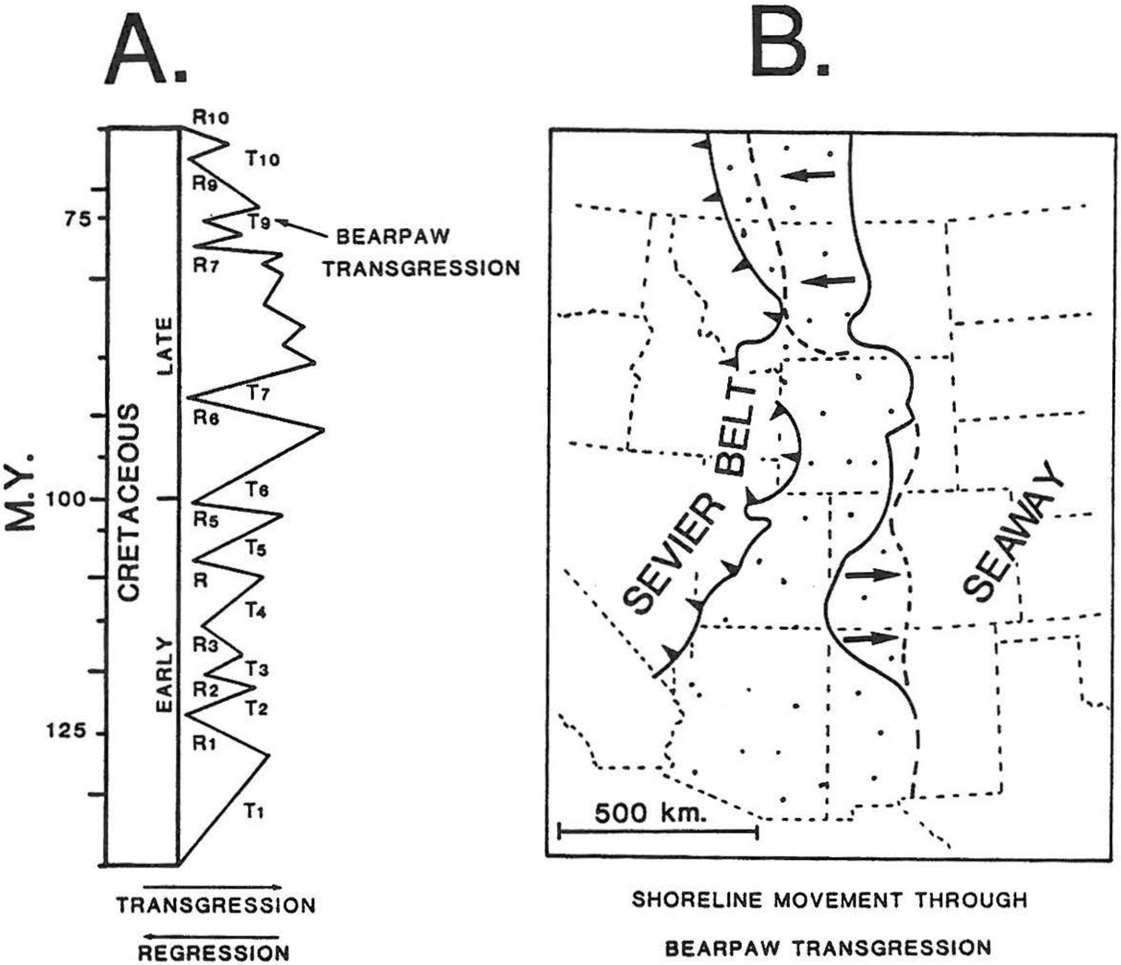
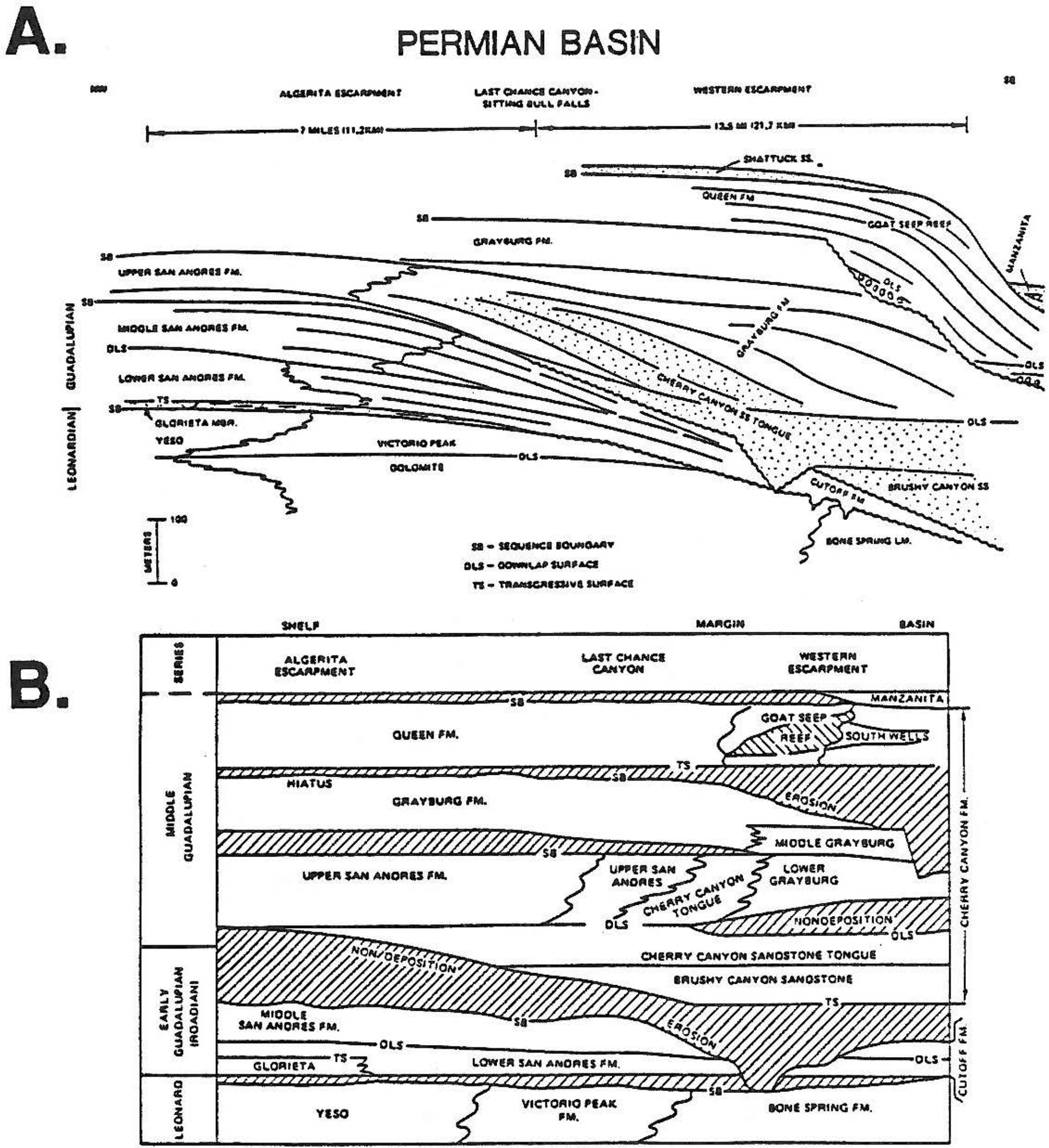


Figure 7.3 (A) Cretaceous transgressive (T) - regressive (R) history for the western interior from Kauffman (1977). (B) Shoreline movement during the Bearpaw Transgression (T9). Solid line shows shoreline position at start, and dashed line shows position at end of Bearpaw Transgression. Modified from Lillegraven and Ostresh (in press).



SARG & LEHMANN, 1986

Figure 7.4 Nortwest-southwest stratigraphic cross-section of the Guadalupian time interval in part of the Permian Basin of west Texas. The stratigraphy is shown in terms of (A) sequence boundaries and (B) in a time-space diagram (from Sarg and Lehmann, 1986).

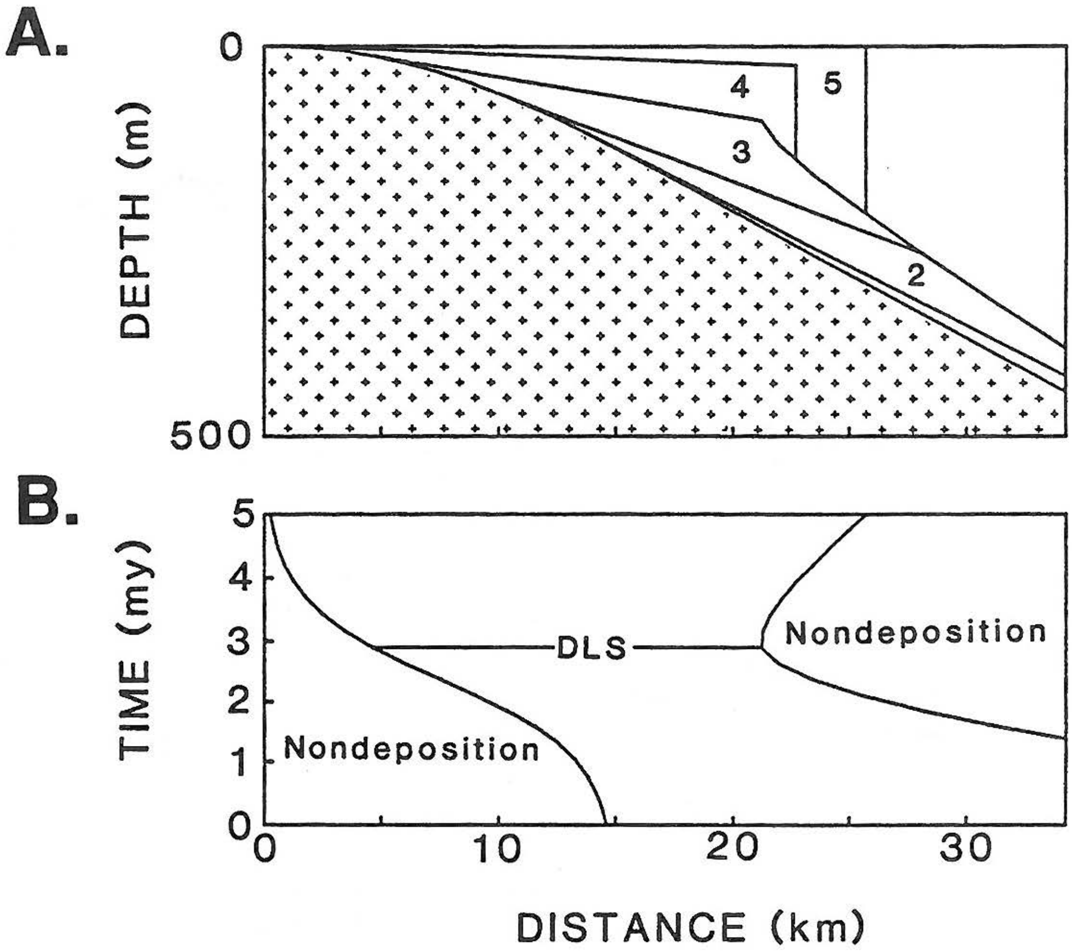


Figure 7.5 Synthetic stratigraphic cross-section of a distal margin of a foreland basin (thrust belt would lie to the right) showing stratigraphic development that would occur if subsidence rate slowed down following a thrust-loading event. Sediment supply rate and eustatic sea level are held constant. The stratigraphy is shown in terms of (A) sequence boundaries (for five units) and (B) in a time-space diagram. Compare with figure 7.4.

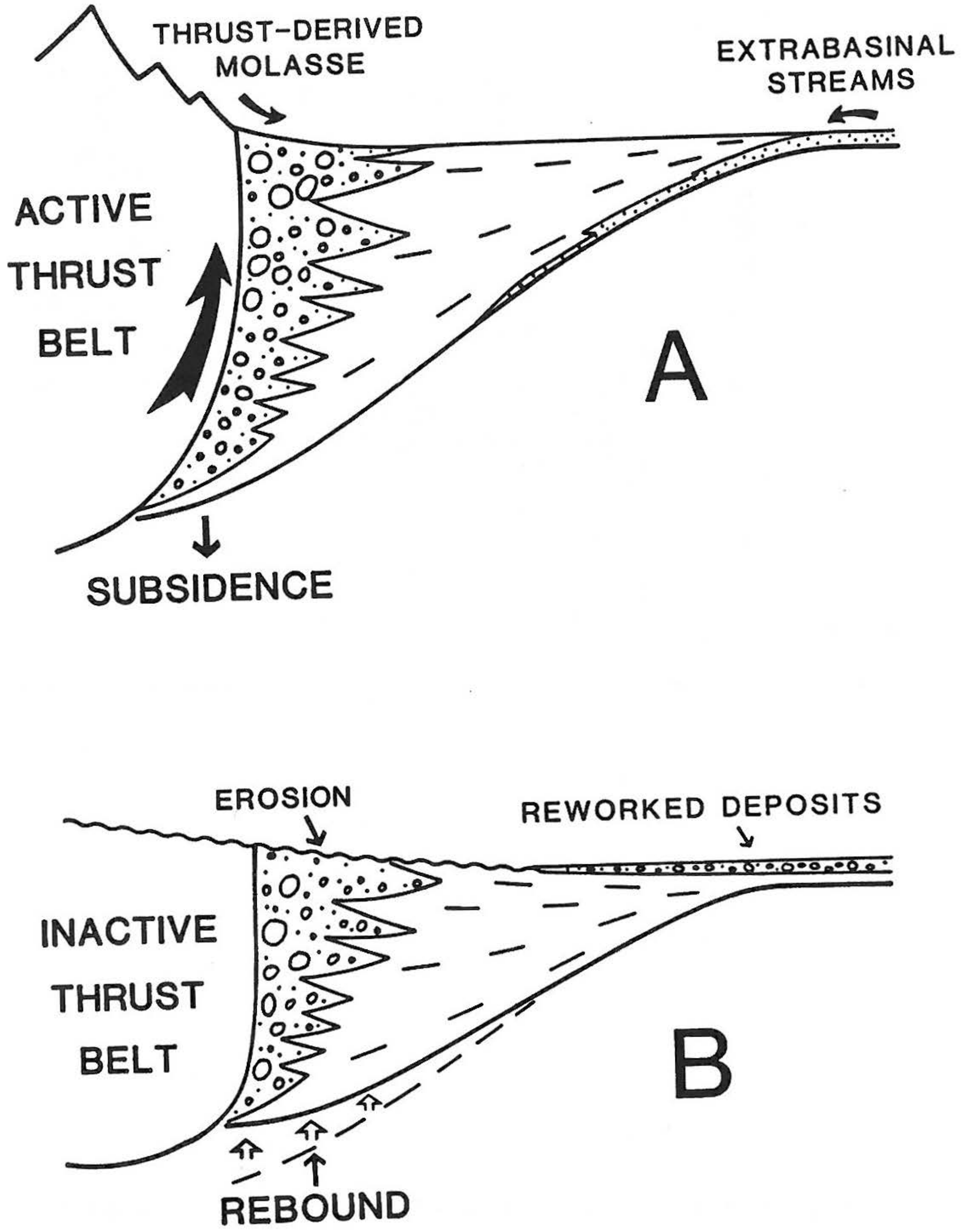


Figure 7.6 Depositional models for synorogenic (A) and post-orogenic (B) phases of foreland basin evolution. From Heller et al. (1988).

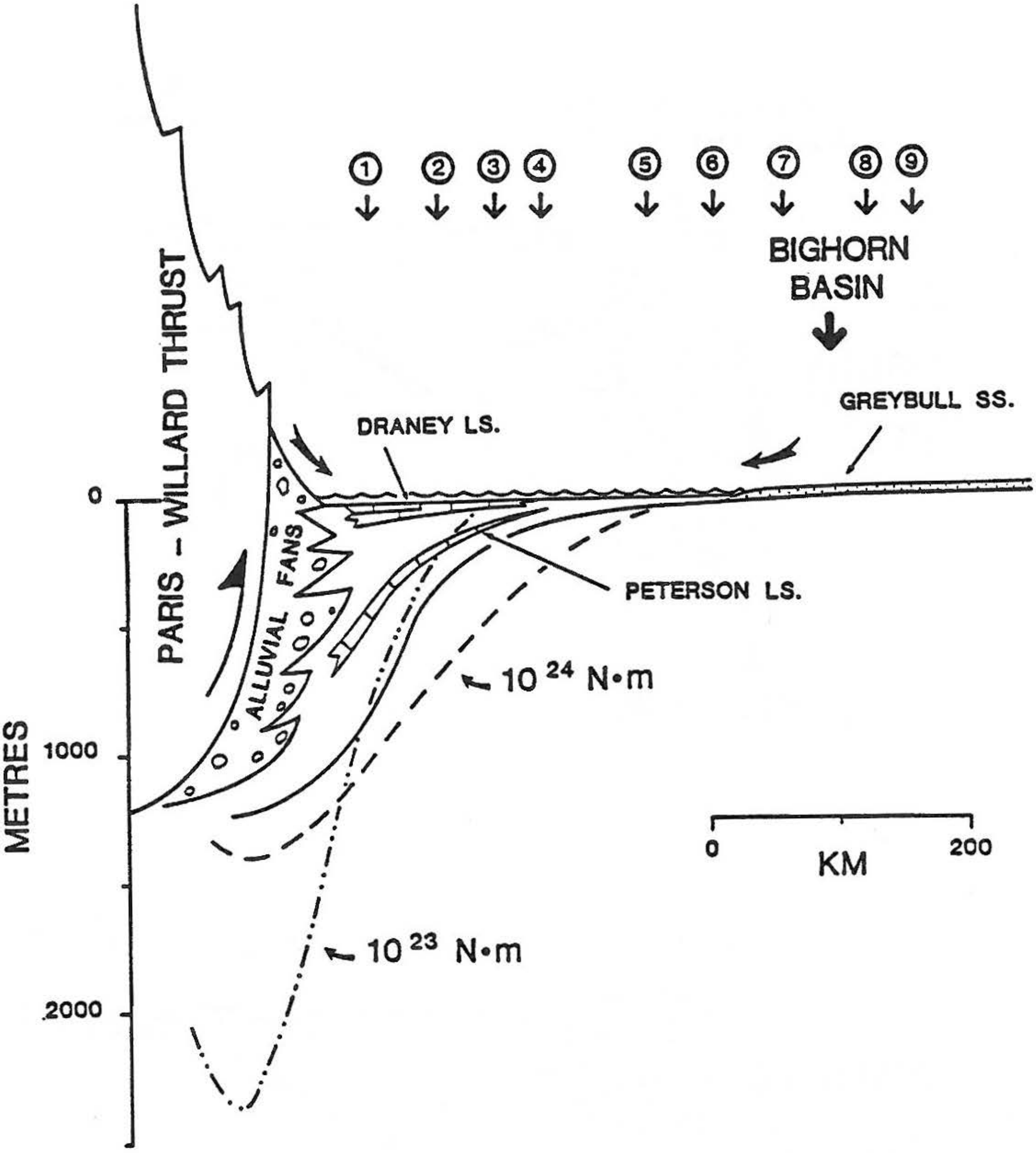


Figure 7.7 Diagrammatic cross-section of deposition in Wyoming foreland during emplacement of Paris-Willard thrust sheet. Theoretical configurations of sediment-filled basin resulting from emplacement of thrust plate with flexural rigidities of 10^{23} N·m and 10^{24} N·m are shown for comparison. Thicknesses shown are from measured sections (circled numbers): 1 - section 26 of Eyer (1969); 2,3, and 5 - D.L. Blackstone (1949, unpublished data); 4 - section 6V-1 of Furer (1970); 6 - section 5 of Love (1945); 7 and 8 - from Winslow (1986); 9 - section 11 of Mirsky (1962). From Heller et al. (1988).

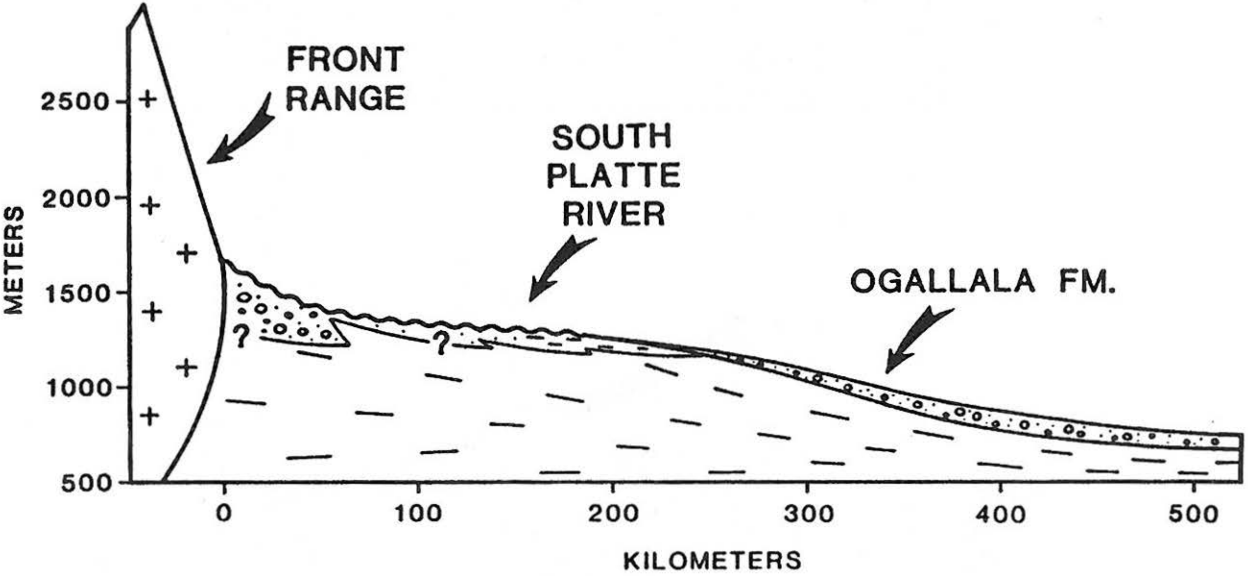


Figure 7.8 Cross-section of Neogene depositional system across north-east Colorado and western Nebraska. Compiled from Stanley and Wayne (1972) and Babits (1987). From Heller et al. (1988).

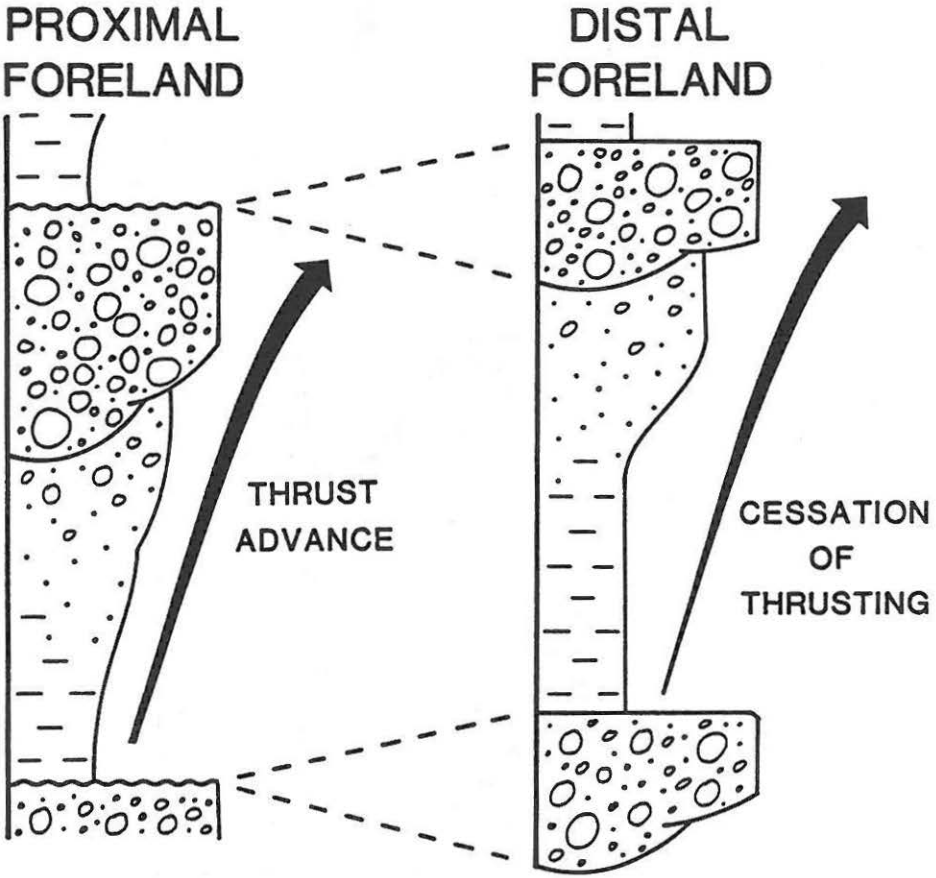


Figure 7.9 Correlation and interpretation of coarsening-upward sequences from proximal and distal parts of a foreland-basin, based on two-phase model explained in text. From Heller et al. (1988).

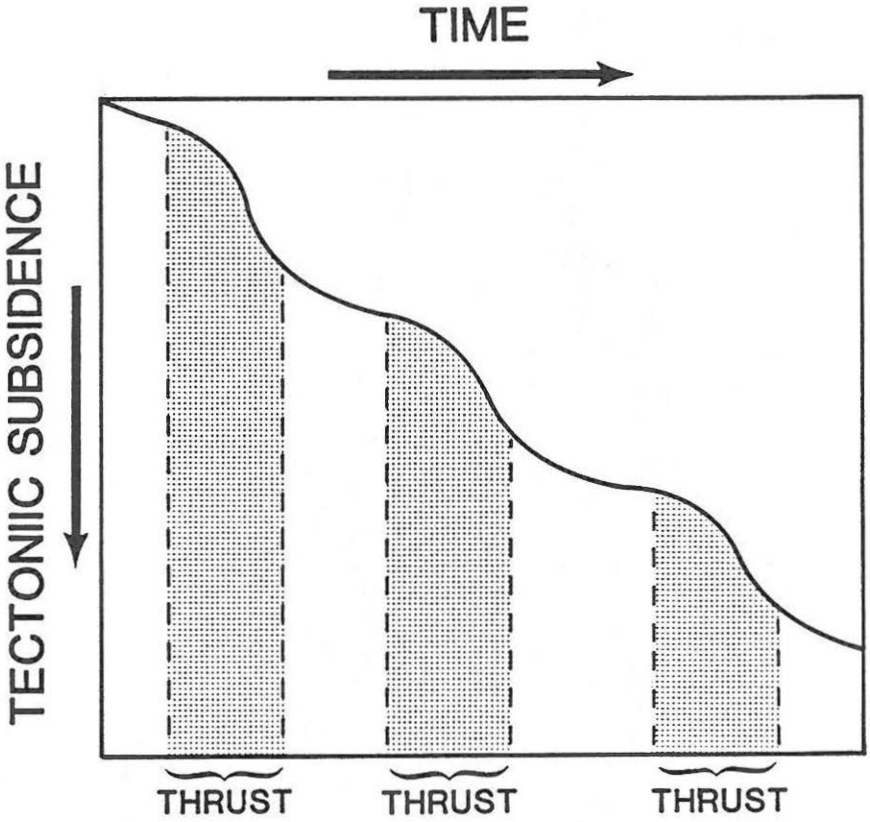


Figure 7.10 Idealized subsidence curve for foreland basins showing increased subsidence rate during thrusting events. Compare with figure 6.1.

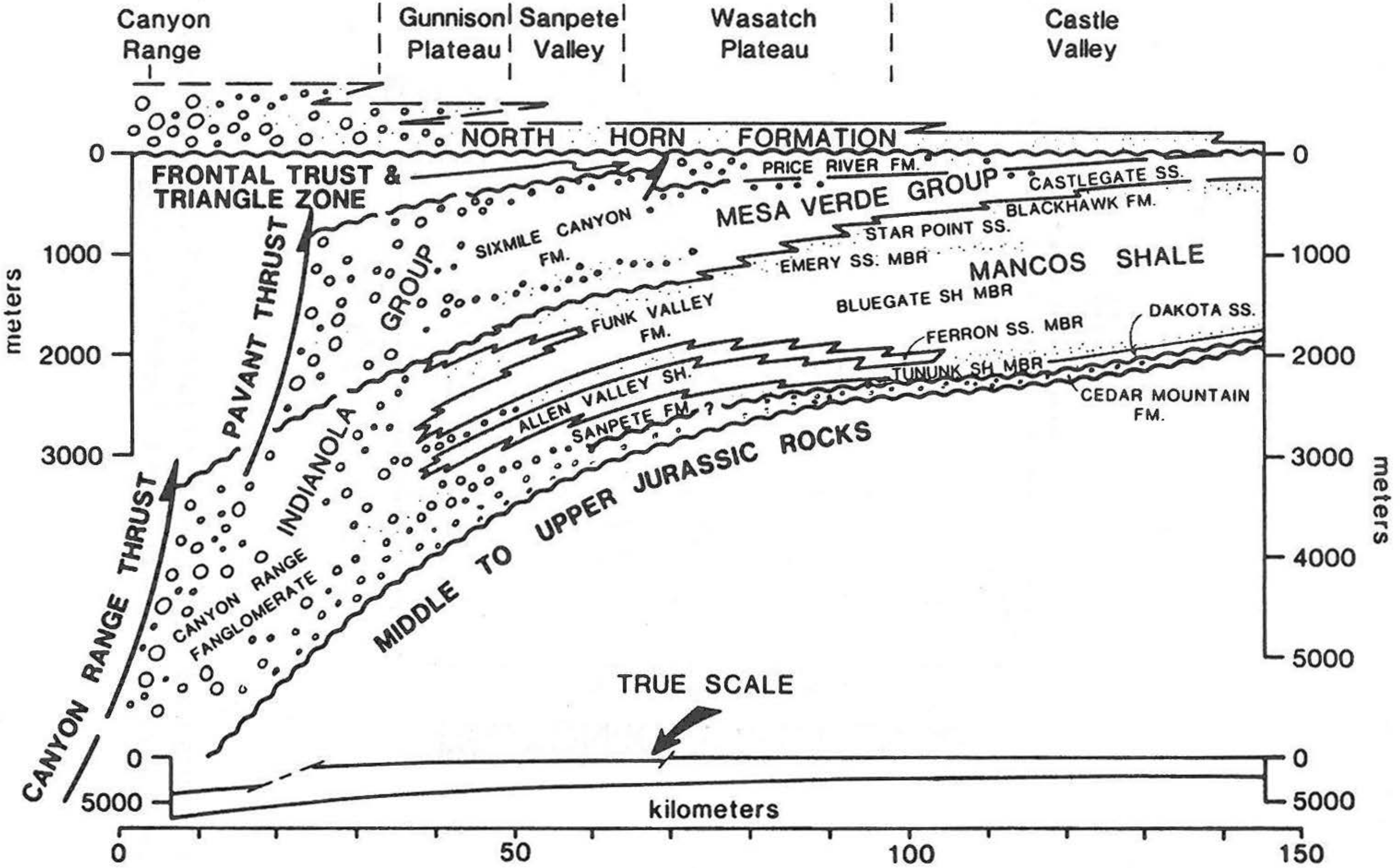


Figure 7.11 East-west (left to right) cross-section of central Utah showing Upper Cretaceous stratigraphy adjacent to the Sevier thrust belt (from Lawton, 1986).

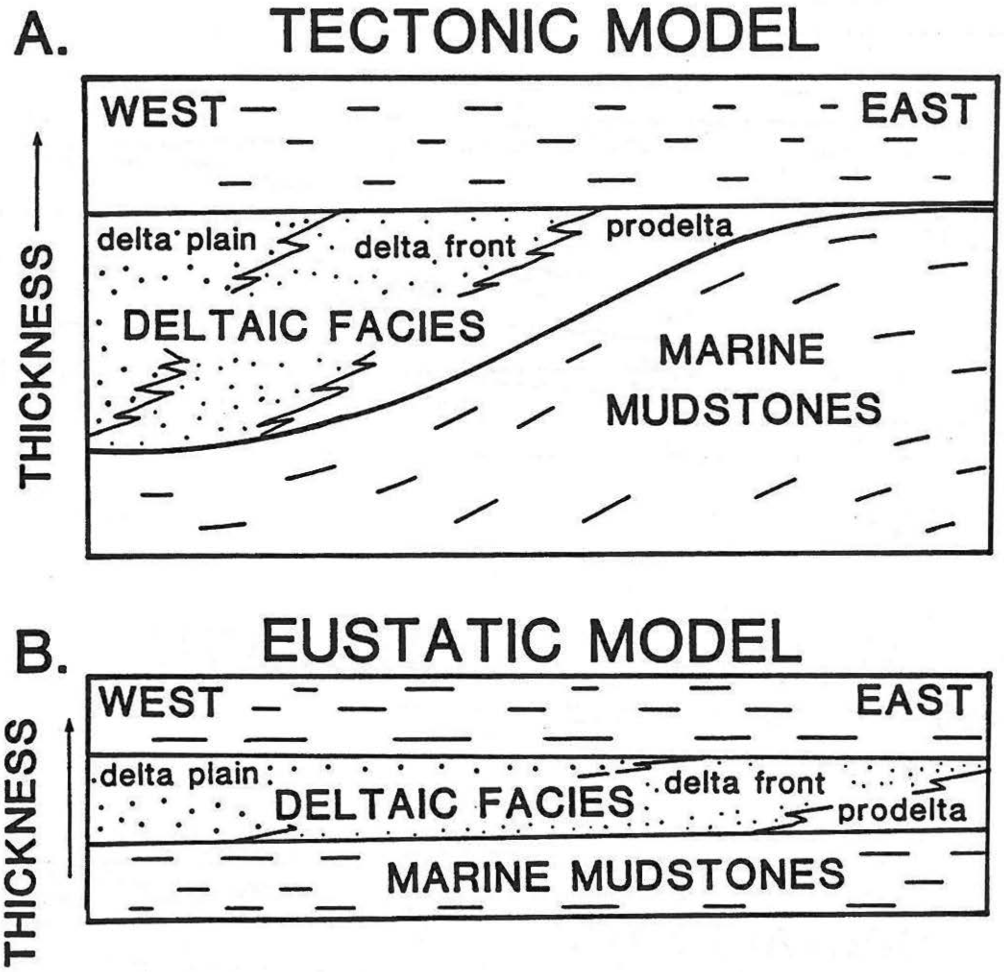


Figure 7.12 Idealized stratigraphic cross-sections of foreland basin deposits resulting from (A) a thrust loading event (to the west) and (B) a eustatic event (landward is to the west).

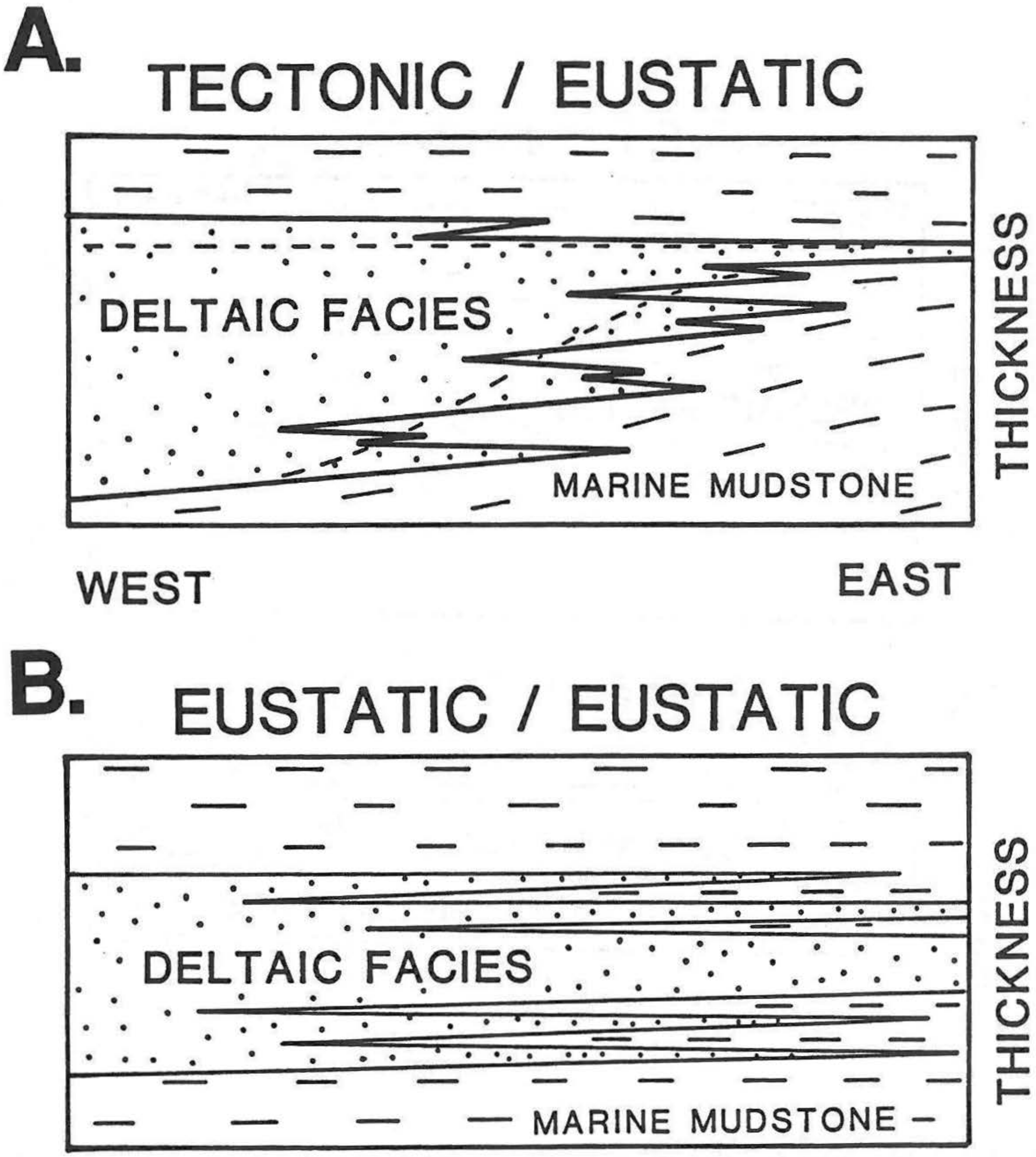


Figure 7.13 Idealized stratigraphic cross-sections of foreland basin deposits resulting from (A) short-term eustatic fluctuations superimpoesd on an overall tectonically-induced deltaic progradation, and (B) short-term eustatic fluctuations superimposed on an overall eustatically-induced deltaic progradation.

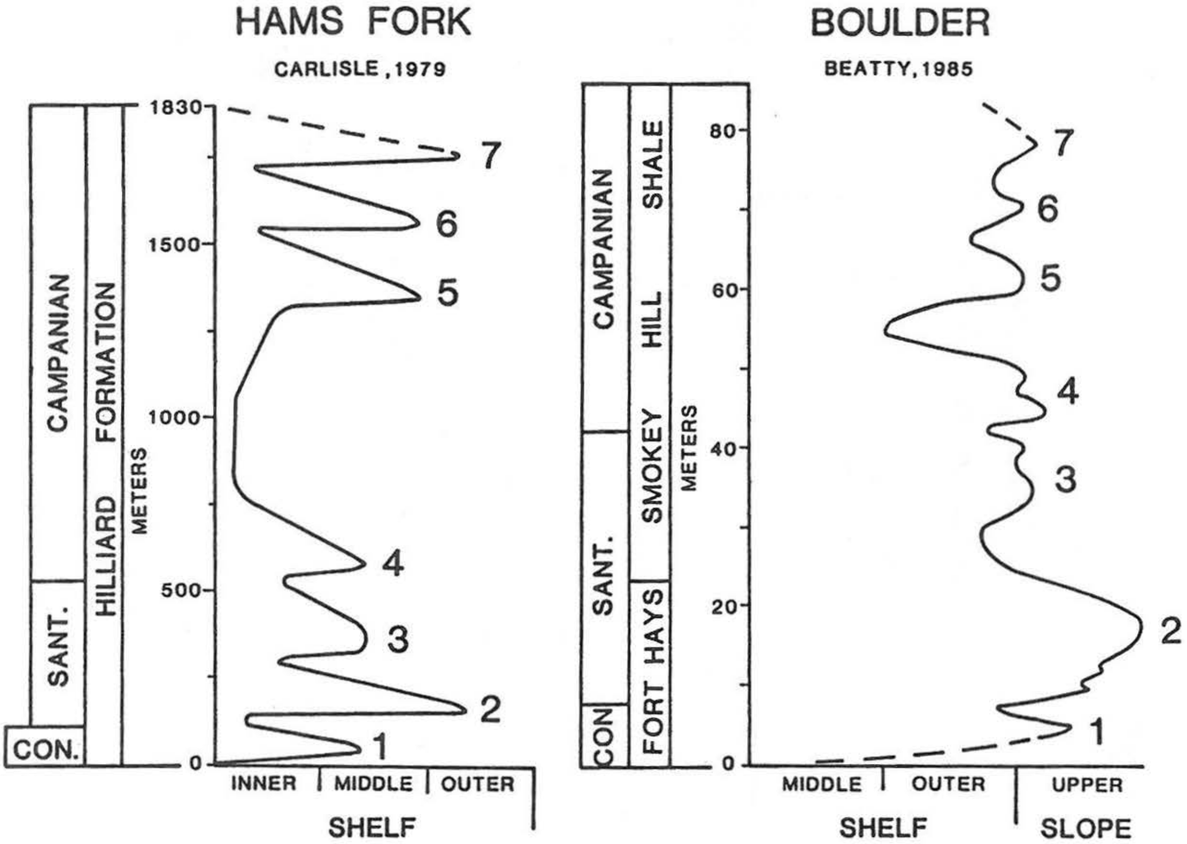


Figure 7.14 Coniacian to early Campanian stratigraphic records from Hams Fork, Wyoming (Carlisle, 1979) and Boulder, Colorado (Beatty, 1985) showing thickness and paleobathymetric data. Seven deepening events (T_{1-7}) are shown.

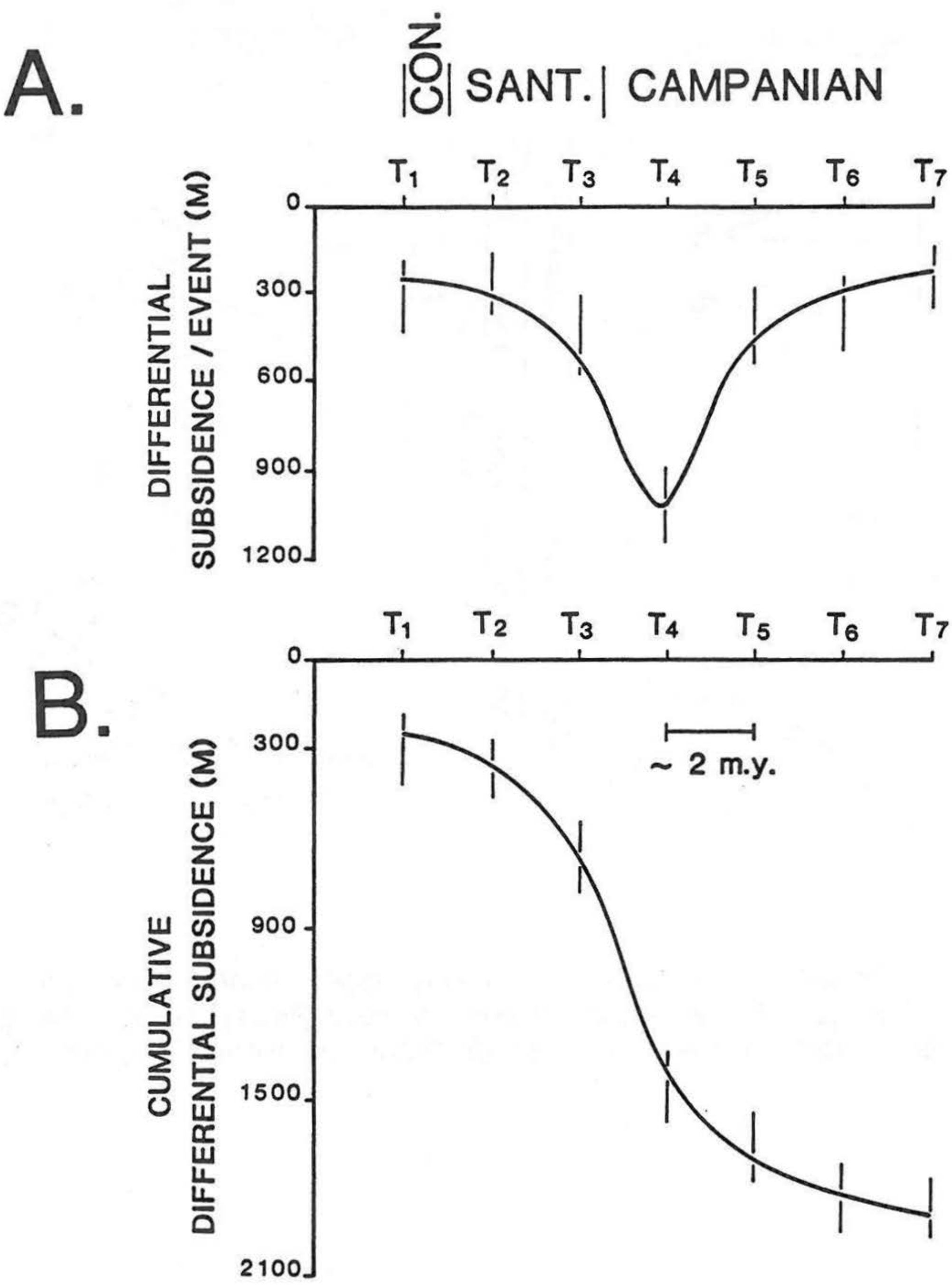


Figure 7.15 (A) Differential subsidence and (B) cumulative subsidence histories for events T_1 through T_7 between the Hams Fork and Boulder sections. Note increase in differential subsidence rate going into event T_4.