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SEQUENCE STRATIGRAPHY

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INTRODUCTION

Sequence stratigraphy is the study of sediments and sedimentary rocks in terms of repetitively arranged facies and associated stratal geometry (Vail 1987; Van Wagoner et al 1988, 1990; Christie-Blick 1991). It is a technique that can be traced back to the work of Sloss et al (1949), Sloss (1950, 1963), and Wheeler (1958) on interregional unconformities of the North American craton, but it became systematized only after the advent of seismic stratigraphy, the stratigraphic interpretation of seismic reflection profiles (Vail et al 1977, 1984, 1991; Berg & Woolverton 1985; Cross & Lessenger 1988; Sloss 1988; Christie-Blick et al 1990; Van Wagoner et al 1990; Vail 1992). Sequence stratigraphy makes use of the fact that sedimentary successions are pervaded by physical discontinuities. These are present at a great range of scales and they arise in a number of quite different ways: for example, by fluvial incision and subaerial erosion (above sea level); submergence of nonmarine or shallow-marine sediments during transgression (flooding surfaces and drowning unconformities), in some cases with shoreface erosion (ravinement); shoreface erosion during regression; erosion in the marine environment as a result of storms, currents, or mass-wasting; and through condensation under conditions of diminished sediment supply (intervals of sediment starvation). The main attribute shared by virtually all of these discontinuities, independent of origin and scale, is that to a first approximation they separate older deposits from younger ones. The recognition of discontinuities is therefore useful because they allow sedimentary successions to be divided into geometrical units that have time-stratigraphic and hence genetic significance.

Precise correlation has of course long been a goal in sedimentary geology, and the emergence of sequence stratigraphy does not imply that existing techniques or data ought to be discarded. Instead, sequence stratigraphy provides a

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unifying framework in which observations of intrinsic properties such as lithology, fossil content, chemistry, magnetic remanence, and age can be compared, correlated, and perhaps reevaluated. With the possible exception of sedimentary units characterized by tabular layering over large areas and by the absence of significant facies variation (for example, some deep-oceanic sediments), it is hard to imagine attempting to interpret the stratigraphic record in any other context. We make this point because criticisms leveled at sequence stratigraphy have tended to lose sight of the essence of the technique.

In this regard, it is unfortunate that the development of sequence stratigraphy has coincided with the reemergence of the notion that in marine and marginal marine deposits sedimentary cyclicity is due primarily to eustatic change (Vail et al 1977; Haq et al 1987, 1988; Posamentier et al 1988; Sarg 1988; Dott 1992). Eustasy (global sea-level change) may in fact have modulated sedimentation during much of earth history but, as a practical technique and in spite of terminology currently in use, sequence stratigraphy does not actually require any assumptions about eustasy (Christie-Blick 1991). Indeed, one of the principal frontiers of this discipline today is the attempt to understand patterns of sediment transport and accumulation as a dynamic phenomenon governed by a great many interrelated factors.

In most cases, specific attributes of sedimentary successions (for example, the lateral extent and thickness of a sedimentary unit, the distribution of included facies, or the existence of a particular stratigraphic discontinuity) cannot be ascribed confidently to a single cause. In particular, the roles of "tectonic events," eustatic change, and variations in the supply of sediment can be partitioned only with difficulty (Officer & Drake 1985, Schlanger 1986, Burton et al 1987, Cloetingh 1988, Kendall & Lerche 1988, Galloway 1989, Cathles & Hallam 1991, Christie-Blick 1991, Reynolds et al 1991, Sloss 1991, Underhill 1991, Kendall et al 1992, Karner et al 1993, Steckler et al 1993, Driscoll et al 1995). Each of these factors operates at a broad range of time scales (cf Vail et al 1991), and none is truly independent owing to numerous feedbacks. For example, the accumulation of sediment produces a load, which in many cases significantly modifies the tectonic component of subsidence (Reynolds et al 1991, Steckler et al 1993, Driscoll & Karner 1994). The space available for sediment to accumulate is therefore not simply a function of some poorly defined combination of subsidence and eustasy (the now-popular concept of "relative sea-level change") because that space is influenced by the amount of sediment that actually accumulates.

As a result of feedbacks, there are also inherent leads and lags in the sedimentary system; these influence the timing of the sedimentary response to any particular driving signal in ways that are difficult to predict quantitatively (Jordan & Flemings 1991, Reynolds et al 1991, Steckler et al 1993). This phenomenon is particularly significant for efforts to sort out the role of eustasy (e.g. Christie-Blick 1990, Christie-Blick et al 1990, Watkins & Mountain 1990, Loutit 1992). If the phase relation between the eustatic signal and the resulting stratigraphic record varies from one place to another, then the synchrony or lack thereof of observed stratigraphic events may prove to be less useful than previously thought as a criterion for distinguishing eustasy from other controls on sedimentation. At the very least, the comparison of sites needs to take into account the other important variables.

Recognition of these inherent difficulties has led to a gradual shift in research objectives away from such goals as deriving a "sea-level curve," and toward studies designed to investigate the effects of specific factors known to have been important in governing sedimentation in a particular sedimentary basin or at a particular time in earth history. Among the most important factors are the rates and amplitudes of eustatic change, subsidence patterns in tectonically active and inactive basins, sediment flux or availability, the physiography of the depositional surface (for example, ramps vs settings with a well-developed shelf-slope break), and scale. The subdiscipline of high-resolution sequence stratigraphy has emerged in the course of this research partly in response to the need for detailed reservoir stratigraphy in mature petroleum provinces and partly because many of the interesting issues need to be addressed at an outcrop or borehole scale (meters to tens of meters) rather than at the scale of a conventional seismic reflection profile (Plint 1988, 1991; Van Wagoner et al 1990, 1991; Jacquin et al 1991; Leckie et al 1991; Mitchum & Van Wagoner 1991; Posamentier et al 1992a; Flint & Bryant 1993; García-Mondéjar & Fernández-Mendiola 1993; Johnson 1994; Posamentier & Mutti 1994). Another frontier in sequence stratigraphy is the application of sequence stratigraphic principles to the study of pre-Mesozoic successions (e.g. Sarg & Lehmann 1986; Lindsay 1987; Christie-Blick et al 1988, 1995; Grotzinger et al 1989; Sarg 1989; Ebdon et al 1990; Kerans & Nance 1991; Levy & Christie-Blick 1991; Winter & Brink 1991; Bowring & Grotzinger 1992; Holmes & Christie-Blick 1993; Lindsay et al 1993; Sonnenfeld & Cross 1993; Southgate et al 1993; Yang & Nio 1993). In spite of its roots in Paleozoic geology (Sloss et al 1949), sequence stratigraphy has been undertaken primarily in Mesozoic and Cenozoic deposits owing to the greater economic significance, more complete preservation, and amenability to precise dating of sediments and sedimentary rocks of these eras. However, applications to older successions within the past decade have provided important new perspectives about the development of individual sedimentary basins, as well as data relevant to many of the issues outlined above.

Standard concepts and the basic methodology of sequence stratigraphy are described in numerous articles, especially those by Haq et al (1987), Vail (1987), Baum & Vail (1988), Loutit et al (1988), Van Wagoner et al (1988, 1990), Posamentier et al (1988), Sarg (1988), Haq (1991), and Vail et al (1991). In this review, we have chosen to emphasize areas of disagreement or controversy, especially with respect to the origin of stratigraphic discontinuities, which we think is one of the most interesting general issues in sedimentary geology.

CHOICE OF A FRAMEWORK FOR SEQUENCE STRATIGRAPHY

The objective of sequence stratigraphy is to determine layer by layer how sedimentary successions are put together, from the smallest elements to the largest. We are thus interested in all of the physical surfaces that at different scales separate one depositional element from another, and it could be argued that disagreements about how elements are defined and combined are secondary to the overall task at hand. Indeed, different perceptions are in part a product of real differences that have emerged in the study of contrasting examples. However, it is clear that stratigraphy represents more than a series of random events. In many cases there exists a definite hierarchy in layering patterns. In choosing a framework for sequence stratigraphy, it is therefore important to select elements that are as far as possible genetically coherent and not merely utilitarian. Currently, at least three schemes are being used (Figure 1). Here we briefly make the case for the form of sequence stratigraphy that emerged from Exxon in the 1970s and 1980s (Vail et al 1977, 1984; Vail 1987; Sarg 1988; Van Wagoner et al 1988, 1990), in preference to "genetic stratigraphy" (Galloway 1989) and "allostratigraphy" (NACSN 1983, Salvador 1987, Walker 1990, Blum 1993, Mutti et al 1994).

Sequence stratigraphy and genetic stratigraphy differ primarily in the way that fundamental depositional units are defined (Figure 1). In the case of sequence stratigraphy, the *depositional sequence* is defined as a relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities (Mitchum 1977, Van Wagoner et al 1990, Christie-Blick 1991). In the most general sense, an unconformity is a buried surface of erosion or nondeposition. In the context of sequence stratigraphy, it has been restricted to those surfaces that are related (or are inferred to be related) at least locally to the lowering of depositional base level and hence to subaerial erosion or bypassing (Vail et al 1984, Van Wagoner et al 1988). According to this definition, intervals bounded by marine erosion surfaces that do not pass laterally into subaerial discontinuities are not sequences. The fundamental unit of genetic stratigraphy, the *genetic stratigraphic sequence*, is bounded by intervals of sediment starvation (Galloway 1989). These correspond approximately with times of maximum flooding and their significance is therefore quite different from that of subaerial erosion surfaces.

Both kinds of sequence are recognizable in seismic reflection and borehole data. The principal argument for adopting the genetic stratigraphic approach is utilitarian: Intervals of sediment starvation are laterally persistent and paleontologically useful. However, the boundaries of genetic stratigraphic sequences are

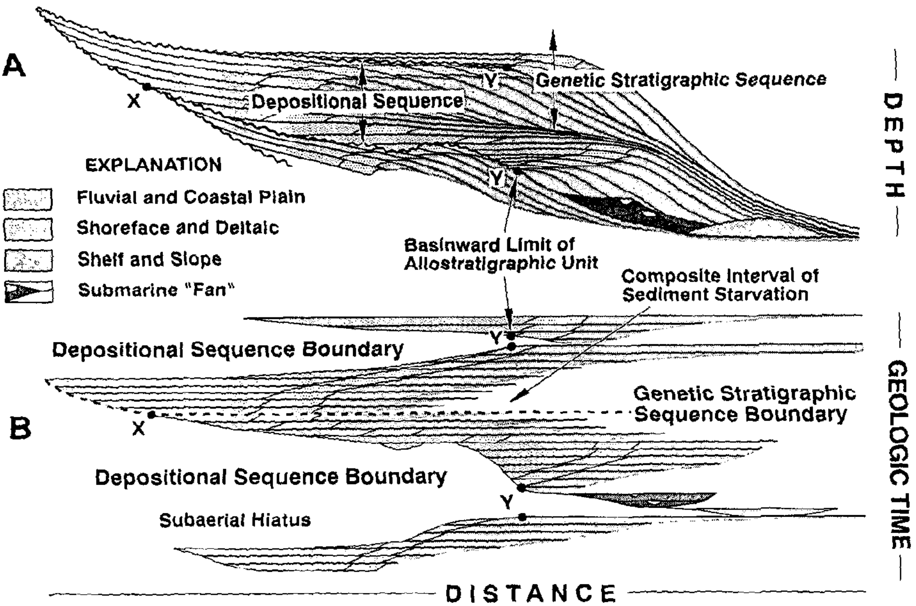


Figure 1 Conceptual cross sections in relation to depth (*A*) and geological time (*B*) showing stratal geometry, the distribution of siliciclastic facies, and competing schemes for stratigraphic subdivision in a basin with a shelf-slope break (from NACSN 1983, Galloway 1989, Vail 1987, Christie-Blick 1991, Vail et al 1991). Boundaries of depositional sequences are associated at least in places with subaerial hiatuses, and they are the primary stratigraphic discontinuities in a succession. Boundaries of genetic stratigraphic sequences are located within intervals of sediment starvation, and they tend to onlap depositional sequence boundaries toward the basin margin (point X). Allostratigraphic units are defined and identified on the basis of bounding discontinuities. Allostratigraphic nomenclature is not strictly applicable where a bounding unconformity passes laterally into a conformity or where objective evidence for a stratigraphic discontinuity is lacking (basinward of points labeled Y).

located somewhat arbitrarily within more-or-less continuous successions. In some cases, no distinctive surfaces are present. In others, intervals of starvation may contain numerous marine disconformities or hardgrounds (lithified crusts, commonly composed of carbonate). Objective identification of the maximum flooding surface is usually difficult, and so genetic stratigraphy is especially limited in high-resolution subsurface and outcrop studies. Undue focus on intervals of starvation also makes it possible to ignore the presence of prominent unconformities and to conclude (perhaps incorrectly) that sedimentary cyclicity is due primarily to variations in sediment supply (Galloway 1989), when the very existence of subaerial unconformities probably requires some additional mechanism (Christie-Blick 1991). The sequence stratigraphic approach can also be problematic: The boundaries of depositional sequences tend to be of variable character, subject to modification during transgression, and difficult to recognize once they pass laterally into fully marine successions. Yet sequence stratigraphy is preferable to genetic stratigraphy because in many settings sequence boundaries related to subaerial erosion are the primary stratigraphic discontinuities and therefore the key to stratigraphic interpretation (Figure 1*B*; Posamentier & James 1993).

Allostratigraphy differs from sequence stratigraphy and genetic stratigraphy by taking a more descriptive approach to physical stratigraphy (NACSN 1983, Salvador 1987, Walker 1990, Blum 1993, Mutti et al 1994). As formalized in the North American Stratigraphic Code, allostratigraphic units are defined and identified on the basis of bounding discontinuities; in this respect they are fundamentally similar to those of sequence stratigraphy (Figure 1). Different terminology is justified by Walker (1990) on at least two counts. Sequence stratigraphic concepts are regarded as imprecise, especially with respect to scale and the meaning of such expressions as "relatively conformable" and "genetically related." It is also argued that sequence stratigraphy is not universally applicable—for example, in unifacial or nonmarine successions or where uncertainty exists about the origin of a particular surface. Allostratigraphic nomenclature is, however, rejected here for several reasons, including historical priority. The designations of "alloformations," "allomembers," etc are unnecessary and overly formalistic; and these terms are not strictly applicable where a bounding unconformity passes laterally into a conformity (Baum & Vail 1988) or where objective evidence for a discontinuity is lacking (basinward of points labeled Y in Figure 1). As with sequence stratigraphy, allostratigraphy involves making judgments about the relative importance of discontinuities (and hence the degree of conformability or genetic relatedness), but it does not *require* an attempt to distinguish surfaces of different origin. Nor does this nongenetic terminology help much where sequence stratigraphic interpretation is admittedly difficult (for example, in deposits lacking laterally traceable discontinuities).

DEPARTURES FROM THE STANDARD MODEL

In the standard model for sequence stratigraphy (Figure 2), unconformity-bounded sequences are composed of "parasequences" and "parasequence sets," which are stratigraphic units characterized by overall upward-shoaling of depositional facies and bounded by marine flooding surfaces and their correlative surfaces (Vail 1987; Van Wagoner et al 1988, 1990). These depositional elements are themselves assembled into "systems tracts" (Brown & Fisher 1977) according to position within a sequence and the manner in which parasequences or parasequence sets are arranged or stacked (Posamentier et al 1988, Van Wagoner et al 1988). Bounding unconformities are classified as type 1 or type 2 according to such criteria as the presence or absence of incised valleys, the prominence of associated facies discontinuities, and whether or not lowstand deposits (LST in Figure 2) are present in the adjacent basin. This particular view of stratigraphy is best suited to the study of siliciclastic sedimentation at a differentially subsiding passive continental margin with a well-defined shelf-slope break, and under conditions of fluctuating sea level. As with any sedimentary model, it represents a distillation of case studies and inductive reasoning, and modifications are therefore needed for individual examples, for other depositional settings, and as concepts evolve (Posamentier & James 1993).

Parasequences

Upward-shoaling successions bounded by flooding surfaces (parasequences) are best developed in nearshore and shallow-marine settings in both siliciclastic

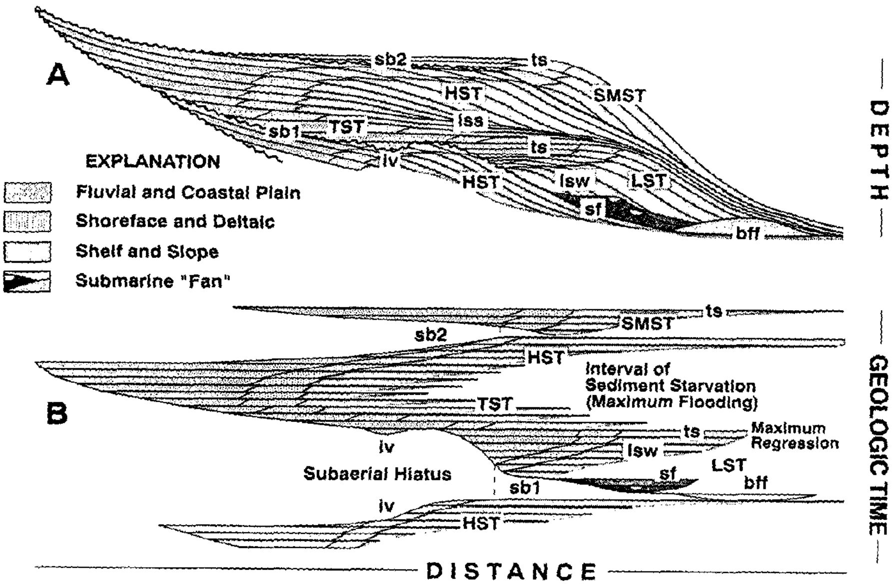


Figure 2 Conceptual cross sections in relation to depth (*A*) and geological time (*B*) showing stratal geometry, the distribution of siliciclastic facies, and standard nomenclature for unconformity-bounded depositional sequences in a basin with a shelf-slope break (modified from Vail 1987 and Vail et al 1991, specifically to include offlap). Systems tracts: SMST, shelf margin; HST, highstand; TST, transgressive; LST, lowstand. Sequence boundaries: sb2, type 2; sb1, type 1. Other abbreviations: iss, interval of sediment starvation (condensed section and maximum flooding surface of Vail 1987); ts, transgressive surface (top lowstand surface and top shelf-margin surface of Vail et al 1991); iv, incised valley; lsw, lowstand progading wedge; sf, slope fan; bff, basin floor fan. Note that in the seismic stratigraphic literature, the term submarine "fan" includes a range of turbidite systems and sediment-gravity-flow deposits that are not necessarily fan-shaped.

and carbonate rocks (James 1984; Van Wagoner et al 1988, 1990; Swift et al 1991; Pratt et al 1992), and their recognition is undoubtedly useful in sequence stratigraphic analysis. However, the tendency for pigeonholing in sequence stratigraphy tends to obscure rather than to illuminate the range of facies arrangements and processes involved. Sedimentary cycles vary from markedly asymmetrical to essentially symmetrical, with the degree of asymmetry decreasing in an offshore direction and depending also on whether parasequences are arranged in a forestepping motif (which favors asymmetry) or a backstepping one. (The term forestepping means that each parasequence in a succession represents shallower-water conditions overall than the parasequence beneath it. Backstepping refers to the opposite motif: overall deepening upward.) Although not strictly included in the definition of the term parasequence, similar sedimentary cycles with the same spectrum of asymmetry are observed in many lacustrine successions (Eugster & Hardie 1975; Steel et al 1977; Olsen 1986, 1990). Moreover, shoaling upwards is not the only expression of depositional cyclicity (for example, in alluvial and tidal deposits and deep-marine turbidites). Objective recognition of a parasequence, as opposed to some other depositional element with sharp boundaries, is therefore tenuous in nonmarine and offshore marine settings unless bounding surfaces can be shown to correlate with marine flooding surfaces.

The concept of parasequences and parasequence sets as the building blocks of depositional sequences is also largely a matter of convention rather than a statement about how sediments accumulate at different scales. There is clear overlap in the length scales and time scales represented by parasequences and high-order sequences (Van Wagoner et al 1990, 1991; Kerans & Nance 1991; Mitchum & Van Wagoner 1991; Vail et al 1991; Posamentier & Chamberlain 1993; Posamentier & James 1993; Sonnenfeld & Cross 1993; Christie-Blick et al 1995). The distinction between sequences and parasequences is therefore primarily a function of whether, at a particular scale of cyclicity, sequence boundaries can or cannot be objectively identified and mapped. An unfortunate by-product of the quest for sequences and sequence boundaries is to impose sequence nomenclature when parasequence terminology would be more appropriate. Flooding surfaces are sometimes interpreted as sequence boundaries when no objective evidence for such a boundary exists (e.g. Lindsay 1987, Prave 1991, Lindsay et al 1993, Montañez & Osleger 1993) or, if a sequence boundary is present, it is located at a lower stratigraphic level.

Systems Tracts

The threefold subdivision of sequences into systems tracts is based on the phase lag between transgressive-regressive cycles and the development of corresponding sequence boundaries (Figure 2*B*). In the standard model for siliciclastic sedimentation with a shelf-slope break, regression of the shoreline continues after the development of the sequence boundary, so that the regressive part of any cycle of sedimentation is divisible into two systems tracts: the highstand below the boundary (HST in Figure 2) and the lowstand (or shelf margin) systems tract above it (LST and SMST in Figure 2). The designation of systems tracts has become standard procedure in sequence stratigraphy as if this were an end in itself but, as with parasequences, subjective interpretation and pigeonholing tend to obscure the natural variability in sedimentary systems. There is no requirement for individual systems tracts to have any particular thickness or geometry, or even to be represented on a particular profile or in a particular part of a basin (Posamentier & James 1993). It is common in deep water for lowstand units to be stacked, with intervening transgressive and highstand units represented by relatively thin intervals of sediment starvation. In shelf areas, the stratigraphy tends to be composed primarily of alternating transgressive and highstand units, but these vary greatly in thickness and they are not necessarily easy to distinguish. Still farther landward, highstands may be stacked with no other systems tracts intervening, a situation that is likely to challenge those intent on assigning systems tract nomenclature in nonmarine successions.

The most troublesome systems tract is the lowstand, which according to the original definition of the term represents sedimentation above a sequence boundary prior to the onset of renewed regional transgression (Figure 2). It is characterized by remarkably varied facies and in many cases by a complex internal stratigraphy that in deep-marine examples continues to be the subject of vigorous debate (Weimer 1989, Normark et al 1993). The lowstand is also the one element of a depositional sequence that separates it from a transgressive-regressive cycle (e.g. Johnson et al 1985, Embry 1988). It is perhaps natural that sequence stratigraphers have attempted to identify lowstand units, even where the presence of such deposits is doubtful (for example, many shelf and ramp settings), because this helps to deflect the criticism that sequence stratigraphy offers nothing more than new terminology for long-established concepts. In shelf and ramp settings, the term lowstand is routinely applied to any coarse-grained and/or nonmarine deposit directly overlying a sequence boundary, especially where such deposits are restricted to an incised valley (e.g. Baum & Vail 1988, Van Wagoner et al 1991, Southgate et al 1993). However, sedimentological and paleontological evidence for estuarine sedimentation within (fluvially incised) valleys (e.g. Hettinger et al 1993) in many cases precludes the lowstand interpretation, because such estuaries develop as a consequence of transgression. JC Van Wagoner (personal communication, 1991) has defended the use of the term lowstand for estuarine valley fills on the grounds that the differentiation of sandstones within incised valleys from those of the underlying highstand systems tract is of practical importance for the delineation of reservoirs for oil and gas. However, such commercial objectives can surely be achieved without fundamentally altering the systems tract concept.

Variations in Sequence Architecture

Case studies in a great variety of settings have led to attempts to develop modified versions of the standard sequence stratigraphic model. Examples include settings where sedimentation is accompanied by growth faulting, terraced shelves and siliciclastic ramps, carbonate platforms and ramps, nonmarine environments (fluvial, eolian, and lacustrine), and environments proximal to large ice sheets (Vail 1987; Sarg 1988; Van Wagoner et al 1988, 1990; Boulton 1990; Olsen 1990; Vail et al 1991; Greenlee et al 1992; Posamentier et al 1992a; Walker & Plint 1992; Dam & Surlyk 1993; Handford & Loucks 1993; Kocurek & Havholm 1993; Schlager et al 1994; Shanley & McCabe 1994). Attempts have also been made to integrate geological studies in outcrop and the subsurface with seismic profiling and shallow sampling of modern shelves and estuaries (Demarest & Kraft 1987; Suter et al 1987; Boyd et al 1989; Tesson et al 1990, 1993; Allen & Posamentier 1993, Chiocci 1994), flume experiments (Wood et al 1993, Koss et al 1994) and small-scale natural analogues (Posamentier et al 1992b), and computer simulations (Jervey 1988, Helland-Hansen et al 1988, Lawrence et al 1990, Ross 1990, Reynolds et al 1991, Steckler et al 1993, Bosence et al 1994, Ross et al 1994). The main emphasis of these studies has been to document variations in the arrangement of facies and associated stratal geometry, but among the more interesting results has been the emergence of some new ideas about the origin of sequence boundaries and other stratigraphic discontinuities.

ORIGIN OF SEQUENCE BOUNDARIES

The conventional interpretation of sequence boundaries is that they are due to a relative fall of sea level (Posamentier et al 1988, Posamentier & Vail 1988, Sarg 1988, Vail et al 1991). For example, a boundary might be said to develop when the rate of relative sea-level fall is a maximum or when relative sea level begins to fall at some specified break in slope, thereby initiating the incision of valleys by headward erosion. The concept of relative sea-level change accounts qualitatively for the roles of both subsidence and eustasy in controlling the space available for sediments to accumulate. However, existing models are fundamentally eustatic because it is invariably the eustatic component that is inferred to fluctuate; the rate of subsidence is assumed to vary only slowly, if at all. Here we draw attention to several ways in which the conventional explanation of sequence boundaries needs to be modified, particularly in tectonically active basins.

Gradual vs Instantaneous Development of Sequence Boundaries

It is widely assumed that sequence boundaries develop more or less instantaneously (Posamentier et al 1988, Vail et al 1991). The main evidence for this is the marked asymmetry of depositional sequences, which in seismic reflection profiles are characterized by progressive onlap at the base and by a downward (or basinward) shift in onlap at the top (Figure 2; Vail et al 1977, 1984; Haq et al 1987). (The term onlap refers to the lateral termination of strata against an underlying surface.) Abrupt downward shifts in onlap are taken to imply rates of relative sea-level change significantly higher than typical rates of subsidence, and that the sea-level change must therefore be due to eustasy, presumably glacial-eustasy (Vail et al 1991).

While it is undoubtedly true that glacial-eustasy has played an important role in modulating patterns of sedimentation since Oligocene time (Bartek et al 1991, Miller et al 1991), and probably during other glacial intervals in earth history, such explanations are not plausible for periods such as the Cretaceous for which there is very little evidence for glaciation (Frakes et al 1992). Nor are such explanations required by available stratigraphic data. In many cases, the highstand systems tract is divisible into two parts: a lower/landward part characterized by onlap and sigmoid clinoforms (clinoforms are inclined stratal surfaces associated with progradation), and an upper/seaward part characterized by offlap and oblique clinoforms (Figure 2; Christie-Blick 1991). Offlap (the upward termination of strata against an overlying surface) is not likely to be due solely to the erosional truncation of originally sigmoid clinoforms. The progressive onlap implied by this interpretation is not possible during a time of increasingly rapid progradation without a marked increase in the sediment supply. Moreover, the inferred erosion is unlikely to have taken place in the subaerial environment because subaerial erosion tends to be focused within incised valleys, and the amount of erosion required in many cases exceeds what can reasonably be attributed to shoreface ravinement during transgression. A more reasonable interpretation is that offlap is due fundamentally to bypassing *during* progradation (toplap of Mitchum 1977), implying that sequence boundaries develop gradually over a finite interval of geologic time (Christie-Blick 1991).

Support for this idea is provided by recent high-resolution sequence stratigraphic studies in outcrop. A remarkable series of forestepping high-order sequences is exposed in the upper part of the San Andres Formation, a carbonate platform of Permian age in the Guadalupe Mountains of southeastern New Mexico (Figure 3*A*; Sonnenfeld & Cross 1993). Individual high-order sequences consist of two half-cycles. The lower half-cycle (primarily transgressive) is predominantly siliciclastic and onlaps the underlying sequence boundary. The upper half-cycle (regressive) is composed mainly of carbonate rocks and is characterized by onlap and downlap (downward termination of strata) at the base and in some instances by offlap at the top. These high-order sequences are themselves oblique to a prominent low-order sequence boundary at the top of the San Andres Formation (top of sequences uSA4 and uSA5). On the basis of karstification, this surface is interpreted by Sonnenfeld & Cross (1993) to have been exposed subaerially. At the resolution of conventional seismic data, only the low-order sequence boundaries in the San Andres Formation would be identified (Figure 3*B*) and the oblique truncation of high-order sequences would be interpreted as offlap (cf figure 11 of Brink et al 1993). However, the siliciclastic half-cycles of the high-order sequences indicate that the platform was bypassed episodically during overall progradation. The sequence boundary at the top of the San Andres was therefore not produced by an instantaneous downward shift in onlap but is instead a composite surface.

Another pertinent example of high-resolution sequence stratigraphy is drawn from the work of Van Wagoner and colleagues in a late Cretaceous foreland-basin ramp setting in the Book Cliffs of eastern Utah and western Colorado (Van Wagoner et al 1990, 1991). Numerous sequence boundaries have been documented in the strongly progradational succession between the Star Point Formation and Castlegate Sandstone (Figure 4*A*). Two of the most prominent boundaries are present at the level of the Desert Member of the Blackhawk Formation and Castlegate Sandstone (Figure 5). These boundaries are characterized by valleys as much as several tens of meters deep incised into underlying

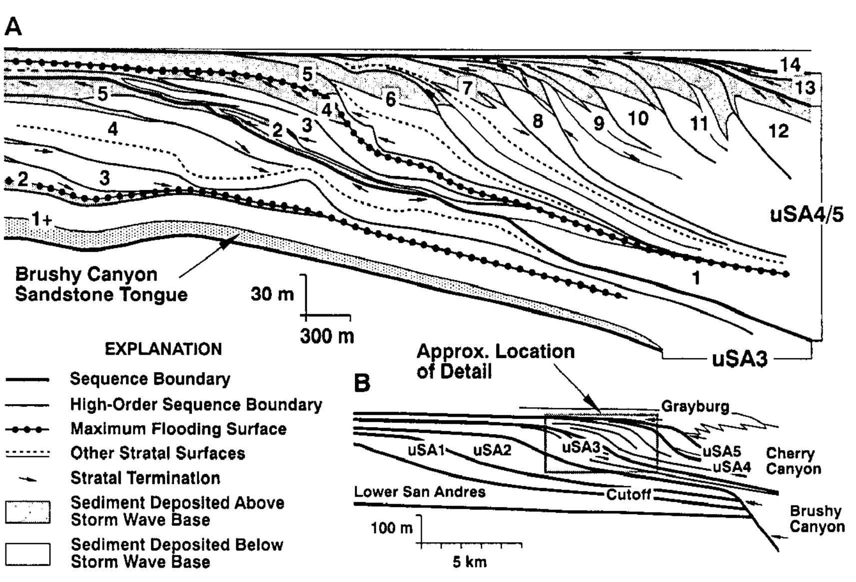


Figure 3 (*A*) Simplified stratigraphic cross section of the upper part of San Andres Formation (Permian) in the Guadalupe Mountains, New Mexico. (*B*) Schematic representation of the broader stratigraphic context of the San Andres Formation at the scale of conventional seismic reflection data. Individual high-order sequences within sequences uSA4 (numbered 1–12) and uSA5 (numbered 13–14) are characterized by stratal onlap and offlap and are themselves oblique to a still lower-order sequence boundary at the top of the San Andres Formation. The datum for cross section *A* is the base of the Hayes sandstone of the Grayburg Formation. Also shown in *B* are the names of other associated lithostratigraphic units. (Modified from Sonnenfeld & Cross 1993.)

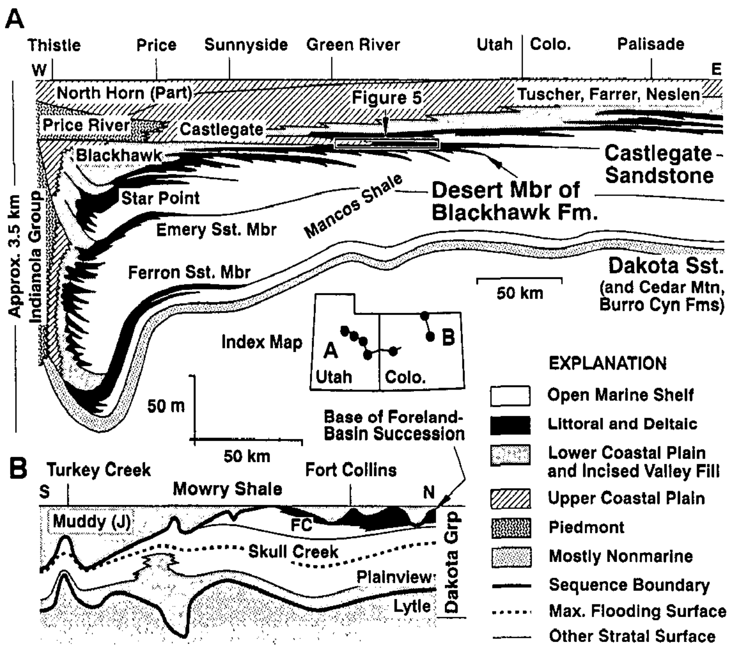


Figure 4 Simplified stratigraphic cross section and lithostratigraphic nomenclature for mid- to upper Cretaceous strata in the Book Cliffs, eastern Utah and western Colorado (*A*; from Nummedal & Remy 1989), with a detail of the Albian sequence stratigraphy (Dakota Group) of north-central Colorado (*B*; from Weimer 1984 and RJ Weimer, personal communication, 1988). A detail of the Desert Member of the Blackhawk Formation and Castlegate Sandstone (box in *A*) is illustrated in Figure 5. The base of the foreland-basin succession is marked approximately by a regional sequence boundary at or near the base of the Dakota Sandstone (in *A*) and at or near the base of the Muddy (or J) Sandstone of the Dakota Group (in *B*). FC refers to the Fort Collins Member, a portion of the Muddy Sandstone that locally underlies the sequence boundary.

littoral sandstones, by the offlap of successive parasequences, and by a marked basinward shift of facies. In the conventional interpretation, the incision of valleys by headward erosion from a break in slope near the shoreline ought to deliver a significant volume of sediment to the adjacent shelf, and prominent lowstand sandstones would be expected (Van Wagoner et al 1988, 1990). Instead, each sequence boundary passes laterally into a marine flooding surface and eventually into the Mancos Shale (Figure 4*A*) with little or no evidence for lowstand deposits as this term is defined above.[1] Our solution to this apparent paradox is that the valleys were not incised as a result of headward erosion.

[1]D Nummedal (personal communication, 1994) has recently identified a possible lowstand deposit basinward of the Castlegate lowstand shoreline on the basis of well-log interpretation. The deposit is perhaps analogous to the relatively thin lowstand units from the Alberta basin described by Plint (1988, 1991) and Posamentier et al (1992a).

A consequence of the idea that sequence boundaries develop gradually during highstand progradation is that incised valleys at the Desert and Castlegate sequence boundaries initially propagated *downstream*, and that most of the eroded sediment accumulated at the highstand shorelines.

If sequence boundaries do not after all develop instantaneously, it is not necessary to call upon rapid eustatic change for which there is no plausible mechanism during nonglacial times. Forward modeling indicates that sequence

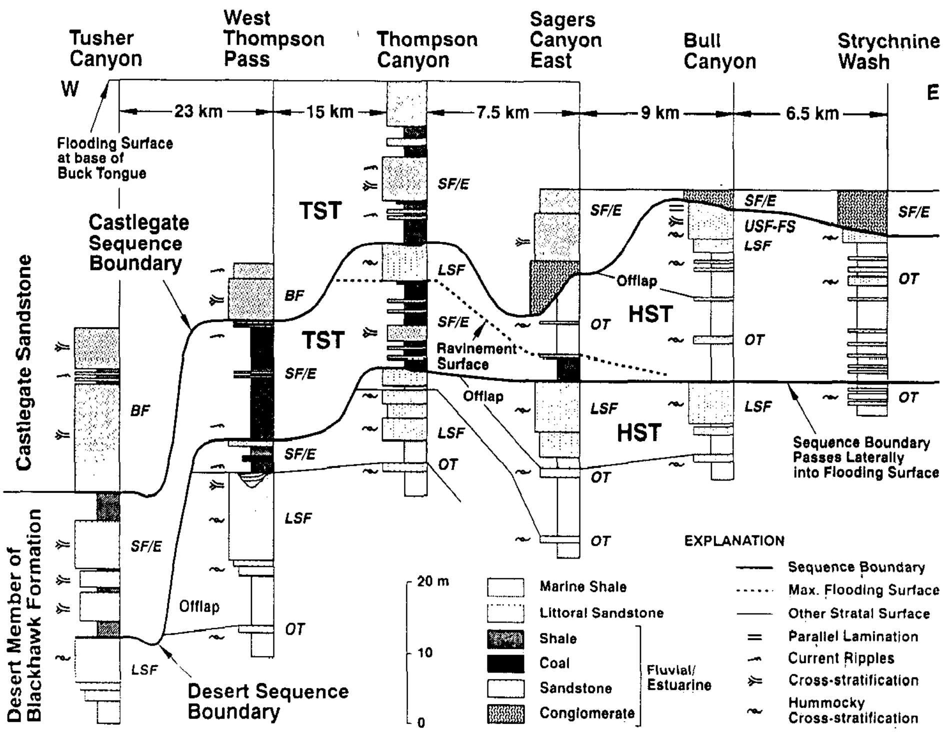


Figure 5 Stratigraphic cross section of the Desert Member of the Blackhawk Formation and Castlegate Sandstone showing depositional facies and sequence geometry (simplified from Van Wagoner et al 1991, Nummedal & Cole 1994). See Figure 4 for location. The two sequence boundaries illustrated are characterized up-dip (west) by well-developed incised valleys. The Desert sequence boundary passes down-dip (eastward) in the vicinity of Sagers Canyon into a marine flooding surface that was probably modified by ravinement during transgression. A similar transition is observed in the Castlegate Sandstone as it is traced farther eastward. Note the presence of offlapping parasequences beneath each sequence boundary. Abbreviations for generalized paleoenvironments: BF, braided fluvial; SF/E, sinuous fluvial/estuarine; FS, foreshore; USF, upper shoreface; LSF, lower shoreface; OT, offshore transition. Systems tracts (modified from the interpretations of Van Wagoner et al 1991 and Nummedal & Cole 1994): HST, highstand; TST, transgressive. Some uncertainty exists about the location of the interval of maximum flooding in the Desert sequence owing to the difficulty of interpreting parasequence stacking trends in thin sections: It is at or slightly above the dashed line labeled *Ravinement Surface* (D Nummedal, personal communication, 1994).

boundaries can be produced by eustatic fluctuations at rates comparable to typical rates of tectonic subsidence and that they do so by gradual basinward expansion and subsequent burial of zones of bypassing and/or erosion (Christie-Blick 1991, Steckler et al 1993). Consequently, if the rate of eustatic change required to generate a sequence boundary is small, there is no reason to assume that sequence boundaries are necessarily due to eustatic change.

Tectonically Active Basins

In the light of these considerations, how do sequence boundaries develop in tectonically active settings such as extensional, foreland, and strike-slip basins? One view, which is almost certainly incorrect, is that the local development of sequence boundaries in such basins may be entirely tectonic in origin (Underhill 1991). Another view is that tectonic processes control long-term patterns of subsidence and that short-term depositional cyclicity is due to eustatic change (e.g. Vail et al 1991, Devlin et al 1993, Lindsay et al 1993). Again, this is an assumption that in many cases is probably not warranted for times in earth history when rates of eustatic change were comparable to rates of tectonic subsidence (e.g. Cretaceous). Sequence boundaries are not merely enhanced or obscured by tectonic activity (cf Vail et al 1984, 1991). Both their timing and their very existence are due to the combined effects of eustasy and variations in patterns of subsidence/uplift and sediment supply. The roles of these factors and the manner in which they interact are admittedly very difficult to sort out, but recent studies provide some useful first-order clues.

An important attribute of tectonically active basins is that it is possible for the rate of tectonic subsidence to increase and decrease simultaneously in different parts of the same basin (Figure 6). Sequence boundaries that are fundamentally of tectonic rather than eustatic origin cannot therefore be attributed satisfactorily to the concept of a relative sea-level fall or even an increase in the rate of relative sea-level fall, because relative sea-level may have been both rising and falling at an increasing rate in different places.

In this regard, patterns of subsidence and uplift in foreland and extensional basins are actually very similar during times of active deformation as well as quiescence (Christie-Blick & Driscoll 1994). In the case of a foreland basin, loading by the adjacent orogen leads to regional subsidence and to uplift in the vicinity of the peripheral bulge, with a wavelength and amplitude that are governed by the flexural rigidity of the lithosphere (Figure 6*A*; Beaumont 1981, Karner & Watts 1983). Uplift may also occur locally at the proximal margin of the basin as a result of the propagation of thrust faults at depth. Similarly, in extensional basins, subsidence and tilting of the hanging-wall block (above the fault in Figure 6) are accompanied by uplift of the footwall (below the fault; Wernicke & Axen 1988, Weissel & Karner 1989). During times of tectonic quiescence, these patterns are reversed, although the amplitudes are small in

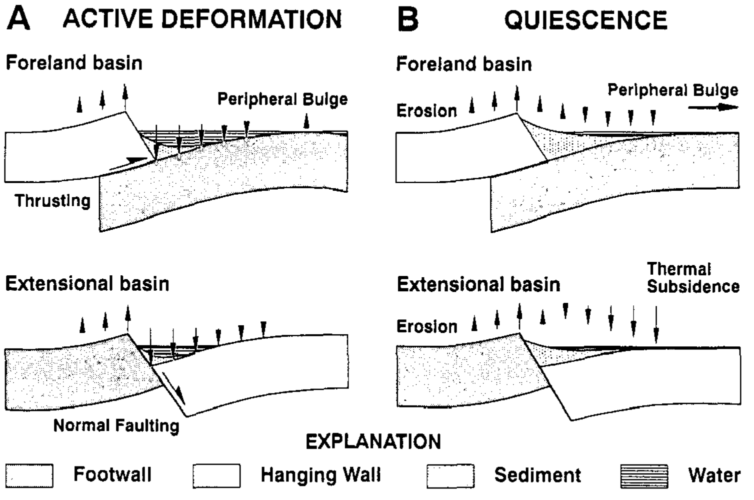


Figure 6 Schematic diagrams comparing patterns of uplift and subsidence in foreland and extensional basins during times of active deformation (*A*) and quiescence (*B*). See text for discussion.

comparison with the deflections engendered by active deformation (Figure 6*B*; Heller et al 1988, Jordan & Flemings 1991). Erosional unloading leads to regional rebound of the orogen and adjacent depocenter and to depression of the peripheral bulge. At the same time, the accumulation of terrigenous sediment derived from the orogen results in additional subsidence at the distal side of the basin and to lateral migration of the peripheral bulge away from the orogen (Jordan & Flemings 1991). A similar pattern of uplift and subsidence may arise during times of tectonic quiescence in extensional basins, through a combination of erosional unloading of the footwall block and thermally driven subsidence, especially when the latter is offset from the original depocenter (as illustrated in Figure 6*B*).

More complicated scenarios can also be envisaged. For example, foreland basins are commonly segmented by block-faulting, and lithospheric extension may be accommodated by a series of tilted fault blocks or distributed inhomogeneously as a function of depth and lateral position within the lithosphere. Subsidence and uplift may also be complicated in three dimensions by the presence of salients in the orogen or of accommodation zones in extensional basins. Patterns of subsidence and uplift in strike-slip basins tend to be even more complicated and subject to marked changes on short time scales (Christie-Blick & Biddle 1985).

The development and characteristics of sequence boundaries in tectonically active basins are directly related to patterns of subsidence and uplift of the sort outlined here and to the fact that the patterns vary between times of active deformation and quiescence. At a regional scale, the progradation of sedimentary systems, the filling of available accommodation with sediment, the lowering of depositional base level, and the incision of valleys are favored during times of tectonic quiescence (e.g. Heller et al 1988). In contrast, active deformation is associated with regional subsidence and tilting, flooding and backstepping of sedimentary systems, an increase in topographic relief along the faulted basin margins, and a narrowing of the depocenter (e.g. Underhill 1991). In both foreland and extensional basins, the most prominent sequence boundaries therefore are expected to correspond with the onset of deformation. These boundaries consist of subaerial exposure surfaces that pass laterally into marine onlap/downlap surfaces of regional extent (Jordan & Flemings 1991, Underhill 1991, Driscoll 1992, Christie-Blick & Driscoll 1994, Driscoll et al 1995). In contrast to the standard model, the formation of a sequence boundary is not necessarily associated with a basinward shift in facies, and where present, such facies shifts may be restricted to areas that were subaerially exposed. The development of topographic relief may in some cases lead to the accumulation of thick successions of turbidites in deep water. However, contrary to the conventional interpretation, these deposits are not strictly "lowstands" if they accumulate during a time of regional flooding [see Southgate et al (1993) and Holmes & Christie-Blick (1993) for a possible example from the Devonian of the Canning basin, Australia].

Several of these points can be illustrated with reference to the late Cretaceous foreland basin of Utah and Colorado (Figure 4). The base of the foreland-basin succession in eastern Utah and Colorado corresponds approximately with a regional sequence boundary at or near the base of the Dakota Sandstone (Figure 4*A*) and at or near the base of the Muddy (or J) Sandstone of the Dakota Group (Figure 4*B*). The succession above this surface is characterized by a marked increase in the rate of sediment accumulation and by an abrupt transition from nonmarine to relatively deep marine sedimentary rocks (Mancos/Mowry Shale; Heller et al 1986, Vail et al 1991). These features are fundamentally attributable to the onset in late Albian time of a phase of crustal deformation and accelerated subsidence; the contribution of eustatic change is indeterminate but is presumed to have been small. We do not preclude the possibility of slightly earlier (late Aptian to Albian, pre-Dakota) foreland-basin development to the west (Yingling & Heller 1992), but the strata are entirely nonmarine and the evidence is equivocal. Within the foreland-basin succession, the origin of other sequence boundaries is less firmly established, but the Blackhawk Formation and Castlegate Sandstone exhibit features that are consistent with our model. The Blackhawk and lower/distal part of the Castlegate (below the Castlegate sequence boundary; Figure 5) are characterized by strong progradation and the development of offlap, consistent with erosional unloading of the orogen during a time of tectonic quiescence (cf Posamentier & Allen 1993). The upper/proximal part of the Castlegate (above the sequence boundary) is transgressive and, as datumed from a flooding surface at the base of the Buck Tongue of the Mancos Shale, it thickens towards the orogen (Figure 5). At a regional scale, it appears to pass laterally into syn-orogenic conglomerate (Price River Formation) and to overstep the Blackhawk Formation, which was uplifted and bevelled to the west (Figure 4). These features make it hard to argue that the Castlegate sequence boundary developed solely or even primarily as a result of eustatic change.

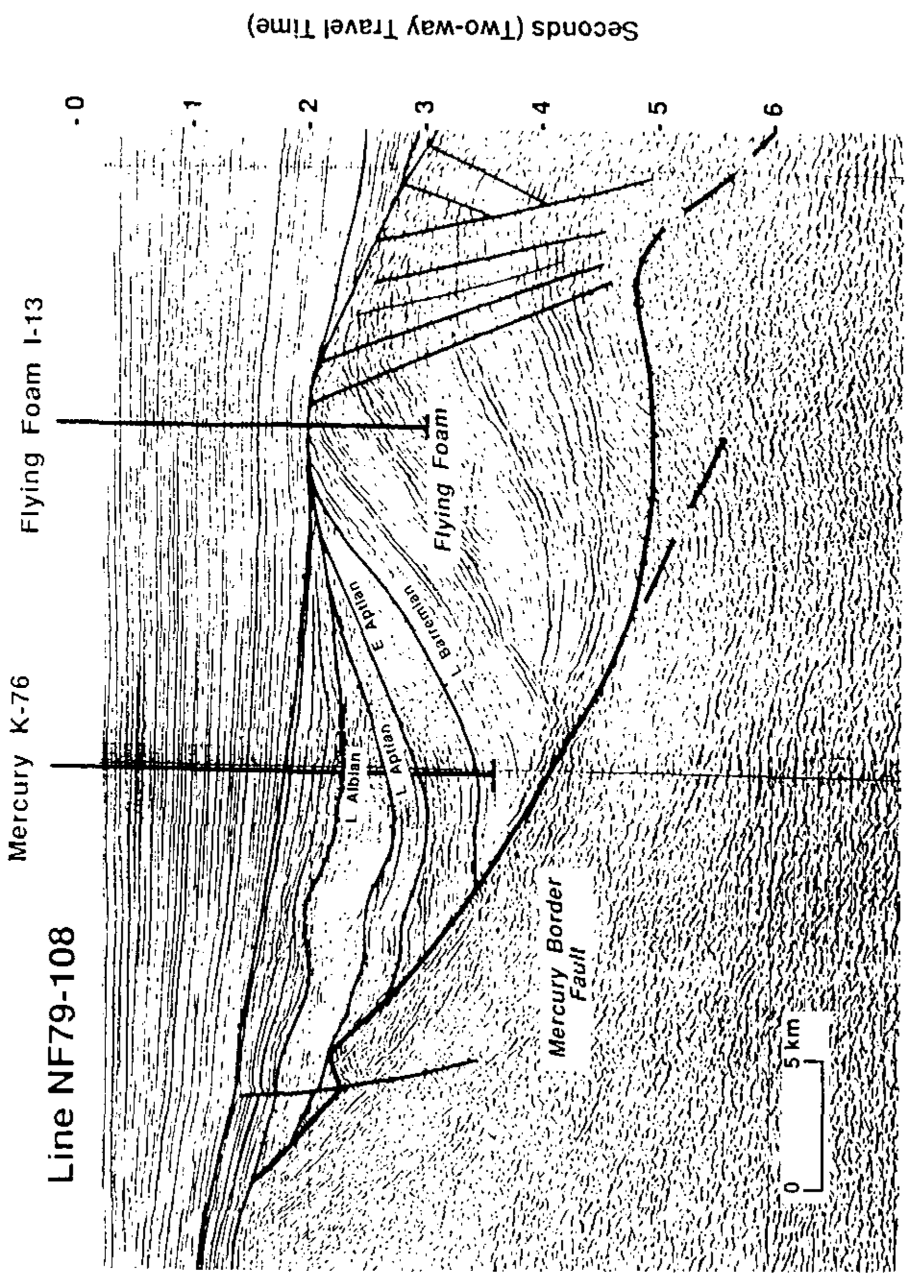
A seismic example of sequence development related to episodic block faulting in an extensional setting is provided by a seismic reflection profile (line NF79–108) from the Jeanne d'Arc basin of offshore eastern Canada (Figure 7). The basin records a series of extensional events between late Triassic and early Cretaceous time (Jansa & Wade 1975, Tankard et al 1989, McAlpine 1991); Figure 7 illustrates the last of these prior to the onset in late Aptian time of seafloor spreading between the Grand Banks of Newfoundland and the Iberian peninsula. Reflections below the late Barremian unconformity are approximately parallel and concordant with the unconformity. Above this surface, the onlap of reflections and their divergence towards the border fault document the onset of crustal extension. Similar reflection geometry is evident at the early Aptian and late Aptian unconformities, although reflections are approximately parallel above the latter. This is taken to indicate that extension had ceased by late Aptian time (Driscoll 1992, Driscoll et al 1995). Evidence from available core indicates that the onlap surfaces are associated with upward deepening of depositional facies, but the surfaces are interpreted as sequence boundaries because they are inferred to pass laterally into subaerial exposure surfaces. The observed geometry requires rifting between late Barremian and late Aptian time. Our preferred interpretation is that each of the mapped sequence boundaries records times of accelerated block tilting. An alternative interpretation, that the early and late Aptian boundaries are due primarily to eustatic fluctuations during a time of more or less continuous block tilting, is not consistent with the absence of anticipated lowstand deposits in closed paleobathymetric lows (for example, at the Mercury K-76 well, Figure 7).

Role of In-Plane Force Variations in the Origin of High-Order Sequence Boundaries

One of the main arguments for interpreting high-order depositional sequences in terms of eustatic change is the absence of another suitable mechanism. We have seen in the Jeanne d'Arc basin example that episodic tilting may provide such a mechanism in extensional basins. However, difficulties arise in foreland basins because, in such settings, subsidence is driven primarily by the integrated vertical load of the orogen. This can change only slowly through a combination of internal deformation, the propagation of thrust faults into the syn-orogenic sediments, and the erosion of topography (Sinclair et al 1991).

An imaginative and somewhat controversial solution to this dilemma has emerged from the recognition and modeling of in-plane force variations in the lithosphere (Lambeck 1983, Cloetingh et al 1985, Karner 1986, Braun & Beaumont 1989, Karner et al 1993). The best evidence for the existence of such forces is the incidence of intraplate earthquakes (e.g. Lambeck et al 1984, Bergman & Solomon 1985). Changes in in-plane force are thought to result from changes in the plate-boundary forces associated with, for example, ridge-push, slab-pull, and collisional tectonics (Sykes & Sbar 1973, Cloetingh & Wortel 1985, Zoback et al 1989). Although uncertainty exists about both the magnitude of the forces and the time scale over which they might vary, it is not unreasonable to think that such forces may be relevant to the development of some sequence boundaries. The response of the lithosphere to in-plane compression consists of two components, one elastic (flexural) and the other inelastic (brittle; Goetze & Evans 1979, Karner et al 1993). The brittle component, which is associated with deformation in the upper part of the lithosphere, is influenced by the preexisting structure of the crust and the orientation of faults with respect to the applied tectonic force. It includes the well-known phenomenon of extensional basin inversion. The flexural response to in-plane compression depends on the shape of any preexisting deflection of the lithosphere, the amplitude of the applied force, and the flexural strength of the lithosphere at the time the force was applied. In the case of foreland basins, the predicted flexural response to compression consists of enhanced subsidence in the depocenter, uplift of the peripheral bulge, and a reduction in the flexural wavelength. The amplitude of subsidence and uplift produced in this way are approximately the same because the wavelengths of features being selectively modified are approximately equal (Karner et al 1993). The predicted flexural response for extensional basins is quite different. In-plane compression results in uplift of the depocenter and increased subsidence of the basin margins. In-plane tension leads to uplift of the basin margins and to enhanced subsidence of the depocenter (Braun & Beaumont 1989, Karner et al 1993). The amplitude and wavelength of the flexural deformation are a strong function of the extensional basin geometry and, in the case of basins undergoing post-rift thermal subsidence, of the spatial relationship between rift basins and any associated passive continental margin.

In view of these considerations, our concepts of active deformation and tectonic quiescence need to be modified. With respect to this second-order effect, panels that in Figure 6 are labeled "active deformation" include times of increased in-plane compression in the foreland basin example and decreased in-plane compression (increased tension) in the extensional basin example. Owing to the relatively short length scales relevant to extensional basins (tens of kilometers), it is anticipated that the stratigraphy of such basins ought to be influenced strongly even at short time scales by episodic block tilting (the brittle


Line NF79-108
Mercury K-76
Flying Foam I-13
Seconds (Two-way Travel Time)
0
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L Barremian
Mercury Border Fault
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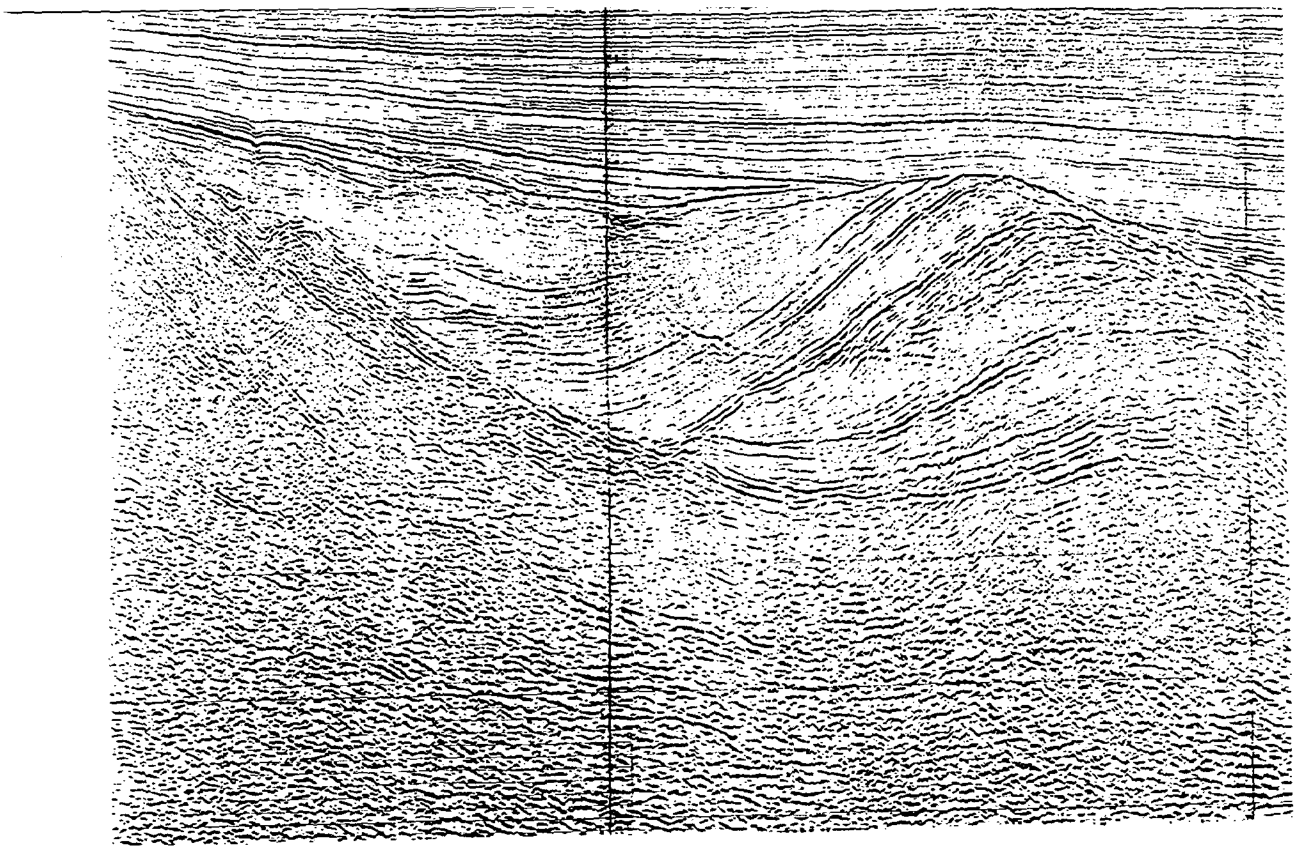

Figure 7 Interpreted and uninterpreted versions of a seismic reflection profile (line NF79-108) from the Jeanne d'Arc basin, offshore eastern Canada (from Driscoll et al 1995). The profile is oriented north-northeast, oblique to the curvilinear Mercury border fault, which strikes approximately eastward at its intersection with the Nautilus transfer zone.

component of the deformation). In contrast, because the integrated vertical load of an orogen changes only slowly (millions to tens of millions of years; Sinclair et al 1991), the stratigraphy of foreland basins ought to be much more sensitive on short time scales to the flexural effects of changes in in-plane compression, providing that the forces involved are sufficiently large. A possible test of this idea is to compare the stratal geometry of sequences of different scale in the same basin. If high-order sequences are due to eustatic change, as many have inferred (e.g. Posamentier & Allen 1993), their geometry ought to reflect overall patterns of subsidence. If they are instead a result of changes in in-plane compression, high-order transgressive systems tracts ought to thicken preferentially toward the orogen relative to associated highstand units (as appears to be the case in the Castlegate example), and backstepping sequences ought to thicken toward the orogen relative to forestepping sequences. The complications associated with lateral changes in facies, compaction, and water depth can be reduced by studying transects parallel to shorelines across foreland basins with axial drainage (for example, the Dunvegan Formation of the Alberta basin; Plint 1994).

CONCLUSIONS

Sequence stratigraphy is concerned with the analysis of sediments and sedimentary rocks with reference to the manner in which they accumulate layer by layer. As a practical technique and in spite of existing terminology, it requires no assumptions about eustasy. One of the principal frontiers of the discipline is an effort to develop an understanding of the many interrelated factors that govern sediment transport and accumulation in a great range of depositional settings and environments. The conventional interpretation of sequence boundaries is that they are due to a relative fall of sea level and that they develop more or less instantaneously. In this paper we argue that in many cases such boundaries form gradually over a finite interval of geologic time. The widely employed concept of relative sea-level change provides few insights into how sequence boundaries actually develop, especially in tectonically active basins.

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