8. PASSIVE MARGIN STRATIGRAPHY

A. THIRD-ORDER SEDIMENTARY CYCLES

One of the more long-lived debates in the earth sciences concerns the origin of sedimentary sequences observed on continental margins (Barrell, 1917; Pitman, 1978; Seuss, 1906; Sloss, 1962; Vail et al., 1977; Vella, 1965; Watts, 1982). As defined by Vail et al. (1977), these sequences are packages of conformable sediments, representing a time span of 1 to 10 My, that are bounded by unconformities or horizons that can be correlated with unconformities. Strata within individual sequences show continuous onlap onto the continental margin, whereas sequence boundaries mark abrupt seaward shifts in coastal onlap. Originally, Vail et al. (1977) argued that onlapping sediments record a marine transgression, whereas sequence boundaries signal an abrupt regression. Subsequently, more detailed studies of a small number of sequences (Vail et al., 1984) showed that onlapping sediments in the upper parts are nonmarine, suggesting that regressions are gradual rather than abrupt. Based on the apparent global synchroneity of many sequence boundaries, Vail et al. (1977) and Haq et al. (1987) propose that transgressions and regressions are caused by oscillatory eustatic variations. Some workers (Hallam, 1984; Miall, 1986) question whether or not the resolution of biostratigraphic correlations is sufficient to show that sequence boundaries are truly synchronous between basins.

A fundamental difficulty in understanding the origin of sedimentary sequences is that transgressions and regressions can be caused by changes in the rate of tectonic subsidence or sedimentation, as well as eustatic variations (cf. Curray, 1964; Sloss, 1962; and Chapter 7). Hubbard et al. (1985) argue that on the Newfoundland and Beaufort Sea margins of Canada many sequence boundaries are associated with delta-lobe switching and nondeposition rather than sea-level fluctuations. Pitman (1978) and Pitman and Golovchenko (1983) demonstrate that on subsiding passive margins, a decreasing sea-level history with abrupt changes in the rate of fall can produce an onlap history similar to that described by Vail et al. (1977). We will look at their work in more detail below. Variations in tectonic subsidence rates may ensure that sedimentary sequences differ in age between margins (Parkinson and Summerhayes, 1985; Thorne and Watts, 1984). Watts (1982), Watts et al. (982), and Watts and Thorne (1984) show that a significant fraction of coastal onlap may be explained by the increase in lithosphere rigidity that occurs as margins cool following rifting. As we saw in Chapter 5, part O, recent work by Cloetingh et al. (1985), Cloetingh (1986), and Karner (1986) shows that modest fluctuations in intraplate stress levels can cause vertical deflections of the lithosphere, thereby providing a tectonic explanation for cycles of transgression and regression.

Studies of oxygen isotopes in foraminifera from deep-sea sediments suggest that glacioeustatic variations have occurred during much of the late Tertiary time (Keigwin and Keller, 1984; Matthews and Poore, 1980; Miller et al., 1985; Moore et al., 1987). Figure 8.1 shows one recent attempt (by Miller et al., 1985) to correlate the glacioeustatic record with coastal onlap curves. Individual lowstands in the sea-level record are constrained by $\delta^{18}O$ measurements, but other parts of the curve are based on the assumption that short-term glacioeustatic variations cause symmetric rises and falls. The long-term sea-level fall is based on a reconstruction by Kominz (1984). Onlap curve is based on seismic stratigraphy and well data. Miller et al. (1985) correlate the mid-Oligocene unconformity with the maximum rate of sea-level fall. However, erosion during regressions, possible delays in the onset of sedimentation during the subsequent transgression, and uncertainties in biostratigraphic correlations all conspire to make precise dating of unconformities difficult.

Miller et al. (1985) estimate an uncertainty of 1 to 2 My in the ages of the Oligocene unconformities. Such a large uncertainty makes it equally possible to correlate the mid-Oligocene unconformity with the glacioeustatic lowstand. It is important to note that not all Tertiary unconformities are associated with $\delta^{18}O$ shifts (Miller et al., 1987). In the following sections, we will look at why Miller et al. (1987) correlate unconformities to rapid eustatic falls, and we will determine the conditions under which this correlation is legitimate.

B. PITMAN'S SEDIMENTATION MODEL

One of the great events of 1978 was the publication of a model for clastic sedimentation on subsiding continental margins by Walter Pitman. The treatment of passive margin sedimentation in this model is much more sophisticated than anything you will see in the complicated basin evolution models produced by the geophysics community. Unfortunately, the model's predictions have largely been misunderstood. As a side note, Pitman's model is an example of a geometric model, one in which the system is assume to be in grade at all time. In Chapter 9, we will discuss a newer class of models that explicitly account for sediment transport; these models are sometimes called dynamic stratigraphy models.

A diagram of Pitman's model is shown in Figure 8.2. Subsidence is modeled as a rigid rotation about a landward hinge line with a maximum subsidence rate of R_{SS} at the shelf edge. The width of the shelf is D. A graded slope α is maintained on the shelf and coastal plain. R_{SL} denotes the rate of eustatic sea-level fall, and X_L is the shoreline location. The rate of shoreline migration (in the seaward direction) is related to the rate of water-depth increase (R_{WD}) by:

$$\frac{dX_L}{dt} = -\frac{R_{WD}}{\alpha} \tag{8.1}$$

We note that the rate of water-depth increase is a function of the rate of subsidence (R_T), the sedimentation rate (R_S), and the rate of eustatic sea-level fall (R_{SL}):

$$R_{WD} = R_T - R_S - R_{SL} \tag{8.2}$$

where the subsidence rate increases linearly with distance from the hingeline

$$R_T(x) = R_{SS}\frac{x}{D} \tag{8.3}$$

The clever aspect of the model is the way that sedimentation and subsidence rates are related. Assume that erosion occurs landward of the shoreline, while deposition occurs seaward. At the shoreline, the sedimentation rate is zero (this assumption can be relaxed, the important point is that the sedimentation, or erosion, rate at the shoreline must remain constant). To maintain a graded slope, sedimentation (or erosion) rates at any location on the margin must equal the difference in subsidence rates between that location and the shoreline:

$$R_S(x) = R_{SS}\left[\frac{x - X_L}{D}\right] \tag{8.4}$$

Substituting equations 8.4, 8.3, and 8.2 into equation 8.1 and rearranging terms, we find an expression for the shoreline location as a function of time:

$$\frac{d X_L}{d t} + \frac{R_{SS}}{D \alpha} X_L = \frac{R_{SL}}{\alpha} \tag{8.5}$$

An important point to notice is that the quantity $D\alpha / R_{SS}$ forms a natural time constant, which we will call the tectonic time constant (τ_T). This time constant is a measure of the time required for the shoreline to move to its equilibrium location. We can, thus, rewrite the shoreline equation as:

$$\tau_T \frac{d X_L}{d t} + X_L = \frac{R_{SL} D}{R_{SS}} \tag{8.6}$$

In the next two sections, we will look at solutions for equation 8.6 to monotone and periodic sea-level variations.

C. MONOTONE SEA-LEVEL FALLS

Pitman (1978) shows that if sea level falls at a constant rate (R_{SLO}) for a long period of time compared to τ_T, then the shoreline will move to an equilibrium location X_L^{EQ} where the rate of tectonic subsidence is equal to the rate of sea-level lowering:

$$X_L^{EQ} = \frac{R_{SL0} D}{R_{SS}} \tag{8.7}$$

This result is obtained by dropping the derivative that appears in equation 8.6. (The equilibrium shoreline location is, by definition, the place where the rate of change of the shoreline location is zero.) Equation 8.7 predicts that transgressions are greatest when sea level is falling fastest. Now you can understand why people correlate major unconformities with rapid eustatic lowerings (as in Fig. 8.1). We (following Pitman, of course) have justified the correlation using simple mathematics. Or have we?

Pitman was very careful to state the conditions under which equation 8.7 is valid. Eustatic sea-level has to fall at a constant rate for a time long compared to τ_T. In Figure 8.1, we see an oscillatory sea-level variation. For that case, the rate of sea-level change is not constant. In fact, R_{SL} is constantly changing. But let's not overreact and throw out the correlation altogether. We will see in the next section that the common correlation is approximately valid as long as the tectonic time constant is small compared to the period of the sea-level oscillation.

Before moving on, take a look at Figure 8.3 (after Pitman, 1978). It shows how the shoreline location (dashed line) will vary in response to abrupt changes in the rate of sea-level fall (solid line). The shoreline location is found by integrating equation 8.6 through time with D = 250 km, α = 0.2 m/km, and R_{SS} = 25 m/My (τ_T = 2 My). This figure is important because it shows that both transgressions and regressions can occur at a time when eustatic sea level is falling.

Figure 8.4 shows results from a similar thought experiment. Sea level is allowed to fall at an average rate of 5 m/My for 80 My. The rate of sea level fall is not constant but shifts back and forth between 8 and 2 m/My in a periodic way. Passive margin parameters are D = 250 km, R_{SS} = 10 m/km, α = 0.2 m/km; for these parameters, the characteristic time (τ_T) is 5 My. When the rate of sea level fall is 8 m/My (2 m/My), the equilibrium position is at 200 km (50 km). If the period of the eustatic-rate fluctuation is long (40 My; Fig. 8.4A) there is enough time, almost, for the shoreline to reach equilibrium. When the period is shorter, 10 or 4 My (Fig. 8.4B and C), there is no longer sufficient time for the shoreline to reach its equilibrium position. In these cases, the magnitudes of the regressions and transgressions are diminished, and the maximum regression no longer occurs when sea level is falling fastest.

You can see that the magnitude and timing of the maximum regression depends on the period of the eustatic rate-of-change history.

D. PERIODIC SEA-LEVEL FLUCTUATIONS

Equation 8.7 was derived under the assumption that sea level falls at a constant rate. Consequently, it cannot be used to understand how a shoreline shift in response to a continuously varying sea level, like the one shown in Figure 8.1. A reasonable approximation to the Oligocene glacioeustatic variation is shown in Figure 8.4. The sea-level history consists of a short-term component of magnitude Δh and period T superposed on a long-term fall. The rate of sea-level change is:

$$R_{SL} = R_{SL0} + \frac{2\pi \Delta h}{T} \cos\left(\frac{2\pi t}{T}\right) \tag{8.8}$$

Solving equation 8.6 using equation 8.8, we (Angevine, 1989) find that the shoreline location will vary according to:

$$X_L(t) = X_L^{EQ} - \frac{\Delta h \, \xi_T}{\alpha \sqrt{1 + \xi_T^2}} \sin\left[\frac{2\pi t}{T} + \tan^{-1}\left(\frac{1}{\xi_T}\right)\right] \tag{8.9}$$

where ξ_T ($= 2\pi \tau_T/T$) is a dimensionless tectonic parameter. The shoreline simply oscillates about the equilibrium position predicted by equation 8.2. Although the periods of the shoreline and sea-level oscillations are identical, there is a phase shift between the two. This phase shift translates into a time lag (τ) that ranges from 0 to $T/4$, depending on the value of ξ_T:

$$\tau = \frac{T}{2\pi} \tan^{-1}\left(\frac{1}{\xi_T}\right) \tag{8.10}$$

Figure 8.5 shows that regressions are synchronous with lowstands when ξ_T is large compared to one, and synchronous with the maximum rate of fall when ξ_T is small compared to one. The magnitude of the shoreline oscillation, a function of ξ_T also, varies from 0 to $\Delta h/\alpha$. Regressions and transgressions are most extensive when ξ_T is large, but they vanish as ξ_T decreases toward zero. This solution can be extended to more complicated sea-level histories using standard Fourier-series techniques.

It is of interest to understand how much the tectonic parameter can vary by, from one passive margin to another. As was discussed earlier, the model's applicability is restricted to older margins at which the width and subsidence rate are relatively constant. On older margins, subsidence rates are low; a reasonable range is R_{SS} = 5 to 30 m/My (Pitman and Golovchenko, 1983). The widths of margins, measured from hinge line to shelf edge, vary between 100 and 300 km. Although the choice of a margin slope is difficult, Pitman (Pitman, 1978) uses α = 1 to 0.1 m/km. Taking extreme values for each parameter, ξ_T can be as large as 56.5 or as small as 0.2, for third-order eustatic fluctuations (T = 1 to 10 My). Thus, transgressions or regressions on two different margins can differ in timing by as much as 2.5 My. Because the time lag can vary from one margin to another, there is no reason to expect regression-related unconformities to be globally synchronous. To summarize this discussion, Figure 8.6 shows how the magnitude and time lag of the shoreline oscillation varies with the period of the eustatic oscillation, for T = 1 to 100 My, and τ_T. In this figure, a nondimensional shoreline oscillation of one is equivalent to a regression (or transgression) of magnitude $\Delta h/\alpha$. A nondimensional time lag of one is equivalent to a time lag of $T/4$ between the shoreline and eustatic oscillations.

E. EXAMPLE I--U.S. ATLANTIC MARGIN

For the mid-Oligocene regression to occur when eustatic sea level was falling fastest, as proposed by Miller et al. (1985); also, see Fig. 8.1) and according to equation 8.10, ξ_T must be less than 0.2. Assuming that the period of the sea-level oscillation is 6 My and that the margin's width is 250 km, then the ratio of R_{SS}/α must be greater than 1.31 x 10^6 m/My to meet this condition. Even with a relatively gentle slope of α = 0.1 m/km, the subsidence rate must be unreasonably large: R_{SS} = 131 m/My. We conclude that either the glacioeustatic fluctuation is not sinusoidal, or the sequence boundary is younger than suggested by Miller et al. (Miller et al., 1985). A more realistic value for ξ_T can be found by assuming reasonable values for R_{SS} and α. Taking R_{SS} = 25 m/km and α = 0.5 m/km, then:

$$\tau_T = 5.\ my \quad and \quad \xi_T = 5.2 \tag{8.11}$$

suggesting that the mid-Oligocene sequence boundary should be correlated to the eustatic lowstand. The actual time lag calculated from equation 8.10 is only 180 ky. If the amplitude of the sea-level oscillation is 20 to 40 m (see Fig. 8.1), then the shoreline could migrate 40 to 80 km seaward of its equilibrium position during the regression. Choosing a smaller slope for the shelf or a larger sea-level fluctuation will yield a more extensive regression. For example, if we reduce the slope to $\alpha = 0.2$ m/km, then we increase the amplitude of the shoreline oscillation to 90 to 180 km and the time lag to τ = 425 ky. Doubling the magnitude of the sea-level oscillation will double the size of the shoreline oscillation but leave the time lag unchanged.

F. TIME DELAYS IN SEDIMENTATION

The most unrealistic aspect of Pitman's model may be the assumption that sedimentation rates adjust instantaneously to variations in the rate of sea-level change (the assumption of constant α runs a close second). Studies of the U.S. continental margin (Curray, 1964; Pitman, 1978; Swift, 1970) suggest that much of the sediment carried to the margin by river systems is being deposited in river valleys and canyons that were flooded during the Holocene sea-level rise. Consequently, on some parts of the shelf, clastic sedimentation rates are low, and relict sediments are exposed. A necessary consequence of low sedimentation rates is a gradual increase in water depth as tectonic subsidence continues.

Any lag in sedimentation rates (relative to a changing sea level) will tend to reinforce the near synchroneity of the shoreline and sea-level oscillations for third-order eustatic cycles. Under certain circumstances, the magnitudes of transgressions and regressions may even be amplified. This rather surprising result can be understood by considering the response of the shoreline to a sudden increase in the rate of sea-level fall. The shoreline will migrate toward the new equilibrium position predicted by equation 8.2, seaward of its previous location. Because sedimentation rates increase seaward of the shoreline but lag behind the sea-level change, the sedimentation rate at the new equilibrium position will initially be too high, and the shoreline must migrate even farther seaward. As sedimentation rates equilibrate, the shoreline will migrate back to its predicted position.

G. CONCLUSIONS

Cycles of transgression and regression on passive margins need not be associated with rises and falls of eustatic sea level. They may be due to variations in the rate of eustatic fall. Some independent data such as the $\delta^{18}O$ record may be required to decide which style of sea level variation actually occurred. A second point is sedimentary sequences are unlikely to be globally synchronous, even though they may be caused by eustatic sea-level variations. On subsiding continental margins, transgressions and regressions are influenced not only by eustatics but also tectonics and sedimentation. The apparent synchroneity of sedimentary sequences noted by many workers probably reflects the short duration of third-order cycles and the uncertainty in correlations between basins.

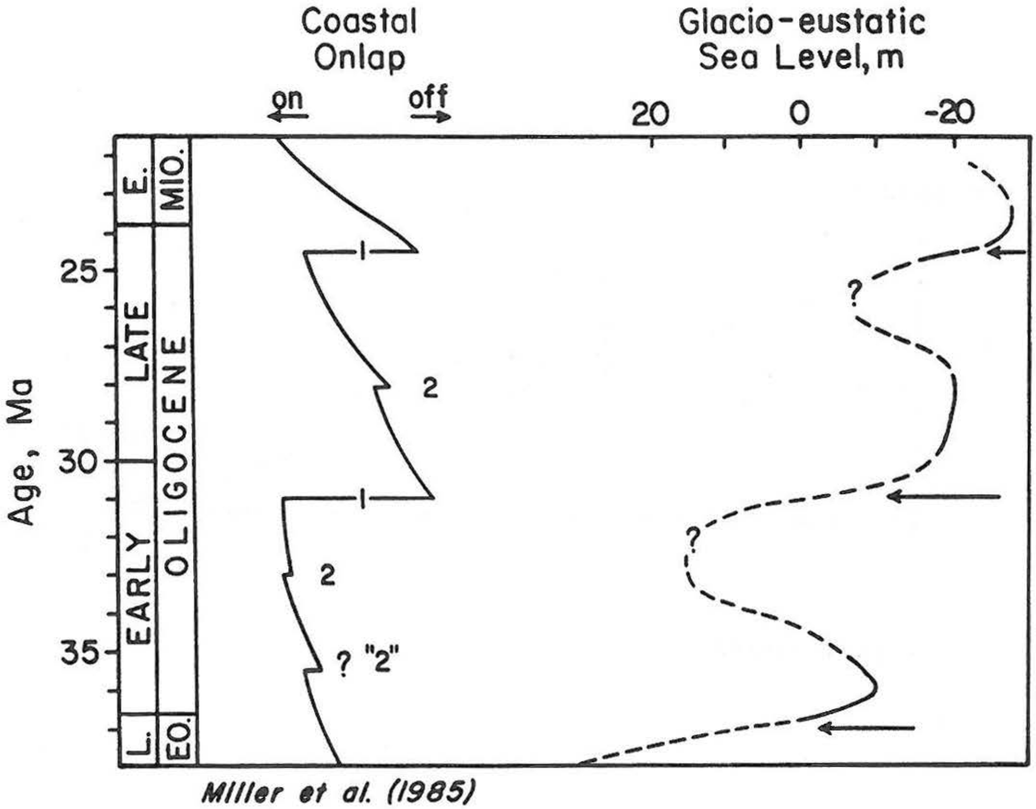

Figure 8.1 Proposed correlation (after Miller et al., 1985) between Atlantic passive-margin stratigraphy and glacioeustatic oscillations during Oligocene time. Eustatic curve is based on $\delta^{18}O$ data and presumes uniform rises and falls. Major sequence-bounding unconformities (indicated by 1; 2 indicates a minor event) are portrayed as being synchronous with maximum rates of sea-level lowering (shown by arrows).

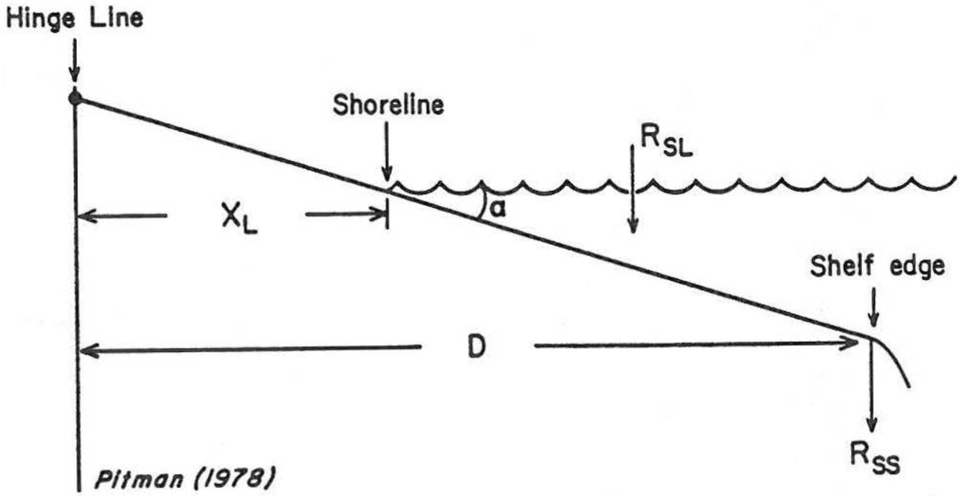

Figure 8.2 Simplified model of a passive continental margin (after Pitman 1978). Subsidence rates increase linearly from zero at the hinge line to a maximum (R_{SS}) at the shelf edge. Erosion and sedimentation keep pace with subsidence and eustatic variations, maintaining a graded slope α on the coastal plain and continental shelf. R_{SL} is the rate of sea-level change (positive for falls). The shoreline location, X_L, is measured from the hinge line, and D is the width of the margin. The model possesses a natural time constant τ_T (= $D\alpha/R_{SS}$).

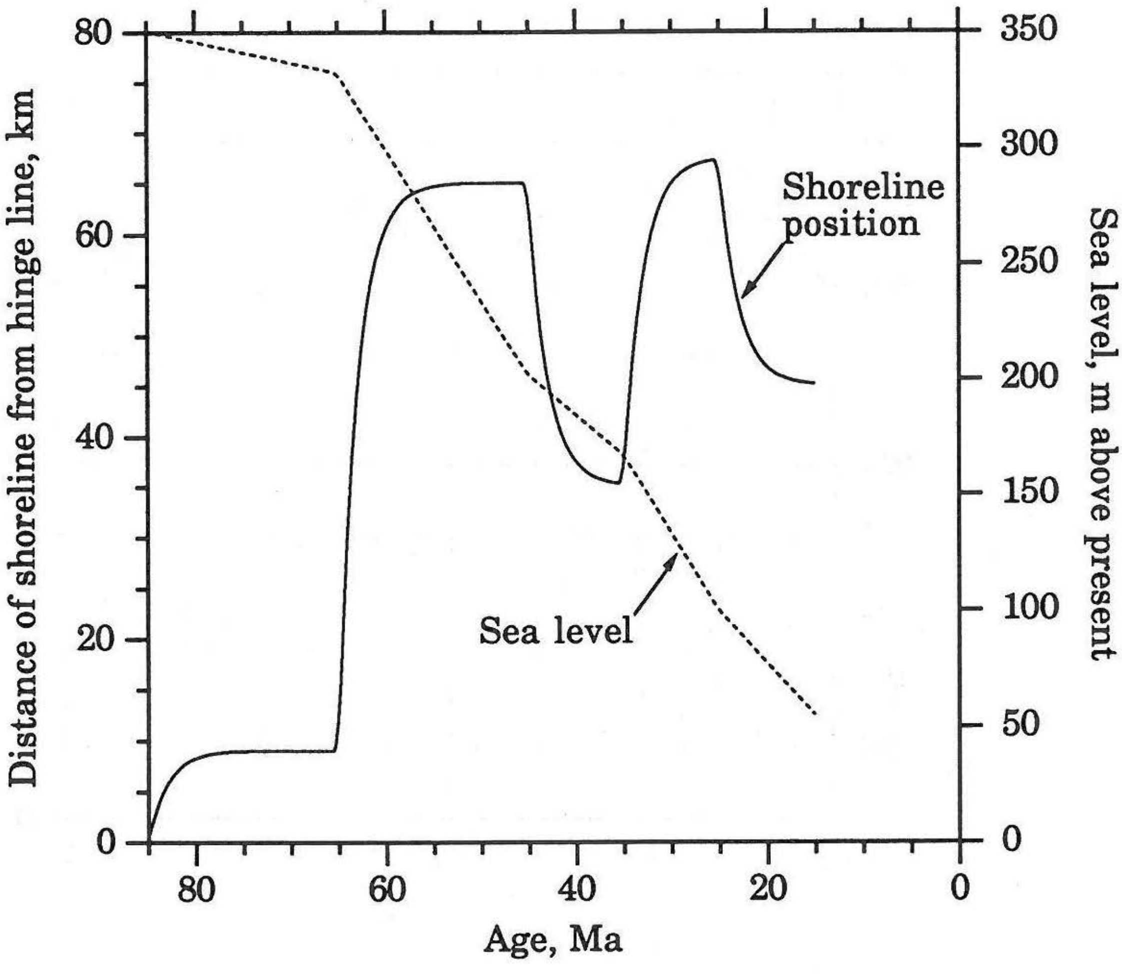

Figure 8.3 Displacement of shoreline (solid line) in response to abrupt variations in rate of sea-level fall (dashed line, after Pitman, 1978). Cycles of transgression and regression occur even though sea level is only falling. The time constant for this margin is $\tau_T = 2$ My.

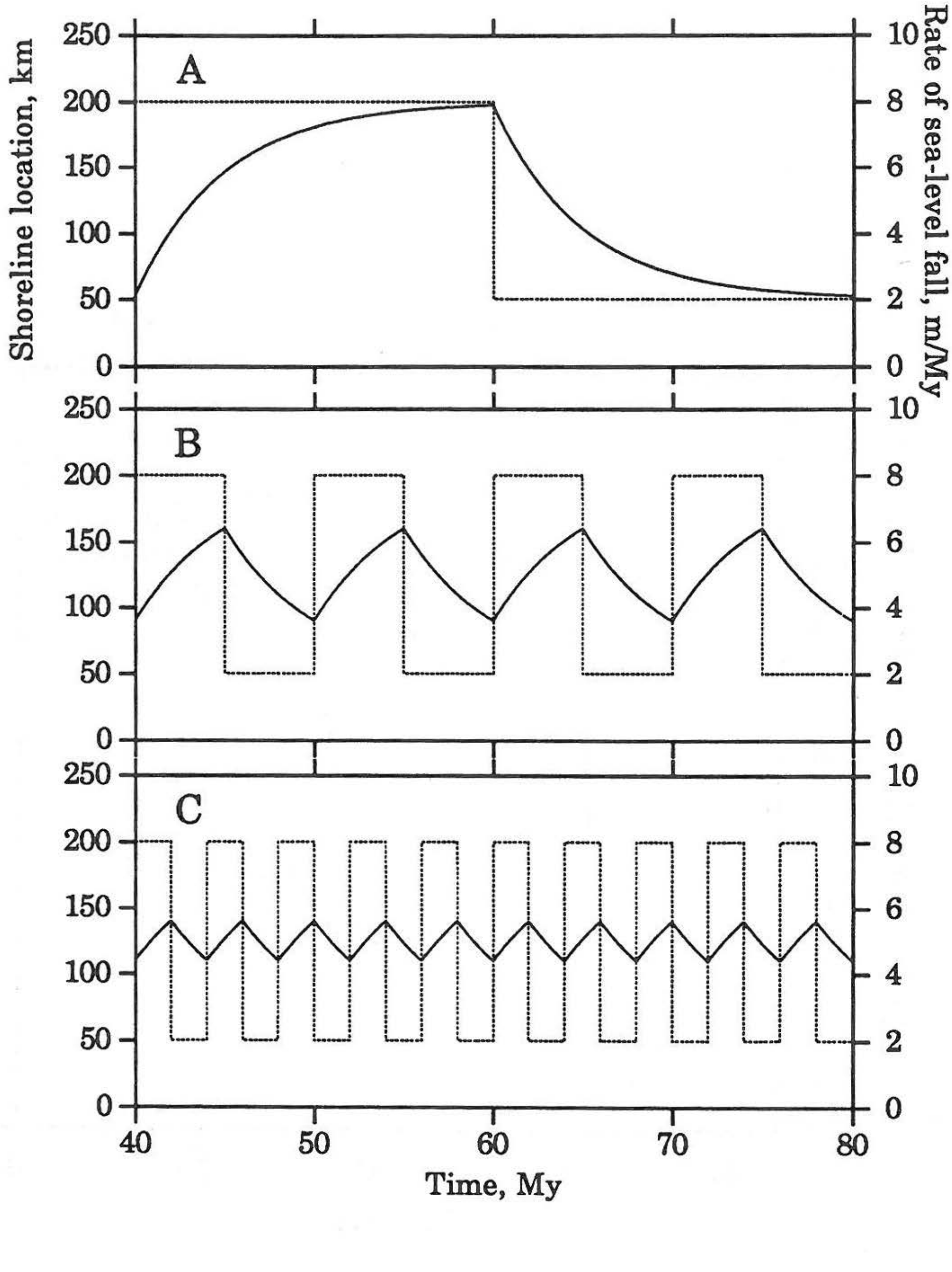

Figure 8.4 Displacement of shoreline (solid line) in response to a periodic change in rate of sea-level fall (dashed line) for periods of (A) 40 My, (B) 10 My, and (C) 4 My. Magnitude of shoreline displacement decreases as the period decreases. Note that, in (B) and (C), the maximum regression does not occur when sea level is falling fastest.

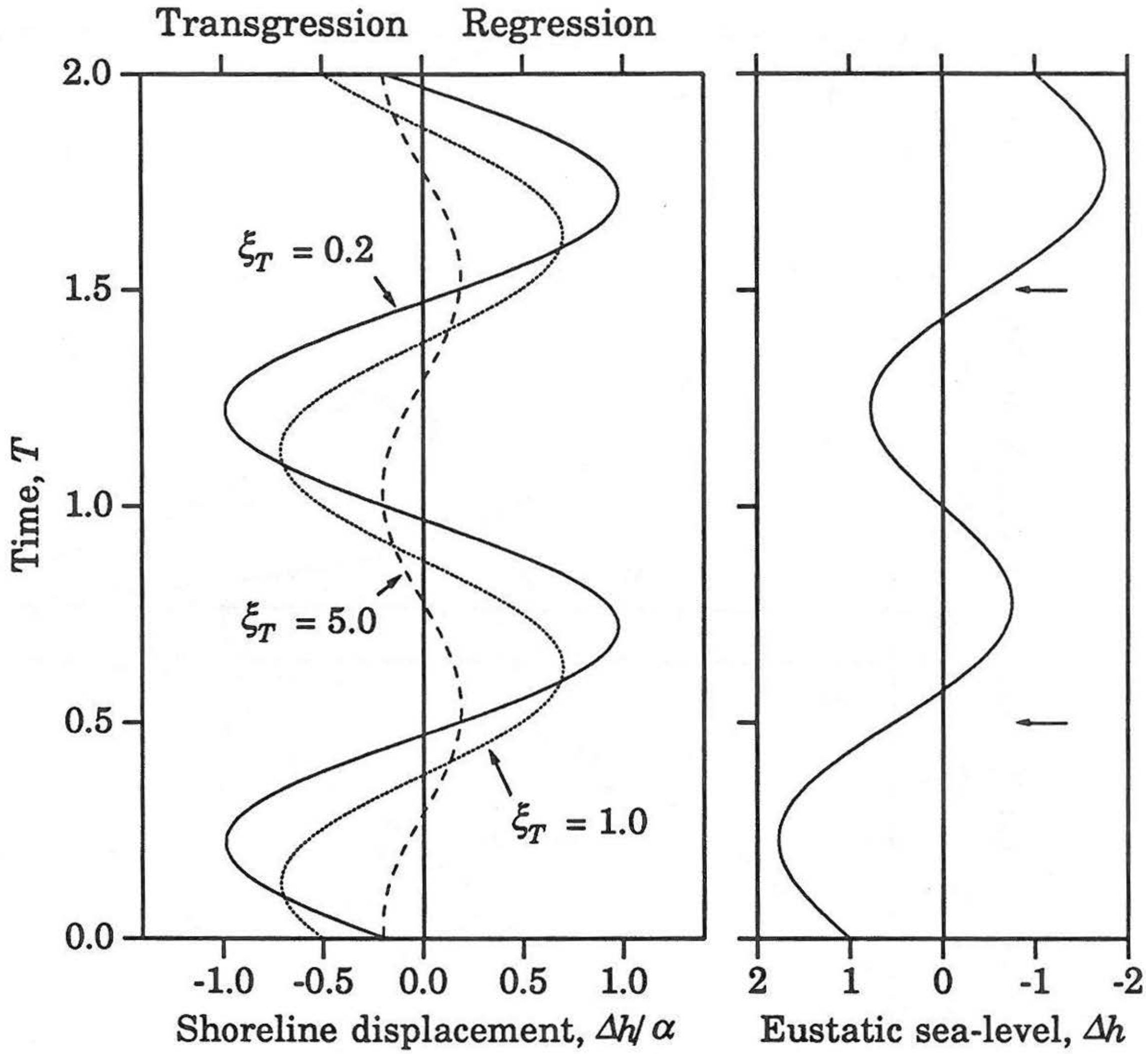

Figure 8.5 Displacement of shoreline from equilibrium position, caused by a sinusoidal eustatic oscillation of period T and amplitude Δh (from equation 8.9). The amplitude and phase of the shoreline oscillation vary with ξ_T (= $2\pi\tau_T/T$) a dimensionless tectonic parameter that measures the influence of the tectonic time constant (τ_T). When ξ_T is small, the maximum regression occurs as sea level is falling most rapidly. For large values of ξ_T, the maximum regression is synchronous with the eustatic lowstand. From one margin to another, the timing of transgressions and regressions can vary by as much as one-fourth the period of the eustatic oscillation due to variations in subsidence rate and geometry.

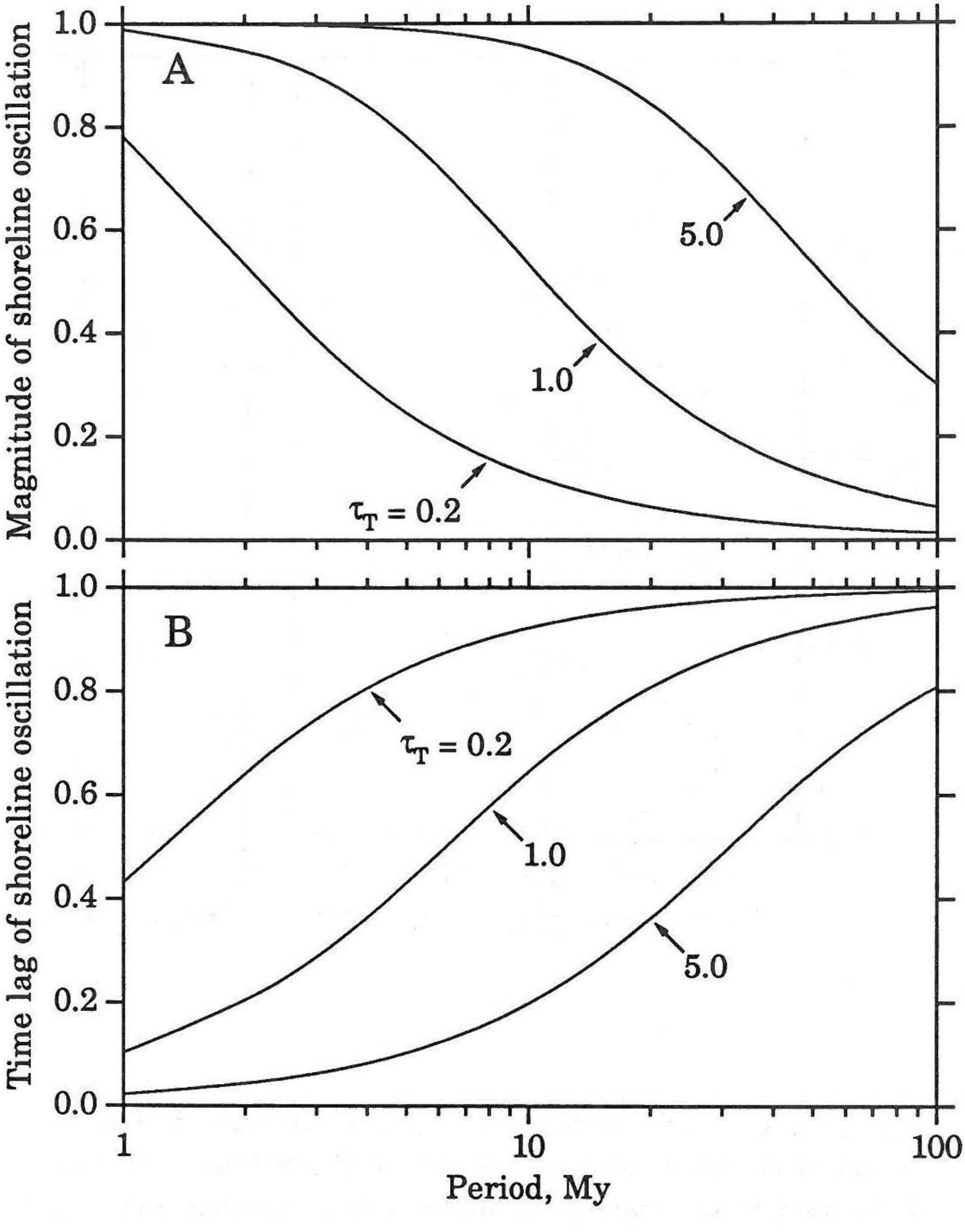

Figure 8.6 (A) Dimensionless magnitude (from equation 8.9) and (B) dimensionless time lag (from equation 8.10) of the shoreline oscillation as a function of the period of the eustatic oscillation (T) and tectonic time constant (τ_T). See text for more information.