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PRE-QUATERNARY SEA-LEVEL CHANGES

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INTRODUCTION

While the paramount importance of global sea-level changes has been widely accepted by Quaternary geologists throughout this century, it is only within the last decade that such changes in the rest of the Phanerozoic have received much attention. There are two principal reasons for this arousal of interest. First, increased exploration of the stratigraphic record across the Earth, both on land and under the sea, and significant improvement in biostratigraphic correlation have drawn attention to what appear to have been globally synchronous events. Second, oceanographic research has demonstrated the importance of tectonically controlled changes in the cubic capacity of the ocean basins, notably variations in ridge volume, which provide a ready mechanism for sea-level changes even at nonglacial times.

The adjective *eustatic* was first proposed for global changes of sea level by Suess (1906) in his great treatise *The Face of the Earth*. (The corresponding noun is *eustasy*, not eustacy as it is commonly misspelt; the unconvinced reader need only think of ecstasy.) Eustatic sea-level changes can be inferred, as Suess recognized, by (*a*) plotting the temporal spread of marine sediments over the continents and (*b*) estimating the depositional depth variations in marine stratal sequences, which can be correlated across and between continents. To these methods can be added seismic stratigraphy, in which biostratigraphically age-determined seismic sequences on the continental margins are analyzed in terms of cycles of coastal onlap and offlap.

When the results of these methods are broadly in accord, confidence in a eustatic interpretation is increased, but the range of confidence can vary considerably. It is widely accepted that there was an exceptionally high sea

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level in the late Cretaceous and an equally marked low one at the Paleozoic-Mesozoic boundary, but smaller-scale changes and short-term oscillations have proved more controversial. This is because of the complicating effects of regional tectonics and the limits of biostratigraphic resolution in the all-important matter of intra- or intercontinental correlations.

If global "signals" can be satisfactorily disentangled from local or regional tectonic "noise," eustatic changes can be identified and accepted because well-understood processes are known to produce such variations. The alternative suggestion that different continents have moved up and down in concert (Sloss & Speed 1974) is not only less economical but demands recourse to otherwise unsupported speculations about the thermal and tectonic behavior of both the continents and the upper mantle.

In this review, eustasy is first treated in terms of shorter-phase oscillations and then in terms of changes through the whole Phanerozoic, involving both major oscillations and longer-term secular trends. (The Precambrian must obviously be excluded because of the inability to obtain sufficiently refined correlation and the uncertainty as to which strata are marine in origin.) Correlation of sea-level variations with changes in sedimentary facies, isotope ratios, and faunas, and the underlying causes of these variations, are then briefly discussed.

SEA-LEVEL OSCILLATIONS

Facies Analysis

The most obvious and longest-practiced method of inferring eustatic changes is to analyze the facies of continental sequences in terms of alternating marine and nonmarine strata in more marginal marine situations, and shallower- and deeper-water deposits in more offshore, fully marine situations. If shallower-water marine phases of sedimentation can be correlated over considerable distances with regressive episodes, and deeper-water marine phases with transgressive episodes in the marginal environments, then purely local controlling factors can be ruled out, and events of at least regional significance are indicated. Similarly, the correlation of shallowing events from terrigenous siliciclastic to calcareous facies within marine deposits can be used to exclude an interpretation involving merely increased influx of sands and muds from some adjacent land area, leading perhaps to a regression by sediment progradation.

Figure 1 illustrates this point. The sandier units on the left side of the diagram, with, for instance, trough cross-bedding and oscillation ripples, exhibit facies characteristics suggesting shallower-water deposition than the intervening more argillaceous units. They also correlate biostratigraphically with the erosional events and shallow-water facies of the purely carbonate sequence on the right (such as bored and truncated hardgrounds, intraformational conglomerates, dolomites, benthos-rich, stromatolitic, and birdseye limestones), better than, say, with intervening deeper-water micrites with pelagic fauna. The former area could signify an environmental setting marginal to a delta, the latter an isolated carbonate bank, such as the Bahamas, or a marginal marine platform free from terrigenous siliciclastic influx, such as the Yucatan or the southern side of the Persian Gulf. Note that the intervening area, a deeper-water basin characterized by a relatively high subsidence and sedimentation rate, contains a uniform argillaceous facies. Siliciclastic muds may be deposited over a considerable depth range, and regional or eustatic sea-level changes may find no expression in the sedimentation. Furthermore, subsidence in a basin will counteract sea-level fall and may conceal it if the rates of vertical movement match each other.

The degree of confidence in eustatic interpretation clearly depends on (*a*) the reliability of bathymetric estimates, (*b*) the precision of biostratigraphic correlation, and (*c*) the extent of such correlation, preferably from continent to continent and at least across large areas of continent. Whereas there is in many if not most cases a high measure of consensus about the *relative* depositional depths of particular marine facies, based on an array of sedimentological and paleoecological evidence, *absolute* determinations are at best only very approximate (Hallam 1981a). The quality of correlation obviously varies with the available fossils, with the most precise age determinations probably being given by ammonites and planktonic foraminifera. Unfortunately, the deposits of the shallower or more marginal

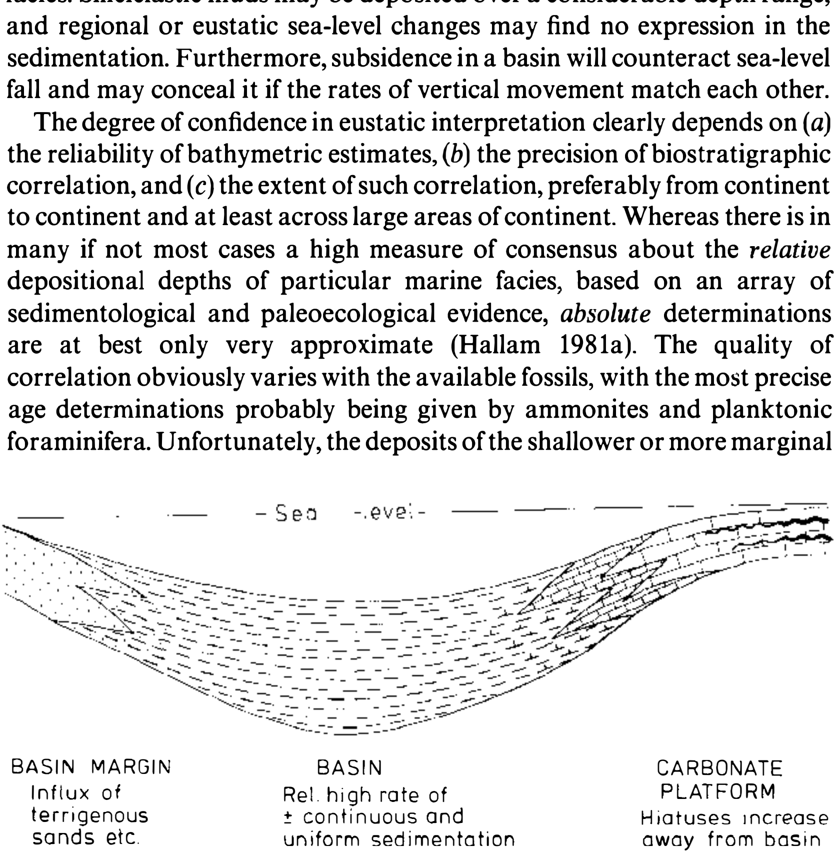

Figure 1 Illustration of how sea-level fluctuations affect different sedimentary regimes. Two episodes of shallowing give rise to spreads of sand into the left-hand basin margin, and corresponding spreads of shallower and deeper neritic limestone facies into marls and shales of the right-hand basin margin. These episodes of sea-level fall are represented by hiatuses in the carbonate platform sequence on the right and lack any representation in terms of facies changes in the basin center.

marine environments, which are best for depth estimates, are frequently the least likely to contain good biostratigraphic indices. Allowance must also be made for at least a limited amount of diachronism in following transgressions laterally because continents are unlikely ever to have been as flat and smooth as a billiard table.

Examples of facies analysis leading to eustatic interpretations, covering a wide stratigraphic range, are given by Brenchley & Newall (1980), Lenz (1982), Ramsbottom (1979), Hallam (1978, 1981b,c), Hancock & Kauffman (1979) and Olsson et al (1980).

Seismic Stratigraphy

Seismic stratigraphy, as developed in recent years by the major oil companies, is essentially a geological approach to the interpretation of subsurface data produced by seismic reflection profiling, and has been widely used in the interpretation of continental margins. Its application to the study of eustasy has been pioneered by Vail et al (1977).

The primary seismic reflections evidently tend to follow bedding surfaces and unconformities with velocity-density contrasts. One might naturally assume that such reflections would normally bound major lithological units, even when these cut across stratification surfaces, but Vail et al, on the basis of careful studies well controlled by borehole logging, insist that in fact they follow chronostratigraphic boundaries that parallel such surfaces. This is a crucial assumption in all that follows, but the supporting evidence has to be taken largely on trust because little of it has been published.

In what might conveniently be called the Vail technique, major stratigraphic units consisting of a relatively conformable succession of strata, with upper and lower boundaries defined by unconformities, are classified as *depositional sequences*. The age of the boundary unconformities is determined by tracing them laterally into conformable successions. Within a given region, a relative sea-level rise is inferred from the progressive landward onlap of littoral and/or coastal nonmarine deposits in a marine sequence, which is termed *coastal onlap*, and a relative fall is inferred from the downward shift of coastal onlap. Regional sea-level curves are drawn on the basis of this type of analysis, from which chronostratigraphic correlation charts are constructed. If cycles of relative rise and fall can be correlated in several widely separated regions across the world, the underlying control is held to be eustatic.

The great strength of seismic stratigraphy is that vast and otherwise inaccessible terrains can be rapidly analyzed in terms of shifting packages of sediment; given adequate borehole coverage for good facies and biostratigraphic control, there is no reason why valid and important results should not be obtained. One should take full account of the possible drawbacks, however. In practice it is doubtful if the chronostratigraphic precision (which in the pre-Cenozoic is based mainly on palynology) usually matches that achievable in studying sections exposed on land, and information on pre-Mesozoic rocks is scanty. Likewise, facies details are not normally available, and the fundamental assumption that impedance contrasts invariably follow chronostratigraphic surfaces is open to question. Most important, details of the evidence supporting the eustatic claims of the Exxon group (Vail et al 1977) are not published, and hence their claims cannot be checked directly. One wonders if teams from other oil companies would interpret the same data in a similar way, or how useful the Vail curves have proved in prediction within the industry.

Until the supporting evidence becomes publicly available, probably the best one can hope for is to match the Vail eustatic curves against those produced using other techniques. Conformity in results should increase confidence in interpretation.

Distinction of Eustatic from Local and Regional Tectonic Events

It is obviously of great importance to be able to disentangle the effects of local and regional tectonics when evaluating the role of eustasy in a given case. Thus a change in a sequence from shallower- to deeper-water facies could be the result either of a rise in sea level or an increase in the rate of subsidence uncompensated by an increased sedimentation rate. In extreme cases the influence of local tectonics is obvious, as evidenced, for instance, by sharp angular unconformities, fault-bounded basins with stratigraphically thick but laterally restricted scarp-front conglomerates, and abrupt lateral variations in stratal facies and thickness. Over extensive areas, however, such telltale features are absent (most notably in cratonic regimes), and facies changes tend to be more subtle and widespread. These cases require careful and detailed facies analysis utilizing the widest possible array of sedimentological, paleoecological, and stratigraphic evidence (e.g. Corbin 1980, Loughman 1982, Phelps 1982).

As regards *local* changes, involving areas on the order of thousands to tens of thousands of square kilometers, the technique proposed by Hallam & Sellwood (1976) may prove helpful. Most cratonic regimes can readily be divided into a series of basins and swells, respectively characterized by greater and lesser thicknesses of stratal sequences of a given age, which correspond to higher and lower mean sedimentation rates. For a given regime, the mean thicknesses of successive stratigraphic intervals are determined, as is the percentage deviation from the mean for each time interval for the various basins and swells. Areas subjected to minimal local tectonic disturbance should record a more or less constant basin or swell tendency over the time period in question, whereas areas where sedimentation has been strongly affected might be expected to show frequent deviations, both positive and negative, from the mean.

An example from the Jurassic of England is presented in Figure 2. In this case the dominant tectonic influence is taphrogenic, with the rate of subsidence being controlled primarily by vertical movements of the horst-and-graben type, either within the Hercynian basement or the Jurassic itself. The best candidates for sequences recording eustatic signals in terms of shallowing and deepening events are clearly those with little evidence of local tectonic disturbance, such as North Creake in Figure 2.

It is also desirable to distinguish *regional* events of subsidence or uplift, involving areas on the order of hundreds of thousands or more square kilometers. Thus, Officer & Drake (1982) point out a 140-m relative change in elevation northeastward along the east coast of North America, over a

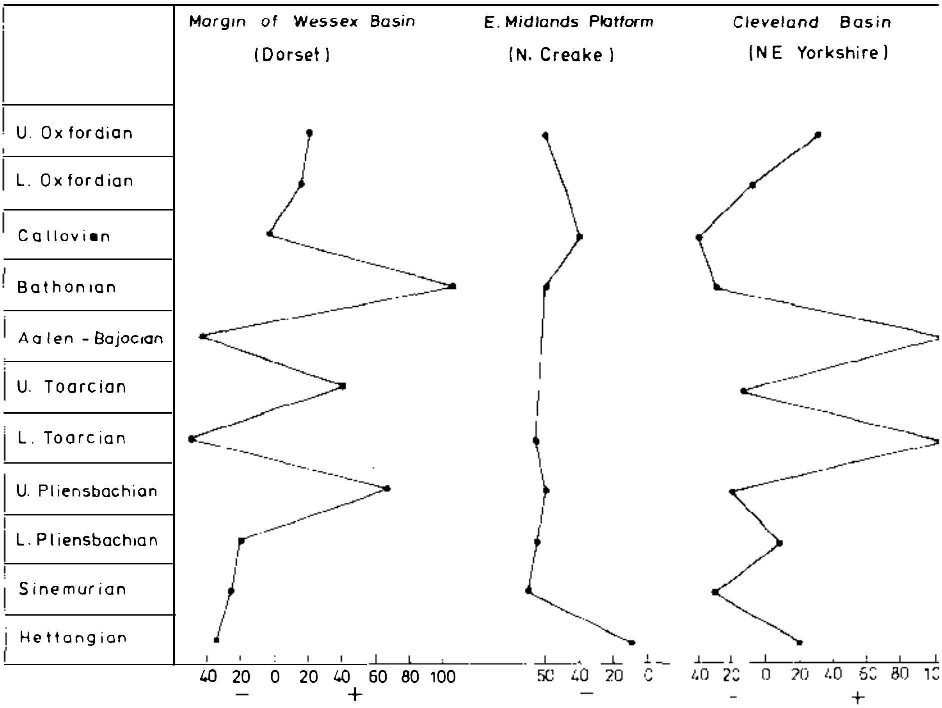

Figure 2 Graphical plots illustrating fluctuations in the rate of subsidence in different English localities throughout the Jurassic. Numbers at bottom are percentage deviation from mean stage thickness of Anglo-Welsh sections. Note the striking lack of correlation between Dorset and Yorkshire (based on data presented in Hallam & Sellwood 1976).

distance of 3000 km. This has apparently taken place over the last 18,000 years, giving a rate of elevation of Florida with respect to the Scotia Shelf of 0.8 cm yr^{-1}. Bally (1981) relates widespread unconformities in the stratigraphic record to major plate reorganizations that are not necessarily orogenic events, and he questions the eustatic interpretation of Vail et al (1977).

Watts (1982), following McKenzie (1978), has developed a model to account for the subsidence of passive continental margins, in which thermal contraction follows initial rifting and extension, and sediment loading occurs with the flexure of a progressively more rigid basement subsequent to the rifting phase. There is initially a significant onlap of sediments onto the basement due to the abrupt transition from fault-controlled Airy-type subsidence to flexure-controlled subsidence, followed by a slower rate of progressive onlap due to an increase in flexural rigidity with age. This pattern of onlap closely matches that attributed by Vail to the rise of sea level, and Watts claims a correlation between the beginning of Vail's supercycles and the age of the rift-drift transition at the continental margins formed by the breakup of Pangea. Watts concedes that a few transgressions are too widespread, however, to be accounted for solely by this regional subsidence.

One effective, though imprecise, way of disentangling regional epeirogenic effects from eustatic ones is to construct a general eustatic curve by plotting the areal distribution of marine sediments and making best estimates of shoreline positions for successive time intervals on a global scale; these are then matched against the pattern of transgression-regression and shallowing-deepening episodes in a given region. A marked disparity signifies regional epeirogenic involvement. Thus, the regressive character of the Bajocian (Middle Jurassic) of the North Sea is anomalous in global terms and indicates significant regional tectonism involving both uplift and subsidence; this interpretation is supported by the evidence of contemporary faulting and volcanicity (Hallam 1978).

Bond (1976) proposes a method for assessing the role of regional subsidence. If we assume that the sediment supply keeps the depositional interface at depositional base level, the maximum thickness of sediment is given by

$$2.4t = h + 3.4t - 3.4h,$$

where h is rise of sea level, t is the thickness, and 2.4 and 3.4 g cm^{-3} are the mean sediment density and mantle density, respectively. Any excess sediment thickness must be attributed to subsidence.

Bond concludes from his analysis that the late Cretaceous sea-level rise that flooded nearly half of North America, implying a rise of over 300 m, was insufficient to account for the thickness of nearly half of the Upper Cretaceous transgressive deposits of the Gulf and Atlantic coastal plains and the Western Interior. A corresponding amount of subsidence must therefore be invoked in addition.

The immense thickness of deposits in the western part of the Western Interior Cretaceous seaway province has long indicated a high rate of subsidence and, hence, a foredeep adjacent to the rising Laramide mountains. Hancock & Kauffman (1979) show that the increasing proportion of regressive facies from Santonian times onward is anomalous in global terms and relates to the Laramide uplift to the west. Australia is likewise anomalous in that the sea retreated from most of the continent following a widespread Aptian-Albian transgression (Brown et al 1968), whereas in the rest of the world the sea-level maximum was not reached until Campanian times. Although this does not necessarily indicate uplift of the whole of Australia, because the sea could have reached the interior through one or more confined areas or straits, epeirogenic movements on the regional rather than the local scale seem to be implied.

The Pliensbachian and Toarcian stages of the Lower Jurassic provide an especially instructive case history in how regional tectonics may be disentangled from eustatic factors in accounting for geographic and temporal variations in marine sedimentation (Hallam 1978, 1981b; Figure 3). In the British area terrigenous sands and silts were introduced for the first time in the Jurassic into what had been a predominantly argillaceous sequence approximately at the Lower-Upper Pliensbachian boundary; and the top of the Pliensbachian is marked everywhere either by sandstones or oolitic ironstones that give clear evidence of having been deposited in shallower water than the more argillaceous or locally calcareous strata above and below. Relatively shallow-water sandstones or limestones also mark the top of the stage in western France (Normandy) and the western parts of the Iberian peninsula. The British deposits are overlain in most areas by a Lower Toarcian unit of laminated organic-rich shales, which is widespread in northwest Europe and is universally accepted as a relatively deep-water deposit.

Therefore, in terms of a purely eustatic model, the top of the Pliensbachian (Spinatum Zone) would naturally be taken as marking the lowest stand of sea level. There is, however, an important regional unconformity directly *below* the Spinatum Zone, recorded almost everywhere in Great Britain (Phelps 1982), that does not accord with this simple picture.

Studies elsewhere serve to resolve this problem. Southeastward into eastern France and southern Germany the arenaceous deposits disappear, replaced by calcareous and argillaceous deposits. These indicate a predominantly westerly land source, but the Lower Toarcian organic-rich shale is still present (schistes cartons in France, Posidonienschiefer in Germany). Facies analysis suggests a general pattern of marine deepening through the later Pliensbachian continuing to an early Toarcian maximum, followed by shallowing. Such a pattern corresponds more closely to the world picture as revealed by the deposits of western North and South America and eastern Asia, the only other regions outside Europe where adequate stratigraphic information is available (Hallam 1981c). The

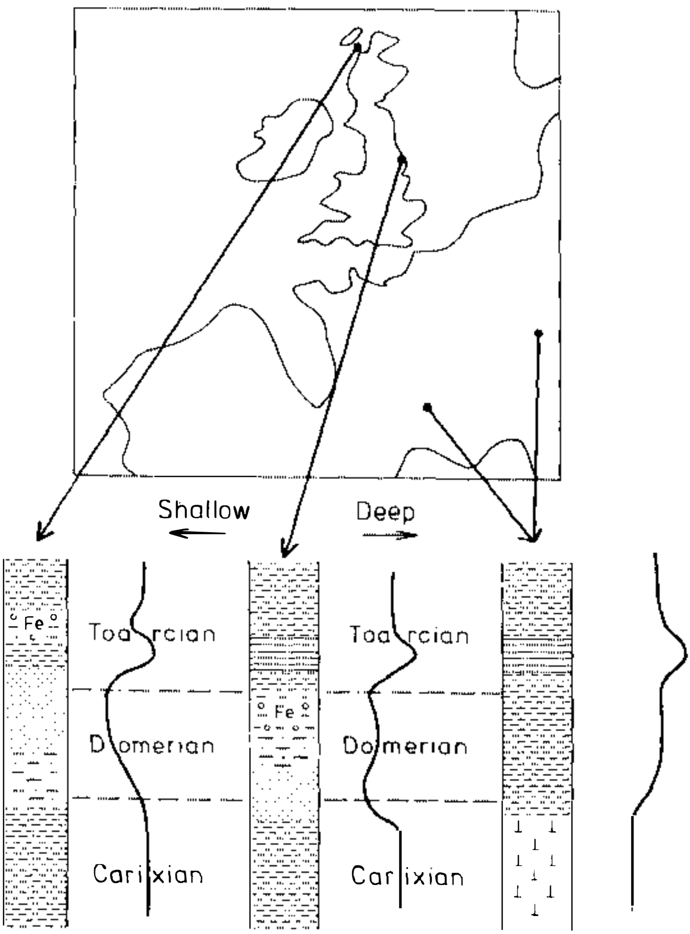

Figure 3 Schematic sections of Lower Jurassic in different parts of western Europe, with accompanying inferred bathymetric curves. Carixian = Lower Pliensbachian, Domerian = Upper Pliensbachian.

deposits on the western margins of Europe therefore appear to reflect regional uplift to the west, which can plausibly be related to taphrogenic horst-and-graben phenomena anticipating the later opening of the Atlantic (Hallam & Sellwood 1976). It is important to note, however, that the eustatic signal is still recorded in the British area, with the shallowest-water deposits of the youngest Pliensbachian recording a transgressive episode following widespread uplift and erosion.

The Lower Jurassic of northwest Europe provides another illuminating example of the interaction of regional tectonic and eustatic factors. It is the only region outside the Pacific margins with marine Hettangian, and global analysis suggests a Hettangian rise of sea level following an end-Triassic fall (Hallam 1981b,c). The sea apparently flooded into northwest Europe as a result of both this rise and the widespread crustal attenuation due to extension (Hallam & Sellwood 1976).

Rates and Amount of Sea-Level Change

While it is clearly important to determine as precisely as possible the rates of oscillatory sea-level rise and fall, we are still sadly lacking in reliable quantitative estimates because of various methodological difficulties. Rises are easier to document than falls, provided the biostratigraphic correlation is good. Marine transgressions and the correlative deepening phases in fully marine sequences can often be precisely dated, but the regressive facies commonly associated with eustatic falls may lack stratigraphically useful fossils. Additionally, erosion following regression may lead to the removal of part of the sedimentary sequence. There is furthermore the problem that coarsening-upward siliciclastic sequences can result either from sea-level falls or from the filling in of a marine basin by sediments at a time of stillstand. In extreme cases, this may even occur during times of sea-level rise if the rate of shallowing due to sediment influx exceeds the rate of rise (Vail et al 1977).

There are a number of ways of circumventing this last difficulty. If shallowing-upward siliciclastic sequences can be correlated with shallowing-upward carbonate sequences elsewhere, then a good case for sea-level fall on at least a regional scale can be established (Hallam 1978).

Another approach is indicated by the work of Thiede (1981). The occurrence of neritic fossils in pelagic sediments of the Deep-Sea Drilling Project (DSDP) cores from the Pacific documents the repeated injection of skeletal debris derived from shallow-water provinces into adjacent deep-sea basins through the late Mesozoic and Cenozoic, during short-lived intervals when little or no benthic sediment was so transported. Thiede demonstrates that such episodes correlate in almost all cases with low stands of sea level in the scheme of Vail et al (1977). Short-phase eustatic falls, leading to transient land connections, are also the most plausible explanation for the pronounced similarities of North American and European late Jurassic terrestrial vertebrate faunas, because paleogeographic data point to shallow marine separation of the two continents at this time (Prothero & Estes 1980).

In the facies analysis method of studying eustasy, approximate order-of-magnitude estimates of the extent of sea-level oscillations can be made by assigning likely depths of deposition to the deepest- and shallowest-water strata of given marine cycles and then by obtaining best estimates of age from biozonal and radiometric data. By assuming depth ranges of a few tens of meters, which seem plausible for the deposits in question, Hallam (1978) and Ramsbottom (1979) both obtained figures of up to a few centimeters per 1000 yr of Jurassic and Carboniferous age. The data of Olsson et al (1980) indicate a major short-phase sea-level fall at a comparable rate of $\sim$3 cm per 1000 yr during the early Oligocene.

A more direct method of estimation, proposed by Vail et al (1977), is to measure the vertical and horizontal components of coastal onlap from seismic stratigraphic data from the continental margins. However, Watts (1982) points out that they did not correct for flexural effects that vary as a function of time and position. Furthermore the Vail techniques for estimating sea-level rise and fall are quite different from each other.

There is as yet no accord about the relative rates of sea-level rise and fall and the proportion of time occupied by stillstand, although there is of course no reason why these should not vary significantly in particular instances. On the basis of facies analysis, Hallam (1978) argued that the predominant mode for the northwest European Jurassic was probably a rapid rise and fall interrupted by a longer phase of stillstand, while Ramsbottom (1979) suggested that the mode for the Carboniferous of the same region was a gradual or rapid rise, followed by brief stillstand and a very rapid fall (Figure 4).

Vail et al (1977) proposed a general Phanerozoic mode of gradual or moderately rapid rise decelerating with time, minimal stillstand, and extremely rapid, "geologically instantaneous" fall. This was based on a direct correspondence between relative changes of coastal onlap with relative changes of sea level. It is now admitted (Vail & Todd 1981) that this mode is erroneous because it takes no account of the facies differences between coastal plain (marine) and alluvial plain (continental) facies, and in fact the landward boundary that should be plotted is the upper limit of the coastal plain. Where an alluvial plain is present, the relative sea-level boundary corresponds with the facies change between the coastal and alluvial plain. Unfortunately, seismic stratigraphic techniques do not often permit this identification. Charts of relative changes of coastal onlap

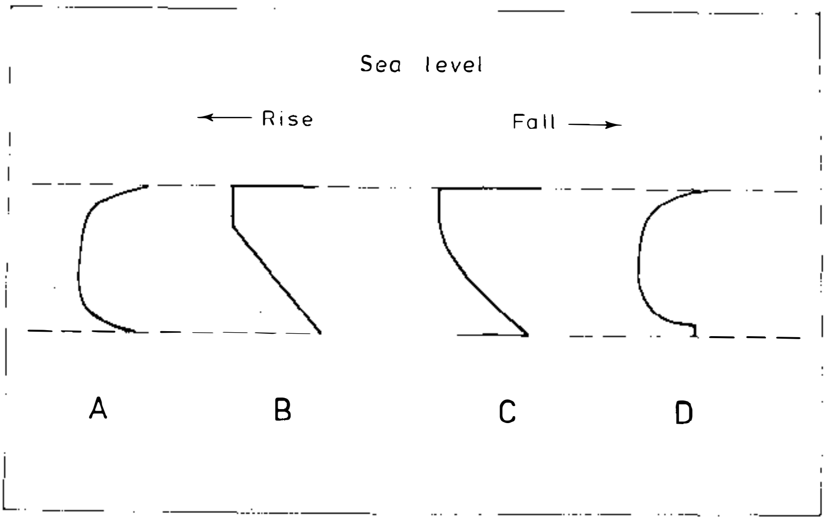

Figure 4 Diagrams illustrating the character of sea-level rises and falls through time in eustatic cycles. *A*. The most probable mode for Jurassic cycles, based on European sections (Hallam 1978). *B*. The Carboniferous mode, based on British sections (Ramsbottom 1979). *C*. The mode of second- and third-order cycles, based on seismic stratigraphy (Vail et al 1977). *D*. Revised mode for Jurassic cycles, based on seismic stratigraphy (Vail & Todd 1981).

typically show abrupt shifts from widespread to restricted, whereas charts of relative changes of sea level commonly show more-gradual shifts. Vail & Todd's amended oscillatory eustatic curves for the Jurassic, based on North Sea data, correspond quite closely to those inferred independently from my facies analysis of marine sections exposed on land (Figure 4).

CHANGES THROUGH THE PHANEROZOIC

The only comprehensive Phanerozoic eustatic curve in existence is that of Vail et al (1977, Part 4, Figure 1). This shows both first- and second-order cycles. The first-order cycles are defined by low sea-level stands at the beginning of the Cambrian and in the Permian-Triassic and Quaternary, and high stands in the Ordovician-Silurian and late Cretaceous, with the late Cretaceous sea level attaining the highest value of any in the Phanerozoic. Superimposed on this gross cyclicity are a series of second-order cycles with a characteristic sawtooth pattern: gradual, decelerating rises followed by sudden falls. Smaller third-order cycles with a similar shape are indicated in other more detailed diagrams.

The Vail Phanerozoic curve is only partly based on continental margin seismic stratigraphy; the Paleozoic part is based on the North American cratonic sequences of Sloss (1963). Thus its quality and reliability are variable through time, even accepting the criteria and assumptions of the Vail team. It has been pointed out above that the sawtooth shape of the second- and third-order cycles is no longer accepted by Vail, and the lack of third-order cycles in the Cretaceous part of the detailed Jurassic-Tertiary curve (Vail et al 1977, Part 4, Figure 2) is strictly a manifestation of Exxon management policy in releasing data. Furthermore, although data from widely separated regions across the world are considered, there is a strong weighting in favor of the Atlantic margins of Europe and North America and the Gulf Coast region.

The broad trends of the curve can be tested by the independent technique of estimating the proportion of continents covered by the sea at successive time intervals. This method is thought to give reliable first-order approximations, but claims to high precision should be viewed with reservation. Because of subsequent erosion, the feather edge of marine sedimentary units does not necessarily correspond closely with the original limits of deposition, and as Cogley (1981) has pointed out, most paleogeographic analyses have ignored the 16% of continents covered by present-day shelf seas because of inadequate data. If only present land areas are considered, then unless ancient and modern shorelines coincide everywhere, estimates of the extent of ancient epicontinental seas must exceed the extent of modern seas.

Areal plots of the degree of continental inundation, based on the paleogeographic maps of Strakhov (1948) and Termier & Termier (1952), show a significant departure from the smoothed Vail curve, insofar as the extent of marine cover of the continents is considerably greater in the early to mid-Paleozoic than in the late Mesozoic (Egyed 1956, Holmes 1965, Hallam 1977a). On the other hand, Wise (1974) used Schuchert's (1955) paleogeographic maps of North America to argue for a condition of essentially constant continental freeboard throughout the Phanerozoic, interrupted by short-term oscillations of sea level, 80% of which remained within about 60 m of a normal freeboard level ~20 m above the present level. The maps were drawn several decades ago and do not take into account extensive modern work that demonstrates a much greater spread of sea over North America in Paleozoic times. My areal plot of Schuchert's maps produces a curve closely resembling the smoothed first-order cycle curve of Vail et al, but by taking into account the new data, the revised curve strongly resembles my plots of the global Strakhov and Termier & Termier data (as well as those of the Soviet Union) based on a comprehensive series of paleogeographic maps (Hallam 1977a).

These results suggest that Phanerozoic sea level was at a maximum not in the late Cretaceous, as proposed by Vail et al (1977), but in the late

Ordovician, when nearly two thirds of the continents were inundated. In addition, the late Devonian and early Carboniferous seas were also slightly more extensive than those of the late Cretaceous, which covered about one third of the present continental area. Because the probability of losing the stratigraphic record through subsequent deep burial, metamorphism, and erosion must increase with time, paleogeographic maps for older periods are likely to underestimate the former extent of marine cover. Moreover, the proportion of carbonates to terrigenous siliciclastics on the cratons is significantly higher in the Paleozoic than in the Mesozoic and Cenozoic. Since erosion rates apparently increase exponentially with elevation (Garrels & McKenzie 1971), this implies sediment sources that were both more areally restricted and topographically subdued (Hallam 1977a).

The hypsometric curve of continental elevations can be used to estimate the sea-level change that will flood varying portions of the continents, on the assumption that the ancient hypsometric curves were similar to the present one (Forney 1975). Forney's method was developed by Bond (1978), who used the scatter of data for different continental elevations to distinguish between sea-level changes and the vertical motions of large continental surfaces. Because of post-Cretaceous continental uplift, use of the present hypsometric curve in conjunction with data on marine spread for earlier times is likely to overestimate the amount of sea-level rise required by a variable amount.

The reliability of the hypsometric method can be evaluated against other methods for the late Cretaceous high stand, which has received much attention in the last few years. Plots of the data derived from Strakhov's global maps indicate that a maximum of ~40% of the continental area was flooded, indicating by the hypsometric method a rise of ~350 m; the corresponding Termier & Termier figure is ~220 m (Hallam 1977a). Strakhov's figure compares much more closely than Termier & Termier's to my (42%) and Bond's (1976) figure (45%) for North America, and for a variety of reasons Strakhov's data are here considered to give a more accurate, though less precise, picture of Phanerozoic sea-level change in general.

Sleep (1976) observed that the inferred late Cretaceous coastline in Minnesota changed little in elevation, which implies only a slight amount of local tectonic disturbance. If epeirogenic "noise" is minimal, the height of sea level may be estimated directly; thus, Sleep arrived at a figure of 375 m for late Turonian–early Coniacian times, which compares very well with the Strakhov-derived figure and with Bond's (1976) figure of 390 m for the highest Cretaceous stand. Bond (1978) subsequently arrived at a lower global estimate of 150–200 m as a result of taking into account variations in continental hypsometries, while Watts & Steckler (1979) proposed an even lower figure of ~ 100 m as a result of their study of the subsidence history of the continental margin of eastern North America.

By using the entirely independent method of measuring the changing volumes through time of the oceanic ridge system, Pitman (1978) obtained a figure of 350 m for 85 m.y. B.P., which is in close agreement with the earlier figures cited above.

Hancock & Kauffman (1979) recognized from their study of the late Cretaceous deposits of western Europe and the United States Western Interior a late Campanian–mid Maastrichtian maximum of ~650 m. Their method involves first determining the present height of sea level for a given stage in tectonically undisturbed areas, then adding the thickness of Upper Cretaceous marine sediments up to that stage and the assumed depth of sea at the time of deposition. Since their figure is considerably higher than the others, it seems possible that Hancock & Kauffman might have (*a*) overestimated the depth of deposition or (*b*) underestimated the amount of subsequent epeirogenic uplift of certain areas such as the Western Interior. On the other hand, the value of Watts & Steckler (1979), based only on a regional study, seems too low; hence the figure of ~350 m, which falls almost halfway between the extreme values, is here accepted as being the most reasonable, having been arrived at, more or less, by three independent methods.

The general smoothed trend of the eustatic curve presented in Figure 5*A* has been determined on the basis of the extent of marine epicontinental cover as estimated from paleogeographic maps, with the sea-level values estimated by the method of Forney (1975), using his hypsometric curve. For the Ordovician maximum, the Strakhov-derived global figure is ~50% marine cover (~450 m above present), whereas the corresponding figures derived from the more detailed USSR and North American data are ~60% and ~600 m (Hallam 1977a). While there is an admitted danger of overestimating the ancient sea level because of subsequent epeirogenic uplift, the Strakhov maps, which are several decades old, are thought to underestimate the areal spread of sea in the early Paleozoic to a significantly greater extent than for later periods (for the reasons outlined above), and the higher figure of 600 m is here accepted as being of the right order of magnitude.

As regards the second highest sea-level stand, in the late Devonian and early Carboniferous, the Strakhov and North American figures correspond closely (~40%, 380 m; ~43%, 400 m, respectively), although the results for the USSR are somewhat higher (Hallam 1977a).

Apart possibly from the earliest Cambrian, the lowest stand was at the Permian-Triassic boundary, for which time Forney (1975) estimates a value ~40 m below the present. This is supported by the analysis of Cogley

(1981), who draws attention to the existence of continental Permo-Triassic deposits beneath large areas of present shelf seas.

Superimposed on the general trend are a series of oscillations corresponding to Vail's second-order cycles and to successive marine transgressions and regressions over continental margins and interiors. In many

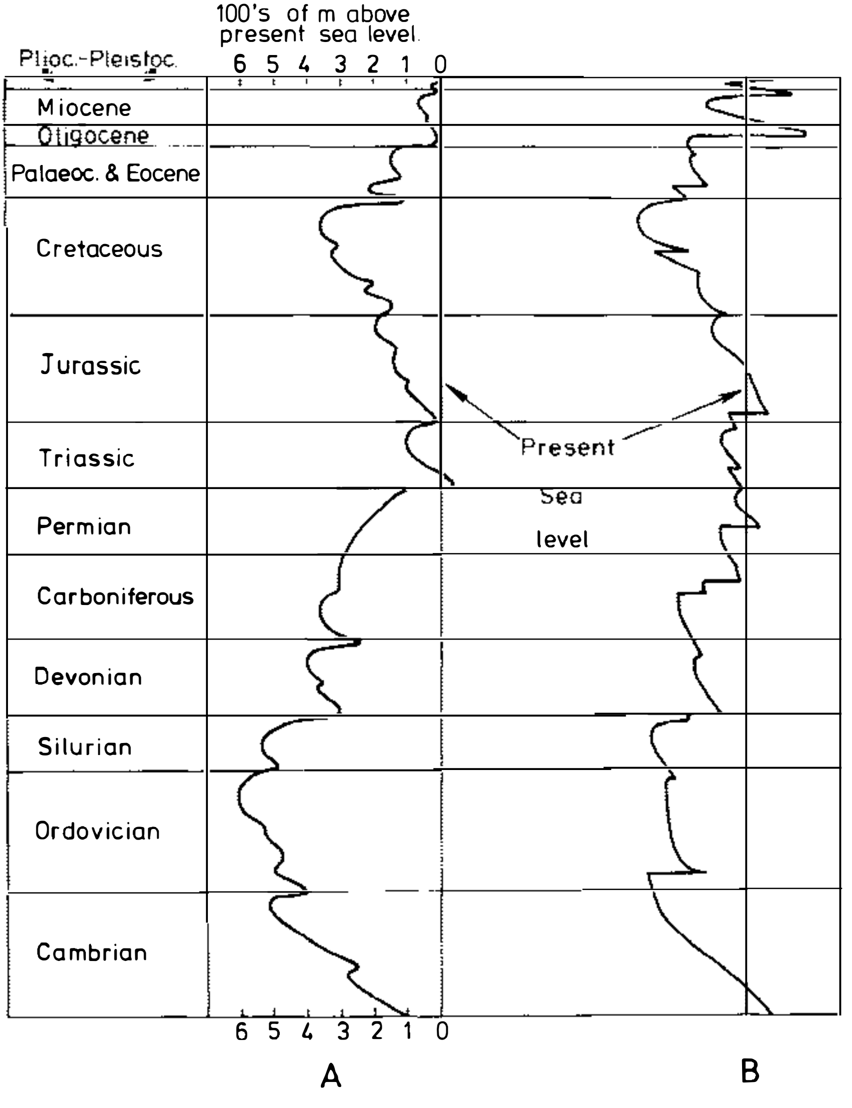

Figure 5 Eustatic curves for the Phanerozoic. *A*. This paper. *B*. After Vail et al (1977).

cases, data are presently inadequate to determine the extent of such sea-level changes, so that there is here a considerable element of "guesstimation" in producing the curve. The evidence used in producing the curve is based not only on reports addressed specifically to eustasy that deal with more than one continent, but also on general stratigraphic accounts of cratonic regions across the world: in Eurasia (Ager 1980, Nalivkin 1973, Lee 1939), Africa (Furon 1963), Australia (Brown et al 1968), and North America (Levin 1978). Only the more important events are noted.

Paleozoic

The Vail curve (Figure 5*B*) shows a progressive rise of sea level through the Cambrian from a low stand slightly lower than today at the Precambrian-Cambrian boundary. The sea-level rise is supported by stratigraphic evidence of transgressive deposits on cratons across the world (Matthews & Cowie 1979). The recognition in recent years of Tommotian deposits in a number of places suggests that perhaps the earliest Cambrian seas were more extensive than indicated in the Vail curve. Oscillations within the Cambrian have not yet been clearly established; the likeliest candidate is a regression followed by a transgression across the early–mid-Cambrian boundary, the evidence for which is especially well marked in Europe. The Vail curve indicates that the Paleozoic seas reached their maximum extent in the late Cambrian (or the beginning of the Ordovician). This is not accepted here; instead, the late Ordovician seas are considered to have been slightly more extensive.

Ordovician and Silurian eustasy is discussed by Leggett et al (1981), Lenz (1982), and Fortey (1984); where there is disagreement, the results of Fortey are favored because his evidence spans the largest number of regions across the world. There is general support for Vail of a significant fall of sea level at the end of the Cambrian, followed by an early Ordovician (Tremadoc) rise. Minor sea-level falls also mark the Tremadoc-Arenig and Arenig-Llanvirn boundaries. While the early part of the mid-Ordovician (Llanvirn) was a time of relatively low Ordovician sea level, there was a notable rise later on, in the Llandeilo, which is marked by an especially wide spread of sea in North America. The biggest Ordovician transgression however, was in the Caradoc, when the seas reached their greatest extent. (In eastern North America this phenomenon is obscured by the effects of the Taconic orogeny.) There is general agreement that sea level dropped sharply at the end of the period (Ashgill) and was followed by a rapid rise at the start of the Silurian (Llandovery). In the Vail curve only a very minor fall is indicated, shortly *before* the end of the Ordovician. From their facies analysis in Norway, Brenchley & Newall (1980) suggest that the end-Ordovician sea-level fall was 50–100 m.

The Silurian appears to correspond with one major eustatic cycle, the Llandovery transgression continuing to a maximum at about the Llandovery-Wenlock boundary, and then followed by a progressive fall to the end of the period, although this latter trend might have been interrupted by a minor Ludlow rise. The indication in the Vail curve that sea level fell at the end of the Silurian to its then lowest stand since the early Cambrian is well supported by cratonic stratigraphy across the world, but the sharpness of the fall is greatly exaggerated.

The lowest sea-level stand might actually have been in the late Gedinnian, shortly after the beginning of the Devonian, because there are a number of areas in Europe and North America where marine conditions may have continued from the Silurian (House 1975a). House's (1975a,b) analysis of Devonian facies and faunas concentrates on Europe and North America but also takes into account evidence from other continents. He infers eustatic control for major early mid-Devonian (Eifelian) and early late Devonian (Frasnian) transgressions, leading to progressive onlap and a Frasnian maximum spread of sea. A number of minor transgressive phases of less certain origin are recorded from the Siegenian to the early Famennian, and major eustatically induced regressions in the early (Gedinnian and Emsian) and late (Famennian) Devonian. There is broad agreement with the Vail curve insofar as progressive (but stepwise) sea-level rise was followed near the end of the period by a more rapid sea-level fall.

Sea level began to rise again at or slightly before the Devonian-Carboniferous boundary (House 1975b) and reached a Carboniferous maximum in the late Dinantian (Visean), which marks the last really extensive global spread of neritic carbonates. Areal plots of the spread of marine sediments, either on the global or continental scale, show a significant reduction from the Lower to the Upper Carboniferous (Hallam 1977a), supporting a sea-level fall in the Vail curve from the Mississippian to the Pennsylvanian. That this was a sudden and considerable fall, the fourth biggest in the Paleozoic according to the Vail curve, is based on evidence of significant regression in the North American midcontinent. According to Ramsbottom (1981), this is more likely to be the result of epeirogenic uplift on a regional scale than a sea-level fall. His own analysis of British Carboniferous deposits, using the Bond (1976) technique for eliminating the effects of regional tectonism, suggest a ~300-m rise of sea level through the Dinantian and Namurian, followed by a period of stillstand in the Westphalian. He notes, however, evidence of widespread nonsequences at the Lower-Upper Carboniferous boundary in several continents. Here, a more modest sea-level lowering from the Lower to the Upper Carboniferous than that proposed by Vail is accepted. It is likely that the regressive nonmarine facies widely developed in the late

Carboniferous is associated with the onset of a major phase of the Hercynian Orogeny in several continents. That these deposits are frequently as widespread as the underlying marine deposits argues against a significant sea-level drop, because they must have been laid down close to base level.

There has been no modern analysis of oscillatory eustatic changes across the Carboniferous-Permian boundary, but Schopf's (1974) analysis of the areal spread of marine sediments indicates a continuing regression through the Permian, accelerating toward the end of the period. Forney's (1975) analysis, using Schopf's data and the hypsometric method, suggests a late Permian sea-level fall of between 125 and 225 m. While early Permian epicontinental seas were evidently more extensive than late Permian ones (which were probably even less extensive than today), there seems to be little support for the Vail proposal of a sudden and marked mid-Permian sea-level fall, followed by a slight rise to the end of the period. Sea level must have been at its lowest at the very end of the period; and even in those few areas where there is an apparently continuous marine transition to the Triassic, as in the Caucasus, the Salt Range of Pakistan, East Greenland, and South China, argument persists about whether or not nonsequences occur at the boundary.

Mesozoic

Like the Silurian, the Triassic exhibits a simple first-order cycle of transgression followed by regression that intercontinental analysis suggests was under eustatic control (Schopf 1974, Hallam 1981b). The sea began to spread in the early Triassic and withdrew from East Greenland and the West Australian coastal zone at the end of this time, not to return until the Jurassic. There was a rapid transgression in the early mid-Triassic (Anisian), with the seas spreading again after a minor interruption to a maximum in Ladinian-Carnian times, when about 17% of the continental area was inundated. Thereafter, sea-level lowering is indicated by shallowing of facies within the Tethyan zone and passage to nonmarine facies elsewhere.

According to Vail et al (1977), sea level reached its Triassic maximum height in the Norian; this is presumably based on the fact that the spread of both marine and continental sedimentary deposits reached a maximum at that time. This is certainly true of the Atlantic margin regions of Europe and North America. There are, however, two alternative explanations for the apparent paradox in these areas of onlap of more regressive facies; both are thought to be more plausible explanations than Vail's because there is no indication from either facies or stratigraphy that shallow epicontinental seas were choked by heavy terrigenous sediment influx. One explanation is that the spread of sediments increased through Triassic times as Hercynian mountains (probably horst blocks) were progressively worn down by erosion and intervening depressions filled up, leading to sediment "spill-over" onto the adjacent peneplains (Hallam 1981b). The second alternative involves flexural onlap on structural highs in regions of stretched lithosphere (Dewey 1982, Watts 1982).

There was a relatively sharp sea-level fall at the end of the Triassic, immediately preceded over a wide area by a transgression that might have been eustatically controlled (Hallam 1981b). Thus, across the Triassic-Jurassic boundary the only epicontinental sea outside the Pacific margins was in northwest Europe, and its presence there was, as already noted, at least partly related to subsidence of continental crust as a consequence of tectonic stretching. There is clear evidence, furthermore, of a regression followed by a hiatus at the system boundary. Thereafter, the *leitmotif* of Jurassic eustasy was one of interrupted sea-level rise to an Oxfordian-Kimmeridgian maximum, when epicontinental seas covered a quarter of the present continental area (Hallam 1978). The revised Jurassic eustatic curve of Vail & Todd (1981), based on a detailed analysis of North Sea data, shows broad overall agreement with mine, but there are important differences of detail. The most glaring is the sudden eustatic fall seen in the Vail & Todd curve at the end of the Hettangian, which was so large that sea level was not restored to its Norian and Hettangian height until mid-Jurassic (Bajocian) times. This is quite unacceptable on grounds of distribution and character of marine early Jurassic facies on the continents (Hallam 1981c, Loughman 1982). The early Sinemurian in fact was a time of significant eustatic rise, not fall, and the sea reached areas never covered in Hettangian or Norian times.

Whereas the different eustatic curves agree about important transgressive pulses in the Toarcian, Bajocian, and Callovian and about regression in the Aalenian, the Kimmeridgian, rather than the Oxfordian, is accepted by Vail & Todd (1981) as the time of the most significant late Jurassic sea-level rise. However, this relates at least partly to a major regional event in northwest Europe leading to the deposition of the Kimmeridge Clay. Global data suggest a sea-level fall near the end of the Jurassic (Hallam 1978), but according to Vail & Todd sea level did not fall significantly until the end of the Berriasian (early Cretaceous), an event corresponding in Europe to the rapid change from calcareous (Purbeck) to siliciclastic (Wealden) facies. This is evidently the result of disregarding upward facies changes indicating regression in favor of determining the time of downward shift of coastal onlap. Vail & Todd's curve shows a succession of short-phase eustatic oscillations for Kimmeridgian-Tithonian times, but this is more likely to relate to North Sea tectonics.

The Tithonian part of the Jurassic eustatic curve is, together with the Bathonian, the least well established because of difficulties of biostratigraphic correlation, but there was probably a transgressive pulse early in the stage. This is clearly suggested by data from the Andes, where marine Tithonian succeeds regressive evaporites and red beds of Upper Oxfordian and Kimmeridgian age. While the earlier regression, markedly counter to the world trend, clearly relates to Andean tectonics, it is worth noting that the Sinemurian, late Pliensbachian–early Toarcian, Bajocian, and Callovian pulses of sea-level rise are all clearly recorded by transgressive events in the southern Andes. Thus, the early Tithonian transgression may also have a eustatic component, especially as it appears that it correlates at least approximately with a relatively deep-water organic-rich shale unit in northwest Europe and western Siberia (Upper Kimmeridgian in the English sense, Lower Tithonian in the French sense).

Using the hypsometric method, a sea-level rise through the Jurassic of slightly under 200 m is indicated. The results of three different methods of estimating the amount of eustatic rise through the early Jurassic are presented here. The hypsometric figure is $\sim$110 m (Hallam 1981c), while the Vail & Todd figure, based on seismic stratigraphy, is $\sim$75 m. The Bond (1976) technique can be applied to the English Lower Jurassic best in an area of stable platform comparatively resistant to subsidence, using data from the North Creake borehole mentioned earlier; the ammonite evidence indicates that no significant hiatuses are present (Hallam & Sellwood 1976). The value arrived at is $\sim$70 m, but since the sequence is mainly argillaceous, some increase should be allowed for compactional effects. The level of agreement between the three methods is encouraging.

The Vail curve for the Cretaceous shows a general rise to a Campanian-early Maastrichtian maximum (followed by a sharp and considerable end-Maastrichtian fall) from a low early Valanginian stand after a sudden end-Berriasian fall. A minor mid-Aptian and major mid-Cenomanian sudden fall interrupt the general trend.

In the early Cretaceous, the important early Hauterivian rise claimed by Vail is supported by evidence of a major transgression at this time (Cooper 1977). Cooper also argues for a notable eustatically controlled regression within the Aptian. For the younger Cretaceous there is a broad agreement between the eustatic curves of Vail et al (1977) and Hancock & Kauffman (1979). Hancock & Kauffman conclude that sea level rose from an early Albian stand close to that of today, to a late Campanian–early Maastrichtian peak of $\sim$650 m above present; this was followed by a rapid fall considerably thereafter in the late Maastrichtian. The only notable interruption was a sharp fall in the late Turonian (not the mid-Cenomanian as maintained by Vail). Five eustatically controlled transgressive peaks are recognized: late Albian, early Turonian, Coniacian, mid-Santonian, and late Campanian–early Maastrichtian.

Tertiary

Within the Tertiary, the most important change concerns the Oligocene. According to Vail et al (1977), there was a huge sea-level fall of nearly 400 m, by far the biggest in the Phanerozoic, that took place suddenly in the late Oligocene in a mere geological instant. This appears to be based on data from four regions: the North Sea, northwest Africa, the San Joaquin Basin of California, and the Gippsland Basin of Australia (Vail et al 1977, Part 4, Figure 5). Of these, only the North Sea and northwest Africa show a pronounced drop, and the Gippsland Basin hardly records the change at all. Yet the Vail curve is heavily weighted, without explanation, in favor of the North Sea.

Although there is strong evidence for a major Oligocene regression, the Vail estimates of the timing, extent, and speed of the eustatic fall are open to question. Thus, Olsson et al (1980), on the basis of a facies and biostratigraphic study of the United States Atlantic coastal plain, argue cogently for a moderately rapid fall of perhaps as much as 150 m in the *early* Oligocene to account for a widespread gap between youngest Eocene and early late Oligocene. This timing of the major fall and subsequent rise is supported by evidence from the Gulf Coast (Murray 1961), western Europe (Pomerol 1973), Australia (Carter 1978), and southern Africa (Siesser & Dingle 1981).

The *leitmotif* of Tertiary eustasy is, of course, one of more or less progressive fall through the era, with Paleocene seas being more extensive than those of the Eocene, and Oligocene seas less extensive still. Because, however, of the dramatic early Oligocene fall, Miocene seas were *more* extensive than the Oligocene, reaching their maximum toward the middle of the period, or just after (Serravalian), suggesting a sea-level peak at this time. Among the more important oscillatory changes, it is generally agreed that there was a rapid end-Miocene fall of perhaps as much as 100 m (Adams et al 1977, Loutit & Keigwin 1982), followed by an early Pliocene rise. It is here suggested that the end-Paleocene regression was more important than the mid-Paleocene one proposed by Vail et al. It is uncertain whether the Eocene transgressive cycle reached its eustatic peak early or late in the period.

Smaller-scale eustatic cycles than those portrayed in Figure 5, such as Vail's third-order Mesozoic and Cenozoic cycles or those reported by Cooper (1977), Hallam (1978), and Ramsbottom (1979) among others, are naturally more difficult to establish because a smaller eustatic change is more likely to be swamped by local or regional tectonic processes. Each individual case must be scrutinized carefully, using the most refined biostratigraphic data available and the largest possible number of regions, together with an analysis directed at eliminating the effect of noneustatic events.

CORRELATION WITH SEDIMENTARY FACIES AND ISOTOPIC CHANGES

The correlation between spreads of regressive siliciclastic, carbonate, and evaporite facies and sea-level low stands is too obvious to pursue further, but some other, less obvious relationships warrant brief attention.

A strong association between the stratigraphic distribution of finely laminated organic-rich shales and marine transgressive episodes has been recognized throughout the Phanerozoic (Hallam & Bradshaw 1979, Jenkyns 1980, Leggett et al 1981). Notable examples include widespread deposits in the Caradoc, the Llandovery, the Devonian-Carboniferous boundary, the early Sinemurian and Toarcian, the Callovian, the early Tithonian, and the Aptian-Albian and Cenomanian-Turonian boundaries. The last two have been related to oceanic "anoxic events" (Jenkyns 1980), but the lack of DSDP data for times prior to the Callovian rules out the confirmation of this for older deposits. The most important reason for the association is probably that the early stages of transgression over the continents are characterized by broad stretches of poorly oxygenated shallow water (with restricted circulation with the open ocean) that provide a short transit for organic matter from productive surface water to the bottom sediments. Consequently, there is less oxidation and greater retention of organic matter (Hallam 1981a, Arthur & Jenkyns 1981).

Whereas phosphorite genesis correlates in general with times of high sea level and warm climate, it correlates more specifically with marine transgressions, whereby a number of shallow pericontinental and epicontinental sites are available for phosphate fixation. In other words, such shallow seas tend to sequester nutrients and in consequence are highly fertile. The resultant high plankton productivity will in turn lead to reduced oxidation in this particular setting and therefore to the greater initial retention of organic phosphorus. On the other hand, oceanic anoxic events correlate with high sea-level stands (Jenkyns 1980) but do not generally correlate well with major phosphorogenic episodes. This may be because abundant phosphorus is fixed in the deep ocean by organic carbon and is therefore relatively unavailable in the epicontinental seas (Arthur & Jenkyns 1981).

Scholle & Arthur (1980) found that the Aptian-Albian and Cenomanian-Turonian anoxic events correlated with episodes of heavy $\delta^{13}C$ carbonate carbon. This was tentatively attributed to the increased preservation by burial of light $\delta^{13}C$ organic carbon during anoxic episodes, so that the dissolved carbon remaining in the oceanic reservoir became progressively heavier isotopically. An alternative explanation is proposed by Loutit & Keigwin (1982), whose isotopic study of carbonate carbon records an abrupt shift coinciding with the previously reported end-Miocene eustatic fall. They argue that the shelves have acted as carbon sinks during regressions, when erosion supplies previously deposited organic matter, depleted in ^{13}C, to the open ocean. Thus oceanic $\delta^{13}C$ values can be decreased by introducing isotopically light organic matter during sea-level low stands.

As regards regressive episodes, Hallam & Bradshaw (1979) noted for the western European Jurassic a common association with oolitic ironstones, which frequently cap shallowing-upward marine sequences. Van Houten & Karasek (1981) report a similar relationship for the Devonian of Libya. On occasions, Jurassic ironstones may actually occur at the base of a transgressive sequence, but the facies relationships always indicate that they are among the shallowest-water, most "regressive" deposits. In general, the formation of oolitic ironstones is related to episodes of relatively high sea level, mild climate, and tectonic stability accompanied by reduced influx of siliciclastic detrital sediment. The facies associations of the ironstones indicate repeated shallowing-upward sequences, abrupt waning of sediment supply, and rapid renewed sea-level rise; the ironstones tend to occur at the sharpest lithofacies discordance (Van Houten & Bhattacharyya 1982).

Another highly intriguing relationship, in this case between sea level and the $^{87}Sr/^{86}Sr$ ratio of seawater, has been pointed out by Spooner (1976). The ratio of seawater strontium (0.7091) is lower than the ratio of dissolved strontium delivered to the oceans by continental runoff ($\sim$0.716). This difference can be accounted for by the exchange with the isotopically lighter Sr of oceanic crust during the process of hydrothermal convection within spreading ridges. Spooner demonstrated a good correlation between estimates of the increase in land area in the last 70 m.y., related to fall of sea level, and the strontium isotope ratio. This implies that there is covariation between the increase of seawater $^{87}Sr/^{86}Sr$ and the increase in the continental runoff flux.

Spooner assumed that land area and continental runoff are directly proportional, but consideration of the Phanerozoic as a whole, based on the marine carbonate analyses of Veizer & Compston (1974), suggests a more complicated picture. Their data indicate variation in the sea-water ratio from a maximum in the early Cambrian (0.7093) that is comparable to the present-day value (0.7091) to a minimum in the late Jurassic (Oxfordian; 0.7069).

In Figure 6*B* the broad trend of Phanerozoic change in the strontium isotope ratio is plotted adjacent to the eustatic curve of Figure 5*A*. Whereas there is good general agreement between the trend and falling sea level since the Cretaceous, as Spooner pointed out, the minimum is in the late Jurassic, not the late Cretaceous, as might have been predicted from Spooner's hypothesis. Similarly, the Permo-Triassic interval fails to show the

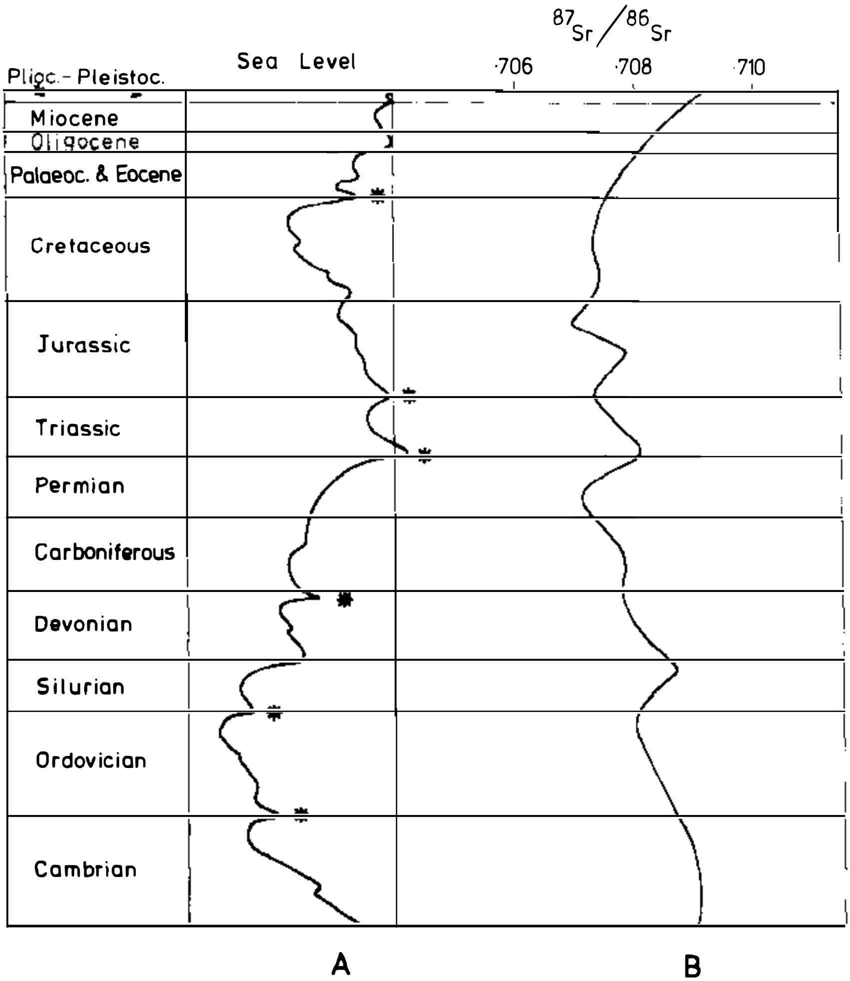

Figure 6 *A*. The Phanerozoic eustatic curve of Figure 5*A*, with asterisks signifying the mass-extinction phases of Newell (1967). *B*. Curve representing the changing $^{87}Sr/^{86}Sr$ ratio through the Phanerozoic, based on Veizer & Compston (1974, Figure 5). In the Veizer & Compston diagram the curve is represented by a band signifying a modal tendency of a wide scatter of data points.

conspicuous peak that would be expected from a simple runoff–land area correlation, with the curve from the Carboniferous rising, not falling. If, however, climatic variations are considered, some form of runoff hypothesis can be salvaged. Thus the Permo-Triassic was a time of widespread aridity after the humidity of the Carboniferous (Frakes 1979), so that runoff would have been correspondingly reduced. Furthermore, there is good evidence that runoff increased from the Jurassic to the Cretaceous. Thus, contrary to Frakes' (1979) claim that a trend of increasing global aridity through the Jurassic continued into the Cretaceous, Middle and Upper Jurassic evaporite-bearing red beds in the United States Western Interior are succeeded by Cretaceous coal measures, while in large areas of western and southern Europe, carbonate- and evaporite-bearing "Purbeck" facies are succeeded by coarse siliciclastic paralic and fluvio-deltaic "Wealden" facies. A similar pattern of change in the Middle East persuaded Murris (1980) of a Jurassic to Cretaceous change to a more humid climate. More generally, a global diminution in evaporite deposition (Meyerhoff 1970, Gordon 1975) is consistent with a change from a more arid Jurassic to a more humid Cretaceous climate.

In earlier times, the late Silurian–early Devonian isotope peak corresponds well with the conspicuous sea-level low in the eustatic curve and thus with a simple runoff–land area relationship, as does the fall through the Devonian and the fall from the Cambrian to a late Ordovician trough. The general level of the early Paleozoic curve is anomalously high compared with the rest of the Phanerozoic, since the eustatic curve indicates a minimum of exposed land. It should, however, be borne in mind that continental runoff must have been higher before the Devonian because of the absence or sparsity of vegetation cover. (This is an evolutionary phenomenon, of course, and not a function of the degree of aridity.)

Thus, while a good correlation may exist between the amount of continental runoff and the strontium isotope ratio of seawater, this amount is a function of at least three variables: area of exposed land, climate, and extent of vegetation cover. Further isotope data are required to confirm the reliability of the Veizer & Compston curve. If the effect of sea-level change can be partitioned out, there is promise of an independent monitor both of climatic change and of the evolution of terrestrial vegetation in the Paleozoic.

CORRELATION WITH BIOGEOGRAPHY AND ORGANIC EVOLUTION

Compared with the open ocean, epicontinental seas are much more likely to have been affected by changes in the physical environment. Even a quite modest change of sea level could have had significant environmental consequences, not only in the extension of marine habitats but in variations of temperature, salinity, and oxygen content within large bodies of shallow water with somewhat restricted interchange with ocean water (Hallam 1981a).

In terms of biogeography one might reasonably predict that organisms occurring in the shallower-water deposits of, for instance, cratonic interiors should exhibit a greater tendency to endemism than deeper-water, more offshore organisms because of their greater tendency to isolation at times of regression. That this is true for groups as different, and as widely separated in time, as Ordovician trilobites (Fortey 1975, 1984) and Jurassic bivalves (Hallam 1977b) suggests the existence of a phenomenon of more general importance.

Furthermore, it follows that times when epicontinental seas were widespread should correlate with times when the incidence of cosmopolitan faunas was relatively high; conversely, provincial marine differentiation of neritic organisms should be high at times of regression. This prediction is borne out for Jurassic mollusks (Hallam 1977b), and, more generally, for the Paleozoic. Thus there was a relatively high incidence of endemism in the early Devonian, late Carboniferous, and late Permian, and a low incidence in the late Ordovician, mid-Silurian, and late Devonian to early Carboniferous (Ziegler et al 1981). The degree of endemism does not, however, invariably fall with sea level, or at least with the reduced extent of sea; much depends on the particular geographic situation. Thus the creation of a new seaway in the North American Western Interior in late Cretaceous times led to the evolution of endemic ammonites (Kennedy & Cobban 1976) and bivalves (Kauffman 1973).

The grosser patterns of evolution can be considered in terms of extinctions and radiations. Since episodes of mass extinction and radiation tend to coincide for a diversity of groups of different biology and habitat, this argues against the Darwinian view that emphasizes the paramount importance of biotic competition, and in favor of first-order control by the physical environment (Hallam 1983b).

Newell (1967) was the first to relate explicitly episodes of mass extinction among diverse animal families to eustatic falls of sea level. Of the major extinction phases he recognized (at or close to the end of the Cambrian, Ordovician, Devonian, Permian, Triassic, and Cretaceous), all but the first have recently been confirmed by the statistical analysis of Raup & Sepkoski (1982). Details of the groups concerned are given in Newell (1967) and Hallam (1981a).

Newell's extinction phases are indicated in Figure 6*A*, which confirms his conclusion about the correlation with sea-level falls. The detailed analysis by Schopf (1974) and Simberloff (1974) demonstrates that the fall in marine invertebrate diversity through the Permian correlates well with the inferred fall of sea level and accords well with ecological predictions based on the so-called "species-area relationship." As the area of epicontinental sea habitat declines, so the extinction rate increases; hence, the fall in diversity. There are some statistical grounds for believing that the species-area relationship may hold for the Phanerozoic as a whole (Sepkoski 1976, Flessa & Sepkoski 1978).

Doubt remains in particular cases about whether the extinction rate increased directly because of regressions, or whether it was because of the widespread anoxic bottom conditions that characteristically mark the beginning of the succeeding transgression (e.g. Hallam 1981b). The latter phenomenon was clearly more significant in causing the mass extinction of marine invertebrates in Europe during the early Toarcian (Hallam 1977b). Many species endemic to Europe disappeared, and the seas were subsequently repopulated by immigrants from the proto-Pacific. This example shows that physical events on a regional scale may be significant promoters of extinction where endemism is involved, although the overall controlling event was in fact eustatic. Whether, more generally, the immediate cause of extinction was regression leading to local emergence and more variable water temperatures and salinities in the shallower seas that remained, or instead was widespread anoxia, the end result would clearly have been increased environmental stress and drastic reduction in habitable area. As might be anticipated, the invertebrates most vulnerable to extinction, apart from ammonites and planktonic forams, tended to be those adapted to shallower-water reefal or perireefal habitats (Hallam 1981a).

The full ecological implications of sea-level fall have yet to be worked out. A brief consideration of the well-documented Pleistocene sea-level oscillations caused by glaciation and deglaciation suggests that these events do not by and large correlate with episodes of pronounced extinction or speciation. Indeed, the characteristic response of both terrestrial and marine organisms to the pronounced climatic changes of the Pleistocene has been to migrate to ecological refuges, in effect to track their environment. It might therefore be wondered why organisms did not respond in a similar way to the much slower changes of sea level in the lengthy periods of climatic equability.

At least two possibilities suggest themselves. Perhaps the Pleistocene sea-level falls, though dramatically rapid in geological terms, were too short-lived to have the kind of environmental impact required to cause extinction. Or perhaps the increasingly unstable environments of the late Cenozoic associated with climatic deterioration caused a selection for eurytopic organisms well adapted to withstand environmental instability. In contrast, the comparative stability of, for instance, Mesozoic environments might have allowed the establishment of complex ecosystems characterized by comparatively stenotopic organisms, which would have been vulnerable to even modest environmental vicissitudes.

Another point to be noted from Figure 6*A* is that not all major sea-level falls correlate with significant mass extinction events at the family level, although this does not of course rule out extinctions at generic and species levels. This is perhaps most clear in the case of the late Silurian–early Devonian eustatic low stand. Perhaps the most critical event is a relatively sudden fall, as appears to have happened at the end of the Ordovician, Permian, Cretaceous, and probably Triassic, that is too rapid for many or most organisms to adapt to.

It is especially intriguing to observe that the end-Permian, end-Triassic, and end-Cretaceous mass extinctions of marine groups coincide with mass extinctions of large terrestrial reptiles (Bakker 1977, Tucker & Benton 1982, Cooper 1982). Another such correlation is in the late Eocene, a time of mass extinction of archaic land mammals (Colbert 1955) and marine microfauna (Benson 1975, Corliss 1979). The controlling event in this case was probably a pronounced climatic decline associated with Antarctic glaciation, which had as one by-product a marked sea-level fall (see below), but such a significant glacially induced event can be ruled out for the other three correlated extinction events. A climatic factor may, however, be involved, because marine regressions will serve to increase the seasonal extremes of continental climate. In particular, Pangea at the end of the Paleozoic must have experienced a climate of extreme continentality, not only because of its coherence (Valentine & Moores 1972) but also because of the high albedo of extensive low-latitude deserts (Barron et al 1980).

Attention has recently been focused on an extraterrestrial catastrophe to account for the end-Cretaceous extinction of the dinosaurs and other organisms, with either asteroid or cometary impact giving rise, respectively, to inhibition of photosynthesis by ejected dust (Alvarez et al 1980) or oceanic poisoning of microplankton leading to increased CO_2 production and hence atmospheric temperature rise (Hsü et al 1982). The end-Cretaceous event was by no means catastrophic to many marine and continental groups, however, least of all apparently to tropical land plants, the group that should have been most affected by photosynthesis inhibition (Hancock 1967, Bakker 1977, Clemens et al 1981, Birkelund & Hakansson 1982). The end-Permian and end-Triassic reptile extinctions indicate that the dinosaur extinction was not unique, and in general the end-Cretaceous extinction event affected the world's fauna much less than the end-Permian one, when it is estimated that perhaps as many as 96% of marine species died out (Raup 1979).

The extinction of calcareous plankton at the end of the Cretaceous is really a greater enigma than the contemporary or near-contemporary extinction of the dinosaurs in terms of plausible terrestrial events. In particular it is difficult to see how such pelagic groups as globigerinid forams should have been affected by epicontinental sea regression. Nevertheless, Hart (1980) has proposed a factually well-supported model that relates their explosive diversification in the late Cretaceous to the increasing depth of epicontinental seas. The obvious implication, surprising as it may seem, is that the *decreasing* depth of such seas might have had a deleterious influence.

Not only is there in general a good correlation between eustatically induced regressions and increased extinction rates, but the converse is also true: major transgressions frequently correlate with mass radiations of marine organisms, presumably as a consequence of improvements in habitat area and quality. Thus, both the Cambrian and Ordovician invertebrate radiations have been related to contemporary sea-level rise (Brasier 1979, House 1967). Among many other examples, the successive Triassic and Jurassic eustatic rises correlate with two successive phases of ammonite radiation. Most spectacular of all is the explosive diversification of a large number of groups in the mid-Cretaceous (such as coccolithophores, foraminifera, diatoms, dinoflagellates, deep-sea ichnofauna, veneroid bivalves, neogastropods, crabs, and teleost fish) near the start of the biggest rise of sea level since the mid-Paleozoic (Hallam 1983b).

Smaller-scale evolutionary effects should not be neglected. Fortey (1984) has observed that the generic longevity of trilobites that lived in deeper-water, extracratonic habitats is greater than shallower-water, more inshore forms. This is presumably a consequence of the lower predictability and higher stress of the shallower-water habitats, promoting higher extinction and hence higher speciation rates at times of sea-level change on either the regional or global scale. A similar pattern has long been recognized among Jurassic ammonites. The deeper-water suborder Phylloceratina had higher generic longevity and are hence much less useful stratigraphically than the shallower-water Ammonitina. Even within the Ammonitina, there may be depth-related differences. Thus, Phelps (1982) found that within the Pliensbachian family Liparoceratidae, the species longevity of *Liparoceras*, inferred on facies grounds to have lived in a deeper-water habitat, is greater than contemporary shallower-water *Androgynoceras*. I have recognized a similar pattern among Sinemurian Arietitidae. The large genera of this family, such as *Arietites* and *Coroniceras*, are common only in shallower-water facies and had only brief time spans, whereas the small *Arnioceras* is abundant only in deeper-water shale facies and had an appreciably greater range in time.

There is increasing evidence that for many species, population increases ("radiations"), migrations, and extinctions can be related to transgressions and regressions on a regional or global scale in a way parallel to that for mass radiations and extinctions (e.g. Hallam 1983a). This suggests that events such as those at the end of the Paleozoic and Mesozoic may be merely the end members of a whole spectrum of terrestrially induced physical events that affected the biosphere.

CAUSES OF SEA-LEVEL CHANGE

Of the various possible causes of eustatic changes, only two are of any significance: melting and freezing of polar ice caps, and changes in the volume of the ocean basins (Donovan & Jones 1979). These have been, respectively, termed glacio- and tectonoeustasy (Fairbridge 1961).

With regard to glacioeustasy, the complete melting of all present land ice should cause a sea-level rise of 40 to 50 m (Pitman 1978). Sea-level depression at the time of the maximum volume of Pleistocene ice sheets is less certain, but Donovan & Jones (1979) make an approximate estimate of 100 m. For the late Tertiary, oxygen isotope data have suggested to Shackleton & Kennett (1975) that the Antarctic ice cap was first established about mid-Miocene times. The end-Miocene eustatic fall and subsequent early Pliocene rise can therefore plausibly be attributed to glaciation and deglaciation phenomena, respectively (Adams et al 1977).

There is abundant evidence from terrestrial plants, marine macrobenthos, and microplankton and oxygen isotopes of a dramatic fall in global temperatures across the Eocene-Oligocene boundary. Together with evidence of increased current scouring on the deep ocean floor, the marine data indicate the formation of the layer of cold, deep water known as the *psychrosphere* and an increased rate of oceanic circulation (Hallam 1981a). Sedimentary evidence from the peri-Antarctic sea floor, such as dropstones, has been held to indicate the initiation of sea (but not land) ice at this time (Kennett 1977). The evidence cited here shows that the most dramatic sea-level fall in the Tertiary took place in the late Eocene to early Oligocene, implying rather strongly that substantial glacial buildup had occurred on the Antarctic continent, even though the oxygen isotope data have been held not to support this. In fact, a shift to heavier isotopes can indicate *either* colder water *or* raised salinity due to the extraction of isotopically lighter water to produce polar ice, *or* both.

Further back in time there is no evidence of polar ice caps, and abundant evidence of global climatic equability, until the mid-Permian (Frakes 1979); consequently, any sea-level changes must have been tectonoeustatic in origin. Even for the time when the Gondwana ice sheets were in existence, purely glacioeustatic interpretations pose problems. Thus, on a small scale, the frequently assumed minor eustatic fluctuations produced by waxing and waning of ice sheets during alternating glacial and interglacial episodes carry implications for sedimentation in, for instance, the English Upper Carboniferous Coal Measures, deposited in a paralic setting close to sea level. In such a setting, eustatic lows should be signified by horizons of deep erosional channeling, but such features are rare (A. P. Heward, personal communication). On a larger scale, the waning and ultimate disappearance of the Gondwana ice sheets in the mid- to late Permian coincided with a *fall*, not a rise, of sea level, implying that glacioeustatic effects were swamped by tectonoeustatic effects.

It has been widely assumed that the end-Ordovician sea-level fall and the early Silurian rise were glacioeustatic in origin, with both related to the growth and decay of the Saharan ice sheet (Frakes 1979), but no other convincing case has been made for glacioeustasy in the Paleozoic; thus, tectonoeustasy must have been the process of overriding importance.

Turning then to tectonoeustasy, attention has been concentrated on the late Cretaceous rise and subsequent fall. Hays & Pitman (1973) correlated the rise with an episode of accelerated seafloor spreading, which would have caused a significant expansion in volume of the oceanic ridge system. This interpretation is based on the age-depth relationship of mid-oceanic ridges, which approximately follows a time-dependent exponential cooling curve (Pitman 1978). However, a more recent analysis of Cretaceous magnetic anomalies and a revised chronology significantly reduce the rate of acceleration from the early to the late Cretaceous demanded by the quantitative analysis of Hays & Pitman (1973) and Pitman (1978) (Hallam et al 1984).

Two other processes have been proposed to account for these events. The mid- to late Cretaceous was a time of significant increase in the length of the ocean ridge system, with the Atlantic Rift extending north and south and the Gondwana continents disintegrating. This must have caused a rise of sea level even without any acceleration in the spreading rate. Post-Cretaceous fall could, to some extent, relate to the progressive consumption through subduction beneath Asia and the Americas of ridges separating the Kula from the Farallon, and the Farallon from the Phoenix plate, together with the Pacific-Kula ridge (Hallam 1977a, 1980).

Schlanger et al (1981) argue that the late Cretaceous sea-level rise was mainly the result of mid-plate volcanism producing swells on the Pacific and Farallon plates, which began to subside about 70 m.y. ago, thereby causing sea-level fall.

All these tectonoeustatic processes, which have a common cause in variations of heat flow from the mantle, could have operated to variable extents throughout the Phanerozoic, and a condition of stable sea level for geologically significant periods of time should in consequence be considered an exceptional circumstance. It is sometimes possible to propose a plausible link between plate tectonic events and eustasy. Thus, Anderton (1982) explicitly relates the Cambrian sea-level rise to the opening of the Iapetus Ocean, associated with the creation of a spreading ridge. Valentine & Moores (1972) argued that the late Paleozoic sea-level fall was the result of the suturing of several continents to produce the supercontinent of Pangea, which caused the cessation of spreading, the collapse of ocean ridges, and the deformation and underthrusting during continent-continent collisions leading to emergence.

The Mesozoic rise of sea level corresponds broadly to the subsequent disintegration of Pangea and the concomitant growth of new spreading ridges, but new DSDP evidence (Sheridan 1983) supports the contention made on other grounds that the oldest, central sector of the Atlantic did not commence opening until late mid-Jurassic times (Hallam 1980). Thus the mid-Triassic and early Jurassic sea-level rises are unaccounted for by the Pangea disintegration.

According to Pitman (1978), the maximum rate of sea-level change producible by the growth and decay of ocean ridges is ~ 1 cm/10^3 yr, about three orders of magnitude slower than Pleistocene glacioeustatic change. This rate is slightly slower than the very approximate figures for short-term eustatic oscillations of Hallam (1978) and Ramsbottom (1979) cited earlier, and much slower than that for eustatic falls derivable from the data of Vail et al (1977). As noted earlier, the Vail "geologically instantaneous" eustatic falls appear to be an artifact of the method of analysis, but there remains a problem concerning the rates of sea-level oscillations, nevertheless.

Pitman's (1978) solution, based on a quantitative analysis of the interaction of sea-level change with sedimentation and subsidence on passive continental margins, is to demonstrate that transgressive and regressive events may not be simply indicative of eustatic rise and fall but of changes in the rate of sea-level change. Thus a decrease in the rate of sea-level rise and an increase in the rate of sea-level fall may result in regressions, while an increase in the rate of sea-level rise and a decrease in the rate of sea-level fall may produce transgressions.

By checking, however, continental margin stratigraphy with correlative events in the deep sea (Thiede 1981) or cratonic interiors, it is sometimes possible to eliminate such "hinge" phenomena from consideration and to indicate genuine sea-level oscillations. If the inferred rates of sea-level change appear too high, it could be that (*a*) the facies depth range in inferred eustatic sedimentary cycles has been overestimated, (*b*) the time interval has been underestimated, and/or (*c*) undetected regional tectonic factors have intervened. Only detailed facies analysis in a refined biostratigraphic framework will convincingly resolve such questions.

There remains the problem of how to account for the long-term change. When the considerable changes related apparently to the successive formation and disintegration of Pangea are allowed for, there appears a secular trend toward fall in sea level through the course of the Phanerozoic. To some extent this can be accounted for by successive phases of orogeny ("Caledonian," "Hercynian," etc), resulting in marginal continental accretion along subduction zones leading to increased mean continental thickness. This cannot, however, be the whole story because shield areas, unaffected by Phanerozoic orogeny, also exhibit evidence of the more or less secular regression (Hallam 1977a).

Since a loss of ocean water is extremely unlikely, as is extensive underplating of continental crust away from subduction zones, the most plausible explanation involves a gradual reduction through time of heat flow from the mantle (Turcotte & Burke 1978, Brown 1984). As the oceanic lithosphere has cooled, thermal expansion has diminished and hence the ocean basins have deepened by a modest amount through the course of the Phanerozoic, causing seawater to be drained off the continents.

Such an interpretation carries with it, of course, the implication that Precambrian seas were generally more extensive than even those of the Paleozoic. The late Precambrian regression, giving rise to a hiatus sometimes known as the *Lipalian interval*, and subsequent Cambrian transgression are more likely to have a tectonoeustatic origin rather than a glacioeustatic one (Matthews & Cowie 1979). The regression is perhaps analogous to that which took place at the end of the Paleozoic.

SUMMARY

Eustasy can be studied using a variety of methods, including areal plots of the changing temporal distribution of marine deposits, facies analysis of stratigraphic sequences, and seismic stratigraphy, allied with the best available means of biostratigraphic correlation. The results of these various methods are then compared for use in eliminating the complicating effects of local and regional tectonics in the interpretation of sea-level oscillations. The determination of the rate and amount of sea-level change is also discussed.

Use is made of areal plots, in conjunction with hypsometric data and a variety of stratigraphic sequence evidence, to produce a eustatic curve for the pre-Quaternary Phanerozoic. Notwithstanding its necessarily tentative and provisional nature, this curve is considered to be a more accurate representation of Phanerozoic eustasy than that of Vail et al (1977). There is

an overall trend of declining sea level from a late Ordovician high stand ~600 m above the present, on which several major and a larger number of minor oscillations are imposed. The most important oscillations are (*a*) a late Paleozoic fall to an end-Permian low stand and (*b*) a late Mesozoic rise culminating in the late Cretaceous, when sea level was ~350 m higher than today. A more accurate eustatic curve will only be produced by a collaborative effort involving detailed biostratigraphic correlation, stage-by-stage facies analysis, paleogeographic map production, seismic stratigraphy, and regional tectonic analysis.

A strong correlation exists between marine transgressions and the distribution of organic-rich shales and phosphorites, as does a somewhat weaker correlation between regressions and oolitic ironstones. Correlations with $^{13}C/^{12}C$ and $^{87}Sr/^{86}Sr$ are also noted. As regards the latter, there appears to be covariation between the increase in the $^{87}Sr/^{86}Sr$ ratio in seawater and the amount of continental runoff. Data for the whole Phanerozoic suggest that runoff is a function of the continental area, the degree of aridity, and the evolutionary development of a mantle of terrestrial vegetation in the early to mid-Paleozoic.

There is also a good correlation between eustatic changes and events in the biosphere. Times of low sea level correlate with episodes of increased endemicity and extinction of marine organisms, while times of high sea level are associated with episodes of increased pandemicity and radiations. The celebrated end-Cretaceous mass extinction is not unique but simply one of the most spectacular end-members of a whole series of extinction events that appear to relate to global or regional regressions of corresponding magnitude. This makes it questionable that extraterrestrially induced catastrophes need be invoked.

Changes in the volume of the ocean basins (tectonoeustasy) are generally more important than glaciation and deglaciation in controlling sea-level variations prior to the Quaternary, but the early Oligocene and late Ordovician regressions are probably due to the growth of polar ice caps. The gross features of the Phanerozoic eustatic curve are primarily the result of plate tectonic processes involving continental collision and disintegration (with concomitant changes in ocean ridge volume) and a secular withdrawal of epicontinental seas due to ocean-basin deepening related to the slow cooling of the lithosphere. Smaller-scale oscillatory changes are more controversial and demand more intensive study using a wide array of stratigraphic, sedimentological, and paleoecological methods.

Bearing in mind one of the morals of the continental drift controversy, and the hostility of an earlier generation of geophysicists to Wegener's proposals for the controlling processes, we should seek to establish eustatic oscillations firmly on empirical grounds and make the results as quantitative as possible, rather than dismiss them as unlikely because no appropriate controlling force can be conceived. At least three tectonoeustatic processes can be proposed, but too little is known of their relative importance in the past to permit categorical statements at the present time. On the other hand, the effects of regional as opposed to local epeirogenic movements have not always been adequately appreciated by students of eustasy; these need much more research.

According to Brown (1984), "Throughout the Earth's history, internal heat production has provided the dominant constraint on lithosphere evolution." It is becoming increasingly evident that it has also exercised a major effect on biosphere evolution, thereby helping to vindicate the claim made many years ago by Chamberlin (1909) that significant organic changes through time have ultimately been under the control of diastrophism.

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