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7 Seismic Stratigraphic Record of Sea-Level Change

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ABSTRACT

Seismic stratigraphy is a technique for stratigraphic correlation, which involves the identification of regional unconformities (sequence boundaries) in seismic reflection profiles. These surfaces form during times of relative sea-level fall as a result of abrupt basinward shifts in sites of sediment accumulation, and many have been interpreted to have a eustatic origin. Patterns of progressive onlap punctuated by downward shifts in onlap have been quantified in numerous petroliferous basins and form the basis for a eustatic sea-level curve. The interpretation of sea-level change from seismic stratigraphic data is controversial, however, and the purpose of this chapter is to evaluate critically the method and assumptions by which the sea-level curve has been obtained.

The geometry of stratal surfaces can be determined from seismic sections because these surfaces are among the most abundant reflectors of seismic energy in sedimentary basins. Interference between reflections from different surfaces limits vertical resolution to about one-quarter of the acoustic wavelength (typically on the order of tens of meters), and because the seismic pulse travels as a spherical wave front, there are also limits to horizontal resolution, that is, to the accurate determination of the spatial location and size of features that generate acoustic reflections. Sequence boundaries are recognized by the oblique termination of seismic events (onlap, downlap, toplap, and erosional truncation), and if appropriate velocity and density logs or vertical seismic profiles (VSPs) are available from boreholes, it may be possible to locate a given unconformity in the rock stratigraphy to well within the level of seismic resolution (on the order of meters). Nevertheless, owing to the limits of resolution, many more unconformities are present in a sedimentary basin than can be confidently traced with seismic data. Not all seismic events have primary stratigraphic significance, but nonstratigraphic events generally do not interfere seriously with stratigraphic interpretation. The assumption that unconformities have chronostratigraphic significance is generally a good approximation at an intrabasinal scale for coastal areas and shallow continental shelves. In this chapter, two examples of diachronous unconformities are presented that raise questions about the universal application of seismic sequence analysis to the study of eustatic changes.

Unconformities associated with downward shifts in coastal onlap result primarily from an increase in the rate of sea-level fall or a decrease in the rate of subsidence. A prominent shift in onlap can occur when the rate of sea-level fall is comparable to the rate of tectonic subsidence (<1 cm/1000

yr). Such an onlap shift is not necessarily accompanied by rapid regression of the shoreline. The controversial issue of whether sea level was oscillatory or monotonically decreasing from the Late Cretaceous to the onset of glaciation in the mid-Cenozoic is unresolved by published modeling studies, which did not account for the magnitude of short-term changes in paleobathymetry or sediment supply. The formation of sequence boundaries of probable eustatic origin at times of long-term sea-level rise (such as the Early Cretaceous) suggests that sea-level changes may generally be oscillatory, though of small amplitude. The approximate correspondence of several second-order sequence boundaries with times of major plate-boundary reorganization may reflect regional bias in the data, but in part is probably the result of tectonoeustasy. While the idea of global tectonic events cannot be entirely discounted, no viable mechanism has yet been established for producing correlative unconformities on a global scale. Most tectonic mechanisms for generating sequence boundaries predict dissimilar ages in different basins.

The key to distinguishing boundaries of eustatic and local origin is geochronological resolution. Although there are inherent uncertainties in the time scale, and in the correlation of seismic and rock sections with each other and to the time scale, many second-order boundaries appear to be very persistent from one basin to another. The occurrence of globally synchronous sea-level change is also corroborated by abundant conventional stratigraphic evidence for much of the Phanerozoic. The spacing of third-order boundaries is close to or finer than biostratigraphic resolution, and it may never be possible to resolve the ages of many of these boundaries with sufficient precision for objective correlation between basins.

Amplitudes of eustatic fluctuations cannot be inferred from seismic stratigraphic data alone because coastal aggradation (the vertical component of onlap) is primarily a result of basin subsidence, not sea-level rise; and downward shifts in onlap reflect only the rate of sea-level fall in relation to the rate of basin subsidence. Variations of coastal encroachment (the horizontal component of onlap) are sensitive to lateral gradients in the rate of subsidence, which are in general time-dependent and not necessarily linear. Whichever method is used to gauge changes in coastal onlap, the large component of subsidence cannot be easily or objectively removed to derive the smaller eustatic signal, and similarities in the patterns of coastal onlap for different basins for the most part indicate a similar overall subsidence history. For the purpose of deriving a eustatic sea-level curve, the global onlap chart provides information only about the times of sea-level fall and rise. We conclude that the main limitations of the eustatic curve derived from seismic stratigraphy are (1) the uncritical interpretation of all second- and third-order boundaries as eustatic, (2) the uncertainties about the calibration of many boundaries to the geological time scale, and (3) the largely conjectured inference of amplitudes.

INTRODUCTION

Conventional Stratigraphic Record of Sea-Level Change

Oscillatory changes in sea level relative to the continents, on time scales of <10 m.y., have long been inferred from paleobathymetric variations in facies successions, and from stratigraphic evidence for transgressions and regressions, that is, the alternating landward and seaward migration of the shoreline (e.g., Suess, 1906; Hallam, 1984). If consistent changes in water depth or shoreline location are determined on a regional to intercontinental scale, in a range of siliciclastic to carbonate facies, a eustatic (i.e., global) control is suggested, with the degree of confidence depending on the reliability and resolution of the facies interpretation, the precision of the biostratigraphic correlation, and the extent of such correlation. Examples of this procedure are described by McKerrow (1979; Ordovician and Silurian), Lenz (1982; Ordovician to Devonian), M. E. Johnson *et al.* (1985; Silurian), J. G. Johnson *et al.* (1985; Devonian), Ramsbottom (1979; Carboniferous), Ross and Ross (1985; Carboniferous), Hallam (1981, 1988; Jurassic), and Hancock and Kauffman (1979; Cretaceous). Longer term eustasy and differential vertical movements of the continents (>10 m.y.), as indicated by the degree of continental flooding and the elevation of ancient shorelines, have been discussed by Bond (1978a,b, 1979), Harrison *et al.* (1981, 1983), and Sahagian (1987) and are summarized in Chapter 8 of this volume by C. G. A. Harrison.

Water depth and shoreline location are sensitive over a wide range of time scales to the rates of tectonic subsidence and clastic sediment supply (or carbonate sediment production) as well as to eustasy. Even if depositional base level is modulated by eustasy, a given eustatic event may not be evident in the facies that are preserved, and times of maximum or minimum water depth or transgression/regression may not be precisely synchronous at different localities (e.g., Parkinson and Summerhayes, 1985).

For example, the time of maximum water depth in a subsiding basin starved of sediment is clearly not the time of maximum transgression because of the time lag between the onset of subsequent regression and the arrival of significant amounts of sediment in the deep basin. Where sedimentation generally keeps pace with subsidence, and sea-level changes are oscillatory, maximum transgression corresponds with times at which the rate of eustatic rise is fastest; where the sediment supply is low, transgression may occur even during a eustatic fall, providing that the rate of net subsidence exceeds the rate of eustatic fall (Pitman and Golovchenko, 1983). Analyzing the effects of sea-level change on shoreline position in differentially subsiding passive continental margins, Pitman (1978) and Pitman and Golovchenko (1983) argued that under conditions of falling sea level and constant shelf gradient, a phase lag exists between the time of an instantaneous change in the rate of sea-level fall and the time at which the shoreline reaches a new equilibrium position. According to their analysis, lags are on the order of millions of years for a typical shelf, with longest phase lags corresponding to broad shelves of steep gradient and to ones characterized by slow subsidence.

In view of these considerations, the degree of synchroneity evident in the conventional stratigraphic record of short-term sea-level change is remarkable. Such synchroneity reflects a bias toward data from shallow-water continental platforms of low gradient. It also suggests that sea level varies continuously at a broad range of frequencies rather than episodically and that the equilibrium shoreline positions inherent in the model of Pitman (1978) and Pitman and Golovchenko (1983) are rarely attained. Moreover, the assumption that the point of zero sedimentation rate either coincides with the shoreline (Pitman and Golovchenko, 1983) or lies at a fixed distance landward from it (Pitman, 1978) is unrealistic. The estimates of lag times are therefore hard to evaluate.

Seismic Stratigraphic Record of Sea-Level Change

Seismic stratigraphy is an approach to the investigation of sea-level fluctuations that is less sensitive than conventional stratigraphy to variations in sediment supply (Vail *et al.*, 1977, 1980, 1984; Vail and Hardenbol, 1979; Vail and Todd, 1981; Vail, 1987). Developed for interpreting seismic reflection profiles (Figure 7.1), seismic stratigraphy makes use of regional surfaces of erosion or nondeposition known as unconformities or sequence boundaries. These form during times of relative sea-level fall, when patterns of sedimentation shift abruptly, generally toward the basin (near-horizontal bold lines in Figure 7.1). Vail *et al.* (1977) suggested that many sequence boundaries are of the same age in different parts of the world and are therefore due primarily to a global process, eustasy. They also developed a technique for quantifying the amplitude of relative sea-level change from the sawtooth patterns of

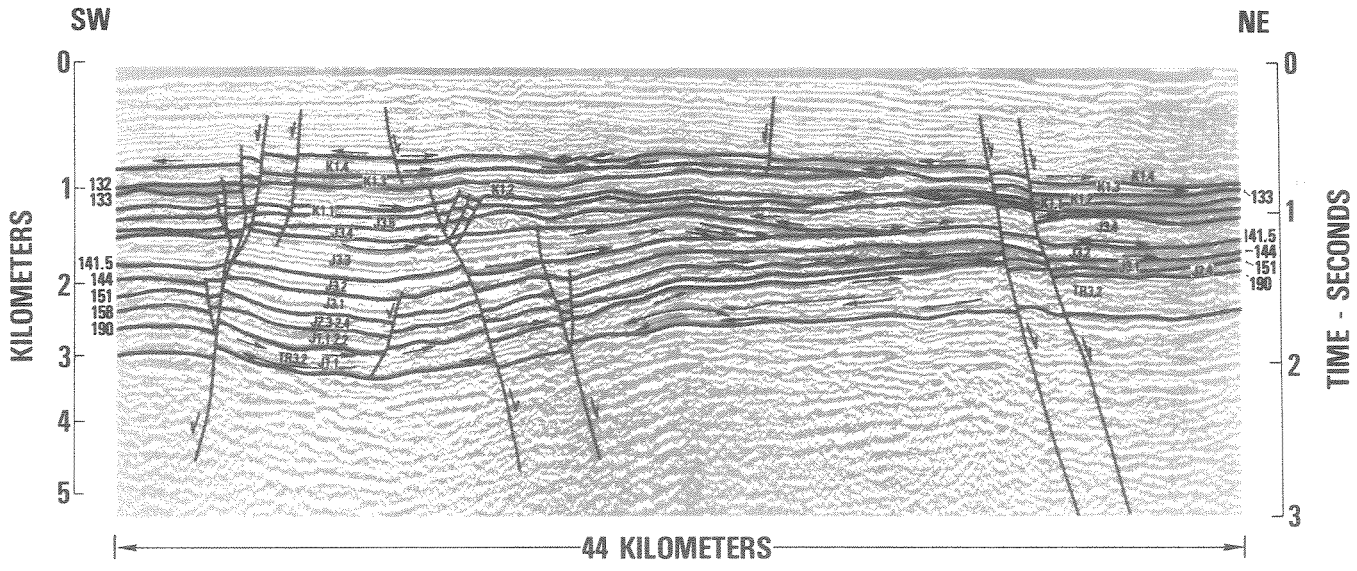


FIGURE 7.1 Seismic section northeast of Beatrice Field, Inner Moray Firth, North Sea, United Kingdom, showing interpretation of seismic sequences defined by the termination of seismic events (full arrows). Numbers with TR, J, and K prefixes identify Triassic, Jurassic, and Cretaceous sequences. Numbers on either side of the section are estimated ages of sequence boundaries in millions of years. Depth is given in kilometers and two-way travel time for seismic waves. Cross-cutting bold lines are inferred faults, with half arrows indicating apparent sense of displacement. From Vail *et al.* (1984).

onlap observed on seismic sections (see Figure 7.10). By comparing results from different basins, they obtained an estimate of global sea-level changes popularly known as the Vail sea-level curve. This led to considerable discussion as to why sea level should rise slowly and fall quickly. In subsequent publications, the original curve is termed a chart of "relative change of coastal onlap" (e.g., Vail and Todd, 1981; Vail *et al.*, 1984) and is the basis for a smoothly varying "eustatic curve," which takes into account the fact that discontinuous changes in onlap are accompanied by gradual changes in shoreline position. The most recent version, synthesizing for the first time the entire Mesozoic and Cenozoic, and incorporating the results of detailed well-log and outcrop studies, is that of Haq *et al.* (1987).

The so-called sea-level curve has been widely quoted in the geological literature and in recent textbooks (e.g., Kennett, 1982; Miall, 1984; Boggs, 1987), but apart from the obvious criticism that much of the supporting data has not yet been published, the possible limitations of the curve are not generally appreciated. In particular, the original coastal onlap chart is still commonly but incorrectly reproduced as the "sea-level curve." Recent articles by Brown and Fisher (1980), Watts (1982), Hallam (1984, 1988), Steckler (1984), Thorne and Watts (1984), Watts and Thorne (1984), Parkinson and Summerhayes (1985), Miall (1986), Summerhayes (1986), Burton *et al.* (1987), and Hubbard (1988) have raised several important issues for seismic stratigraphic interpretation, such as the origin and chronostratigraphic significance of seismic reflections, the precision with which unconformities can be identified and calibrated, the influence of tectonics in the formation of sequence boundaries, the significance and quantification of onlap (especially to derive amplitudes of eustatic fluctuations), and regional bias in the "global" curve. In the light of these criticisms, the purpose of this chapter is to evaluate the seismic stratigraphic method as a tool for investigating sea-level change. The main conclusion of this chapter is that seismic stratigraphy provides important information about the timing of sea-level fluctuations, on a time scale of millions of years, but little about magnitudes.

SEISMIC IMAGING OF STRATAL GEOMETRY

The gross geometry of unconformities and other stratal surfaces can be determined in seismic sections because these surfaces are commonly an important source of acoustic impedance contrasts. Acoustic impedance, or the product of rock density and the velocity of seismic waves traveling through the rock, determines to what extent seismic energy is reflected from a given surface. The range of velocities in common sedimentary rocks is greater than the range of densities, and velocity is consequently more important in controlling the strength of reflections. For example, lithified sandstone and shale are characterized by densities of approximately 2.32 and 2.42 g/cm^3, respectively, and by corresponding velocities of 3.6 and 4.2 km/s Nafe and Drake, 1963; Dobrin, 1976). The amplitude of a reflection from a planar contact between these two rock types is about 10 percent of that of the incident wave. Nonlithified deep-sea sediments commonly possess velocity and density contrasts more subtle than these values, and the amplitudes of reflections from the more prominent stratal boundaries in such settings are consequently much weaker, perhaps 1 to 2 percent that of the incident wave. For weak reflections, the detection of stratal boundaries is thus critically dependent on the distinction of reflected seismic energy from background noise. This is a function of geologic setting, the quality of the recording and processing system, and the experience of the interpreter.

Most reflection events seen on a seismic section are composites of reflections from individual interfaces, but a comparison between seismic sections and corresponding well-log cross sections suggests that in many cases the configuration of reflections mimics the configuration of stratal surfaces at the level of resolution permitted by the seismic data (Vail *et al.*, 1977). A common but generally incorrect expectation is that reflections correspond to the boundaries of major lithological units, even where these cut across stratal surfaces (e.g., Hallam, 1984). Most lateral changes in sedimentary facies involve such gradual variations in acoustic impedance that they are not a significant source of reflections.

SEISMIC RESOLUTION

Despite major advances in the recording and processing of seismic reflection profiles over the last two decades, basic laws of physics limit the precision with which acoustic images portray the geometry of subsurface stratal boundaries. To gauge the limits of reflection profiling in practical terms, it is helpful to understand the processes that govern vertical and horizontal resolution. Vertical resolution concerns the ability to distinguish two closely spaced reflecting surfaces, regardless of whether these are boundaries of different beds or the upper and lower surfaces of the same bed. Horizontal resolution concerns the ability to detect narrow features and then to image them in their proper location.

Vertical Resolution

The limits to vertical resolution, thoroughly discussed by Widess (1973), Neidell and Poggiagliolmi (1977), Sheriff (1977, 1985), and Mahradi (1983), can be illustrated by considering reflections from a thin tabular layer

of shale enclosed in a thick sandstone layer of lower impedance. Assuming the density and velocity of shale given above, a wave of frequency 20 Hz passing through the shale has a wavelength of 210 m. Experiments and synthetic models have shown that for layers as thin as one-quarter wavelength (about 50 m for the shale), local peaks corresponding to reflections from the upper and lower surfaces are discernible, and the time separation between the peaks is an accurate measure of layer thickness (Widess, 1973). However, for layers thinner than 1 wavelength, addition of the two reflections results in a composite wave whose peak amplitude is sensitive to bed thickness (Neidell and Poggiagliolmi, 1977). As bed thickness is decreased, the amplitude attains a minimum value at a thickness of one-half wavelength, and it increases again to a maximum at one-quarter wavelength. For thicknesses less than one-quarter wavelength, the distinct contributions from the upper and lower surfaces cannot be identified, and the shape of the composite wave continues to change until the layer thickness is one-eighth wavelength. For still thinner layers, the wave shape stabilizes but peak amplitude decreases uniformly with decreasing layer thickness. It does not matter whether a thin layer is a single unit or a composite of numerous thin beds. As long as the aggregate thickness is less than one-eighth wavelength, the reflected energy is characterized by a nonunique composite waveform. For example, if the shale layer were 1 m thick, it would produce a reflection whose amplitude was roughly 0.5 percent that of the incident wave. This bed would produce the same reflected waveform if its thickness were 5 m or 50 cm. Only the amplitude would vary. Unfortunately, to use these amplitude variations as measures of layer thickness, a nearby contact with a layer thicker than 1 wavelength is needed for calibration. This is rarely possible.

The limit to vertical resolution generally increases with acoustic velocity and burial depth and decreases with increasing acoustic frequency. Acoustic velocity tends to increase during burial as a result of progressive compaction and cementation. Seismic energy is also attenuated by reflection from successive interfaces, so that deep reflectors are less easily distinguished from noise than shallow ones. Assuming a monotonic seismic source of frequency 20 Hz, the practical one-quarter wavelength limit to resolution cited above for shale is 50 m. For rocks with high acoustic velocity, such as limestone (about 5.0 km/s), the one-quarter wavelength limit is higher (about 65 m). Fortunately, tuned arrays of marine seismic sources in common use are of relatively broad band and generate components of many tens of hertz. To the extent that sufficient high-frequency signal is generated and recorded, the practical vertical resolution is on the order of 10 to 50 m.

Horizontal Resolution

A convenient way of understanding the concept of horizontal resolution is to think of acoustic pulses traveling in the Earth as spherical wave fronts. Each impedance change encountered by a downgoing pulse acts as a point-source reflector that returns energy as a similar three-dimensional wave front. It is therefore possible to receive an echo off a buried feature at a considerable horizontal distance from the feature. In marine seismology, streamers consisting of hydrophones summed together in a linear array minimize engine noise and reflections returned from features directly ahead of or behind the streamer, but they do not eliminate side echoes.

The problem of the resolution of size and geometry can be explained by reference to Fresnel zones. The pulse from a seismic source is typically several cycles long, and the reflections originating from early and late portions of the pulse therefore interfere with each other. Early arrivals add constructively, but later arrivals are increasingly out of phase and approach total cancellation when the time lag corresponds to one-half wavelength. Arrivals at greater time delays alternately add and subtract, but their contribution is relatively small. Because the seismic pulse is a three-dimensional spherical wave front, the initial part of a pulse that reflects constructively from a planar surface has a measurable cross-sectional area called the first Fresnel zone (Sheriff, 1977).

As a result of this phenomenon, the seismic pulse can be regarded as having a "footprint" whose size depends on (1) the fundamental frequency of the source, which determines the critical one-half wavelength distance, and (2) the depth to the reflector, which determines the radius of the wave front, and hence the radius of the first Fresnel zone. For a seismic source centered at 20 Hz, and a stratigraphic section with an average acoustic velocity of 2.25 km/s, a reflection from a depth of 4 s (two-way travel time) has a diameter of about 500 m (Sheriff, 1985). This simple example shows how objects not directly beneath the seismic streamer can reflect considerable energy. Because the width of the footprint depends on the pulse frequency, sources of low frequency tend to detect reflectors out of the plane of the seismic section more readily than those of high frequency.

The radius of the first Fresnel zone also governs the width of subbottom reflectors that can be accurately detected. Synthetic models have shown that strong returns can be expected from features less than one Fresnel zone wide, but little information is preserved about the size or shape of such bodies (Sheriff, 1985). All objects appreciably narrower than one Fresnel zone have nearly identical reflections.

RECOGNITION OF UNCONFORMITIES IN SEISMIC SECTIONS

Unconformities are recognized in seismic sections by the oblique termination of seismic events (see Figure 7.1). Events that terminate against an underlying boundary indicate stratal onlap or downlap (Figure 7.2). Onlap involves updip terminations, unless the stratigraphic section has been subsequently tilted; downlap is generally downdip. Events terminating against an overlying boundary indicate toplap or erosional truncation. Toplap arises from sediment bypassing across the top of a prograding sedimentary wedge, whereas erosional truncation involves the removal of previously deposited sediment. These stratigraphic relations commonly change laterally along a boundary, and seismic events locally parallel unconformities, especially at correlative conformities, where there is effectively no depositional hiatus. A reflection is generated by the unconformity itself only if a significant impedance contrast is present. Such reflections may be discontinuous, especially where discordance of overlying and/or underlying reflectors leads to lateral phase changes (Vail *et al.*, 1977, 1980; Vail and Todd, 1981).

Unconformities possessing the fundamental geometrical properties of onlap, downlap, toplap, and erosional truncation vary considerably in lateral extent, from hundreds of meters to hundreds or thousands of kilometers, and they separate depositional sequences ranging in thickness from meters to thousands of meters. Observations in outcrop indicate that large-scale sequences commonly contain other sequences of smaller scale or "higher order," although there is probably a continuum of possible scales (Ryer, 1983; Busch and Rollins, 1984). Owing to the restrictions of resolution described above, not all instances of stratal termination against a given unconformity are imaged in seismic data, so that reflections may locally appear concordant even where the corresponding strata are slightly discordant. In addition, apparent seismic toplap and downlap can arise where clinoforms merge into shelf or basin deposits that are too thin to be resolved acoustically (see Figure 8 of Tucholke, 1981). The number of unconformities identified in a seismic section is therefore limited by vertical seismic resolution, and in general many more unconformities are present in a sedimentary basin than can be confidently traced with seismic data. It is usually not possible with seismic data to resolve a given unconformity over the entire area in which it is present.

These limitations are overcome in practice by selection of unconformities on the basis of their lateral persistence and by calibration against the rock record. Although vertical seismic resolution is typically on the order of tens of meters, seismic sections acquired during exploration for petroleum are routinely and reliably tied to borehole or well data by means of synthetic seismograms derived from velocity and density logs, or by using vertical seismic profiles (Sheriff, 1977; Badley, 1985). Where biostratigraphically resolvable or associated with a distinct change in facies or stratal dip, it may be possible to locate a given

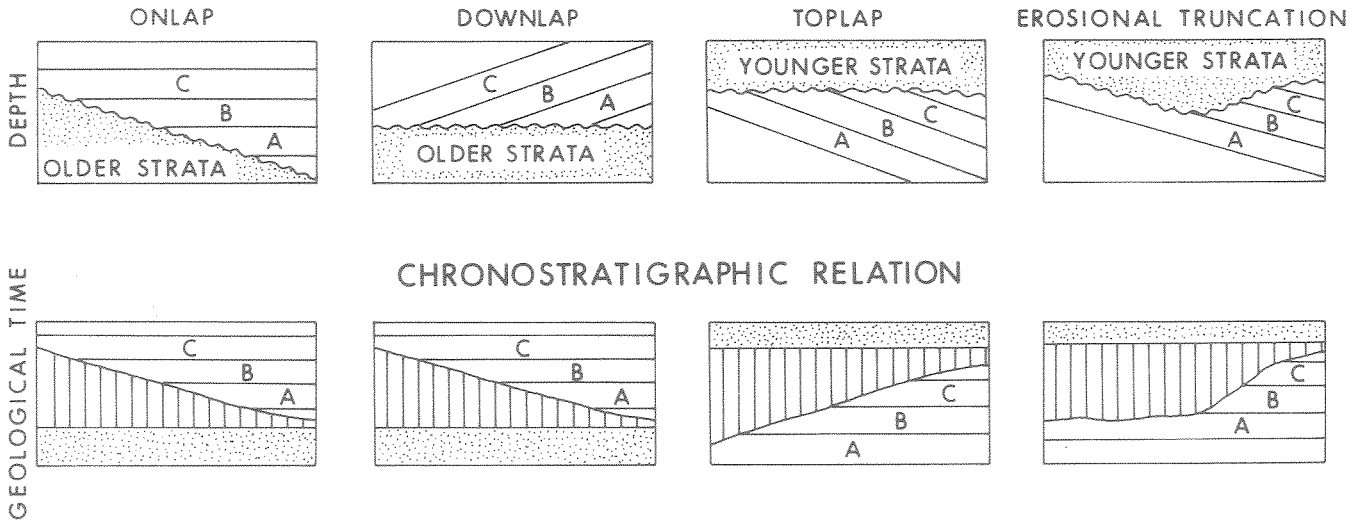


FIGURE 7.2 Stratigraphic and chronostratigraphic relations of onlap, downlap, toplap, and erosional truncation. Vertical ruling indicates the duration of the hiatus represented by the unconformity. After Ramsayer (1979).

unconformity in the rock stratigraphy to at least an order of magnitude better than seismic resolution (see Vail and Todd, 1981). The most common error in seismic stratigraphy arises where such well control is lacking, and an unconformity reflection is traced laterally into a prominent reflection that actually onlaps the unconformity (Vail *et al.*, 1980).

In seismic stratigraphic work, unconformities are delineated primarily by a downward (basinward) shift in the position of coastal onlap. This is not because boundaries with such geometry are necessarily more important than boundaries lacking such a shift in onlap, but for the most part because such boundaries can be traced with the greatest confidence. Indeed, for seismic interpretation, Vail *et al.* (1984) restricted the term unconformity to those surfaces involving local erosional truncation or subaerial exposure, phenomena commonly associated with a downward shift in onlap. According to this usage, marine surfaces resulting from deep-water sediment starvation and/or dissolution, and representing significant hiatuses but lacking evidence for erosion, are not unconformities. The delineation of unconformities on the basis of onlap and downlap is especially appropriate where "follow cycles" are present. A follow cycle is commonly associated with a strong reflection along an erosional surface, and consists of a second peak on the waveform beneath the principal reflection (Vail and Todd, 1981). Where the follow cycle masks underlying reflections, the unconformity may appear to be stratigraphically lower than its true position.

A downward shift in onlap would seem to be most easily resolved where (1) the gradient of the depositional surface is large (e.g., the continental slope); (2) there is significant differential subsidence; (3) the hiatus represented at a given locality is long; and (4) overall sedimentation rates are high, as in young, rapidly subsiding margins. However, as discussed below, the situation is complicated by the fact that an unconformity produced by a given eustatic event is of greatest regional extent in old, slowly subsiding margins. By considering rates of tectonic subsidence and seismic resolution, Thorne and Watts (1984) concluded that for the passive margin of the eastern United States (an "old" margin), it is unlikely that seismic stratigraphy can resolve unconformities representing a hiatus of less than 4 m.y. This may be overly pessimistic for seismic stratigraphy in general if surfaces as close as one-quarter acoustic wavelength can be resolved, and if unconformities can be traced toward the continent from areas of high subsidence rate, where the reflection terminations are most obvious. Assuming a practical limit to vertical resolution of 50 m and a relatively slow sediment accumulation rate of 2 cm/1000 yr, it should be possible to resolve a hiatus of 2.5 m.y. For broadband seismic sources, the resolution may be considerably better.

SEISMIC REFLECTIONS LACKING PRIMARY STRATIGRAPHIC SIGNIFICANCE

Although seismic sections resemble geologic cross sections, not all seismic events have primary stratigraphic significance. Examples of nonstratigraphic events are those produced by low-angle faults and diagenetic boundaries, together with such features as multiples, coherent noise, diffractions not migrated during data processing to their proper position, and sideswipe (energy returned from outside the plane of the section; see Tucker and Yorston, 1973). Low-angle normal faults and thrust faults in places geometrically resemble some stratigraphic boundaries, but they are not a significant source of confusion in most of the basins used to derive information about sea-level change. Diagenetic boundaries appear to be best developed in fine-grained marl-limestone successions, in which much of the "bedding" may of diagenetic origin, or at least significantly modified by diagenesis (Hallam, 1986; Ricken, 1986). Fortunately, such diagenetic layering mainly parallels depositional layering, and usually does not generate artificial reflection terminations that could be confused with sequence boundaries. Multiples, noise, diffractions, and sideswipe can generally be recognized by an experienced interpreter because they tend to cut across events of stratigraphic origin. Of course, the interpretation of seismic data is not always straightforward, and an example from USGS seismic line 25 of the passive margin of the eastern United States has been discussed by Thorne and Watts (1984).

Line 25 passes through DSDP Site 612 and within 11 km of the COST B-3 well (Figure 7.3). Between these two holes, well-defined reflection terminations on line 25 are present at depths of between 2.3 and 2.0 s (two-way travel time), with at least four instances of apparent onlap against a prominent reflection (event 6 in Figure 7.3) in a lateral distance of only 8 km. In spite of this geometry, no hiatus has yet been detected with biostratigraphic data in COST B-3 at the level of these terminations (Poag, 1980), and only a minor hiatus has been observed at Site 612 (upper lower to lower middle Eocene; Miller and Hart, 1987; Poag and Low, 1987). Moreover, the high amplitude of event 6 at Site 612 can be related to a postdepositional diagenetic front, which is slightly oblique to stratal surfaces, and which separates siliceous nannofossil chalk above from porcellanite-bearing nannofossil chalk below. Two interpretations are possible. (1) The boundary is not a sequence boundary, but a diagenetic front, and the apparent termination of reflections is an artifact of poor seismic resolution (Thorne and Watts, 1984). (2) The boundary is indeed a sequence boundary, but one on which a diagenetic front has been superimposed and for which the biostratigraphic evidence is inconclusive. The existence of a

prominent reflection immediately beneath and parallel to event 6 favors the second interpretation. If the boundary were diagenetic, onlapping reflections should continue across it. In addition, most of the biozonation in the COST B-3 well (for which the greatest hiatus should be observed) was based on rotary cuttings rather than cores, and a marked decrease in sediment accumulation rate in the lower and middle Eocene strata encountered in this well is consistent with a hiatus (Poag and Schlee, 1984).

CHRONOSTRATIGRAPHIC SIGNIFICANCE OF UNCONFORMITIES

An unconformity is a buried surface of erosion or nondeposition, whose principal significance for seismic stratigraphy is that it separates younger sediments or sedimentary rocks from older sediments or rocks below (Figure 7.4). The term unconformity thus refers to surfaces not only at a great range of scales but also involving hiatuses of markedly different duration, so that an unconformity at a small scale might be regarded as a conformity at a larger scale. An assumption of the technique of seismic interpretation promoted by Vail *et al.* (1977, 1984) is that a given surface has the same chronostratigraphic significance throughout a given basin; that is, although the duration of the hiatus represented may vary laterally, the sediments above are everywhere younger than the sediments below (Figure 7.4a). The possibility that an unconformity might be diachronous, in the sense that the sediments above the surface at one locality (e.g., D_1 in Figure 7.4b) might be older than those below it at a different locality (e.g., D_2 in Figure 7.4b), is generally excluded both for observational reasons and because sequence boundaries are regarded by Vail *et al.* as primarily a response to eustatic fluctuations, with rates comparable to or greater than the rate of tectonic subsidence. It is doubtful that high rates of sea-level change are necessary, but the chronostratigraphic assumption is probably a good approximation in coastal areas and on the continental shelves of passive margins. In such settings, the formation of unconformities is largely controlled by variations in the rates of subsidence and sea-level change, and during time intervals equivalent to a depositional sequence (1 to 10 m.y.), subsidence occurs at a relatively uniform rate. The assumption of chronostratigraphic significance may not be valid in the deep oceans, where unconformities are not necessarily related to variations in depositional base level (Tucholke, 1981; Tucholke and Embley, 1984), and in some tectonically active areas, where the rate of tectonic subsidence is not only spatially and temporally variable but changes diachronously within the basin. We do not believe that diachronous unconformities constitute a significant difficulty for seismic stratigraphic interpretation in most of the basins used by Vail *et al.* (1977, 1984) to derive the sea-level curve, but here we briefly discuss two examples of diachronous stratigraphic boundaries because the existence of such boundaries is not recognized by most seismic stratigraphers.

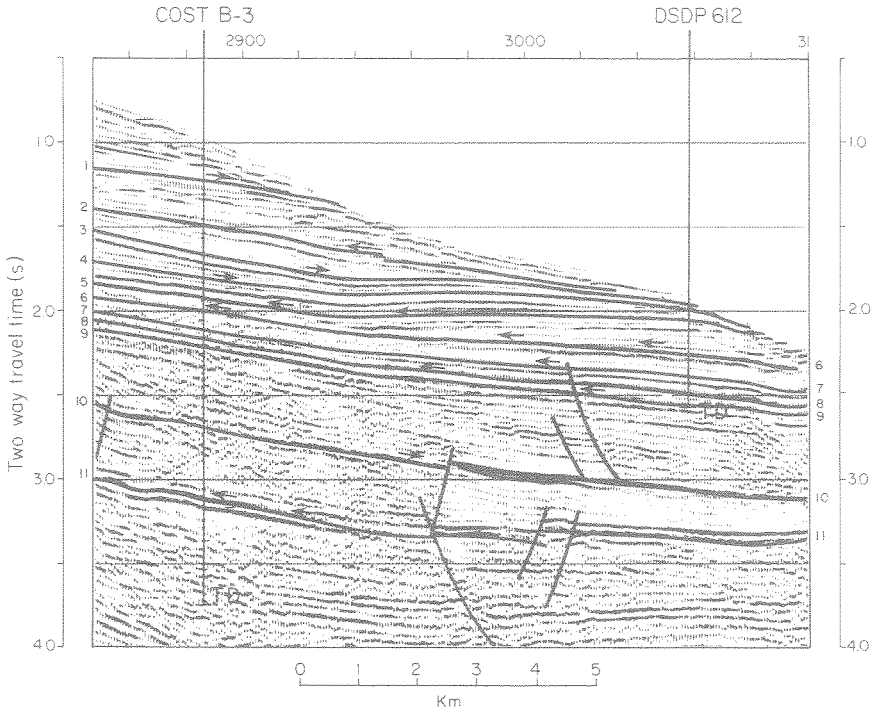


FIGURE 7.3 U.S. Geological Survey seismic line 25 between shot points 2847 and 3100. DSDP Site 612 was drilled at 3558 on line 25. COST B-3 has been projected from 11 km to the northeast to 2885. The numbered bold lines indicate tentative interpretations by A. B. Watts and N. Christie-Blick of sequence boundaries. These are based entirely on reflection terminations (arrows) observed in this single very short segment of the profile, and are subject to revision. After Thorne and Watts (1984).

Diachronous Unconformity on the Blake-Bahama Outer Ridge

Deposition and erosion in the deep ocean are controlled by many processes, including bottom-water current speed, bottom-water chemistry, sediment composition, and sediment cohesion (Tucholke and Embley, 1984). Of these, changes in bottom-water currents appear to be primarily responsible for a diachronous unconformity on the Blake-Bahama outer ridge (BBOR).

Deep-ocean currents are driven by subtle density differences related to temperature and salinity distributions, and they are capable of eroding the sea floor and transporting particles as large as silt and fine sand in suspension (Richardson *et al.*, 1981). A particularly well studied current, the western boundary undercurrent (WBUC), flows south and west along the continental rise of the eastern United States (Heezen *et al.*, 1966). The Coriolis force, which tends to turn the flow to the right in the Northern Hemisphere, is balanced by an opposing pressure gradient, and this results in a quasi-steady geostrophic current that does not necessarily flow in a straight line.

The pressure gradient in a geostrophically balanced system can be supplied by several factors, including seafloor topography. The BBOR on the continental rise off Georgia is a topographic obstruction that exerts a substantial pressure gradient on the WBUC. As the water is diverted around the ridge toward the east (to the left when looking downcurrent), the current speed increases to maintain a constant volume rate of transport, and this leads to an increase in the Coriolis force. The increased Coriolis force in turn increases the tendency for the water mass to turn to the right against the BBOR. This self-regulating system has operated in dynamic balance over the past 25 m.y., and little sedimentation has taken place beneath the core of the flow. In contrast, the region of more gentle seafloor gradient along the crest of the BBOR has not experienced a history of topographically intensified bottom currents. Depositional conditions have been maintained, and so much sediment has fallen out of suspension that the BBOR is now over 2 km high.

The basic elements of the inferred history of the BBOR are shown schematically in Figure 7.5a, which represents a view downcurrent. The BBOR is on the right, and current strength is portrayed by contours of equal speed. Sediment carried in suspension by the WBUC accumulates most rapidly in areas away from the core of the flow, while nondeposition or even erosion takes place beneath the region of greatest current strength. As sediments lap onto the base of the BBOR, the zone of nondeposition migrates upslope (Figure 7.5b). The long-term effect of this process is to produce a diachronous unconformity (Figure 7.5c). Without borehole control, incorrect age relations might be inferred because all of the strata lapping onto the base of the BBOR appear to be younger than any of the strata exposed along the erosional upper flank. In fact, they are of the same age. The error lies in assuming that the erosional surface was created at the same time everywhere.

Past contour-following geostrophic currents are likely to have been important along lower continental slopes and upper rises. In these regions, unconformities may be related both to changes in depositional base level, as in the case of the continental shelves, and to margin-parallel oceanic currents such as those of the BBOR. Some unconformities of the continental slopes may therefore be diachronous, and this possibility needs to be investigated by further seismic stratigraphic studies of sediments deposited in this setting.

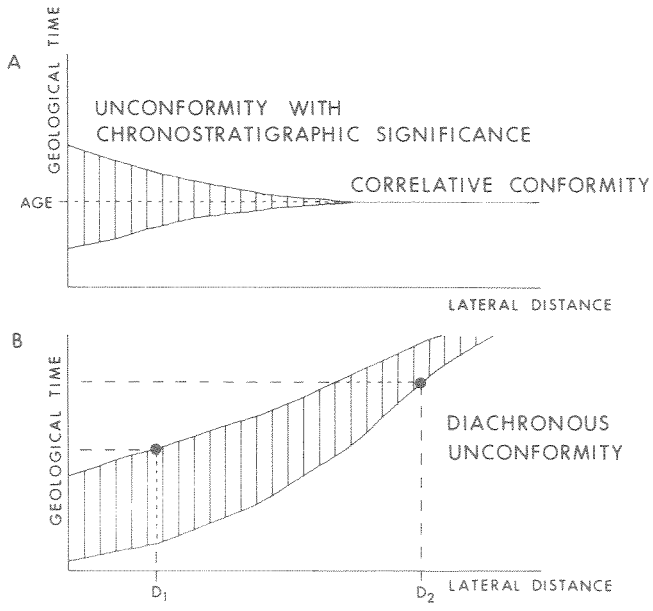


FIGURE 7.4 Chronostratigraphic cross sections of (A), an unconformity with chronostratigraphic significance; and (B), a diachronous unconformity.

Diachronous Unconformity in Alluvial-Fan Sediments Along the San Andreas Fault

An example of a diachronous unconformity in an active tectonic setting has been documented by Weldon (1984) in the vicinity of the strike-slip San Andreas Fault in southern California (Figure 7.6). The Harold Formation, Shoemaker Gravel, and Older Alluvium are Pleistocene alluvial and alluvial-fan sediments, derived from the San Gabriel Mountains southwest of the San Andreas Fault, and deposited northeast of the fault within the Mojave

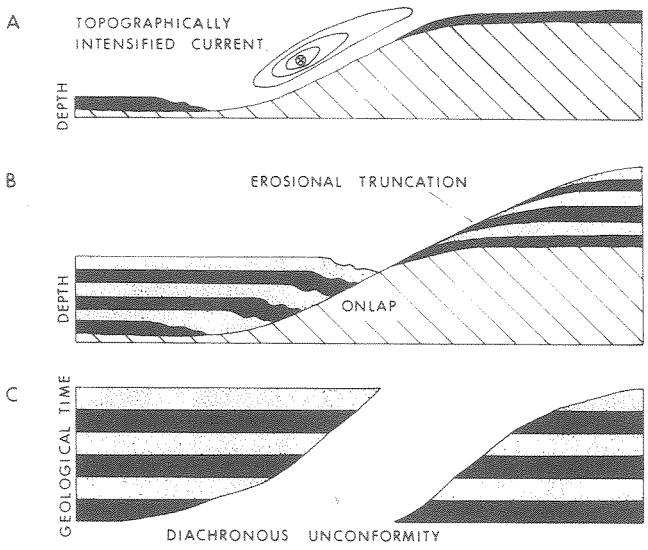


FIGURE 7.5 Development of a diachronous unconformity along the east side of the Blake-Bahama outer ridge (BBOR). (A) View downcurrent, with current strength portrayed by contours of equal speed. (B) Sediment carried in suspension by the western boundary undercurrent (WBUC) accumulates most rapidly in areas away from the core of the flow, while the region of greatest current strength is characterized by nondeposition or erosion. (C) Migration of the zone of nondeposition upslope produces a diachronous unconformity. Conventional seismic sequence analysis would incorrectly predict that the onlapping strata on the left are entirely younger than the truncated strata on the right.

Desert. Sedimentation was accompanied by uplift and tilting, with most of the deformation concentrated in the time interval represented by the angular unconformity between the Shoemaker Gravel and Older Alluvium (Miesling, 1984). Magnetostratigraphic results show that the prominent unconformity between the Shoemaker Gravel and the Older Alluvium is markedly diachronous, occurring within the Matuyama reversed interval in the vicinity of Crowder Canyon, but within the Brunhes normal interval at Phelan Peak and near Puzzle Creek, approximately 23 km northwest of Crowder Canyon. We should expect similar diachronous unconformities on a variety of scales wherever blocks are uplifted or folded, or basins subside in a diachronous fashion, as is common in strike-slip basins such as those of southern California (Christie-Blick and Biddle, 1985), and in basins where sedimentation is accompanied by the propagation of thrust faults (J. Suppe, Princeton University, personal communication, 1986).

ORIGIN OF UNCONFORMITIES

Even the most widespread unconformities are of finite areal extent because they tend to pass laterally into correlative conformities (Figure 7.4a). The formation of an unconformity can therefore be considered in terms of the expansion and subsequent burial of zones of nondeposition or erosion. At the scale of a seismic section, unconformities in shelf and coastal environments are controlled largely by two factors: (1) changes in depositional base level, an imaginary surface asymptotic to sea level, and above which significant sediment accumulation is not possible; and (2) sediment supply, or production in the case of carbonates. On a smaller scale, of course, factors such as the grain size and cohesion of available sediment, the direction and strength of currents, water depth, depth to wave base, and the geometry of the depositional surface influence the development of unconformities. These are subordinate in comparison with depositional base level and sediment supply and are ignored in the following discussion. We also focus primarily on those shelf and coastal environments in which paleoclimatic and paleoceanographic changes are relatively unimportant in the formation of unconformities.

The elevation of base level at a particular locality is a function of the rates of change of tectonic subsidence and sea level. By definition, points at base level are subject to sediment bypassing, and those above base sea level, to erosion. Expansion of the zone of bypassing is therefore promoted by a decrease in the rate of subsidence and by an increase in the rate of sea-level fall. Unconformities tend to become buried when the rate of subsidence increases or the rate of sea-level fall decreases (or sea level is rising). Nondeposition and the development of condensed intervals in deep water tend to occur at times of sea-level rise because at these times available terrigenous sediment is trapped preferentially in nearshore and coastal areas. A decrease in sediment supply or production also promotes the development of unconformities in coastal regions, as in the familiar example of the switching of delta lobes; and a minor downward shift in the position of coastal onlap can be produced by a decrease in regional sediment supply (see below). However, the lowering of depositional base level is a more effective mechanism for the depression of coastal onlap on a regional scale.

A different interpretation of marine unconformities along continental margins has been suggested by Brown and Fisher (1977, 1980). According to them, marine onlap against the continental slope is commonly initiated not by a relative sea-level fall but during times of diminished sediment supply. At these times, sediment is thought to be eroded from the shelf and redeposited in deeper water. This mechanism was invoked to explain instances of marine onlap where no evidence exists for a downward shift in coastal onlap. However, a decrease in sediment supply and corresponding increase in water depth would seem to be unfavorable for the reworking of shelf sediments. This

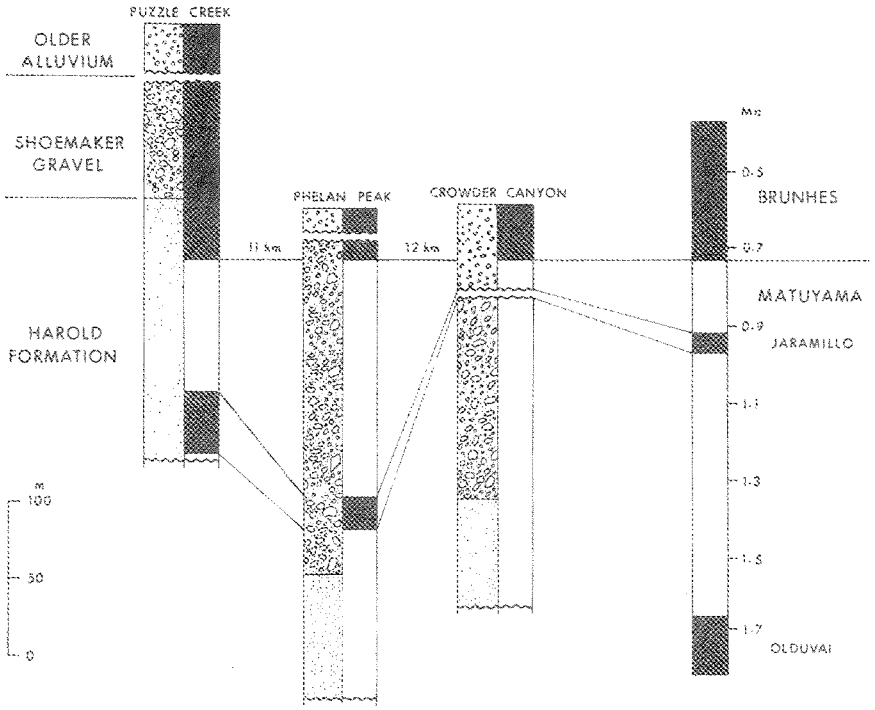


FIGURE 7.6 Magnetic stratigraphy of the Victorville Fan sediments on the northern flank of the San Gabriel Mountains, southern California. Black stripes represent chrons and subchrons of normal polarity; white stripes represent intervals of reversed polarity. All of the units and the angular unconformity between the Shoemaker Gravel and Older Alluvium are younger to the northwest (Puzzle Creek) than to the southeast (Crowder Canyon). After Weldon (1984).

is because storms and tides are less effective in transporting sediment as the shelf becomes deeper. Moreover, as discussed above, not all stratal terminations are acoustically resolved, especially in shelf deposits. Thus the mechanism proposed by Brown and Fisher (1977, 1980) may not be necessary.

Sea-Level Change and Sediment Supply

Conditions for the formation of sequence boundaries in passive continental margin settings have been considered quantitatively by Thorne and Watts (1984). The following is an elaboration of their analysis. To simplify the discussion, we first consider instantaneous changes in the rates of sediment supply and sea-level fall, and time intervals that are sufficiently long for topographic profiles to become dynamically graded. The analysis is then extended to oscillatory sea-level changes and to profiles that are not in dynamic equilibrium.

The rate of tectonic subsidence ($\dot{Y}$) of a basin can be written (modified from Steckler and Watts, 1982):

$$\dot{Y} = \Phi \dot{S}^* \left(\frac{\rho_m - \rho_s}{\rho_m - \rho_w}\right) + \dot{W}_d + \dot{\Delta}_{SL} \left(\frac{\rho_m - \rho_w (1 - \Phi)}{\rho_m - \rho_w}\right), \quad (7.1)$$

where $\dot{W}_d$ is the rate of increase in water depth; $\dot{\Delta}_{SL}$ is the rate of eustatic sea-level fall (using the sign convention of Thorne and Watts, 1984); $\dot{S}^*$ is the rate of increase in the sedimentary thickness, corrected for the effects of compaction; ρ_m, ρ_s, and ρ_w are the densities of mantle, sediment, and water; and Φ is the basement response function relating sediment and water loads to tectonic subsidence. Note that ρ_m and ρ_w are constants but that ρ_s actually changes with time as the sediment becomes lithified. If we assume Airy isostasy, $\Phi = 1$, Eq. (7.1) simplifies to

$$\dot{Y} = \dot{S}^* \left(\frac{\rho_m - \rho_s}{\rho_m - \rho_w}\right) + \dot{W}_d + \dot{\Delta}_{SL} \left(\frac{\rho_m}{\rho_m - \rho_w}\right). \quad (7.2)$$

For a point above sea level, Eq. (7.2) must be modified to correct for the sediment load above the datum ($-\dot{h}$):

$$\dot{Y} = \dot{S}^* \left(\frac{\rho_m - \rho_s}{\rho_m - \rho_w}\right) + (\dot{h} + \dot{\Delta}_{SL}) \left(\frac{\rho_m}{\rho_m - \rho_w}\right), \quad (7.3)$$

where $\dot{h}$ is the rate of decrease of elevation with respect to sea level. If the lithosphere has strength, and loads are compensated regionally rather than locally, the basement response function (Φ) in Eq. (7.1) assumes a time-dependent value that varies from less than to greater than unity, but this does not change the overall conclusions of the following discussion.

Equations (7.2) and (7.3) can be rearranged as follows:

$$\dot{\Delta}_{SL} = \dot{Y}\left(\frac{\rho_m - \rho_w}{\rho_m}\right) - \dot{S}^*\left(\frac{\rho_m - \rho_s}{\rho_m}\right) - \dot{W}_d\left(\frac{\rho_m - \rho_w}{\rho_m}\right) \quad (7.4)$$

and

$$\dot{\Delta}_{SL} = \dot{Y}\left(\frac{\rho_m - \rho_w}{\rho_m}\right) - \dot{S}^*\left(\frac{\rho_m - \rho_s}{\rho_m}\right) - \dot{h}. \quad (7.5)$$

The ratio $(\rho_m - \rho_w)/\rho_m$ is approximately equal to 0.7; and $(\rho_m - \rho_s)/\rho_m$ is approximately 0.3.

The expression of a sequence boundary for a ramp setting in a differentially subsiding basin under conditions of falling eustatic sea level ($\dot{\Delta}_{SL} > 0$) is illustrated in Figure 7.7. The point of onlap A (Figure 7.7a) lies on what is here termed a line of critical bypassing, deposition taking place basinward of A and bypassing or erosion occurring on the landward side. At A, $\dot{S}^* = 0$, and

$$\dot{\Delta}_{SL} = \dot{Y}\left(\frac{\rho_m - \rho_w}{\rho_m}\right) - \dot{h}. \quad (7.6)$$

Stratal surfaces within the accumulating wedge of sediment (shown by dashed lines) are sigmoid and approximately concordant with the upper and lower bounding surfaces (terminology from Vail *et al.*, 1977). The depositional slope changes in the vicinity of the shoreline because sedimentary processes in subaerial and shallow marine environments are different (Swift, 1970, 1976). This slope change corresponds with the depositional coastal break of P. R. Vail (Rice University, personal communication, 1987) and van Wagoner *et al.* (1987), and is typically located at depths of as much as a few meters below sea level. Here we assume for simplicity that the depositional coastal break coincides with the shoreline, although we recognize the conceptual difference between these two features. In the case of a graded topographic profile, and a stationary shoreline (L_1), $W_d = h = 0$ at all points seaward of A, and

$$\dot{\Delta}_{SL} = \dot{Y}\left(\frac{\rho_m - \rho_w}{\rho_m}\right) - \dot{S}^*\left(\frac{\rho_m - \rho_s}{\rho_m}\right). \quad (7.7)$$

For a profile that is initially graded to a fixed shoreline, a decrease in sediment supply at constant depositional base level leads to transgression of the shoreline [$\dot{W}_d > 0$; Eq. (7.4)]. This in turn results in steepening of the subaerial profile. Dynamic equilibrium is reestablished through enhanced erosion and a shift of the line of critical bypassing (A) toward the basin. A similar argument shows that an increase in sediment supply leads to regression, and a shift of A away from the basin. Thus small shifts in the position of onlap can be produced by changes in sediment input, without changing either $\dot{\Delta}_{SL}$ or $\dot{Y}$. In this simple case, the distance between A and the shoreline is related to the slope of the depositional surface, the lateral gradient of Y, the magnitude of $\dot{\Delta}_{SL}$, the abundance of available sediment, and the time since the system was last perturbed.

A small increase in the rate of sea-level fall results in a downward shift in the position of onlap to B, where the rate of tectonic subsidence ($\dot{Y}$) is greater than at A (Figure 7.7b). The shift is geologically instantaneous because all points landward of B are now subject to erosion ($\dot{S}^* < 0$), or at least to relative uplift ($\dot{h} < 0$), which favors erosion [Eq. (7.5)]. The downward shift in onlap is not a response to a fall in sea level, but to an increase in the rate of fall in sea level. Hence the distance between A and B, however measured, by itself provides no information about the amplitude of sea-level change, contrary to the methodology originally proposed by Vail *et al.* (1977) for measuring sea-level falls from seismic stratigraphy. It is also clear from Eq. (7.6) that prominent shifts in onlap can result from modest changes in the rate of sea-level fall, comparable to typical rates of tectonic subsidence (<1 cm/1000 yr). The increase in Δ_{SL} and increased sediment supply (from erosion) together cause regression or regression at an increased rate, but slower than the rate of change of the position of onlap. A new equilibrium shoreline position (L_2) is reached when the rate of sedimentation at the shoreline is just sufficient to balance the rate of net subsidence [Eq. (7.7)]. Erosion landward of B does not necessarily produce any marked (acoustically resolvable) discordance, because the surface is initially concordant with the underlying strata, but erosion does result in an apparent shift in the position of A to the erosional edge, A′, reducing the coastal encroachment (the horizontal component of coastal onlap; Vail *et al.*, 1977) apparent in the underlying sequence. The line of critical bypassing (corresponding to points A and B) and the shoreline are thus related but different entities, which move laterally at different rates, and not necessarily in the same direction (see Vail *et al.*, 1977, 1984; Vail and Todd, 1981). We emphasize this important conceptual difference between onlap shifts and transgressions/regressions because it is commonly blurred in the literature (e.g., Pitman and Golovchenko, 1983; Parkinson and Summerhayes, 1985; Summerhayes, 1986).

Figure 7.7c shows the effect of a more rapid increase in the rate of sea-level fall, where the new equilibrium line of bypassing (C) occupies a position that was formerly marine not subaerial (cf. Figure 7.7a). In this case, the line of bypassing shifts instantly to the shoreline (L_1) because all points between A and L_1 are immediately subject to relative uplift and erosion [Eq. (7.5)]. Bypassing then extends rapidly toward C, at a rate that depends on the slope of the depositional surface, the relative magnitudes of Y and Δ_{SL}, and the sediment supply. The point C again corresponds

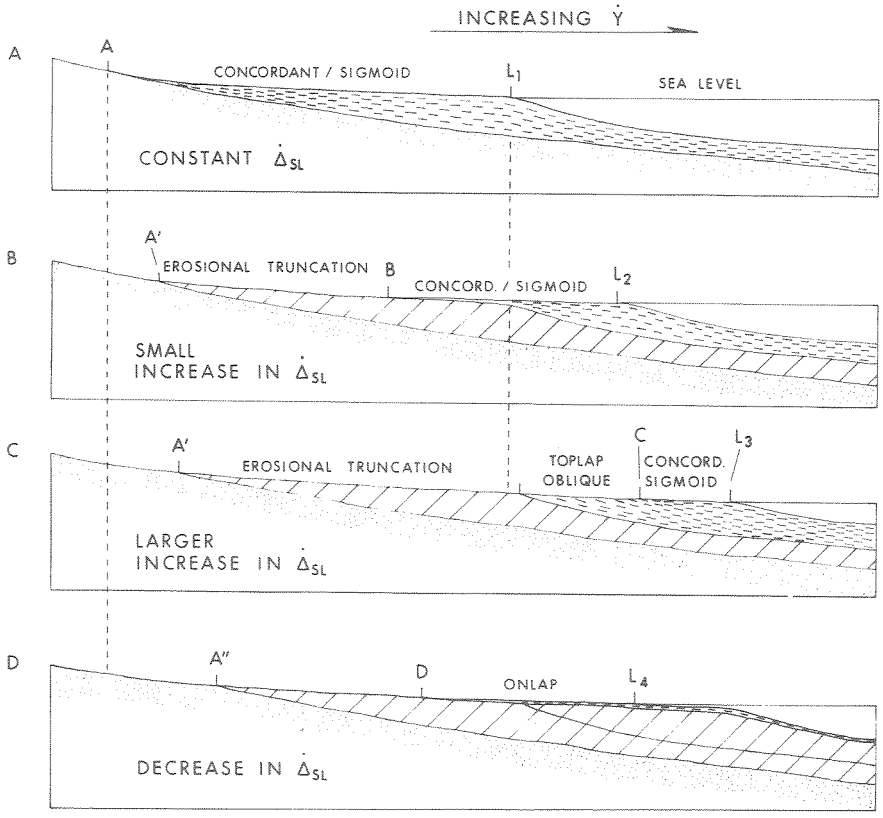
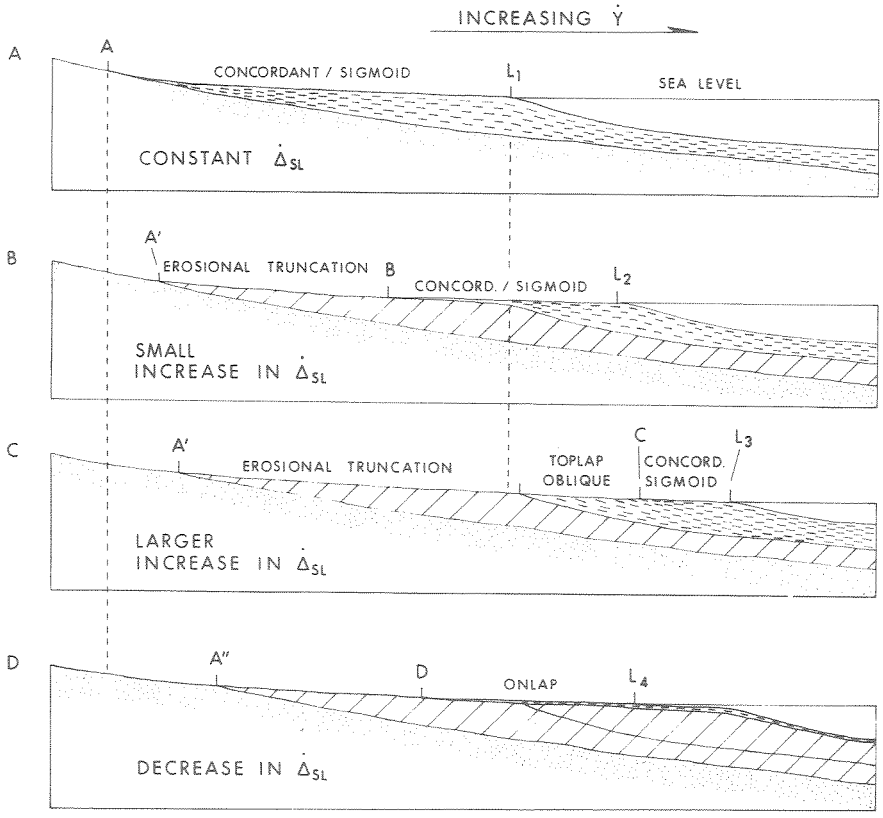


FIGURE 7.7 Formation of an unconformity (sequence boundary) under conditions of falling sea level ($\Delta_{SL} > 0$), for a ramp setting in a differentially subsiding basin, in which the rate of tectonic subsidence (Y) increases to the right. In each panel, the line of critical bypassing (corresponding to points A, B, C, and D) separates areas of deposition from areas of nondeposition and erosion. Points L_1, L_2, L_3, and L_4 indicate the location of the shoreline and approximately the depositional coastal break. Dashed lines within sequences are drawn parallel to stratal surfaces, and diagrammatically indicate concordant and discordant relations with bounding unconformities. When the rate of sea-level fall is constant [see (A)], sediments onlap at the line of critical bypassing; the stratal geometry is sigmoid, and strata are approximately concordant with upper and lower bounding surfaces. An increase in the rate of sea-level fall [see (B) and (C)] produces a downward shift in onlap whose magnitude is determined by the rate of fall and the rate of tectonic subsidence. A small increase in the rate of sea-level fall is shown in (B) (type 2 sequence boundary), and a larger increase, in (C) (type 1 sequence boundary). A decrease in the rate of sea-level fall leads to progressive onlap away from the basin [see (D)]. See text for additional explanation.

to $S^* = 0$ [Eq. (7.6)], and bypassing between L_1 and C leads to the development of toplap and an oblique tangential stratal pattern. Continued regression of the shoreline toward the new equilibrium position L_3 results in erosion landward of C, and deposition of sigmoid strata between C and L_3.

In Figure 7.7b,c, the sequence boundary forms at the time of increased rate of sea-level fall and corresponds to the contact between the units indicated by hatched and dashed-line patterns. The case shown in Figure 7.7b is equivalent to a type 2 sequence boundary, overlain by a shelf-margin systems tract (van Wagoner *et al.*, 1987). The case shown in Figure 7.7c, in which the line of critical bypassing extends beyond and below the initial depositional coastal break (approximately L_1), is equivalent to a type 1 sequence boundary overlain by a lowstand systems tract (van Wagoner *et al.*, 1987).

The depositional coastal break was formerly termed the shelf edge (Vail and Todd, 1981; Vail *et al.*, 1984), and this led to confusion with the shelf break, the familiar physiographic boundary between the shelf and slope of modern passive continental margins (van Wagoner *et al.*, 1987). Seismic stratigraphic interpretations involving lowering of coastal onlap below the shelf edge during times of minimal continental glaciation were questioned by Pitman and Golovchenko (1983) and by Thorne and Watts (1984) because the rate of tectonic subsidence at the shelf break commonly exceeds the rate of sea-level fall that can be sustained by nonglacial mechanisms for sea-level change. This ceases to be a problem for most ramp settings, such as that illustrated in Figure 7.7, in which the depositional coastal break is at a considerable distance inboard of the shelf break. There may or may not be a problem in the case of continental margins at which the depositional coastal break and shelf break coincide, depending on the rate of tectonic subsidence, and on whether observed onlap below the edge of the shelf is truly coastal or is marine onlap. Pronounced shelf bypassing and submarine-fan sedimentation, explicitly associated with type 1 unconformities, do not require complete exposure of the shelf (May *et al.*, 1983). Significant bypassing probably takes place as soon as the line of critical bypassing intersects the upper parts of submarine canyons, through which sediments can be efficiently transported to the deep marine environment. The shelf may also become subaerially exposed without the line of critical bypassing reaching the



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shelf break. The current Mississippi delta stands close to the shelf break of the Gulf of Mexico in spite of the Holocene sea-level rise as a result of the extremely high rate of sediment supply. However, in this case progradation was accompanied by significant aggradation, and the line of critical bypassing lies well landward of the shelf break.

Thorne and Watts (1984) concluded that during nonglacial intervals, variations in the rate of sea-level fall are insufficient to produce any unconformities unless the rate of tectonic subsidence is small, as in old passive margins. However, their analysis refers to the outer parts of margins, where the rate of tectonic subsidence is generally large. We have shown above that unless there is a synchronous increase in the rate of tectonic subsidence of appropriate magnitude, an increase in the rate of sea-level fall always produces an unconformity in a passive margin, but the unconformity may not extend very far onto the continental shelf (Figure 7.7b).

After a graded profile has been established (Figure 7.7d), a decrease in the rate of sea-level fall leads to progressive onlap away from the basin (to point D), and to transgression of the shoreline (to L_4). The rates of onlap and transgression may differ because the position of the shoreline is influenced by the rate of sediment supply.

In the discussion above, we have considered only the case of episodic changes in the long-term rate of sea-level fall, and rates of sea-level change no greater than typical rates of tectonic subsidence in passive margins. Quantitative modeling by Pitman (1978), in Figure 4 of Steckler (1984), and by Watts and Thorne (1984) indicates that the gross features of Cenozoic stratigraphy of the Atlantic margin of the United States can be satisfactorily explained by assuming such long-term sea-level changes rather than the higher-amplitude oscillatory rises and falls depicted by Vail *et al.* (1977, 1984) and Haq *et al.* (1987; see Figure 7.8). The efficacy of small sea-level changes in producing sequence boundaries is undeniable. Yet the question of whether Cenozoic sea level was oscillatory or monotonically decreasing prior to the onset of glacioeustasy is not resolved by the modeling studies (e.g., Watts and Thorne, 1984), which did not account for the magnitude of short-term changes in paleobathymetry and sediment supply [Eq. (7.4)]. Sequence boundaries of probable eustatic origin are known not only from times of long-term sea-level fall (such as the Cenozoic) but also from times of long-term rise (such as the Early Cretaceous; Bond, 1979; Watts and Steckler, 1979; Haq *et al.*, 1987). A sequence boundary of early Aptian age is cited by Haq *et al.* (1987) as particularly prominent. Cyclical changes in paleobathymetry and shoreline position can occur during sea-level rise, but prominent downward shifts in coastal onlap require falling sea level ($\dot{\Delta}_{SL} > 0$) or uplift [$\dot{Y} < 0$, Eq. (7.6)]. To the extent that Early Cretaceous downward shifts in onlap were of eustatic origin, eustatic sea-level fluctuations during that interval must have been oscillatory with the rate of short-term fall at times exceeding the rate of long-term rise. This leads to the conclusion that sea-level changes may generally be oscillatory, although of small amplitude.

The discussion is now extended to consider the effects of oscillatory sea-level change and profiles that do not attain dynamic equilibrium. Under these conditions, sequence boundaries correspond approximately to times of fastest sea-level fall (see Figure 7.8), and the assumption of global synchroneity is still a good approximation (Vail *et al.*, 1984). With reference to Figure 7.7b, expansion of the zone of bypassing begins as soon as the rate of sea-level fall is greater than about 0.7 times the rate of subsidence at A [Eq. (7.6)]. After a finite interval of time, Δ_{SL} attains a maximum value, and the line of critical bypassing reaches B. A subsequent decrease in $\dot{\Delta}_{SL}$ leads to renewed onlap away from the basin. We conclude that type 2 sequence boundaries of eustatic origin should be globally synchronous for all basins connected with the open ocean. For faster rates of sea-level fall, the line of critical bypassing may intersect the depositional coastal break before the rate of sea-level fall reaches its maximum value. We therefore expect type 1 sequence boundaries to be slightly diachronous, although the level of diachroneity (<< one-quater cycle) may be close to or below biostratigraphic resolution even for second-order boundaries. We disagree with the principal conclusion of Summerhayes (1986), reiterated by Hubbard (1988), that most prominent sequence boundaries are likely to have only local significance and not be globally synchronous.

Tectonics

Although this chapter is focused on the seismic stratigraphic record of sea-level change, it is important to consider the influence of tectonics in the development of unconformities, and how the tectonic and eustatic signals may be differentiated. Schwan (1980), Bally (1982), and Watts (1982) noted the approximate correspondence of several of the supercycle (second order) boundaries of Vail *et al.* (1977, 1980, 1984), Vail and Todd (1981), and Haq *et al.* (1987) with times of major plate-boundary reorganization, related to the progressive breakup of the supercontinent Pangea. This correspondence may in part reflect regional bias in the "global" coastal onlap chart, but in part is probably the result of tectonoeustasy associated with changes in the length and rate of crustal accretion at oceanic ridges. Although the full effects of spreading-rate changes are not apparent until more than 70 m.y. after the changes have taken place (Pitman, 1978; Pitman

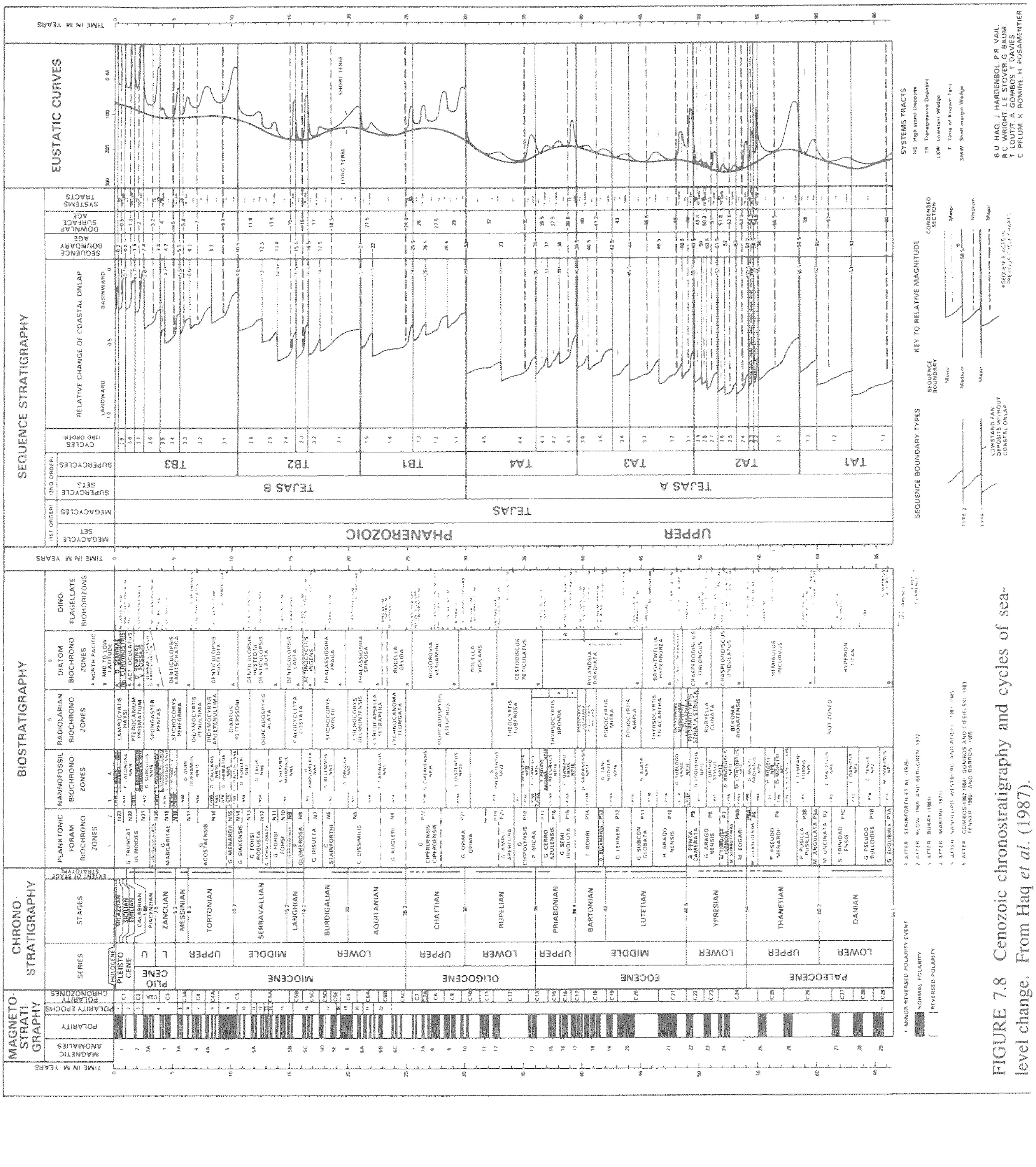

FIGURE 7.8 Cenozoic chronostratigraphy and cycles of sea-level change. From Haq *et al.* (1987).

and Golovchenko, 1983), the rate of change of sea level is affected immediately (Heller and Angevine, 1985). An alternative interpretation of apparently synchronous sequence boundaries on different continents is that they are a result of global tectonic events (Sloss, 1979; Bally, 1982). While this possibility cannot be entirely discounted, no viable mechanism has yet been established. Even if there is communication of stresses between plates, there seems to be little reason for changes in the rate of basin subsidence to be synchronous on a global scale.

Most "tectonic" mechanisms for generating sequence boundaries predict dissimilar ages from one continent to another, or even on different parts of the same continent. To the extent that major boundaries in passive continental margins are globally synchronous, these mechanisms are of secondary importance, so that the precision with which unconformities can be correlated has become a critical issue. Cloetingh *et al.* (1985), Cloetingh (1986), and Karner (1986) suggested that variations in horizontal stresses in the lithosphere of a few hundred bars can induce vertical motions of tens of meters, at rates comparable to typical rates of tectonic subsidence (1 cm/1000 yr). According to these authors, such motions may be responsible for the third-order boundaries on the coastal onlap chart (those representing base-level changes on a time scale of 1 to 3 m.y.). There appears to be good evidence that appropriate reorganization of in-plane stress takes place, but it has not yet been established that stresses vary frequently enough or that stress changes are of sufficient magnitude to produce most or even many of the observed third-order boundaries. According to the model, compressive stress accelerates basin subsidence and produces uplift of the basin margins; tensile stress retards basin subsidence and enhances subsidence of the margins. The rates of uplift and subsidence are sensitive to the magnitudes of stress variations, which vary from place to place within a given lithospheric plate, to flexural rigidity (a function of age), and to basin geometry. In particular, stratigraphically recorded vertical motions are a composite of motions at each of the density interfaces within the existing sedimentary section and beneath the basin (Karner, 1986). For these reasons, no two basins should behave in the same way, and depending on flexural rigidity, the response of a given basin may vary according to the orientation of its margins in map view. For example, in a large basin an individual unconformity might pass to a correlative unconformity not only in the direction of increasing tectonic subsidence but also parallel to the margin of the basin where its orientation changes through 90°. Indeed, several unconformities in one part of a basin might be exactly half a cycle out of phase with unconformities in an adjacent part of the same basin. Such unusual stratal geometry has never been described in published seismic stratigraphic or outcrop studies.

Cloetingh (1986) and Karner (1986) both imply that downward shifts in coastal onlap should be associated with compression, and Cloetingh (1986) even attempts to calibrate the coastal onlap chart of Vail *et al.* (1977) in terms of paleostress fluctuation. It is assumed for this purpose that the chart is weighted in favor of the North Atlantic region, which is regarded as being sufficiently small to have experienced the same stress history. Apart from the obvious hazards of such a correlation, compression does not necessarily produce a downward shift in onlap unless eustatic sea level is near stationary or rising. Under compression, the transition from increased basin subsidence to margin uplift occurs near the flexural node. We have shown above that if sea level is falling, the line of critical bypassing assumes a position within the basin. An increase in the rate of tectonic subsidence under this circumstance would lead not to a downward shift in onlap, but to renewed (or continued) onlap toward the basin margin.

Mörner (1976, 1980, 1981) argued that sea-level changes are markedly diachronous in different parts of the Earth and that they result in part from variations in the configuration of the geoid through geological time. The geoid, or equipotential surface of the Earth's gravity field, contains irregularities as great as 180 m, and undoubtedly has varied in the past. We question Mörner's interpretation of available biostratigraphic data, but more important, we think that his reasoning is flawed because he assumes that geoidal changes affect only the oceans, whereas on geological time scales they result in concomitant adjustments of the solid earth (Steckler, 1984). Thus while geoidal changes affect sea level during intervals of thousands of years, they are not relevant to longer-term eustasy.

A fundamental feature of eustatic unconformities is their global persistence in marine basins. In attempting to distinguish eustatic unconformities from those of local significance (related to tectonics, sediment supply, or paleooceanographic conditions, for example), the apparent absence of a "global" sequence boundary in a given basin is generally regarded as evidence against a eustatic origin (e.g., Thorne and Watts, 1984; Hubbard, 1988). Such a criterion should be applied with caution. Different seismic stratigraphers may subjectively select different unconformities for correlation in the same basin; many unconformities pass laterally into correlative conformities; and a given unconformity may not be seismically resolved even though known to be present (from biostratigraphic data, for example). On the other hand, unconformities of local origin in different basins may fortuitously be of approximately the same age, and thus incorrectly correlated and assumed to represent a eustatic signal. In order

to resolve these problems objectively, it is essential that the stratigraphy of every basin used for comparison is based on an internally consistent detailed interpretation of a grid of seismic sections. Few such interpretations have ever been attempted outside the petroleum industry. It is also imperative to consider the geochronological precision that may be achieved for each sequence boundary.

GEOCHRONOLOGY

Any approach to global seismic stratigraphy requires calibration to geological time through rock stratigraphy, but there are inherent uncertainties in the time scale, and in the correlation of seismic and rock sections with one another and with the time scale. The choice of an appropriate geological time scale is controversial, although every time scale involves similar components. Stratotypes of standard chronostratigraphic units (stages) are correlated with one another in order to establish a chronostratigraphic framework, and these units are then calibrated against a numerical scale. One approach is to use all available radiometric age measurements, including those obtained from "low-temperature" minerals such as glauconite, which commonly differ significantly from ages derived from "high-temperature" minerals (e.g., Odin, 1982; Haq *et al.*, 1987). Another is to correlate stratotypes with changes in geomagnetic polarity, and to calibrate this magnetochronology with the few reliable high-temperature age measurements (e.g., Heirtzler *et al.*, 1968; Berggren *et al.*, 1985; Kent and Gradstein, 1985). Once an age calibration for the chronostratigraphic framework has been chosen, other geological data such as multiple biostratigraphic zonations and geochemical fluctuations can be calibrated to the time scale.

Uncertainties in time scales are caused by errors in isotopic age measurements, and especially by problems in correlation between stratotypes and age measurements. Numerical uncertainties are partly a function of the age of the strata and the techniques employed. For example, the K-Ar technique widely used in Mesozoic and Cenozoic geochronology involves potential errors on the decay constant of less than 2 percent, and a typical accuracy of better than 5 percent (Dalrymple and Lanphere, 1965). Errors in astronomical (Milankovitch) estimates of the ages of boundaries for the latest Quaternary may be less than 5000 years (Imbrie *et al.*, 1984). Comparisons of different time scales suggest that errors are typically about 0.5 to 2 m.y. for the Neogene (cf. Odin, 1982; Berggren *et al.*, 1985), 1 to 7 m.y. for the Paleogene (cf. Odin, 1982; Berggren *et al.*, 1985), 2 to 8 m.y. for the Cretaceous, and as much as 10 m.y. for the Jurassic (Kent and Gradstein, 1985). Numerical calibration of biostratigraphic zones within these intervals is commonly less precise than biostratigraphic correlation. For example, Paleogene foraminiferal zones are typically 1 to 2 m.y. in duration, whereas radiometric precision is about 2 to 3 m.y.

The boundary between the middle and late Miocene provides a good example of correlation and time scale problems. Depending on the author, the age of this boundary varies from 9.5 to 11.5 Ma (Figure 7.9), a range of about 20 percent of the age. Direct calibration of planktonic foraminifera, nannofossils, and magnetostratigraphy has removed some of the ambiguity, and the boundary is now thought to be about 10.4 Ma (Miller *et al.*, 1985a). Differences among earlier estimates were due to miscorrelations. (1) Biostratigraphic correlation of the stratotype Tortonian (basal upper Miocene) was incorrect. Zone NN8 is found in the basal upper Miocene stratotype (Miller *et al.*, 1985a), and this requires Zone NN9 to be well within the upper Miocene (Figure 7.9, column BKV85), and not straddling the boundary as shown in columns VB74, BKD85a, and BKD85b. (2) Correlations of nannofossils and foraminifera with the geomagnetic polarity record were also incorrect. A long normal magnetozone associated with Zone NN9 (Epoch 11, Figure 7.9) was improperly correlated with the Geomagnetic Polarity Time Scale. Instead of being about 11 Ma (columns VB74, BKD85a in Figure 7.9), Zone NN9 must be younger than about 10 Ma (column BKV85).

Virtually all interregional seismic stratigraphic comparisons rely on biostratigraphic correlations. Where synthetic seismograms can be obtained from geophysical logs, or VSPs are available, the major limitations to achievable age resolution have to do with the lack of appropriately positioned boreholes, the use of cuttings rather than cores, and errors or lack of biostratigraphic resolution. Sequence boundaries are most precisely dated at correlative conformities (Figure 7.4), but such conformities may not be penetrated in drilling, or in the absence of sufficient seismic data, may even be misinterpreted as indicating that no unconformity is present. Where two or more unconformities are superimposed, a considerable hiatus may be present, and zonations in such circumstances are usually equivocal. In the case of commercial wells, most biostratigraphic work is based on cuttings, and the stratigraphic significance of such samples is reduced by downhole caving. Problems associated with biostratigraphic errors can be reduced by restricting interregional stratigraphic comparisons to sections that have been studied by the same author, but errors in identification, taxonomy, or calibration of taxa with those studied by others may be significant. Another limitation to biostratigraphic resolution involves the diachrony of taxa (between low and high latitudes, for example). First appearances of taxa are commonly diachronous (Johnson and Nigrini, 1985), and although last appearances are more likely to be syn-

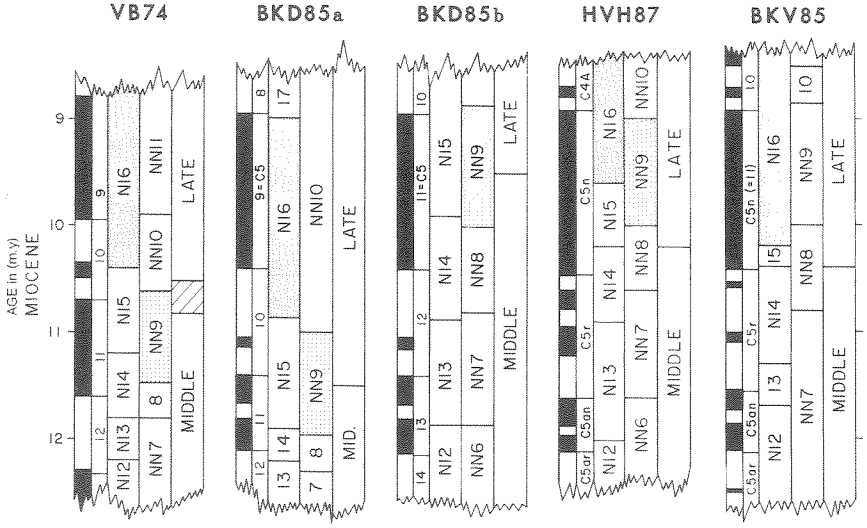

FIGURE 7.9 Comparison of time scales for middle to late Miocene time. Magnetic polarity chronologic units are given in terms of epochs (9 to 14) and chrons (C4 to C5; black, normal polarity; white, reversed polarity). Biostratigraphy: N12 to N16 are planktonic foraminiferal zones; NN7 to NN11 are nannofossil zones. Note the variable position of Zone NN9 (coarse stipple) relative to the inferred age, Zone N16 (fine stipple), and magnetic stratigraphy. Symbols for different time scales: VB74, van Couvering and Berggren (1976) based on biochronology; BKD85a, Barron *et al.* (1985), based on second-order magnetobiostratigraphic correlations and the assumption that Epoch 9 is equivalent to Chron C5n; BKD85b, Barron *et al.* (1985) assuming correlation of Epoch 11 with Chron C5n; HVH87, Haq *et al.* (1987) based upon essentially the same geomagnetic time scale as Berggren *et al.* (1985), and second-order magnetostratigraphic correlations in this time interval; BKV85, Berggren *et al.* (1985) based on first-order magnetobiostratigraphic correlations of Miller *et al.* (1985a), calibrated to the geomagnetic polarity time scale.

chronous, even these may be diachronous between sections at different latitudes (Aubry, 1983). In order to use biostratigraphy confidently, ranges of taxa must be calibrated against an independent chronology. Direct calibration to magnetostratigraphy has potential for greatly improved correlations (e.g., Berggren *et al.*, 1985; Miller *et al.*, 1985a). The precision and reproducibility of biostratigraphic picks are not readily quantifiable. Recent Neogene studies have claimed biostratigraphic resolution as good as 100,000 yr (e.g., Keller and Barron, 1983). Such claims are extravagant (see Berggren *et al.*, 1983), but properly calibrated biostratigraphic ranges have potential for Neogene correlations of better than 0.5 m.y. (Berggren *et al.*, 1985).

Our ability to test the scheme of global unconformities proposed by Vail *et al.* (1977, 1984)—that is, to distinguish between global unconformities and those developed on only a regional or local scale—is limited by our ability to determine the ages of unconformities in continental-margin successions. Biostratigraphic uncertainties of the sort outlined above are often so large that later drilling requires substantial revisions of preliminary findings. A good example is attempts to date the mid-Oligocene unconformity in the continental margin of the eastern United States. Olsson *et al.* (1980) used well cuttings to suggest that the oldest sediments above a prominent unconformity in that region are about 34 to 31 Ma, and therefore not consistent with a major erosional event predicted by Vail *et al.* (1977) at about 30 Ma. Subsequent examination of boreholes on the Irish margin, which indicated a hiatus between 34 Ma and 30 Ma, prompted a reevaluation of the biostratigraphic record from the U.S. margin (Miller *et al.*, 1985b). These workers determined that the hiatus there probably extends to at least 30 Ma, and is thus consistent with the Vail scheme.

In spite of the geochronological problems outlined here and the potential danger for circularity, in which boundaries are assumed incorrectly to be globally synchronous, many second-order sequence boundaries (such as those of the mid-Cenomanian and mid-Oligocene) appear to be very persistent from one basin to another. Although the ages of all boundaries are subject to refinement, the idea that some sequence boundaries record eustatic events remains an appealing working hypothesis. The seismic stratigraphic evidence is supported by a good deal of outcrop evidence, referred to above, for globally synchronous sea-level change through much of the Phanerozoic. It may be difficult to demonstrate a eustatic origin for many third-order sequence boundaries, those derived largely from higher-resolution well-log and outcrop studies (Figure 7.8; Haq *et al.*, 1987). This is because in spite of considerable recent efforts to calibrate the seismic stratigraphic record (Haq *et al.*, 1987), it may never be possible to resolve the ages of many of these boundaries sufficiently well for objective correlation between basins because the spacing of third-order boundaries is close to or finer than biostratigraphic resolution.

INTERPRETATION OF SEA-LEVEL CHANGE

For sequence boundaries and condensed intervals of eustatic origin (perhaps many of the second-order se-

quences), seismic stratigraphy provides information only about times of rapid sea-level fall and rise, respectively. Seismic stratigraphy provides no direct information about times of eustatic highstands and lowstands; they must be interpolated (Figure 7.8; Vail *et al.*, 1984; Haq *et al.*, 1987). Amplitudes of eustatic fluctuations cannot be inferred from seismic stratigraphic data alone because coastal aggradation (the vertical component of onlap) is primarily a result of basin subsidence, not sea-level rise; and downward shifts in onlap reflect only the rate of sea-level fall relative to the rate of basin subsidence [Eq. (7.6)]. Variations of coastal encroachment (the horizontal component of onlap) are sensitive to lateral gradients in the rate of subsidence, which are in general time dependent and not necessarily linear.

Coastal Aggradation

Basin subsidence is primarily a response to tectonic subsidence, amplified by sediment loading and modified to a limited degree by sea-level change and sediment compaction [Eq. (7.1) integrated with respect to time]. Although Vail *et al.* (1977) used coastal aggradation as a direct measure of eustatic sea-level rise (Figure 7.10), it is not an especially good approximation even of a relative sea-level rise (net subsidence plus eustasy). Four difficulties are as follows:

1. As recognized by Vail *et al.* (1984), the upper part of many sequences consists of alluvial as well as coastal plain sediments, so that the observed aggradation exceeds the relative sea-level change.

2. For divergent reflection patterns, to be expected in differentially subsiding basins, estimates of the magnitude of aggradation are critically dependent on the path taken across the seismic section, greater values being obtained in basinward locations (Miall, 1986). Incremental measurements of aggradation near the point of onlap (e.g., Vail *et al.*, 1977) lead to minimum estimates of aggradation, but do not eliminate subjectivity from the procedure.

3. Aggradation varies within a basin according to the local rate of subsidence, basin geometry, and the degree of subsequent erosion and compaction. It is not clear how any measurement on a single seismic section can be objectively regarded as the most representative for inclusion in the coastal onlap chart of the basin.

4. The accurate measurement of aggradation within a sequence requires reliable estimates of interval velocities for each of the stratal segments used.

Ideally, coastal aggradation could be corrected for subsidence, compaction, and water-depth changes (e.g., Watts and Steckler, 1979; Hardenbol *et al.*, 1981), but magnitudes of eustatic fluctuations are difficult to estimate even at a single site. This is due to uncertainties in estimating paleobathymetry, particularly where sediments accumulated in water more than 200 m deep; in determining how sediments compacted; in making corrections for sediment loading on lithosphere with finite but poorly known flexural rigidity; and in separating tectonic subsid-

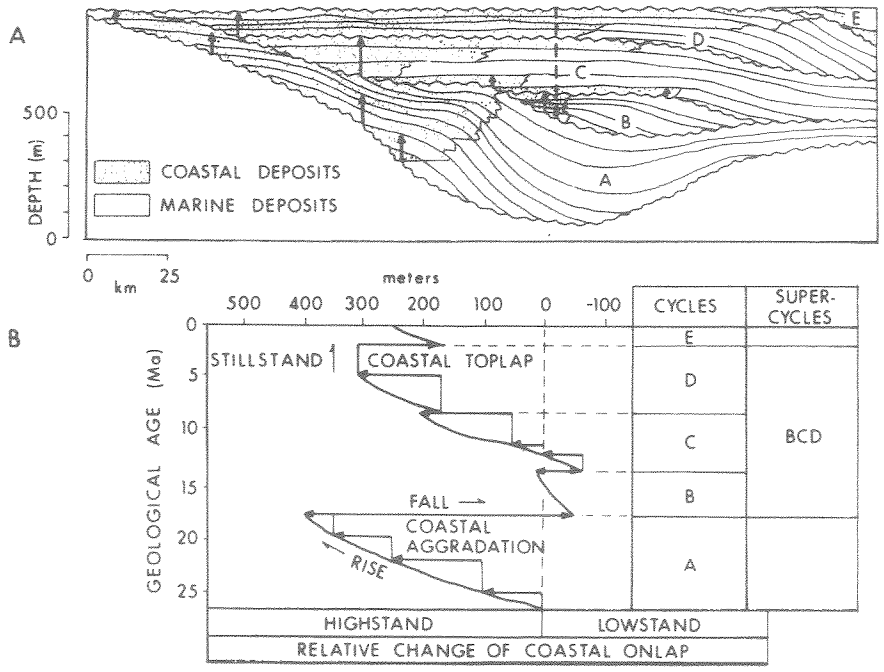

FIGURE 7.10 Procedure for constructing regional chart of relative changes of coastal onlap from estimates of coastal aggradation and downward shifts in coastal onlap—(A) stratigraphic cross section and (B) regional chart of cycles of relative change of coastal onlap. The letters A to E are arbitrary labels for five depositional cycles shown. A supercycle is a group of cycles during which there are only minor downward shifts in onlap. After Vail *et al.* (1977, 1984). See text for an evaluation of this procedure.

ence from eustasy. Once observed stratigraphic thicknesses have been corrected for water-depth changes, compaction, and sediment loading, the derived subsidence curve can be compared with a best-fit model curve for tectonic subsidence. If high-frequency deviations from smoothly varying long-term tectonic subsidence are largely of eustatic, not local tectonic, origin, the misfit between the two curves is a first approximation to the eustatic signal (e.g., Watts and Steckler, 1979; Hardenbol *et al.*, 1981). However, the estimated amplitudes of eustatic oscillations vary according to assumptions involved in selecting the best-fit model. The model curve may also be biased by long-term eustatic effects, and is not necessarily a true measure of tectonic subsidence. These problems for a single site are compounded if the stratigraphic input is coastal aggradation measured incrementally along a surface, and the errors involved are difficult to assess. For all of the reasons summarized above, the measurement of coastal aggradation is an inappropriate method for estimating eustasy.

Miall (1986) questioned the method for quantifying onlap changes for the additional reason that it appears to ignore the fact that reflections are dipping (Figure 7.10a). Though contrary to geological intuition, for unmigrated seismic data, stratigraphic thickness is most accurately measured on a vertical scale (two-way travel time multiplied by appropriate velocities, or vertical thickness in a depth section). Two-way travel time is the time required for the most direct reflection, which is perpendicular to dipping strata. As a consequence, no correction for stratal dip is required. In migrated data, subsurface points are corrected for spatial mislocation, and aggradation is more properly measured perpendicular to reflections.

Downward Shifts in Onlap

The sawtooth asymmetry of the coastal onlap chart reflects the tendency for intervals of progressive onlap to be punctuated by downward shifts in onlap that appear to be geologically rapid, but the inferred magnitudes of downward shifts have no physical meaning. In Figure 7.10, the coastal aggradation of 400 m measured in sequence A consists largely of differential subsidence during the deposition of A, but the rapid fall of 450 m between cycles A and B includes the differential subsidence during cycles B to D (broken line in Figure 7.10a), which clearly has nothing to do with the formation of the boundary at the top of sequence A. As shown above, a downward shift in onlap is a not a response to a sea-level fall but to an increase in the rate of sea-level fall. Thus even if the downward shift were somehow corrected for the effects of later subsidence and for the various sources of error described in the section above, the shift in onlap would still provide no direct information about the magnitude of sea-level change.

Coastal Encroachment

Recognizing the problems inherent in measuring coastal aggradation, Vail and colleagues have recently begun to use variations of coastal encroachment (the horizontal component of onlap) to construct coastal onlap charts (P. R. Vail, Rice University, personal communication, 1987). Horizontal distances can be measured accurately on seismic profiles, and for superposed sedimentary cycles the degree of coastal encroachment may allow a qualitative comparison of the eustatic fluctuations associated with each cycle. However, the use of coastal encroachment does not remove the ambiguity associated with the fluvial wedge in the upper part of many sequences, and the significance of coastal encroachment in cycles of markedly different age is uncertain. This is because changes of coastal encroachment are sensitive to lateral gradients in the rate of subsidence, which in general are time dependent and not necessarily linear. Identical eustatic fluctuations might produce very different variations of coastal encroachment at different stratigraphic levels within a basin. Moreover, it is unclear how variations of coastal encroachment (measured laterally) can be quantitatively "corrected" for subsidence, compaction, and water-depth changes (measured vertically), and therefore how resulting coastal onlap charts for different basins can be usefully compared to derive a eustatic signal.

Global Onlap Chart

By qualitatively comparing charts of relative change of coastal onlap for different basins, Vail *et al.* (1977, 1980, 1984), Vail and Todd (1981), and more recently Haq *et al.* (1987) derived a global onlap chart for the Mesozoic and Cenozoic (see Figure 7.8). Apart from indicating the timing of global unconformities, assuming that many of the second-order unconformities are global, the significance of such a chart is unclear. It appears to be strongly biased toward North America and Europe (see Figure 11 of Vail *et al.*, 1984), but in our view the most significant problem is that similarities in the patterns of coastal onlap for different basins for the most part indicate similar subsidence history.

Eustatic Curve

In another departure from earlier procedures, attempts have been made recently to estimate the magnitudes of eustatic falls from the degree to which coastal onlap shifts below the depositional coastal break at type 1 unconformi-

ties, making corrections for compaction, loading, and water-depth changes (Greenlee *et al.*, 1988). Some of the problems inherent in measuring downward shifts in coastal onlap are thereby avoided. However, it is critically important to demonstrate that the onlapping strata below the depositional coastal break accumulated near sea level rather than in a deeper marine environment. Only rarely can paleobathymetry at the point of onlap be established with confidence owing to the lack of cores and appropriately positioned wells or boreholes. An additional limitation of this approach is that only part of a eustatic fall is sampled because sea level is already falling before the line of critical bypassing reaches the depositional coastal break, and nearshore sediments begin to onlap the sequence boundary before the onset of the next sea-level rise. In spite of these limitations, we nevertheless expect this approach to yield improved estimates of the magnitudes of sea-level falls if procedures are refined and applied to cored boreholes in transects across continental margins.

The technique of Greenlee *et al.* (1988) does not appear to have been used systematically to derive the eustatic curve published by Haq *et al.* (1987). Indeed, their eustatic curve is strikingly similar to a smoothed global onlap curve, in which the ages of inflection points are constrained by the ages of sequence boundaries and condensed intervals (see Figure 7.8 for the Cenozoic segment). In view of the many arguments outlined above for doubting that the onlap curve is a good measure of the amplitudes of eustatic oscillations, we think that the derived eustatic curve contains information for the most part about the timing of eustatic rises and falls. Large-amplitude eustatic oscillations indicated by Haq *et al.* (as much as 100 m or more for some type 1 unconformities) appear to be inferred largely by analogy with Pleistocene sea-level changes (e.g., Vail *et al.*, 1984), and generally not on the basis of firm evidence for every boundary shown. Amplitudes of sea-level falls associated with type 2 sequence boundaries cannot be determined by Greenlee's method, and these must have been inferred from the shape of the global onlap chart.

In summary, the main limitations of the eustatic curve are (1) that all of the observed sequence boundaries (third order as well as second order) are uncritically assumed to be of eustatic origin; (2) that questions persist about the calibration of many boundaries to the geological time scale; and (3) that the inferred amplitudes of sea-level fluctuations are for the most part conjectural.

SUMMARY AND RECOMMENDATIONS

The technique of seismic stratigraphy has led to a fundamental reevaluation of our approach to stratigraphic and sedimentological studies. It provides a new way of interpreting subsurface stratigraphy, and of comparing the stratigraphic record in different basins. The identification of unconformities of apparently global extent has spawned renewed interest in eustasy and vigorous debate about the interpretation of seismic stratigraphic data.

The key to seismic stratigraphy is the identification and calibration of regional unconformities (sequence boundaries), which in most shelf and coastal areas have chronostratigraphic significance. Sequence boundaries associated with a downward shift in coastal onlap develop in response to changes in the rate of subsidence and rate of sea-level change. The distinction of these phenomena hinges largely on the precision with which individual boundaries can be dated and correlated on a regional to global scale. Amplitudes of eustatic fluctuations cannot be inferred from seismic stratigraphic data alone because coastal aggradation is primarily a result of basin subsidence, not sea-level rise; and downward shifts in onlap reflect only the rate of sea-level fall relative to the rate of basin subsidence. Variations of coastal encroachment are sensitive to lateral gradients in the rate of subsidence. Whichever method is used to gauge changes in coastal onlap, the large component of subsidence cannot be easily removed to derive the smaller eustatic signal.

For continued improvement of the seismic stratigraphic record of eustasy, we recommend objective reevaluation of the ages of sequence boundaries in individual basins, with the aim of distinguishing more confidently boundaries of global extent from those of more restricted distribution. Basins should be selected for study on the basis of stratigraphic completeness; simple tectonic history (e.g., passive continental margins lacking diapirism); and the current or future availability of high-resolution seismic sections, fully cored boreholes, and state-of-the-art geophysical logs and VSPs for calibration of the boundaries. The basins should also be of a range of ages and widely separated to avoid regional tectonic bias. Chronostratigraphic control should be obtained by integrating multiple biostratigraphic, isotopic (stable and radiometric), and magnetostratigraphic criteria. For this purpose, basins at mid-latitudes offer the most favorable trade-off between biostratigraphic and magnetostratigraphic techniques. For improved estimates of the magnitudes of sea-level change, a transect of boreholes is required for two-dimensional analysis of tectonic subsidence. The best resolution of paleobathymetry is obtained from shallow-water carbonate platforms (e.g., Kendall and Schlager, 1981).

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