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.elationship between eustacy and :ratigraphic sequences of passive margins

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3STRACT

It is commonly thought that transgressive or regressive events at may have occurred simultaneously on geographically dispersed ntinental margins have been caused by worldwide sea-level rise fall, respectively. Instead, it is shown here that these events may caused by changes in the rates of sea-level rise or fall. The subsince of an Atlantic-type (passive) margin may be modeled as a ›rdering platform rotating downward about a landward hinge ıe. The rate of subsidence is greatest at the seaward side of the atform and decreases landward to zero at the hinge line. With the :ception of sea-level changes due to glaciation, dessication, and ›oding of small ocean basins and other sudden events, the rate of ıbsidence at the seaward edge of the platform (shelf edge) is ·eater than the rate at which sea level may possibly rise or fall. hus, if sea level is falling, the shoreline will seek that point on the ıbsiding platform at which the rate of sea-level fall is equal to the .te of subsidence minus the sedimentation rate. If the rate of seavel fall decreases, the shoreline will move landward; if the rate in·eases, the shoreline will migrate seaward. If sea level is rising, the ıoreline will move to that point where the rate of sea-level rise is]ual to the sedimentation rate minus the subsidence rate. Thus, if ıe rate of sea-level rise decreases, the shoreline will move seaward; the rate increases, the shoreline will move landward. The position f the shoreline is also a function of the sedimentation rate. These :lationships have been quantified so that the position of the ıoreline and the thickness of the sediments deposited during disrete time intervals may be computed as a function of the rate of :a-level change and the sedimentation rate. A sea-level curve, ased on volume changes of the mid-oceanic ridge system, has been omputed. Sea level is seen to fall persistently from Late Cretaceous › middle Miocene time, but transgressions occur in Eocene and arly Miocene time because the rate of sea-level fall is slower for ıese periods. It is concluded also that the presence of the shoreline eaward of the shelf edge of an Atlantic margin should be ymptomatic of events that may cause rapid sea-level fall, such as lacial build-up or the sudden flooding of large deep basins.

NTRODUCTION

The sediments of continental margins (and cratonic basins) often ecord advances (transgressions) and retreats (regressions) of the eas that are synchronous over widely distributed geographic areas. 3ecause of synchroneity of these events, they have usually been atributed to fluctuations in worldwide sea level. In particular, landvard movement of the shoreline is presumed to have been caused ›y sea-level rise and seaward movement by sea-level fall. However, his is not necessarily the case. Rather, the shoreline tends to stabilize at that point on a margin where the rate of rise (or fall) of sea level is equal to the difference between the rate of subsidence of the platform and the rate of sediment infill. Under these conditions, if sea level is made to rise more rapidly or fall more slowly a transgression will occur. Alternatively, if sea level is made to rise less rapidly or fall more rapidly a regression will occur. In this paper a quantitative relationship between the position of the shoreline and the rates of subsidence, sea-level change, and sedimentation at Atlantic-type margins is developed. The various mechanisms that may cause sea level to change are considered quantitatively. It is shown that volume changes of the mid-oceanic ridge system give potentially the greatest rates and greatest magnitude of sea-level change. A hypothetical sea-level curve for Late Cretaceous to Miocene time is calculated from the changes in volume of the mid-oceanic ridge system. The model developed here quantitatively relates the position of the shoreline to the rates of subsidence, rates of sea-level change, and rates of sedimentation. This model is used with the hypothetical sea-level curve to compute a transgressive-regressive sequence and a lithostratigraphic section for an Atlantic margin.

CAUSES OF SEA-LEVEL CHANGE

Eustatic changes in sea level have been attributed to a variety of processes. Some processes would probably yield slow long-period changes, such as an increase in the volume of ocean water due to generation of juvenile water at mid-oceanic ridge axes and island arcs or changes in the capacity of the ocean basins due to continuing differentiation between continental and oceanic lithosphere. Other causes involve events of shorter duration, which in some cases are reversible; examples of this type are the waxing and waning of grounded continental ice sheets, variations in sediment flux, and changes in the volume of the mid-oceanic ridge system. These various processes are considered quantitatively here, to show that except for glacial variations and phenomena such as the dessication and flooding of small ocean basins, volumetric changes in the mid-oceanic ridge system are potentially the fastest way to change sea level.

Several of the mechanisms to be discussed below would actually cause the depth of the ocean waters adjacent to the continents to change, but some of this change of depth would be compensated by isostatic adjustment of the ocean basins relative to the continents. If the change in water depth is h, the consequent adjustment of this ocean basin (up or down) relative to the continent is d, and the density of the upper mantle is taken to be 3.4 g/cm^3, the change in sea level or freeboard of the continents will be $(h - d) = 0.7\,h$. If we speak of a rate of change of water depth R, then the rate of change of sea level or continental freeboard will be $0.7\,R$.

Geological Society of America Bulletin, v. 89, p. 1389–1403, 8 figs., 2 tables, September 1978, Doc. no. 80911.

Throughout this report, sea-level change will be equivalent to a change of continental freeboard. "Transgression" will mean any landward movement of the shoreline, and "regression" will refer to seaward movement of the shoreline.

EFFECT OF CONTINUING DIFFERENTIATION OF LITHOSPHERE

The plate tectonic process involves the nearly continuous overturn of the oceanic lithosphere. This process, which has been occurring during much of geologic time, is regarded as the most important mechanism for causing differentiation between heavier (oceanic) and lighter (continental and volatile) materials (see Hess, 1962; Hurley, 1968; Hallam, 1971; Wise, 1974). It will be argued below that near-equilibrium conditions were reached perhaps prior to Phanerozoic time and that there has been little or no net differentiation during this interval. However, when and if net differentiation is taking place, the effects of this on sea level might be as follows (Hurley, 1968; Condie and Potts, 1969; Hallam, 1971; Wise 1974). Waters may be added to the oceans by the production of juvenile water at active ridge axes and at island arcs. It also might be removed by the process of hydrothermal alteration of oceanic crust and by subduction of the oceanic lithosphere and part of its sedimentary overburden. If continental material is continuously differentiated from the convecting lithosphere, it may be added to the continents so as to cause them to thicken, thereby increasing the volumetric capacity of the ocean basins, and/or it may be added laterally, causing a decrease in the volume of the ocean basins.

It may be argued that because the overturn has probably been occurring persistently, if at varying rates, for more than 2 b.y. (J. Dewey, 1976, personal commun.), a near-equilibrium condition should have been reached such that net differentiation is nearly zero. If there is any net change in sea level due to these processes it would most likely appear as very gradual emergence or submergence of the continents varying in rate over periods greater than 10^8 yr. Curves for the Phanerozoic giving percentage of land flooded versus time (for example, Wise, 1974, from Schuckert, 1955) suggest persistent emergence from early Paleozoic time to the present. Hallam (1971) has proposed that this might have been caused by continuing differentiation between continental and oceanic materials, but it may be assumed that this emergence represents the net effect of all the processes discussed above. From the Egyed/Termier global percent flooding curve presented by Wise (1974, Fig. 3) and his hypsometric curve (Wise, 1974, Fig. 5) to convert percentage of land flooded to paleo–sea level, the rate of sea-level lowering (emergence) for Phanerozoic time is calculated to be 0.02 cm/1,000 yr. However, Wise (1974) has concluded that this emergence is artificial. The difficulty arises from the fact that the paleogeographic maps from which the curves were made span longer intervals of time for the Paleozoic than for the Mesozoic, and the time span for the Mesozoic maps is larger than for the Cenozoic. Also, areas that are now submerged but may have been exposed in the past are not included in the tabulations. The effect is to make the world of the past appear relatively more wet. Thus, since early Paleozoic time, the net effect of possible continued differentiation on sea level has probably been even less than that calculated above. As is shown below, freeboard area is often more sensitive to the rate of sea level change than to its absolute value.

SEDIMENT BUILD-UP AND REMOVAL

Seuss (1906) suggested sediment infill of the ocean basins as a means of raising sea level. Sea level might also be lowered by sediment removal at subduction zones. It is possible to quantify these processes in an approximate way. The present-day average rate of erosion of land surfaces is about 3 cm/1,000 yr (Holmes, 1965). During Late Cretaceous and early Tertiary time, when there was little or no continental ice, the ratio of sea area to land area would have been greater than 3 to 1. If present-day erosion rates may be regarded as applicable to Late Cretaceous and early Tertiary time, then the amount of terrigenous sediment being transported to the sea would have been equivalent to depositing a layer of continental rock 1 cm thick every 1,000 yr over the oceans. However, part of this would be accommodated by subsidence. If the density of the continental rock is taken to be 2.8 g/cm^3 and the density of the upper mantle is 3.4 g/cm^3, then after subsidence the effect of the influx of sediment would be to raise sea level by 0.25 cm/1,000 yr. The question is whether enough sediment is being transported into the subduction zones to offset the 0.25cm/1,000 yr of sea-level rise due to sediment infill. Sediment contains interstitial water. The rate of sediment infill needed to raise sea level by 0.25 cm/1,000 yr is 0.33 cm/1,000 yr. The volume of the sediment that must be subducted or obducted to maintain a net sea-level change of zero is 1,200 km^3/1,000 yr. The present length of the subduction zone is approximately 40,000 km, and the average rate of convergence is 6 cm/yr (Le Pichon, 1968); thus, the thickness of the sedimentary wedge being squeezed out at the subduction zone would have to be about 0.5 km. The average sediment thickness found on the outer wall of trenches today is about 0.4 km (W. Ludwig, 1976, personal commun.).

The rate of sediment delivered to the oceans is not necessarily in

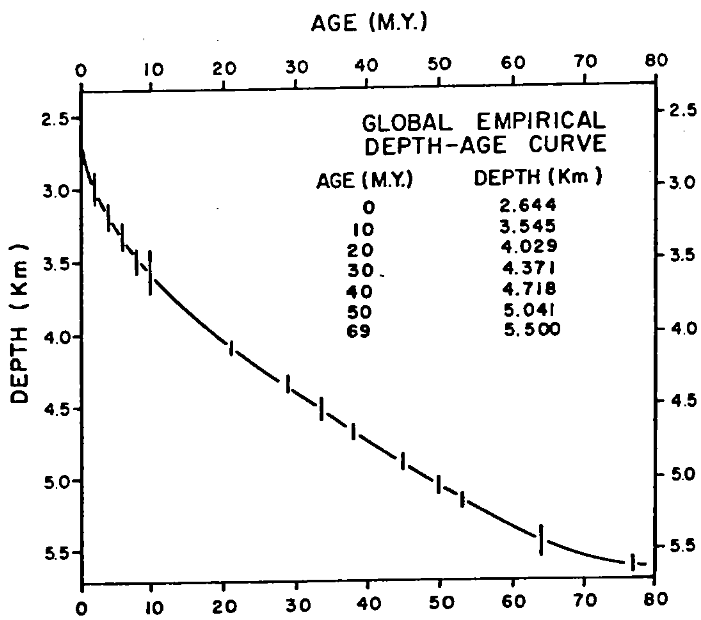

Figure 1. Age versus depth for oceanic crust. Vertical bars show ages where depth data were averaged. Vertical extent of each bar gives standard deviation for each data set. Table gives age versus depth data at 10-m.y. intervals; these data have been interpolated from graph. Data from Sclater and others (1971).

balance with the rate of sediment removal. The former depends on both climatic conditions and the relief of the continents. The latter depends on subduction rates, the length of subduction zones, and the thickness of sediments at the subduction zones; but it seems unlikely that an imbalance could cause sea-level changes at rates in excess of 0.1 cm/1,000 yr.

CRUSTAL SHORTENING AND THERMAL WELTS

Orogeny may cause changes in the volume of the ocean basins and may be considered quantitatively. For example, the Himalayas were uplifted in Tertiary time as a result of collision of the Indian subcontinent with Eurasia which caused extensive crustal shortening. The great-circle length across the compressional front was approximately 3,000 km. The depth of the oceanic basin created south of India was 5.5 km, and the rate of convergence was 5 cm/yr or less (P. Molnar, 1976, personal commun.). The consequent rate of decrease in ocean depth would have been 0.22 cm/1,000 yr, and the rate of change of freeboard of the continents would be 0.15 cm/1,000 yr.

Occasionally ocean basins traverse hotspots, creating volcanic welts such as the Hawaiian seamount chain. If we assume that the welt thus created has the cross section of a trapezoid 500 km wide at the base, 5.5 km high, and 100 km wide at the top and if the hotspot moves at 6 cm/yr, the effect would be to deepen the ocean by 0.025 cm/1,000 yr and raise sea level by 0.02 cm/1,000 yr. This calculation neglects the fact that because of thermal contraction aseismic ridges also subside with time, and hence the rate of 0.02 cm/1,000 yr is a maximum.

VOLUME CHANGES OF MID-OCEANIC RIDGE SYSTEM

Menard (1969), Hallam (1963), Russell (1968), and Valentine and Moores (1972) have all suggested that major fluctuations in sea level may be caused by changes in volume of the mid-oceanic ridge systems. Hays and Pitman (1973) were able to show quantitatively that the Late Cretaceous transgression and subsequent Cenozoic regression could be explained by expansion and contraction of the mid-oceanic ridge system.

The argument is as follows: Age versus depth relationships for all mid-oceanic ridges are the same (Fig. 1) regardless of the spreading history. A curve of age versus depth constructed using data from the equatorial North Pacific, where spreading rates have been fast (fluctuating around 6 cm/yr), will fit quite precisely similar data taken from the central Atlantic, where spreading has been slow, or the Indian Ocean, where rates have been quite variable, or the Labrador Sea, where spreading ceased some tens of millions of years ago (Sclater and others, 1971). The age versus depth relationship approximately follows a time-dependent exponential cooling curve (McKenzie and Sclater, 1969). Newly formed oceanic crust lies at a depth of about 2.7 km, and crust that is 69 m.y. old will have subsided to a depth of approximately 5.5 km. At this depth the ridge flank is often truncated by an abyssal plain. In the calculations, the 5.5-km datum (the average depth of the abyssal plain) will be considered the base level of the ridge system because only the part that protrudes above the abyssal plains affects sea level. By multiplying the time axis of Figure 1 by a spreading rate, the curve may be converted to depth versus distance. It may thus be seen that a ridge that has spread at 6 cm/yr for 69 m.y. (B in Fig. 2, top) will have precisely three times the cross-sectional area of one that has spread at 2 cm/yr for 69 m.y. (see A in Fig. 2, top). It follows that if the spreading rate of the 2-cm/yr ridge were to be increased to 6 cm/yr, the cross-sectional area of the ridge would gradually increase (A in Fig. 2). Conversely, a decrease in spreading rate from 6 cm/yr to 2 cm/yr would cause a decrease in cross-sectional area (B in Fig. 2).

To illustrate the potential effect on eustacy of changes in the volume of the mid-oceanic ridge system, we will calculate the sea-level changes caused by an increase in spreading rate from 2 cm/yr to 6 cm/yr and then a decrease in spreading rate from 6 cm/yr to 2 cm/yr

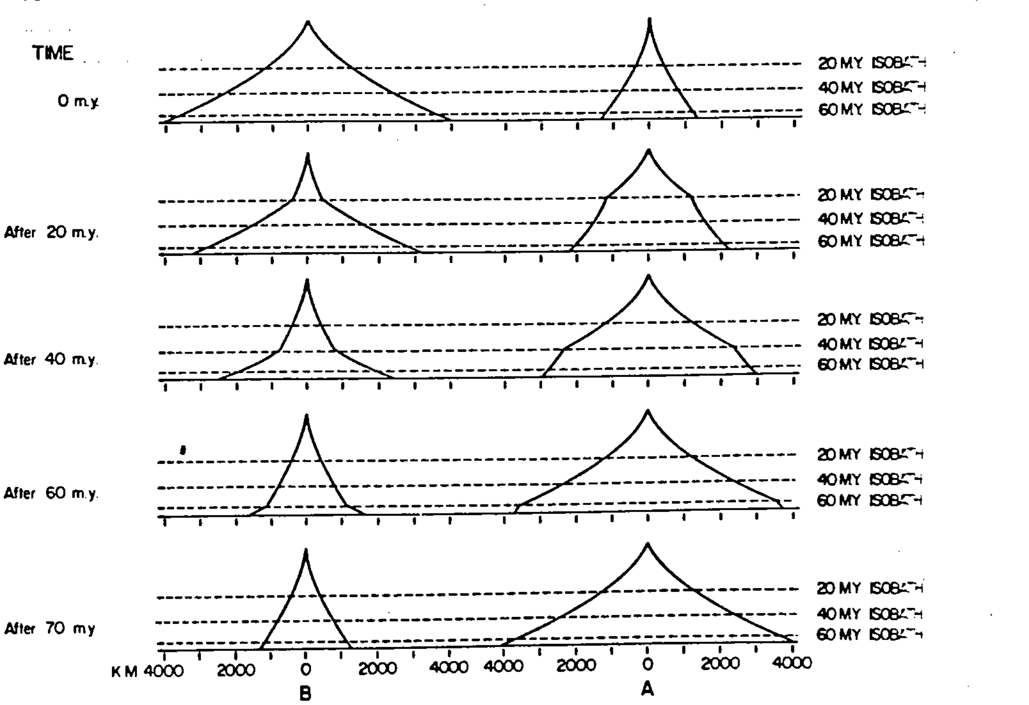

Figure 2. A, top shows profile of ridge that has been spreading at 2 cm/yr for 70 m.y. At time 0 m.y., spreading rate is increased to 6 cm/yr. Sequential stages in consequent expansion of ridge profile are shown: first at 20 m.y. after spreading rate change, at 40 m.y., then at 60 m.y., and finally at 70 m.y. after spreading-rate change. At 70 m.y., ridge will be at a new steady-state profile; cross-sectional area at this time will be three times what it was at 0 m.y. B, top is profile of ridge that has spread at 6 cm/yr for 70 m.y. At 0 m.y., spreading rate is reduced to 2 cm/yr. Sequential stages in subsequent contractions of ridge are shown. At 70 m.y., after change in spreading rate, ridge will be at a new steady-state profile; cross-sectional area will be one-third what it was at 0 m.y.

for a hypothetical ridge segment 10,000 km long. We assume that the Earth is not expanding or contracting. Thus, if worldwide spreading rates increase (or decrease), worldwide subduction rates must increase (or decrease) correspondingly; the mass of the oceanic lithosphere remains constant. To compute the ridge volume and changes in the ridge volume, we have used the data listed in Figure 1 to divide the cross section of the ridge into a set of adjacent trapezoids each 10 m.y. in width. The elevation of the sides of the trapezoids fits the age versus depth curve. Knowing the spreading-rate history and ridge length, the volume and change of volume of the ridge may be computed for each 10 m.y. To convert the ridge volume change to sea-level change (that is, change of continental freeboard), two corrections must be made. First, if the depth of the oceans increases by a thickness h, the ocean basins will subside a distance d, and as shown above the change in freeboard will be 0.7 $h = (h - d)$.

The second correction arises because of the shape of the vessel that contains the ocean waters. As the ocean rises, more surface area is covered. Approximately one-sixth of the Earth's surface area (85×10^6 km^2) lies at elevations between 0 and 500 m (Sverdrup and others, 1942). We assume that as sea level rises the area covered by the sea increases linearly (0.17×10^6 km^2 for each 0.001-km rise in sea level). Thus, the change in continental freeboard $(h - d)$ due to a change in volume of the mid-oceanic ridge system (ΔV) may be calculated from the equation $\Delta V = A_0 \cdot h + (0.7\,h)^2 \times 170/2$, where ΔV Km^3 = the difference between the present ridge volume and the volume at time t, and A_0 = the present-day area of the oceans (360×10^6 km^2).

The change in sea level, due to the expansion and contraction of the hypothetical ridge during a 140-m.y. interval, is shown in Figure 3. It has been assumed that prior to 0 m.y. the ridge has been spreading at 2 cm/yr. for at least 69 m.y. and thus has reached a steady-state profile. At 0 m.y. the spreading rate is increased to 6 cm/yr and stays at this rate until 70 m.y. The sea level rises rapidly at first, then gradually levels off as the ridge expands to its new equilibrium profile. At 70 m.y. the rate of spreading is reduced to 2 cm/yr and stays at the reduced rate until 140 m.y. Sea level drops until a new contracted equilibrium profile is reached at 140 m.y.

The magnitude of the sea-level change is 113 m, and the rate of sea-level change is 0.37 cm/1,000 yr for the first 10 m.y., immediately subsequent to the change in spreading rate. This is greater by a factor of two than rates due to any other cause. The rate of sea-level change is gradually reduced to zero as a new equilibrium profile is reached.

Other ways of expanding or contracting the volume of the ridges are (1) creation of a new ridge system by rifting or (2) destruction of an old ridge system by cessation of spreading or subduction. The ridge volume and hence sea level are a function of spreading rate and ridge lengths. A number of different combinations of changes in both of these parameters will produce the same sea-level curve as shown in Figure 3. Given a ridge system of a constant 10,000-km length, any change in the spreading rate of +4 cm/yr for 70 m.y. and then −4 cm/yr for the next 70 m.y. would yield the same curve. Alternatively, if the ridge length was 40,000 km, a spreading-rate change of +1 cm/yr and then −1 cm/yr would yield the same curve. If the spreading rate increase (or decrease) at a 40,000-km-long ridge were 3 cm/yr for 70 m.y., the ultimate rise (or fall) in sea level would be 339 m, and for the first 10 m.y. the rate of rise (or fall) would be 1.11 cm/1,000 yr. Several other modifications of the ridge system are conceivable. It might be assumed that a maximum rate of sea-level change due to alteration of the geometry of the mid-oceanic ridge system would be about 1 cm/1,000 yr.

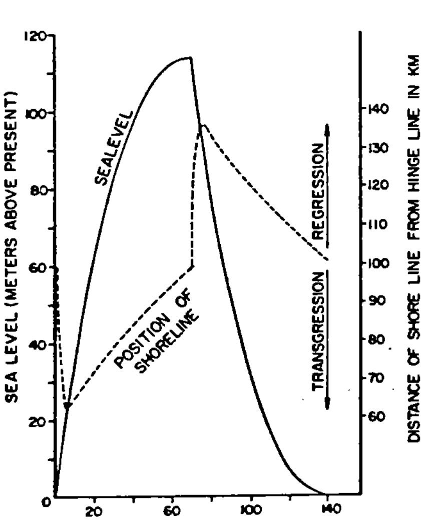

Figure 3. Change in sea level (solid line) is caused by change in spreading rate of ridge 10,000 km long, where ridge has been spreading at 2 cm/yr for a long time. At 0 m.y., spreading rate changes to 6 cm/yr and stays at this rate until 70 m.y., when rate is reduced to 2 cm/yr and stays at this lower rate until 140 m.y. Dashed line shows position of shoreline as a function of rate of sea-level change.

SEA-LEVEL CHANGES, LATE MESOZOIC TO PRESENT

The computations above were made to show that quantitatively large sea-level changes would be caused by geologically reasonable changes in the geometry of the mid-oceanic ridge system. Using spreading-rate data for the mid-oceanic ridge system for the past 180 m.y., Hays and Pitman (1973) demonstrated quantitatively that the Late Cretaceous high sea-level stand and the subsequent fall could be explained by expansion and contraction of the mid-oceanic ridge system in the manner discussed above. This phase of expansion and contraction was caused mainly by a pulse of rapid spreading in Late Cretaceous time (from 110.5 to −84.5 m.y. B.P.; Larson and Pitman, 1972). The calculations of Hays and Pitman (1973) were approximate because much of the necessary data were imprecise, the spreading history of the Indian Ocean ridge system was poorly known, and that of the Tethys was unknown.

The improved data set now available allows significant improvement. The history of the Pacific ridges is now better known (Larson and Hilde, 1974; Cande, 1976; R. L. Larson, 1976, personal commun.), and because of the work of Weissel and Hayes (1971), McKenzie and Sclater (1971), Fisher and others (1971), Sclater and Fisher (1974), and Johnson and others (1976), a reasonably complete picture of the evolution of the Indian Ocean ridges has been developed. As will be seen, the various segments of the ridge system do not necessarily act synchronously; some contract while others expand, but the net effect since Cretaceous time has been contraction in volume and hence sea-level lowering. The ridge volume (and change in volume) is calculated for each 10 m.y. from 85 to 15 m.y. B.P. (at 15 m.y. B.P., middle Miocene, glacial fluctuations may dominate the sea-level curve). To compute the ridge volume at 85 m.y. B.P. the spreading history of each of the ridge segments must be known back to 155 m.y. B.P., but many segments of that age have been subducted. The spreading data for these ridge segments must be obtained by extrapolation and thus could be the source of significant error.

Another source of significant error is in the computed spreading rate, which depends on the reliability of the magnetic polarity time scale (Heirtzler and others, 1968; Larson and Pitman, 1972; Larson and Hilde, 1975). Key parts of the time scale have been tested paleontologically, but calibration with absolute time depends upon the correlation of paleontologic age with absolute age. The time scale used for this purpose by Larson and Pitman (1972) is the Phanerozoic (Anonymous, 1964), modified by Berggren (1969) for Cenozoic time. Berggren and others (1975) and Baldwin and others (1974) have proposed modifications to the paleontologic time scale that would drastically alter the spreading rates. However, in both instances the modifications are dependent upon recalibrations of the paleontologic time scale that are not yet generally accepted.

The major modification of the ridge system of Hays and Pitman (1973) used in this paper is the inclusion of the Indian Ocean ridge system and a ridge system in the Tethys. The data used are listed in Table 1. Ridge lengths and spreading rates are given for sequential 10-m.y. intervals. The ridge axes change length with time as new ridges are created, as ridge axes are subducted, and as the direction of relative motion between plates changes. Flanking parts of ridges also change length as they are subducted. The present-day arrangement of ridges in the Pacific displays all of these characteristics (Pitman and others, 1974).

The computed sea-level change is plotted in Figure 4. Sea level is seen to be 350 m higher than at present in latest Cretaceous time (85 m.y. B.P.). Hays and Pitman (1973) obtained a height of sea level of 521 m above present for 85 m.y. B.P. The value of 350 m obtained in this study is probably more realistic. Eustatic sea-level rise of 350 m would be sufficient to cover about 35% of the present subaerial surface. This is in reasonable agreement with the paleogeographic data of Ronov (1968) and Strakhov (1962; see Hays and Pitman, 1973) and with the elevation of the Late Cretaceous shoreline as determined by Sleep (1976). An early Coniacian shoreline extends from the Mesabi Range to the southwestern boarder of Minnesota, a distance of 160 km (Sleep, 1976). It lies at about 375 m above present sea level. Its elevation is fairly uniform, indicating that local tectonism has been insignificant. Regionally there has been post-Cretaceous erosion, which may have caused about 50 m of uplift (Sleep, 1976). Sleep (1976) concluded that the sea probably was at 325 m above present in Coniacian time, which is in good agreement with the value calculated here. Sea level dropped slowly from 85 to 65 m.y. B.P. During Paleocene time, sea level dropped quite rapidly — the rate of fall increased slightly in early Eocene time, then decreased in the late Eocene, and increased once more in Oligocene time. The rate of sea-level fall is seen to be a maximum of nearly 0.67 cm/1,000 yr, a rate far below that caused by the fluctuations in continental glaciation but greater by at least a factor of three than that caused by any other process. One of the puzzling observations about the sea-level curve is that sea-level fall through early Tertiary time was continuous, although at varying rates. Yet the geologic record consistently shows an Eocene transgression (Hallam, 1963), and it has often been presumed that transgressions of a global nature must be caused by a sea-level rise. The computed rate of sea-level fall for late Eocene time (45 to 35 m.y. B.P.) is 0.37 cm/1,000 yr. It seems unlikely that other mechanisms for raising sea level such as sediment infill or ocean volcanism would have been sufficient to offset this rate of fall. It also seems unlikely that the uncertainties in the data would allow errors of this magnitude. Rather, as will be shown in the next section, the worldwide transgression may have been caused by the reduction in the rate of sea-level fall.

Figure 4. Solid line shows change in sea level due to change in ridge volume for period from 85 to 15 m.y. B.P. It is assumed that by 15 m.y. B.P. glacial build-up in Antarctica may have been sufficient that fluctuations in glacial mass became dominant mechanism causing sea-level changes. Dashed line gives position of shoreline relative to hinge line as a function of rate of sea-level change.

SEDIMENTARY HISTORY OF ATLANTIC-TYPE MARGINS

The problem we are dealing with is the interaction between a slowly subsiding continental margin and a more slowly changing sea level. Many of the arguments presented here are similar to those of Sloss (1962), who pointed out that the occurrence of transgression and regression is in part controlled by the relative magnitude of the rate of sea-level change and the rate of subsidence.

The general structure of Atlantic-type margins is that of a seaward-thickening mass of systematically stratified sediment overlying a deeply subsided, faulted basement platform (see, for example, Drake and others, 1959; McIver, 1972; Rona, 1973; Schlee and others, 1976; Kinsman, 1975). The sedimentary strata consist of seaward-thickening wedges separated, at least in the shallower sections, by remarkably undisturbed planar horizons. The deepest strata are often disturbed by basement horsts and grabens, reefal structures, and diapirs. The thickness of the sediment that has been accumulated at, for example, the shelf edge off the east coast of New Jersey is well in excess of 10 km. Borehole data taken at or near the shelf edge (see, for example, McIver, 1972; Brown and others, 1972) and data from dredges taken at the shelf edge (Fox and others, 1970) show that the entire sedimentary section that underlies the shelf, from the surface to the basement, was deposited at or within several hundred meters of sea level. This means that the basement that floors the section has subsided slowly but persistently through time to a near–sea-level position at the time of the rift-drift transition to its present depth.

Subsidence at these margins continues today. Subsidence appears to be caused by two simultaneously occurring mechanisms (Sleep, 1971; Kinsman, 1975; Watts and Ryan, 1976). The first of these is a driving subsidence, which is attributed to increased density of the rifted basement caused by thermal contraction and/or phase changes. Analysis of the postrift subsidence that has occurred at several margins shows that the driving subsidence decreases exponentially with time, with about a 50-m.y. time constant (Sleep, 1971; Watts and Ryan, 1976). The curve that describes the tem-

TABLE 1. RIDGE LENGTHS, AGES, AND SPREADING RATES

Ridge segment age*	Age of ridge axis (m.y.) 85		75		65		55		45		35		25		15		0	
	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)	L (km)	SR (cm/yr)
Pacific-Farallon																		
0–10	9,000	8.0	9,000	5.0	9,000	5.0	9,000	4.0	9,000	5.0	9,000	6.0	8,500	6.0	8,000	5.0	7,000	5.0
10–20	8,000	8.0	9,000	8.0	9,000	5.0	9,000	5.0	9,000	4.0	8,500	5.0	8,200	6.0	7,900	6.0	7,200	5.0
20–30	7,000	8.0	8,000	8.0	9,000	8.0	9,000	5.0	9,000	5.0	8,000	4.0	7,500	5.0	7,400	6.0	7,500	6.0
30–40	6,000	3.5	7,000	8.0	8,000	8.0	9,000	8.0	8,000	5.0	8,000	5.0	7,200	4.0	6,500	5.0	6,500	6.0
40–50	5,000	3.5	6,000	3.5	7,000	8.0	7,000	8.0	8,500	8.0	7,000	5.0	7,000	5.0	6,300	4.0	5,500	6.0
50–60	4,000	3.5	5,000	3.5	6,000	3.5	6,000	8.0	6,000	8.0	6,500	8.0	6,000	5.0	6,000	5.0	4,500	5.0
60–70	3,000	3.5	4,000	3.5	5,000	3.5	5,000	3.5	5,000	8.0	5,000	8.0	5,000	8.0	5,000	5.0	5,000	4.0
Phoenix-Farallon																		
0–10	4,000	12.0	3,700	5.0	3,400	5.0	3,100	3.0	2,800	4.0	2,500	4.0	2,200	4.0	2,000	3.0	2,000	3.0
10–20	4,000	12.0	3,700	12.0	3,400	5.0	3,100	5.0	2,800	3.0	2,500	4.0	2,200	4.0	2,000	4.0	2,000	3.0
20–30	4,000	12.0	3,700	12.0	3,400	12.0	3,100	5.0	2,800	5.0	2,500	3.0	2,200	4.0	2,000	4.0	2,000	3.0
30–40	4,000	5.0	3,700	12.0	3,400	12.0	3,100	12.0	2,800	5.0	2,500	5.0	2,200	3.0	2,000	4.0	2,000	4.0
40–50	4,000	5.0	3,700	5.0	3,400	12.0	3,100	12.0	2,800	12.0	2,500	5.0	2,200	5.0	2,000	3.0	2,000	4.0
50–60	4,000	5.0	3,700	5.0	3,400	5.0	3,100	12.0	2,800	12.0	2,500	12.0	2,200	5.0	2,000	5.0	2,000	4.0
60–70	4,000	5.0	3,700	5.0	3,400	5.0	3,100	5.0	2,800	12.0	2,500	12.0	2,200	12.0	2,000	5.0	2,000	3.0
Kula-Farallon																		
0–10	1,500	4.0	1,500	4.0	1,500	4.0	900	4.0	450	4:0	450	4.0	..	0.0	..	0.0	..	0.0
10–20	1,500	4.0	1,500	4.0	1,500	4.0	900	4.0	450	4.0	450	4.0	..	0.0	..	0.0	..	0.0
20–30	1,500	4.0	1,500	4.0	900	4.0	450	4.0	450	4.0	450	4.0	450	4.0	..	0.0	..	0.0
30–40	1,500	4.0	1,500	4.0	900	4.0	450	4.0	450	4.0	450	4.0	450	4.0	..	0.0	..	0.0
40–50	1,500	4.0	900	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	0.0	..	0.0
50–60	1,500	4.0	900	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	0.0	..	0.0
60–70	1,500	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	4.0	450	0.0	..	0.0
Pacific-Phoenix																		
0–10	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2	3,350	2.2	3,350	2.2	3,350	2.4	3,350	2.4	3,350	4.7
10–20	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2	3,350	2.2	3,350	2.2	3,350	2.4	3,350	2.4
20–30	3,350	18.5	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2	3,350	2.2	3,350	2.2	3,350	2.4
30–40	3,350	5.0	3,350	18.0	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2	3,350	2.2	3,350	2.2
40–50	3,350	5.0	3,350	5.0	3,350	18.0	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2	3,350	2.2
50–60	3,350	5.0	3,350	5.0	3,350	5.0	3,350	18.0	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5	3,350	2.2
60–70	3,350	5.0	3,350	5.0	3,350	5.0	3,350	5.0	3,350	18.0	3,350	18.0	3,350	18.0	3,350	3.5	3,350	3.5
Pacific-Kula																		
0–10	5,000	4.0	5,000	4.0	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	..	0.0	..	0.0	..	0.0
10–20	5,000	4.0	5,000	4.0	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	..	0.0	..	0.0	..	0.0
20–30	5,000	4.0	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	..	0.0	..	0.0
30–40	5,000	4.0	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	..	0.0	..	0.0
40–50	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	..	0.0
50–60	5,000	4.0	3,000	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	..	0.0
60–70	5,000	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	1,500	4.0	..	0.0
Africa–North America																		
0–10	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0
10–20	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0
20–30	4,400	3.0	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0
30–40	4,400	1.0	4,400	3.0	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0
40–50	4,400	1.0	4,400	1.0	4,400	3.0	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0	4,400	2.0
50–60	4,400	1.0	4,400	1.0	4,400	1.0	4,400	3.0	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0	4,400	2.0
60–70	4,400	1.0	4,400	1.0	4,400	1.0	4,400	1.0	4,400	3.0	4,400	3.0	4,400	3.0	4,400	2.0	4,400	2.0

Africa–South America																		
0–10	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0
10–20	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0
20–30	4,950	5.0	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0
30–40	..	0.0	4,950	5.0	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0
40–50	..	0.0	..	0.0	4,950	5.0	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0	4,950	2.0
50–60	..	0.0	..	0.0	..	0.0	4,950	5.0	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0	4,950	2.0
60–70	..	0.0	..	0.0	..	0.0	..	0.0	4,950	5.0	4,950	5.0	4,950	5.0	4,950	2.0	4,950	2.0
Europe–North America–Greenland–Arctic																		
0–10	..	0.0	..	0.0	5,500	2.0	5,500	2.0	5,500	2.5	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0
10–20	..	0.0	..	0.0	..	2.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0
20–30	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0
30–40	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.0	5,500	2.0	5,500	2.0	5,500	2.0
40–50	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.0	5,500	2.0	5,500	2.0
50–60	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.0	5,500	2.0
60–70	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.0
Indian Ocean																		
0–10	9,000	2.1	8,000	8.4	8,000	8.0	8,000	7.0	9,000	2.8	8,500	2.8	6,000	1.5	5,000	1.5	5,000	1.5
10–20	9,000	2.1	9,000	2.1	8,000	8.4	8,000	8.0	8,000	7.0	8,000	2.8	6,000	2.8	5,000	1.5	5,000	1.5
20–30	9,000	2.1	9,000	2.1	9,000	2.1	8,000	8.4	8,000	8.0	8,000	7.0	6,000	2.8	5,000	2.8	5,000	1.5
30–40	9,000	2.1	9,000	2.1	9,000	2.1	9,000	2.1	8,000	8.4	8,000	8.0	6,000	7.0	5,000	2.8	5,000	1.5
40–50	9,000	2.1	9,000	2.1	9,000	2.1	9,000	2.1	9,000	2.1	8,000	8.4	6,000	8.0	5,000	7.0	5,000	2.8
50–60	..	0.0	9,000	2.1	9,000	2.1	9,000	2.1	9,000	2.1	8,000	2.1	6,000	8.4	5,000	8.0	5,000	2.8
60–70	..	0.0	..	0.0	9,000	2.1	9,000	2.1	8,000	2.1	7,000	2.1	6,000	2.1	5,000	8.4	5,000	7.0
New Zealand–Antarctic																		
0–10	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2	2,700	1.2	2,700	1.2	2,700	1.2	2,700	2.2	2,700	2.8
10–20	..	0.0	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2	2,700	1.2	2,700	1.2	2,700	1.2	2,700	2.2
20–30	..	0.0	..	0.0	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2	2,700	1.2	2,700	1.2	2,700	1.2
30–40	..	0.0	..	0.0	..	0.0	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2	2,700	1.2	2,700	1.2
40–50	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2	2,700	1.2
50–60	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	2,700	4.0	2,700	4.0	2,700	1.2
60–70	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	2,700	4.0	2,700	1.2
Australian-Antarctic																		
0–10	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.5	5,500	2.6	5,500	2.6	5,500	2.6	5,500	4.5
10–20	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.5	5,500	2.6	5,500	2.6	5,500	2.6
20–30	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.5	5,500	2.6	5,500	2.6
30–40	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.5	5,500	2.6
40–50	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	5,500	2.5
50–60	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0
60–70	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0
Tethys–Mediterranean–Red Sea																		
0–10	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0
10–20	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	2,700	3.0
20–30	4,500	3.0	4,500	3.0	4,500	3.0	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	2,700	3.0	..	0.0
30–40	4,500	3.0	4,500	3.0	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	2,700	3.0	..	0.0	..	0.0
40–50	4,500	3.0	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	..	0.0	..	0.0	..	0.0	..	0.0
50–60	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0
60–70	4,500	3.0	2,700	3.0	2,700	3.0	2,700	3.0	..	0.0	..	0.0	..	0.0	..	0.0	..	0.0

Note: L is the length of the segment; SR is the spreading rate (equal to one-half the separation rate).

* Millions of years older than the ridge axis.

poral behavior of this driving subsidence is nearly the same shape as the curve for the oceanic lithosphere (Fig. 1), except that during the first 20 to 30 m.y. after the rift-drift transition, the subsidence of the marginal platform is more rapid than would be predicted from the oceanic subsidence curve (Watts and Ryan, 1976). The sediment that fills the basin because of the driving subsidence causes further subsidence. The total rate of subsidence is generally so slow that sediment influx is sufficient to keep pace. For the first 20 to 30 m.y. after the rift-drift transition, the effect of the sediment load may best be accounted for by assuming a local isostatic (Airy) adjustment. Subsequent to this initial phase of subsidence adjustment, the sediment load is best modeled as the flexure of a rigid beam over a weak substratum. Immediately after the rift-drift transition, the rate of subsidence at the shelf edge may be on the order of 20 cm/1,000 yr (Watts and Ryan, 1976). Although the rate decreases very significantly during the first 20 to 30 m.y., the driving subsidence and thus the subsidence due to the sediment load persists. Thus, for example, at a margin such as the east coast of the United States, where the riftdrift transition probably occurred in Early Jurassic time, the post-Cretaceous subsidence rate at the shelf edge (in the vicinity of Cape Hatteras) has averaged 2.5 cm/1,000 yr (from Rona, 1973). Data from dredge hauls taken at a number of shelf-edge canyons and escarpments around the western central Atlantic show subsidence rates that average between 2 and 4 cm/1,000 yr for post-Jurassic, post-Cretaceous, and post–middle Tertiary time; in no case was the measured rate of subsidence less than 2 cm/1,000 yr (Fox and others, 1970).

Several important points relevant to the subsequent discussion may now be made. First, the rate of subsidence at the shelf edge of all passive margins (young or old) where such measurements have been made is usually greater than 2 cm/1,000 yr and is always greater than the rate at which sea level may normally fall (or rise). This means that even if sea level falls persistently for millions of years, the shoreline will always be upon the shelf; it cannot migrate over the shelf edge. Periods of rapid glacial fluctuation are excepted. It is important to note that during periods when there is very rapid lowering of sea level (~1,000 cm/1,000 yr) because of glacial build-up, the shoreline may migrate over the shelf edge; the entire shelf–coastal-plain platform may become an erosional surface; only the driving subsidence will persist, and this may be offset by rebound due to the erosion. However, in the discussion to follow, geologic periods of significant glacial fluctuation will not be considered.

Second, although the rate of driving subsidence (and hence the rate of subsidence caused by the sediment load) decreases with time, the rate of decrease is quite slow for older margins. Also, the post–Lower Cretaceous sedimentary strata that lie beneath the east coast shelf of the United States thin landward systematically and are separated by remarkably planar surfaces (Schlee and others, 1976). The relatively undisturbed appearance of these horizons suggests that the reaction to the sediment load has been one of flexure of the entire coastal-plain–shelf platform rather than local subsidence. The point of zero vertical motion is at or near the Fall Line (Watts and Ryan, 1976). For the sake of simplicity, then, we will assume that the subsidence of a mature Atlantic-type margin may be modeled as a marginal platform subsiding about a fixed landward hingeline (see Fig. 5). We will further assume that the rate of subsidence at the seaward edge of the platform is constant. Since the subsidence rates at mature passive margins are generally 2 to 4 cm/1,000 yr and since sea-level rise cannot normally exceed 1 cm/1,000 yr, it is reasonable to assume that there is always sufficient sediment to keep pace with any combination of subsidence and sea-level rise. Therefore, it will be assumed that normally the sedimentation rates across the coastal plain and continental shelf are adjusted so as to maintain a quasi-equilibrium profile relative to the sea surface. (The rise in sea level during this past 10,000 yr because of glacial melt has been so rapid, ~1,000 cm/1,000 yr, that most of the shelf areas are not now in equilibrium.) Normally, the excess sediment is deposited over the shelf break onto the slope and rise as progradational wedges; some may reach the abyssal plains. The system is loaded by sediment accumulation not only on the shelf but also on the slope and rise. If the entire system undergoes flexure due to the sediment load, then as the shoreline moves seaward, the axis of sediment loading might also move seaward, as might the effective hinge line or the axis of flexure. Alternatively, a landward migration of the shoreline might cause a landward shift of the effective hinge line. These latter possibilities have not been investigated quantitatively here, but the effect would probably be to amplify transgressions or regressions.

COMPUTATIONS OF POSITION OF SHORELINE AS A FUNCTION OF RATE OF SEA-LEVEL CHANGE

The sedimentation rate is made to vary spatially in such a way that the slope of the coastal-plain shelf is constant in spite of the subsidence. In the equation below the term $X/D \cdot R_{SS}$ gives the subsidence rate at a distance X from the hinge line, and $X_L/D\, R_{SS}$ gives the subsidence rate at the shoreline. By taking the difference of these two terms, the sedimentation rate is adjusted so that erosion ($-\, dsed/dt$) occurs landward of the shoreline and deposition ($+\, dsed/dt$) occurs seaward.

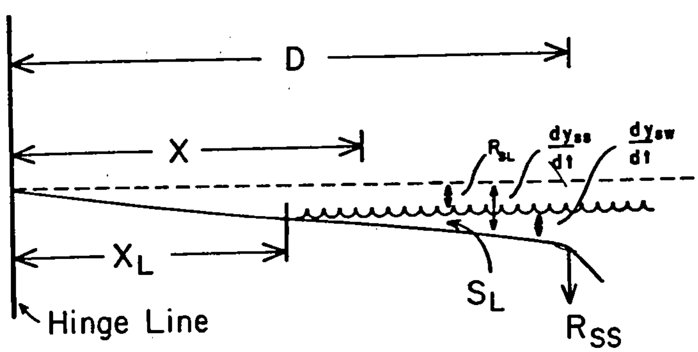

Figure 5. Schematic and simplified model of Atlantic-type margin, modeled as platform subsiding at constant rate about fixed hinge line. Thus, subsidence rate decreases linearly from maximum at shelf edge to zero at hinge line. It is assumed that sedimentation (and erosion) rates are distributed so as to maintain constant slope. D = distance from hinge line to shelf edge; X_L = distance from hinge line to shoreline; X = distance from hinge line to any point on shelf or coastal plain; S_L = shelf and coastal plain slope; R_{SS} = rate of subsidence at shelf edge of basement platform relative to a horizontal plane that extends through hinge line; R_{SL} = rate of sea-level change (positive downward) relative to same horizontal plane; dy_{SS}/dt = rate of vertical movement of shelf surface relative to same horizontal plane; dy_{WS}/dt = rate of vertical movement of sea-level surface relative to shelf surface; dx/dt = rate of movement of shoreline relative to hinge line.

In order to include the possibility that sediment build-up may occur on the coastal plain, an additional uniform sedimentation term S is added.

$$\frac{dsed}{dt} = \frac{X \cdot R_{SS}}{D} - \frac{X_L}{D} \cdot R_{SS} + S . \quad (1)$$

In this case the point of zero sedimentation rate lies landward of the shoreline. However, in spite of the fact that S is uniform over the entire coastal-plain–shelf surface, the effect of the other two terms in the equation is to ensure that the rate of deposition increases seaward of the point of zero sedimentation and that erosion increases landward. At a distance X, the rate of vertical movement (dy_{SS}/dt) of the shelf surface relative to a horizontal plane through the hinge line equals the subsidence rate minus the sedimentation rate:

$$\frac{dy_{SS}}{dt} = \frac{X_L \cdot R_{SS}}{D} - S .$$

R_{SL} = the rate of movement of the sea surface relative to the horizontal plane through the hinge line. Thus, the rate of movement of the sea surface relative to the shelf surface is

$$\frac{dy_{WS}}{dt} = R_{SL} - \frac{X_L \cdot R_{SS}}{D} + S ,$$

and the rate of change of the position of the shoreline along the coastal-plain shelf surface is

$$\frac{dx_L}{dt} = \frac{dy_{WS}}{dt} \Big/ S_L ,$$

$$\frac{dx_L}{dt} = \frac{R_{SL}}{S_L} - \frac{X_L \cdot R_{SS}}{D \cdot S_L} + S .$$

Then

$$X_L = R_{SL}\frac{D}{R_{SS}} + S \cdot \frac{D}{R_{SS}} - e^{-\frac{T \cdot R_{SS}}{D \cdot S_L}} \left(R_{SL}\frac{D}{R_{SS}} + S \cdot \frac{D}{R_{SS}} - X_{LI}\right) . \quad (2)$$

X = the position of the shoreline after time T (in thousands of years). During the time interval T, the rate of sea-level change R_{SL} was constant, and X_{LI} = the position of the shoreline at the beginning of the time interval.

In the case where $S = 0$ and T is large, say 10^7 yr, and where D = 250 km (250×10^5 cm), and the slope $S_L = 1/5{,}000$ and $R_{SS} = 2.5$ cm/1,000 yr, then

$$e^{-\frac{T \cdot R_{SS}}{D \cdot S_L}} = 1/148 ,$$

and

$$X_L \sim R_{SL}\frac{D}{R_{SS}} . \quad (3)$$

Note that in this case the equation makes sense only when sea level is lowering (that is, when R_{SL} is positive). However, it means that when sea level is lowering, the shoreline will move to that point on the subsiding shelf where the rate of sea-level fall is equal to the rate of shelf subsidence.

If S = O, that is, if there is significant uniform sedimentation over the shelf and coastal plain (and T is again large),

$$X_L = (R_{SL} + S)\frac{D}{R_{SS}} . \quad (4)$$

Then if sea level is falling, the shoreline will be shifted a constant distance ($S/R_{SS} \times D$) seaward compared with the position given by equation 3. A rise in sea level (in which case R_{SL} is negative) may be offset by sedimentation, in which case the shoreline will stabilize at an equilibrium point where the rate of subsidence plus the sedimentation rate is equal to the rate of sea-level rise.

Using equation 2, we first calculate the position of shoreline for the sea-level curve given in Figure 3. To quantify the problem, dimensions that might be typical of an Atlantic margin have been selected. D, the distance from the hinge line to the shelf edge, is taken at 250 km, which is approximately the distance from the Fall Line (which might be considered an early Tertiary hinge line) to the shelf edge south of Cape Hatteras. The thickness of Cretaceous strata at the shelf edge extrapolated from the Esso Hatteras well is 1,275 m (Rona, 1973); 350 m must be added because of post-Cretaceous sea-level fall; accordingly, post-Cretaceous subsidence is 1,625 m at a rate (R_{SS}) of 2.5 cm/1,000 yr.

As discussed above, few Atlantic-type shelves have at present an equilibrium profile: sedimentation has not caught up with the postglacial sea-level rise. However, in the vicinity of the Mississippi Delta, the sediment flux has been greater than that needed to maintain an equilibrium profile; progradation over the shelf edge has taken place. This shelf slope (S_L), which might be considered more typical for early Tertiary time, is 1/5,000 (S_L).

For the first set of computations, S is equal to 1 cm/1,000 yr. R_{SL} has been averaged for each 2-m.y. interval from the sea-level curve of Figure 3. It is assumed that prior to 0 m.y., R_{SL} has been zero for many millions of years; thus, from equation 4, X_{LI} = 100 km. At 0 m.y., sea level begins to rise rapidly, and the shoreline at first moves rapidly landward; but as the rate of sea-level rise decreases with time, the shoreline, which reaches a maximum landward position at about 6 m.y., begins to move seaward. Because the rate of sea-level rise continues to decrease until 70 m.y., the shoreline continues to move seaward. Thus, even though during this first 70 m.y. sea level has been continuously rising; because the rate of rise has continuously decreased we have seen a rapid transgression followed by a slow regression. Note that the exponential term of equation 2 is an expression of the delay that must take place between the time that a new rate of sea-level change occurs and the time when the shoreline reaches its new equilibrium position. At 70 m.y., sea level begins to fall rapidly at first and then progressively more slowly. As a consequence of this, we see first a rapid regression, but as the rate of sea-level fall decreases, the shoreline must move landward, and we see a slow transgression.

The example above illustrates the point that a regression may occur when sea level is rising simply because of a decrease in the rate of rise, and transgressions may occur where sea level is falling because of a decrease in the rate of fall. In this case, times of maximum transgression and regression coincide with time of most rapid sea-level rise and fall, respectively, and not with periods of maximum or minimum sea-level stand. Of course there is a limit to which sea level may rise without causing radical changes in the geometry of the marginal platform (and interior basin) by drowning sediment-source areas and flooding over former hinge lines.

We now compute the position of the shoreline using the sea-level curve shown in Figure 4. For these computations, D = 250 km (250×10^5 cm); R_{SS} = 2.5 cm/1,000 yr, and S_L = 1/5,000.

For the first set of computations we set S = 0 cm/1,000 yr. R_{SL} has been calculated for each 10-m.y. time interval from the sea-level curve of Figure 4 and is given in Table 2. It is assumed that in Late Cretaceous time, sea level was at a maximum height and at (or even above) the hinge line. In other words, $X_U = 0$ km at 85 m.y. B.P., so that by 65 m.y. B.P. $X_L = 9$ km, which will be X_U for the interval 65 to 55 m.y. B.P., and so forth. The computed position of the shoreline is shown in Figure 4.

The rate of sea-level fall from 85 to 65 m.y. B.P. is 0.09 cm/1,000 yr, so the shoreline moves exponentially, approaching a stable point 9 km from the hinge line. At 65 m.y. B.P., the rate of sea-level fall increases to 0.62 cm/1,000 yr, and so the strandline moves rapidly seaward, exponentially approaching an equilibrium point 62 km from the hinge line. At 55 m.y. B.P., R_{SL} increases to 0.65 cm/1,000 yr, causing a further regression, but at 45 m.y. B.P. (mid-Eocene) the rate decreases to 0.37 cm/1,000 yr, leading to a large transgression. In early Oligocene time (at 35 m.y. B.P.) the rate of sea-level fall increased to 0.66 cm/1,000 yr, with a consequent regression, and in early Miocene time the rate decreased again to 0.40 cm/1,000 yr, causing a further transgression. Since middle Miocene time, sea-level rise and fall has probably been dominated by episodic glaciation.

Again, the volume of the mid-oceanic ridge system as computed from the data of Table 1 decreases throughout latest Cretaceous and Cenozoic time to the present. The effect is that sea level falls continuously through this time, but because of the changing geometry of the ridge system the rate of sea-level fall varies with time. In this case, at any subsiding margin or basin, where the rate of subsidence at the shelf edge or center of the basin is greater than the rate of sea-level fall, the position of the strandline depends on the rate of sea-level fall and will rapidly move landward or basinward as the rate of sea-level fall decreases or increases, respectively. Therefore, the Eocene transgression and Oligocene regression that have been remarked upon so frequently in the literature (see, for example, Hallam, 1963) and appear in the sea-level curves of Holmes (1965), Sloss (1963), Wise (1974), and Vail and others (1978a, 1978b) may have been caused by a period of relatively rapid Paleocene sea-level fall followed by an Eocene period of slow sea-level fall. (Geologic evidence indicates that the Eocene transgression began earlier than shown in Fig. 4. This will be discussed below.)

The computed landward movement of the shoreline during Eocene transgression is about 28 km. Taking the differential form of equation 4, where $S \cdot R_{SS}$ and D are constant and T is assumed to be large,

$$\Delta X_L \cong \Delta R_{SL} \frac{D}{R_{SS}}. \tag{5}$$

It can be seen that the magnitude of the change in the position of the shoreline (ΔX_L) depends upon the magnitude of the change in rate of sea-level rise or fall, but it is also a function of the ratio D/R_{SS} (Fig. 5): the larger the ratio of D/R_{SS} the greater will be the movement of the shoreline.

In the computation above, where $D = 250$ km and $R_{SS} = 2.5$ cm/1,000 yr, $D/R_{SS} = 10^{10}$ yr, but, for example, if $D = 1{,}000$ km and $S = 1$ cm/1,000 yr (dimensions that are perhaps typical of epeiric seas), then $D/R_{SS} = 10^{11}$ yr and ΔX_L for the late Eocene transgression would be 280 km, larger by a factor of ten than shown in Figure 4.

Minor variations in spreading rates, as well as other factors such as orogeny and sediment flux, may affect the rate of sea-level change. For example, the rate of sea-level fall due to ridge contraction was 0.65 cm/1,000 yr from 55 to 45 m.y. B.P. If other causes combine to change the rate of sea-level fall by 0.1 cm/1,000 yr every 2 m.y., such that the rate changes from 0.7 to 0.6 to 0.7 every 4 m.y., the position of the strandline will oscillate between 67.3 and 62.7 km with the same periodicity. When $D/R_{SS} = 10^{11}$ yr, the oscillation will be between 673 and 627 km. The effect of these kinds of oscillations would be to give a saw-tooth shape to the strandline curves of Figure 4, a situation more in keeping with geologic reality (Sloss, 1963).

Changes in the rate of uniform sedimentation will have the same effect as variations in the rate of sea-level change. Taking the differential form of equation 3, but holding R_{SL} as well as D and R_{SS} constant, we get

$$\Delta X_L = \Delta S \frac{D}{R_{SS}}, \tag{6}$$

which is of the same form as equation 5, and thus the same argument and conclusions are applicable.

Accepting for the moment that the model of margin (and basin) subsidence is approximately correct, the greatest source of error in the computation of the position of the shoreline is probably caused by the imprecision of the sea-level curve. The sea-level curve is based solely on computed ridge volume, which itself contains errors; the data has been averaged over 10-m.y. periods. No account has been taken of other mechanisms for changing sea level. An error in the slope of the sea-level curve of 0.1 cm/1,000 yr at any point would result in an error of 10 km in the computed position of the strandline in the particular model discussed. The error would be 100 km in the case where $D/R_{SS} = 10^8$ yr.

TABLE 2. COMPUTATIONS FROM THE DATA OF TABLE 1

Date (m.y. B.P.)	V^* (10^6 km^3)	ΔV (10^6 km^3)	h (m)	$0.7\,h$ (m)	R_{SL} (cm/1,000 yr)
85	363	187	491.5	344.1	
					0.09
75	358	182	479.0	335.3	
					0.09
65	353	177	466.5	326.5	
					0.62
55	318	142	377.9	264.5	
					0.65
45	285	106	285.7	200.0	
					0.37
35	265	87	235.0	164.0	
					0.66
25	227	51	139.4	97.6	
					0.40
15	206	30	82.6	57.8	
0	176	..	..	..	

* It is assumed that from 15 m.y. B.P. back to at least late Mesozoic time the volume of continental glaciers was negligible. Therefore, in order to compute sea level then relative to present sea level, the volume equivalent of the amount of water currently locked up in continental glaciers (~25.0 × 10^6 km^3; Holmes, 1965) has been added to the ridge volume computed for each age interval (except 0 m.y. B.P.) from Table 1 to give the volumes shown.

COMPUTATION OF A THEORETICAL STRATIGRAPHIC CROSS SECTION

By substituting equation 2 into equation 1 for X_L and integrating, equation 8 is obtained, giving the thickness of the sedimentary section deposited (or eroded) during any time interval T at any point X on the margin:

$$\frac{dsed}{dt} = \frac{X \cdot R_{SS}}{D} - R_{SL} + e^{-\frac{T \cdot R_{SS}}{D \cdot S_L}} \left(R_{SL} + S - \frac{X_{LI} \cdot R_{SS}}{D} \right). \quad (7)$$

$$\text{Sed} = \frac{X \cdot R_{SS}}{D} T - R_{SL} T - \frac{D \cdot S_L}{R_{SS}} \left(R_{SL} + S - \frac{X_{LI} R_{SS}}{D} \right) \left(e^{-\frac{TR_{SS}}{DS_L}} - 1 \right). \quad (8)$$

R_{SL} must be constant during the time interval T, X_{LI} is the position of the shoreline at the beginning of time interval T, and all other parameters are also constant and are as defined in Figure 3.

Using this equation and the rates of sea-level change given in Table 2 and Figure 4, hypothetical sedimentary sections have been computed for each 10-m.y. interval from 85 to 15 m.y. B.P. In the first set of computations shown in Figure 6, S = 0 cm/1,000 yr, in which case the moving shoreline is always the boundary between erosion and deposition. Figure 6, part A, shows the sedimentary section at 65 m.y. B.P. The point of pinchout, at 10 km, is where the net sedimentation is zero for the interval 85 to 65 m.y. B.P.; landward the net sedimentation is negative, seaward it is positive. The Paleocene regression (65 to 55 m.y. B.P.; Fig. 6, B) moves the shoreline rapidly seaward. Much of the uppermost Cretaceous littoral facies would have been eroded. Note that the point of pinchout of Paleocene deposits lies landward of the final position of the shoreline. As the shoreline moved from its position at 65 m.y. B.P. to its more seaward position at 55 m.y. B.P., the shelf over which it passed experienced first deposition and then erosion. From the point of pinchout seaward to the shoreline and beyond, the net effect was deposition; all sediments deposited within this time interval should be found in this wedge (except those that are subsequently reworked). From 55 to 45 m.y. B.P. there was a further regression (Fig. 6, C).

During the interval from 45 to 35 m.y. B.P., a large transgression (Fig. 6, D) occurred. Again the moving shoreline acts as a boundary between a zone of erosion and a zone of deposition. The wedge of net positive sedimentation for the interval 45 to 35 m.y. B.P. (shown by the dashed line) is seen to pinch out seaward of the 35 m.y. B.P. shoreline. However, the lens of sediments of 45 to 35 m.y. age extends all the way to the shoreline, because during this transgressive phase the shoreline has moved landward; hence the coastal-plain–shelf surface has experienced first erosion and then deposition. The lower boundary surface of this lens (the solid line)

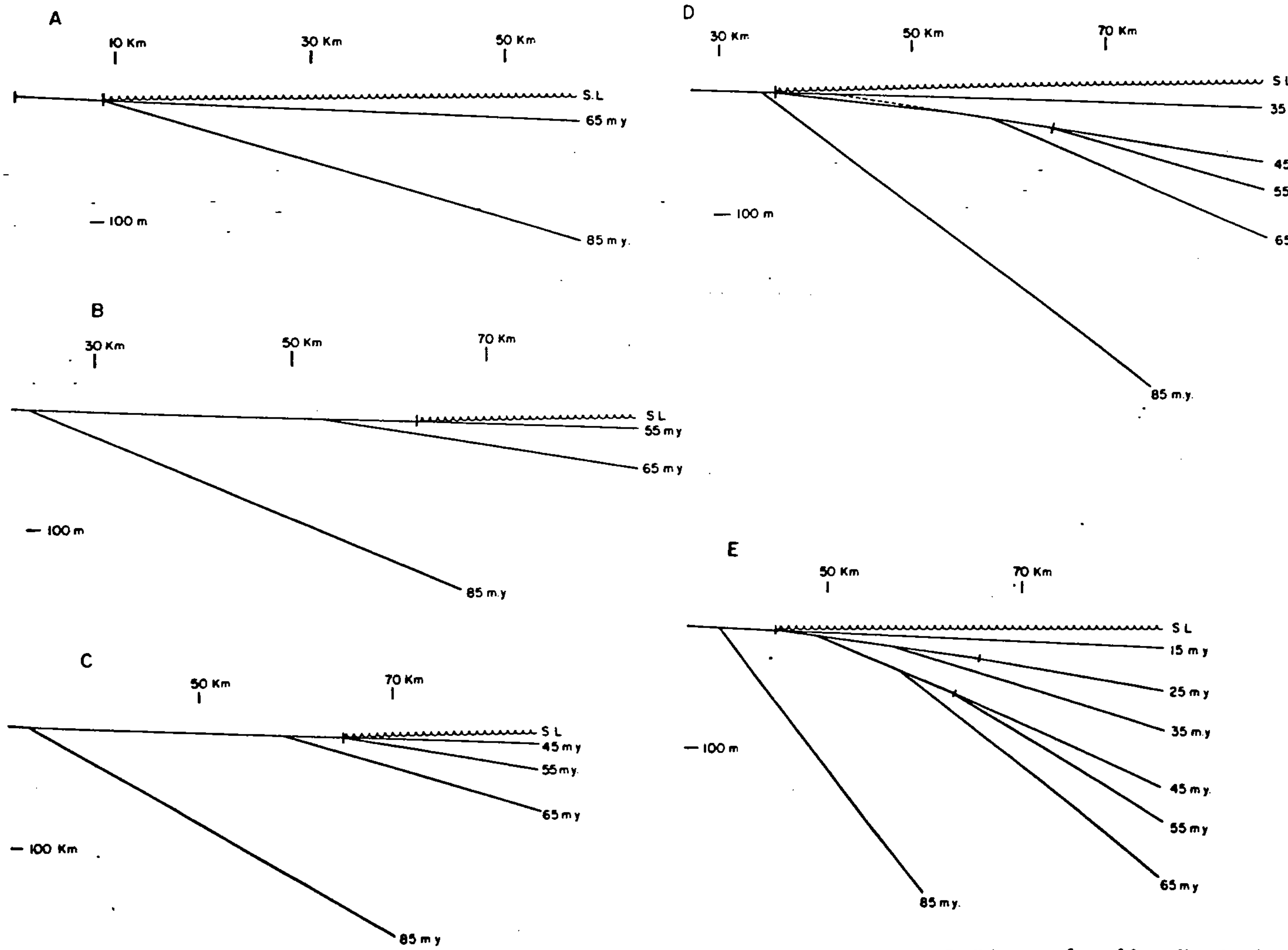

Figure 6. Hypothetical stratigraphic sections for sequence of stages from 85 to 15 m.y. B.P. Distance from hinge line is given in kilometres. Only variable is rate of sea-level fall, given by slope of sea-level curve in Figure 4.

has been determined by calculating at each point on the transgressive surface the depth to which erosion has occurred before the shoreline passed over. This was done by setting the sedimentation rate, *dsed/dt*, equal to 0 in equation 5 and solving for X as a function of T. This gives the time at which the transgressing shoreline passed the point X, but this is also the amount of time since the beginning of the interval during which erosion has taken place at point X. This, then, may be substituted in equation 6 to calculate the depth of erosion at each point X. The erosion surface defined in this way at each point X is the lower boundary of the sedimentary lens of the interval age.

The next sequence of events (Fig. 6, E) is the large Oligocene regression followed by the Miocene transgression. The comments made above about the Paleocene regression and the Eocene transgression are applicable here.

The stratigraphic section has been computed for the case where $S = 1$ to simulate sedimentation occurring on a coastal plain (Fig. 7, A). The geometric configuration of the resultant succession of sediment is the same as when $S = 0$, but in this case the shoreline is always 100 km farther seaward. Even though S is uniform over the entire coastal-plain–shelf platform, the depositional wedge thins landward and pinches out landward of the shoreline. The position of the shoreline zone up through the section is traced by the dashed line. If as discussed above (see equation 6), S is reduced to 0 cm/1,000 yr for 2 m.y. and then increased back to 1 cm/1,000 yr, a brief but extensive transgression occurs (Fig. 7, B). This simply illustrates a fact long observed by sedimentologists (Sloss, 1962), that minor oscillations in sedimentation rates can produce interfingering transgressive and regressive sequences. It is interesting that in this case when S was reduced to zero for the interval from 61 to −59 m.y. B.P., the shoreline moved more than 50 km landward; and at the same time there was more than 50 km of offlap as the point of pinchout moved seaward. At 59 m.y. B.P., S was increased to 1 cm/1,000 yr. This caused the shoreline to move 50 km seaward, but simultaneously 60 km of onlap occurred. During this entire interval, sea level was falling at 0.6 cm/1,000 yr.

Sequences of hotspot activity and Himalayan-type orogeny plus variation in the net sediment being transported to or removed from the ocean can also produce alternating transgressive and regressive phases. The magnitude of the oscillation would depend upon the magnitude of the change in the rate of sea-level rise or fall and on the geometric configuration of the particular margin. The term R_{SL} in the equations is the rate of eustatic sea-level change due to all causes. Although, as in Paleocene time, R_{SL} is dominated by the rate of ridge contraction, secondary causes of sea-level change, such as those listed above, will vary R_{SL} and cause minor transgressive and regressive sequences.

GENERAL RESULTS AND DISCUSSION

The most important conclusion of this work is that transgressive or regressive events recorded at numerous subsiding margins may not be indicative of eustatic sea-level rise or fall, respectively, but may be caused by changes in the rate of sea-level change. A decrease in the rate of sea-level rise will be regressive, as will an increase in the rate of sea-level fall. And conversely, an increase in the rate of sea-level rise or a decrease in the rate of sea-level fall will be transgressive. A key element in this model is the hypothesis that rates of sea-level change, except those due to glacial fluctuations and catastrophes, are always less than 1 cm/1,000 yr and are usually less than the rate of subsidence at the shelf edge of Atlantic-type margins. In fact, the subsidence rate at such margins is often in excess of 2 cm/1,000 yr. It is also assumed that normally sedimentation rates on subsiding margins vary spatially and temporally such that a constant shelf–coastal-plain slope is maintained. In the model, the sedimentation rate seaward of the shoreline is always positive, landward it may be negative or positive (but varies spatially away from the shoreline so as to maintain a constant slope). If

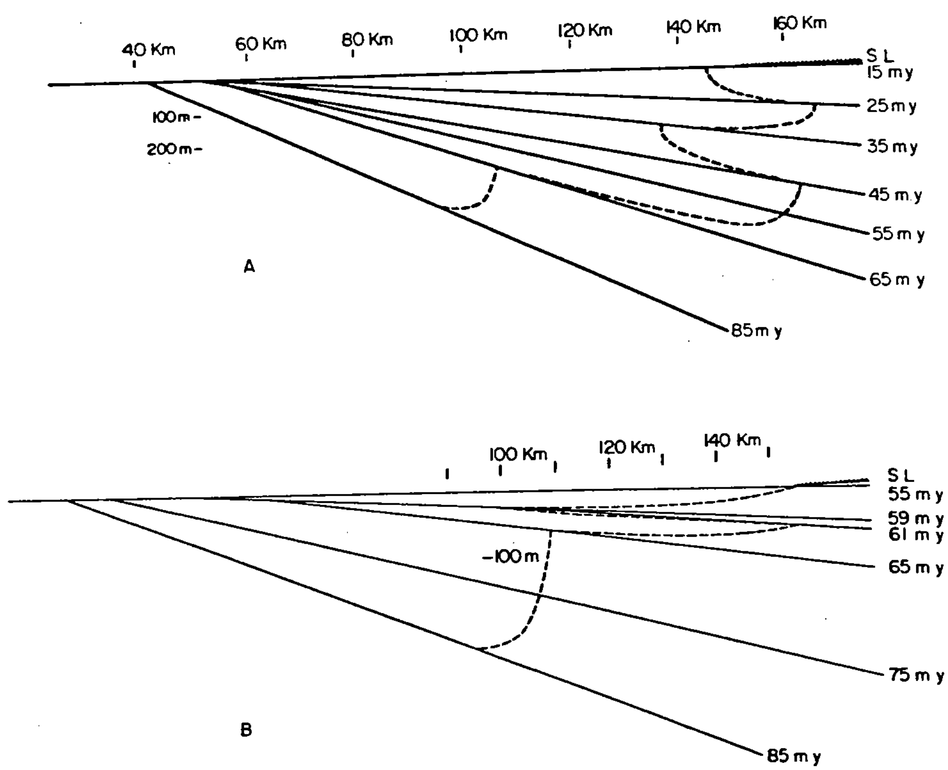

Figure 7. A. Same as in part E of Figure 6, but in this case $S = 1$ cm/1,000 yr. Note that pattern of pinch-out and truncation of various sedimentary wedges is same as in Figure 6, part E, but that position of shoreline (heavy dashed line looping through section) is pushed seaward. B. As in A, $S = 1$ cm/1,000 yr, but in this case S is reduced to zero for interval from 61 to 59 m.y. B.P. and then increased back to 1 cm/1,000 yr.

the rate of sedimentation landward of the shoreline is sufficient when sea level is rising, the shoreline will stabilize at that point where the rate of sedimentation minus the rate of subsidence is equal to the rate of sea-level rise. When sea level is falling, the stable point will be where the rate of subsidence minus the rate of sedimentation is equal to the rate of sea-level fall. Thus, if the rate of sea-level rise or fall changes, the strand will move to a new equilibrium point. Calculations (Fig. 4) indicate that since Late Cretaceous time sea level may have been falling continuously, but it fell slowly during latest Cretaceous, rapidly during Paleocene, less rapidly during late Eocene, more rapidly during Oligocene, and less rapidly during early Miocene time. Thus, there was a minor regression in Late Cretaceous time, a large Paleocene regression, a large Eocene transgression, an Oligocene regression, and an early Miocene transgression. This sequence, shown geographically in Figure 4, is compared with geological data from Africa and North America (Fig. 8). How precisely curves B and C reflect the timing of worldwide transgressive and regressive events is not known. The major point of disagreement is that the main early Tertiary transgression began during Paleocene time rather than in mid-Eocene time as computed. This might be attributable mostly to errors in the sea-level curve.

The sea-level curve has been computed from mid-oceanic ridge volume only. As discussed above, other mechanisms may cause sea level to fluctuate significantly and are not included in the sea-level curve of Figure 4. The most significant source of error in the sea-level curve lies in the inaccuracy of the polarity-reversal time scale by which spreading rates were measured and age versus depth determined. Deep Sea Drilling Project results indicate several modifications that might be made to the magnetic time scale. For latest Cretaceous and earliest Tertiary time, the observed ages are younger by several million years than predicted (see, for example, Larson and Pitman, 1972). This does not necessarily mean that the predicted ages are wrong, for it is possible that several millions of years might pass before sediment is deposited on newly formed crustal material. However, the effect of using the available data to adjust the paleomagnetic time scale is to reduce spreading rates for latest Cretaceous–early Paleocene time, therefore accentuating the rate of sea-level fall and enhancing the regression, and to increase

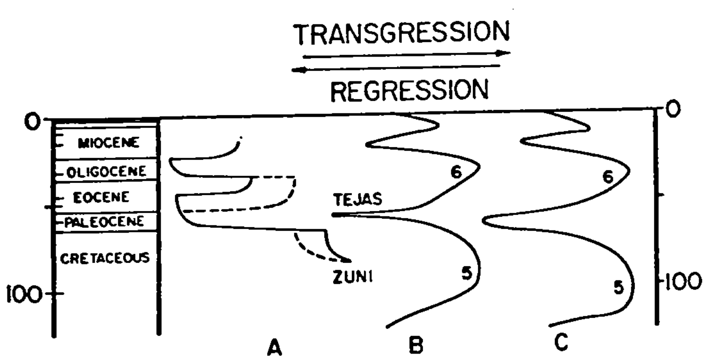

Figure 8. Sequence of late Mesozoic–Tertiary transgressive and regressive events as computed here and determined geologically; scale is arbitrary. A. Solid line shows transgressive and regressive sequence as computed here. Dashed line indicates possible change if DSDP results are used to recalibrate magnetic time scale. B. North American transgressions and regressions from stratigraphic record (Sloss, 1963), as interpreted by Rona (1973). C. African transgression and regression from stratigraphic record (Dearnley, 1966; Haughton, 1963; Reyre, 1966), as interpreted by Rona (1973).

the rate of spreading for late Paleocene and Eocene time, and to reduce further the rates of sea-level fall and thus initiate the transgression sooner (R. L Larson, 1976, personal commun.). The effect of this on the transgressive-regressive sequences might be as shown in Figure 8 by the dashed line. These changes in the paleomagnetic time scale would also necessitate changes in the shape of the age versus depth curve, but the effect would be to reduce slightly the volume of the ridges in proportion to the spreading rates. Note also that the sea-level curve was computed from data averaged for 10-m.y. intervals. As more data relevant to the history of the Indian Ocean becomes available and as the geomagnetic time scale is further refined, this interval might be reduced to 5 m.y.

The model as presented is in essence predictive and is in fact crudely successful. That is, given an approximate sea-level curve computed independently from ridge volume and a simplified model of margin subsidence, an early Tertiary sequence of transgressive and regressive events has been predicted with some success. Also, synthetic stratigraphic and lithologic successions have been constructed (Figs. 6, 7) for an idealized margin. These are too generalized to be compared in detail with the stratigraphic successions from any particular margin, but it does seem possible that by employing a more precise sea-level curve and by modifying the model this method could be used to predict and analyze the lithostratigraphy of particular margins and basins. A possible significant flaw in the model is the simplified concept of marginal (or basinal) subsidence, in which the margin is depicted as subsiding by rotation at a steady rate about a fixed hinge line. In fact, the postrift subsidence history appears to be more complex than assumed here.

Subsidence may be attributed entirely to the combined effects of thermal cooling and sediment loading (Sleep, 1971; Watts and Ryan, 1976). The subsidence due to thermal cooling decreases exponentially with time; the space created by this subsidence is filled with sediment, and this causes further subsidence. The thermal subsidence curve seems to be the same for a number of passive margins (Watts and Ryan, 1976). The subsidence due to the sediment load at a mature margin is probably best synthesized by flexural deformation of a rigid beam. Thus, even under idealized conditions of constant sediment flux, the rate of subsidence will decrease slowly with time. A further complication results because the sedimentation rate is also a function of the position of the shoreline. As the shoreline moves landward, so does the axis of zero flexure; thus, during a transgressive phase, as the shoreline moves landward the redistribution of loading would enhance the transgression. Regressive phases would be enhanced by the seaward migration of the axis of zero flexure. Also, it was assumed above that sedimentation rates are always sufficient to keep pace with subsidence; excess sediments are deposited on the slope and rise. If the sediment flux is just sufficient, there will be little or no deposition on the slope and rise, but when it is very excessive a great deal of sediment may be deposited on the slope and rise. Because the flux of sediment coming into the system may vary with time, the loading of the slope and rise will also vary. All of these factors must eventually be taken into consideration in attempting to construct predictive lithostratigraphic models.

The time-varying elements of subsidence that are common to all passive margins are not yet adequately quantified in such a way that they may be incorporated in the stratigraphic model. When they are and when a better eustatic sea-level curve is available, perhaps more precise productive stratigraphic sequences may be computed for various margins. It is possible that the sea-level curve may be further refined by comparison of computed stratigraphic sequences with geologically constructed sequences. Discrepancies that are common to a number of different margins may be attributed to errors in the sea-level curve. As the sea-level curve becomes more precise, deviations from the computed sequence could be attributed to local tectonic or sedimentary events.

It was remarked above that the present-day shelves (excepting several local areas) are not in equilibrium. Sea-level rise of the past 10,000 yr has been so rapid that sedimentation has not been able to keep up. The average present-day shelf break is at 135 m (D. Hayes, 1976, personal commun.), with quite a range of variations around that value. It seems likely that during early Tertiary time the typical shelf break may have been 40 m. Also at that time there was no significant glaciation; hence, the rate of subsidence at the shelf edge of an Atlantic-type margin was always greater than any possible rate of sea-level fall (catastrophe and local tectonic events excepted). Thus, it was very unlikely that shoreline could move out over the shelf edge, because the shelf edge would always have been subsiding faster than sea level could fall. Under these conditions, canyon cutting from the mouth of a river across a shelf and down a rise and slope would also have been unlikely. The sediment load would have been dropped on the shelf, and when the sediment flux was sufficiently large, deltas would have been formed and progradation over the shelf edge would have occurred.

However, if rapid and significant glaciation were to occur enough to lower sea level, say 50 m in 10^5 to 10^6 yr, then the strandline in many areas would migrate beyond the shelf edge and lie at the top of the slope. Rivers would then rapidly cut channels across the shelf and deposit their sediment load at the top of the slope. Periodic slumping of these deposits would generate turbidity currents that could cut canyons down the slope and across the rise. These canyons would channel the sediments, causing them to be transported to the abyssal plains. Thus, in this case, the rate of subsidence of the shelf-slope-rise complex would be greatly reduced because the sediment load would tend to bypass these areas. Thermally driven subsidence would continue, but an equilibrium situation could not be achieved at least until subsidence and/or glacial melt were suficient to bring the strandline back above the shelf edge.

Events of this kind may have occurred, starting perhaps in earliest Oligocene time. O_{18} data show a dramatic drop in surface and bottom water temperatures (to about present-day values) for Antarctic waters, and these data have been interpreted as indicating the possible beginning of extensive glacial build-up in Antarctica (Shackleton and Kennett, 1975). This would have added to the rate of sea-level fall already occurring due to ridge contraction. If the total were sufficient to cause sea level to fall faster than subsidence could occur, then, as described above, the strand would migrate beyond the shelf edge; canyons would be cut such as to cause sediment to bypass the shelf-slope-rise areas of some margins, hence reducing the rate of subsidence. So long as the rate of sea-level fall (whatever the causes) remained greater than the rate of subsidence at the shelf edge, the strandline would remain below the shelf edge. This may have occurred throughout much of Oligocene time, because sediments of this age are missing or scarce on the margins that surround the Atlantic — in particular, northwest Africa and the east coast of the United States (J. Conolly, 1976, personal commun.).

The waxing and waning of massive continental glaciers has probably been the dominant factor controlling the sedimentary history of the Atlantic-type margins since the Oligocene Epoch.

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