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Table of Contents

SILICICLASTIC SEQUENCE STRATIGRAPHY IN WELL LOGS, CORES, AND OUTCROPS: CONCEPTS FOR HIGH-RESOLUTION CORRELATION OF TIME AND FACIES

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INTRODUCTION

As the search for oil and gas becomes more sophisticated and producing basins and fields become more intensely developed, geoscientists need correspondingly more accurate techniques for stratigraphic analysis. To achieve this accuracy, companies are shooting higher-resolution seismic lines, acquiring 3-D seismic surveys over fields, and coring more to quantify reservoir properties. Exploration and production staff, provided with these more accurate but expensive data, often under-utilize the well log. In basins or fields with a sufficient density of well control, the coupling of conventional well logs and cores with the techniques of *sequence stratigraphy* results in an ultra-high-resolution chronostratigraphic framework for subsurface correlation. Where integrated with seismic and biostratigraphic data, well-log cross sections, interpreted using sequence and parasequence concepts, provide a state-of-the-art framework for analyzing reservoir, source, and seal distribution, whether on a regional or a field-reservoir scale.

Sequence stratigraphy is the study of genetically related facies within a framework of chronostratigraphically significant surfaces. The sequence is the fundamental stratal unit for sequence-stratigraphic analysis. The *sequence* is defined as a relatively conformable, genetically related succession of strata bounded by unconformities or their correlative conformities (Mitchum, 1977). Sequence boundaries form in an increase in water depth. Parasequences and parasequence sets are the building blocks of sequences. A *parasequence* is defined as a relatively conformable, genetically related succession of beds or bedsets bounded by marine-flooding surfaces or their correlative surfaces (Van Wagoner, 1985; Van Wagoner et al., 1988).

A *parasequence set* is defined as a succession of genetically related parasequences that form a distinctive stacking pattern, bounded, in many cases, by major marine-flooding surfaces and their correlative surfaces (Van Wagoner, 1985; Van Wagoner et al., 1988). Parasequence and parasequence set boundaries form in response to an increase in water depth. Under certain depositional conditions, parasequence and parasequence set boundaries may coincide with sequence boundaries. Parasequences are composed of *bedsets, beds, laminasets,* and *laminae* (Campbell, 1967).

These stratal units, ranging from the sequence down to the lamina, are the building blocks of sedimentary rocks; they form a stratigraphic hierarchy and share two fundamental properties: (1) each stratal unit, with the exception of the lamina, is a genetically related succession of strata bounded by chronostratigraphically significant surfaces, and (2) each surface is a single, physical boundary that everywhere separates all of the strata above from all of the strata below over the extent of the surface.

Because of these properties, bounding surfaces that are correlated using well logs, cores, or outcrops, provide a high-resolution chronostratigraphic framework for facies analysis. Vertical facies analysis must be done within conformable stratal packages to accurately interpret coeval, lateral facies relationships along a single depositional surface (Walther, 1894; Middleton, 1973; Reading, 1978; Walker, 1984). Parasequence, parasequence set, and sequence boundaries, occurring with a high frequency in siliciclastic sections, are significant depositional discontinuities;

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in most places, the facies above these boundaries have no physical or temporal relationship to the facies below. Because of this decoupling of facies across these boundaries, vertical facies analysis should be done within the context of parasequences, parasequence sets, and sequences to interpret lateral facies relationships accurately.

Using well logs, cores, or outcrops, each sequence can be subdivided into stratal units called systems tracts, based on their positions within the sequence, the distribution of parasequence sets, and facies associations. Systems tracts are defined as a "linkage of contemporaneous depositional systems" (Brown and Fisher, 1977). Systems tracts provide a high degree of facies predictability within the chronostratigraphic framework of sequence boundaries. This predictability is especially important for the analysis of reservoir, source, and seal facies within a basin or a field.

This book documents the stratal expressions of parasequences, parasequence sets, especially as components of systems tracts, and sequences in well logs, cores, and outcrops. Additionally, the book illustrates well-log, core, and outcrop-recognition criteria for the stratal units from the lamina to the sequence, and demonstrates how the stratal units are used to achieve a high-resolution correlation of time and facies. Finally, the book will relate these stratal patterns to accommodation concepts developed by Jervey (1988), Posamentier et al. (1988), and Posamentier and Vail (1988).

PREVIOUS STRATIGRAPHIC CONCEPTS AND TERMINOLOGY

The sequence as an unconformity-bounded stratal unit was proposed by Sloss in 1948 (Sloss et al., 1949; Sloss, 1950, 1963). Sloss (1963) pointed out, "The sequence concept is not new and was already old when it was enunciated by the writer and his colleagues in 1948. The concept and practice is as old as organized stratigraphy." Nonetheless, Sloss deservedly is given credit for developing the unconformity-bounded sequence as a stratigraphic tool. Sloss (1963) recognized six packages of strata bounded by interregional unconformities on the North American craton between latest Precambrian and Holocene deposits. He called these stratal packages "sequences" and gave them native American names to emphasize their North American derivation (Sloss, 1988). Sloss (1988) used these cratonic sequences as operational units for practical tasks such as facies mapping, although he felt that these sequences "have no necessary applications to the rock stratigraphy and time stratigraphy of extracratonic or extracontinental areas" (Sloss, 1963). Although the concept of the cratonic sequence provided the foundation for sequence stratigraphy, Sloss's ideas had found little acceptance in the 1950s, 1960s, and early 1970s except for Wheeler (1958) and "former students and close acquaintances" (Sloss, 1988).

The next major development in the evolution of sequence stratigraphy occurred when P.R. Vail, R.M. Mitchum, J.B. Sangree, and S. Thompson III of Exxon published the concepts of seismic stratigraphy in the American Association of Petroleum Geologists Memoir 26 (Payton, 1977). In a series of seminal articles these authors presented the concepts of eustasy and resulting unconformity-bounded stratal patterns applied to and documented with seismic-reflection data. Mitchum (1977) sharpened and extended the concept of the sequence by defining it as "a stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities." Vail modified Sloss's (1963) use of sequence in two other important ways. First, the sequence of Vail and Mitchum encompassed a much smaller amount of time than the sequence of Sloss (1963). The original six cratonic sequences were significantly subdivided; Sloss's sequences became supersequences on the Exxon cycle chart. Second, Vail proposed eustasy as the predominant driving mechanism for sequence evolution (Vail et al., 1977). This interpretation alone generated and continues to generate much discussion (Sloss, 1988; Galloway, 1989a). As a result of Memoir 26 (Payton, 1977) and the advent of improved seismic-reflection technology, the sequence as a practical, unconformity-bounded unit for stratigraphic analysis advanced a giant leap beyond Sloss's original concept of cratonic sequences.

Although they represented a major step forward in the application of sequences, seismic-stratigraphy concepts in the late 1970s were applied primarily to basin analysis at the scale of the seismic data. Well logs, cores, and outcrops generally were not used independently to analyze sequences. Seismic stratigraphy did not offer the necessary precision to analyze sedimentary strata at the reservoir scale.

In 1980 the application of seismic stratigraphy was broadened by new accommodation models developed by Jervey (1988) to explain seismically resolvable stratal patterns. The accommodation models quickly led to the realization that the sequence could be subdivided into smaller stratal units, ultimately called "systems tracts" (Brown and Fisher, 1977). In conceptual, 3-dimensional block diagrams developed by Posamentier and Vail, (1988), and Baum and Vail (1988), submarine-fan, lowstand, transgressive, and highstand systems tracts were illustrated in type-1 sequences; shelf-margin, transgressive, and highstand systems tracts were illustrated in type-2 sequences. After 1980 the lowstand system tract of the type-1 sequence was recognized to consist of the basin-floor fan, slope fan, lowstand-prograding wedge, and incised-valley fill (Vail, 1987). Type-1 and type-2 referred to the type of unconformity upon which the sequence rested. Systems tracts and type-1 and type-2 sequences will be explained further in the "Sequence" section, later in the book.

Concurrently with the development of the conceptual models, other Exxon stratigraphers strongly influenced by D.E. Frazier (1974) and C.V. Campbell (1967) began to analyze the stacking patterns of shallowing-upward siliciclastic strata in well logs, cores, and outcrops. The goal of this analysis was to use stacking patterns to improve subsurface correlations of time and facies. These shallowing-upward stratal units are bounded by chronostratigraphically significant marine-flooding surfaces and are composed of laminae, laminasets, beds, and bedsets. Beds, bounded by practically synchronous bedding surfaces, were used as informal time-stratigraphic markers for well-log correlation (Campbell, 1967).

This line of research quickly converged with the conceptual models when it became apparent that the shallowing-upward stratal units and their component sedimentary layers were the building blocks of the systems tracts and sequences. Although shallowing-upward units had been called "cycles" by some other workers (Wilson, 1975; Goodwin and Anderson, 1985), these units were called "parasequences" by Van Wagoner (1985). This usage preserved the dictionary use of the word "cycle" by Vail et al. (1977) to indicate a time in which a regularly repeated event occurs and emphasized the relationship between the parasequence and the sequence.

Groups of associated parasequences were observed to stack into retrogradational, progradational, and aggradational patterns; these distinct associations of parasequences were called "parasequence sets" (Van Wagoner, 1985; Van Wagoner et al., 1988). Each parasequence set approximately corresponded to a systems tract. In addition, each systems tract generally was characterized by a distinct association of facies and by a position within the sequence.

Recognition of parasequences and parasequence sets as the building blocks of the systems tract and the sequence placed them within a chronostratigraphic framework in which their stacking patterns, constituent bedding types, and, to a great extent, their component depositional environments, were predictable. This enhanced their use for the subsurface correlation of time and facies.

The concept of the parasequence, or upward-shoaling cycle as it is commonly named in literature, dates back at least to Phillips (1836) and includes Udden (1912), Weller (1930), Wanless (1950), Duff et al. (1967), Busch (1971, 1974), Wilson (1975), and Einsele and Seilacher (1982). The chronostratigraphic significance of the marine-flooding surface bounding a parasequence was pointed out by Wilson (1975), who stated that carbonate cycles are bounded by widespread transgressive surfaces that may "closely approximate time markers and are more useful as such than the diachronous facies within each cycle." In a review of work by Sears et al. (1941), Krumbein and Sloss (1963) pointed out that the transgressive surface of a progradational-shoreline sandstone approximates a time line. Anderson et al. (1984) and Goodwin and Anderson (1985) also emphasized this importance of cycles for chronostratigraphy, based on work in the carbonate Helderberg Group of New York, and they designated the upward-shoaling carbonate cycle of Wilson (1975) a PAC, an acronym for Punctuated Aggradational Cycle.

By 1983, within Exxon, stratigraphic analysis had evolved beyond parasequence analysis to documentation of the various stratal expressions of siliciclastic sequences and systems tracts in well logs, cores, and outcrops. This represented a major step beyond seismic stratigraphy. Using well logs and cores, a very high-resolution chronostratigraphic framework of sequence and parasequence boundaries, defined solely by the relationships of the strata, could be constructed to analyze stratigraphy and facies at the reservoir scale. Integration of the systematic documentation of siliciclastic sequences, similar advances in carbonate facies (Sarg, 1988), and sequence-keyed biostratigraphy (Loutit et al., 1988) with the methodology of seismic stratigraphy produced the framework and methodology for stratigraphic and facies analysis now known as *sequence stratigraphy*.

As more basins were analyzed with sequence-stratigraphic techniques two important observations were made. (1) Siliciclastic sequences in many parts of the sedimentary record occur with a 100,000- to 200,000-year frequency. This is much higher than has been observed previously by seismic stratigraphers (Goldhammer et al., 1987; Van Wagoner and Mitchum, 1989). (2) The lowstand systems tract is the dominant systems tract preserved in siliciclastic sequences, and on the shelf, its major component is the incised valley.

Examples of incised valleys have been cited in the literature for many years. Fisk (1944) documented the extensive incision in the Mississippi valley in response to the last sea-level fall commencing approximately 27,000 years ago (Williams, 1984). The incised alluvial valley of the Mississippi is, in places, 260 ft deep and 120 mi (193 km) wide (Fisk, 1944). The lower two-thirds of the alluvial fill from Cairo, Illinois, to the present coastline, a distance of approximately 600 mi (963 km), contains gravel and coarse-grained sand. Using high-resolution seismic data, Suter and Berryhill (1985) documented regional incision across the continental shelf of the northern Gulf of Mexico, also in response to the last sea-level fall. Incised valleys in the Albian-aged Muddy Sandstone and its stratigraphic equivalents in the western United States have been studied extensively (Harms, 1966; Stone, 1972; Dresser, 1974; Weimer, 1983, 1984, 1988; and Aubrey, 1989).

Sequence stratigraphy relates the formation of incised valleys to relative changes in sea level and, for the first time, places them in a chronostratigraphic context of parasequence and sequence boundaries. Detailed analysis of sequences in well logs, cores, and outcrops reveals the widespread occurrence in time and space of incised valleys within the updip part of the lowstand systems tract. As a result, the timing and distribution of valley incision and fill becomes more predictable. This, in turn, is critical for understanding:

(1) variations in type-l sequence-boundary expression on the shelf;
(2) regional distribution of shallow-marine and nonmarine depositional environments within each sequence; and
(3) reservoir distribution within the sequence, because on the shelf, incised valleys commonly contain the best reservoirs within each sequence.

THE SEQUENCE AS A TOOL FOR STRATIGRAPHIC ANALYSIS

Application of sequence-stratigraphic analysis depends on the recognition of a hierarchy of stratal units including beds, bedsets, parasequences, parasequence sets, and sequences bounded by chronostratigraphically significant surfaces of erosion, nondeposition, or their correlative surfaces. This method of stratigraphic analysis contrasts with the use of transgressive and regressive cycles of strata for regional correlation of time and facies.

Transgressive and regressive cycles have been used for regional correlation for at least 50 years (Grabau, 1932; Krumbein and Sloss, 1963). Recently, proponents of transgressive and regressive cycles, referred to as T-R units, for regional correlation have included Ryer (1983), Busch and Rollins (1984), Busch et al. (1985), and Galloway (1989a). Galloway (1989a) introduced the "genetic stratigraphic sequence," which is a regressive depositional unit bounded by transgressive surfaces. Although he did not define it specifically, he described it as "a package of sediments recording a significant episode of basin-margin outbuilding and basin filling, bounded by periods of widespread basin-margin flooding."

The genetic stratigraphic sequence is based on Frazier's (1974) concept of depositional episodes patterned after late Quaternary "sequences" deposited during high-frequency "episodes" controlled by glacial cycles. The depositional episodes are bounded by "hiatuses" or flooding surfaces formed during sea-level rise or by shifting delta lobes. Galloway (1989b) applied Frazier's (1974) concept to much larger Cenozoic units of the Gulf of Mexico basin, recognizing about 14 major continental-margin outbuilding episodes, each of which culminated in a major flooding event. Although Frazier's (1974) depositional episodes have frequencies comparable to fourth-order sequences, Galloway's (1989b) units average 4 to 5 Ma in frequency. They commonly include several third-order sequences as defined by Vail et al. (1977).

Both T-R cycle analysis and the nearly identical "genetic stratigraphic sequence" analysis rely on the transgressive surface at the top of a regressive unit or the surface of maximum flooding for regional correlation. We believe that the sequence boundary is a better surface for *regional* stratigraphic analysis than a transgressive surface for the following reasons:

(1) The sequence boundary is a single, widespread surface that separates all of the rocks above from all of the rocks below the boundary. Although all points on the sequence boundary do not represent the same duration of time, one instant of time is common to all points. This synchroneity is basinwide and is interpreted to be global within limits of biostratigraphic dating. For these reasons the sequence boundary has time-stratigraphic significance.
(2) The sequence boundary forms independently of sediment supply. A rapid relative fall in sea level coupled with a large supply of sediment delivered rapidly will result in a sequence boundary strongly marked by truncation. A rapid relative fall in sea level coupled with a minor supply of sediment delivered slowly will result in a sequence boundary marked by widespread subaerial exposure but little truncation. In contrast, transgressions and regressions are strongly controlled by sediment supply and for that reason may not be synchronous, even within a given basin. For example, movements of the shoreline are often due to local differences in sediment supply around a basin rather than sea-level changes, and therefore typically are regionally diachronous.
(3) There are two major transgressive surfaces within the sequence: the first flooding surface forming the upper boundary of the lowstand systems tract and the maximum-flooding surface associated with the condensed section. Typically, several other transgressive surfaces, bounding parasequences within the transgres-

sive systems tract, occur between these major surfaces. All of these surfaces potentially can be confused in regional correlation, especially if the data used to correlate are widely spaced. The age of each transgressive surface within a sequence at different points in a basin may differ significantly depending upon variations in regional sediment supply.

(4) The sequence boundary commonly is marked by significant regional erosion and onlap, which exert a strong control on facies distribution. Transgressive surfaces are characterized by very slow deposition or nondeposition with only relatively minor transgressive scour.

(5) Systems tracts occur predictably within the sequence and are related to the sequence boundary; each systems tract is associated with the boundary at some point. This relationship is not true of the transgressive surfaces.

(6) There is a distinct break in deposition and a basinward shift in facies across the unconformable portion of a type-1 sequence boundary, making it a natural surface for separating relatively conformable facies packages above and below. Commonly, this break occurs within the middle to upper parts of regressive units. If the transgressive surfaces bounding Galloway's (1989a) "genetic stratigraphic sequence" are used to subdivide basin stratigraphy and the sequence boundaries are overlooked, then the basic depositional unit contains a potentially major unconformity within it, making the accurate interpretation of lateral-facies relationships difficult.

(7) Recognizing the unconformable portion of the sequence boundary as part of the hierarchy of chronostratigraphic stratal surfaces and discontinuities described in this book has great significance in working out chronostratigraphy and contemporaneity of facies. However, using only facies boundaries, or subordinating "the stratigraphy of surfaces" (Galloway, 1989a) to facies boundaries that commonly transgress geologic time, may lead to erroneous conclusions about contemporaneity of facies distribution.

As will be discussed throughout this book, the sequence, bounded by unconformities or their correlative conformities, is a highly practical stratal unit for regional stratigraphic analysis with seismic, well log, and biostratigraphic data, as well as for reservoir-scale analysis using well logs, outcrops, and cores. It is most completely understood and used at all scales of analysis by a synthesis of these data bases.

SEQUENCE STRATIGRAPHY AND THE HIERARCHY OF STRATAL UNITS

As already discussed, stratal units from the lamina to the sequence can be grouped into a hierarchy. Recognition of these stratal units and their use in correlating time and facies is the essence of sequence stratigraphy. The following discussion builds upward from the smallest unit in the hierarchy, the lamina, to the largest unit considered in this book, the sequence.

Each stratal unit in the hierarchy is defined and identified only by the physical relationships of the strata, including lateral continuity and geometry of the surfaces bounding the units, vertical-stacking patterns, and lateral geometry of the strata within the units. In addition, facies and environmental interpretations of strata on either side of bounding surfaces are critical, especially for parasequence, parasequence set, and sequence-boundary identification. Thickness, time for formation, and interpretation of regional or global origin are not used to define stratal units or to place them in the hierarchy. In particular, parasequences and sequences can be identified in well logs, cores, or outcrops and used to construct a stratigraphic framework regardless of their interpreted relationship to changes in eustasy.

Documentation of parasequences, parasequence sets, and sequences in this book is primarily from Tertiary strata in the northern Gulf of Mexico and Cretaceous strata of the basins in the western interior of the United States. Examples are exclusively of siliciclastic rocks; however, many of the concepts documented by these examples can also be applied to carbonate strata (Sarg, 1988).

LAMINA, LAMINASET, BED, BEDSET

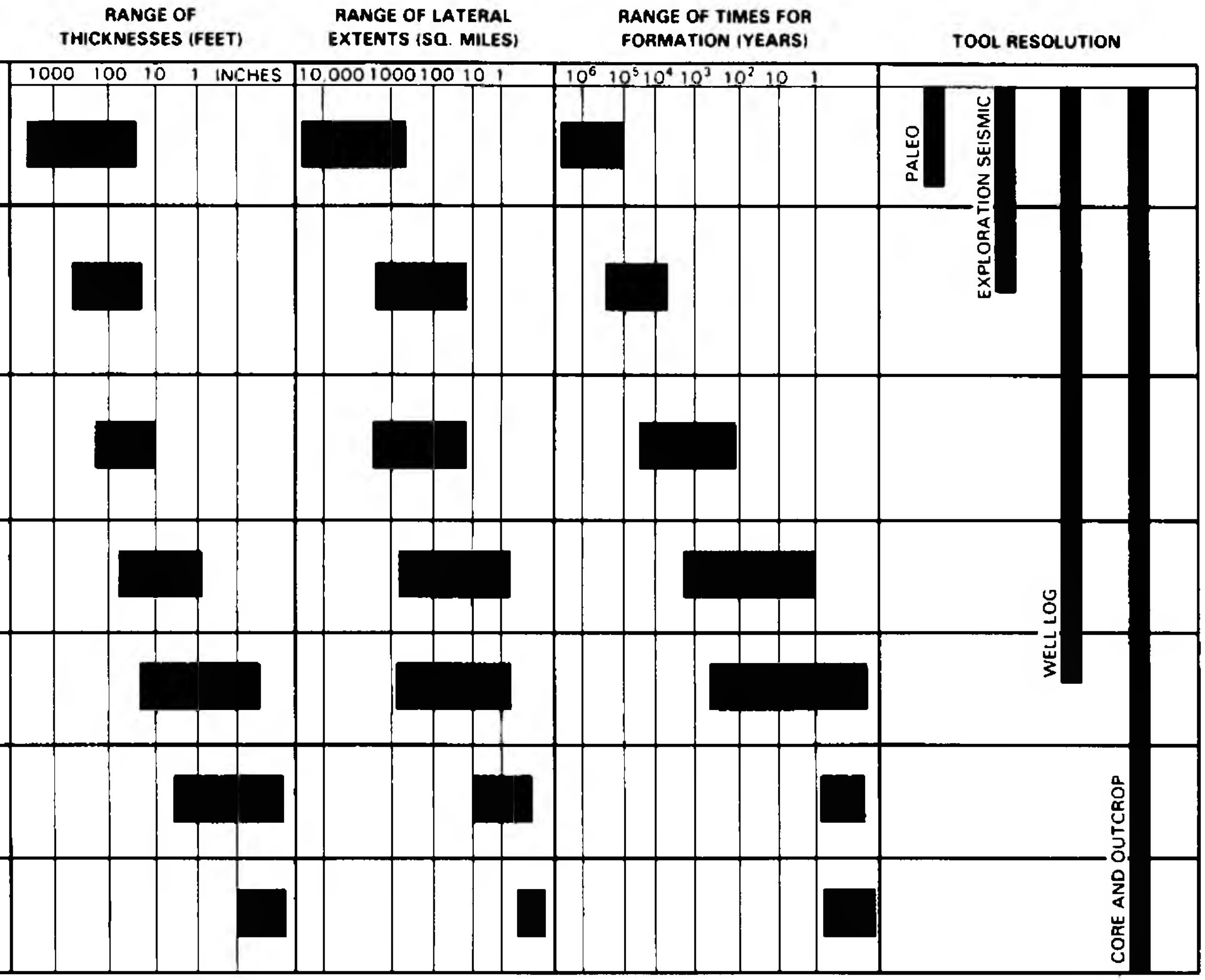
Campbell (1967) identified laminae, laminasets, beds, and bedsets as the components of a sedimentary body; we recognize these stratal units as the building blocks of parasequences. General characteristics of these units are given in Table 1; definitions and more detailed characteristics are given in Table 2. Figure 1 shows these types of strata from delta-front turbidites in cores, outcrops, and well logs from the Panther Tongue of late Santonian age (Fouch et al., 1983) in east-central Utah. Because treatment of these units is not our major thrust, Campbell's (1967) paper is recommended for additional detail.

The four types of stratal units listed above are genetically similar; they differ primarily in the interval of time for formation and in the areal extent of the bounding surfaces. The surfaces bounding the units are defined by (1) changes in texture, (2) stratal terminations, and (3) paraconformities (Dunbar and Rogers,

Stratal Units in Hierarchy: Definitions and Characteristics

TABLE 1

STRATAL UNITS	DEFINITIONS
SEQUENCE	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED STRATA BOUNDED BY UNCONFORMITIES AND THEIR CORRELATIVE CONFORMITIES (MITCHUM AND OTHERS, 1977)
PARA-SEQUENCE SET	A SUCCESSION OF GENETICALLY RELATED PARASEQUENCES FORMING A DISTINCTIVE STACKING PATTERN AND COMMONLY BOUNDED BY MAJOR MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES
PARA-SEQUENCE	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED BEDS OR BEDSETS BOUNDED BY MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES
BEDSET	SEE TABLE TWO
BED	SEE TABLE TWO
LAMINA-SET	SEE TABLE TWO
LAMINA	SEE TABLE TWO



Detailed Characteristics of Lamina, Laminaset, Bed, and Bedset (from Campbell, 1967)

TABLE 2

STRATAL UNIT	DEFINITION	CHARACTERISTICS OF CONSTITUENT STRATAL UNITS	DEPOSITIONAL PROCESSES	CHARACTERISTICS OF BOUNDING SURFACES
BEDSET	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED BEDS BOUNDED BY SURFACES (CALLED BEDSET SURFACES) OF EROSION, NON-DEPOSITION, OR THEIR CORRELATIVE CONFORMITIES	BEDS ABOVE AND BELOW BEDSET ALWAYS DIFFER IN COMPOSITION, TEXTURE, OR SEDIMENTARY STRUCTURE FROM THOSE COMPOSING THE BEDSET	EPISODIC OR PERIODIC. (SAME AS BED BELOW)	(SAME AS BED BELOW) PLUS ● BEDSETS AND BEDSET SURFACES FORM OVER A LONGER PERIOD OF TIME THAN BEDS ● COMMONLY HAVE A GREATER LATERAL EXTENT THAN BEDDING SURFACES
BED	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED LAMINAE OR LAMINASETS BOUNDED BY SURFACES (CALLED BEDDING SURFACES) OF EROSION, NON-DEPOSITION OR THEIR CORRELATIVE CONFORMITIES	NOT ALL BEDS CONTAIN LAMINASETS	EPISODIC OR PERIODIC EPISODIC DEPOSITION INCLUDES DEPOSITION FROM STORMS, FLOODS, DEBRIS FLOWS, TURBIDITY CURRENTS PERIODIC DEPOSITION INCLUDES DEPOSITION FROM SEASONAL OR CLIMATIC CHANGES	● FORM RAPIDLY, MINUTES TO YEARS ● SEPARATE ALL YOUNGER STRATA FROM ALL OLDER STRATA OVER THE EXTENT OF THE SURFACES ● FACIES CHANGES ARE BOUNDED BY BEDDING SURFACES ● USEFUL FOR CHRONOSTRATIGRAPHY UNDER CERTAIN CIRCUMSTANCES ● TIME REPRESENTED BY BEDDING SURFACES PROBABLY GREATER THAN TIME REPRESENTED BY BEDS ● AREAL EXTENTS VARY WIDELY FROM SQUARE FEET TO 1000's SQUARE MILES
LAMINASET	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED LAMINAE BOUNDED BY SURFACES (CALLED LAMINASET SURFACE) OF EROSION, NON-DEPOSITION OR THEIR CORRELATIVE CONFORMITIES	CONSISTS OF A GROUP OR SET OF CONFORMABLE LAMINAE THAT COMPOSE DISTINCTIVE STRUCTURES IN A BED	EPISODIC, COMMONLY FOUND IN WAVE- OR CURRENT-RIPPLED BEDS, TURBIDITES, WAVE-RIPPLED INTERVALS IN HUMMOCKY BEDSETS, OR CROSS BEDS AS REVERSE FLOW RIPPLES OR RIPPLED TOES OF FORESETS	● FORM RAPIDLY, MINUTES TO DAYS. ● SMALLER AREAL EXTENT THAN ENCOMPASSING BED
LAMINA	THE SMALLEST MEGASCOPIC LAYER	UNIFORM IN COMPOSITION/ TEXTURE NEVER INTERNALLY LAYERED	EPISODIC	● FORMS VERY RAPIDLY, MINUTES TO HOURS ● SMALLER AREAL EXTENT THAN ENCOMPASSING BED

1957) marked by burrow, root, or soil zones. Figure 2 illustrates these criteria at the scale of the bed. The bounding surfaces are slightly erosional to nondepositional and separate younger from older strata. The lateral continuity of the bounding surfaces varies from square inches for some laminasets to thousands of square miles for some beds or bedsets. The surfaces form relatively rapidly, ranging from seconds to thousands of years, and so are essentially synchronous over their areal extents (Campbell, 1967). In addition, the time interval represented by the surfaces bounding these layers probably is much greater than the time interval represented by the layers themselves. For all of these reasons, beds and bedsets commonly can be used for chronostratigraphic correlation, over wide areas in many depositional settings. Closely spaced induction logs (0.5 to 2 mi or 0.8 to 3 km apart, especially in marine-shale or mudstone sections) or continuous outcrops provide the most detailed data for a time-stratigraphic analysis based on bed or bedset surfaces.

PARASEQUENCE

Scope of Observations

Parasequences have been identified in coastal-plain, deltaic, beach, tidal, estuarine, and shelf environments (Van Wagoner, 1985). It is difficult to identify parasequences in fluvial sections where marine or marginal-marine rocks are absent, and in slope or basinal sections, which are deposited too far below sea level to be influenced by an increase in water depth. The general concepts presented here apply to all of the depositional environments mentioned above in which parasequences have been recognized; the following discussion illustrates deltaic and beach parasequences because these are common in most basins.

Definitions

We will use the following terms in the described contexts:

Parasequence: A relatively conformable succession of genetically related beds or bedsets bounded by marine-flooding surfaces or their correlative surfaces. In special positions within the sequence, parasequences may be bounded either above or below by sequence boundaries.

Marine-Flooding Surface: A surface separating younger from older strata across which there is evidence of an abrupt increase in water depth. This deepening commonly is accompanied by minor submarine erosion or nondeposition (but not by subaerial erosion due to stream rejuvenation or a basinward shift in facies), with a minor hiatus indicated. The marine-flooding surface has a correlative surface in the coastal plain and a correlative surface on the shelf.

Delta: A genetically related succession of strata deposited at the mouth of a river, causing the coastline to bulge into a standing body of water. The delta can be subdivided into delta-plain and distributary-channel subenvironments dominated by unidirectional, fluvial processes; and stream-mouth bar, delta-front, and prodelta subenvironments dominated by unidirectional or bidirectional processes. The subenvironments of the delta are interpreted from associations of beds and bedsets, sandstone/shale ratios, and sandstone-body geometry.

Beach: A genetically related succession of strata dominated by wave and current processes and deposited as a ribbon of sediment along a coastline of a standing body of water. The beach can be subdivided into backshore, foreshore, upper-shoreface, and lower-shoreface subenvironments based on associations of beds and bedsets, ichnofossil assemblages, and sandstone/shale ratios.

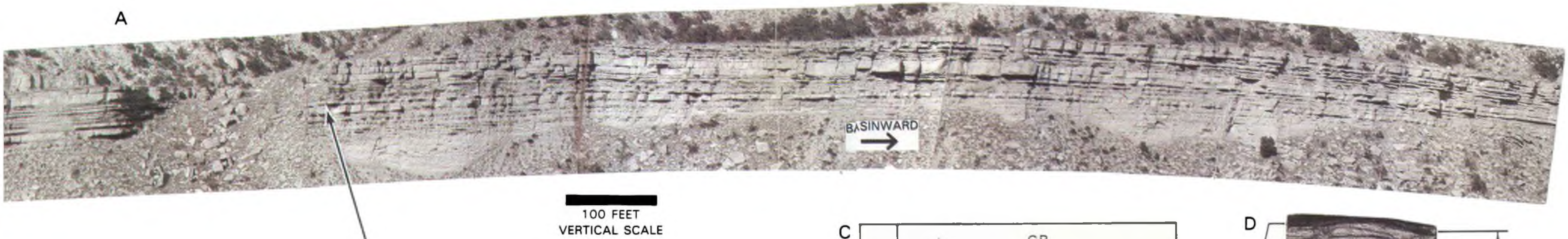
Characteristics

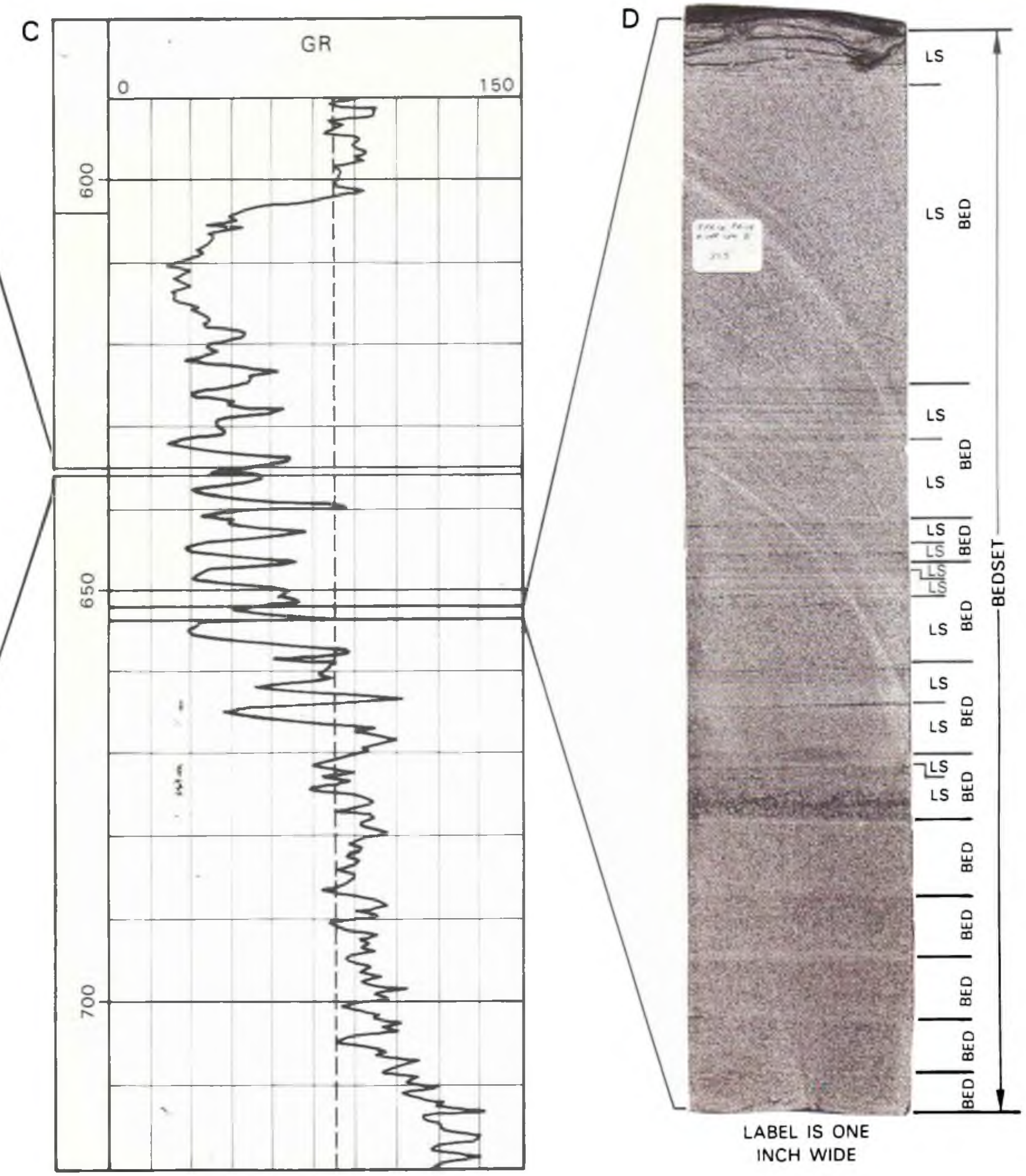
Parasequence characteristics are summarized in Table 1. Most siliciclastic parasequences are progradational, i.e., the distal toes of successively younger sandstone bedsets were deposited progressively farther basinward. This depositional pattern results in an upward-shoaling association of facies in which younger bedsets were deposited in progressively shallower water. Some siliciclastic, and most carbonate, parasequences are aggradational and also shoal upward.

The schematic well-log and stratal characteristics of upward-coarsening and upward-fining parasequences are shown in Figure 3. In the typical upward-coarsening parasequence (Figures 3A–3C), bedsets thicken, sandstones coarsen, and the sandstone/mudstone ratio increases upward. In the upward-fining parasequence (Figure 3D), bedsets thin, sandstones become finer grained (commonly culminating in mudstones and coals), and the sandstone/mudstone ratio decreases upward.

The vertical-facies associations within both the upward-coarsening and upward-fining parasequences are interpreted to record a gradual decrease in water depth. Evidence of an abrupt decrease in water depth, such as foreshore bedsets lying sharply on lower-shoreface bedsets, has not been observed within parasequences. Also, vertical-facies associations indicating a gradual increase in water depth have not been observed within parasequences. If individual "deepening-upward" parasequences do exist, they

Figure 1—Bedset, bed, laminaset, lamina characteristics and relationships.





EXXON PRODUCTION RES. CO
PRICE RIVER COAL NO. 3
NWNW SEC.6-T13S-R10E

A. OUTCROP OF THE PANTHER TONGUE OF THE STAR POINT FORMATION (LATE SANTONIAN-EARLY CAMPANIAN AGE) DEPOSITED IN A DELTA-FRONT ENVIRONMENT. LOCATED IN SESW SEC.12-T13S-R9E EAST-NORTH EAST OF HELPER, UTAH. THE IMBRICATE, BASINWARD-DIPPING STRATA ARE BEDSETS BOUNDED BY BEDSET SURFACES. THE SURFACES ARE CONTINUOUS ALONG THE OUTCROP AND DEFINE THE GEOMETRY OF THE POTENTIAL RESERVOIR ROCKS IN THE DELTA FRONT. THE BEDSETS CHANGE FACIES FROM SANDSTONE TO MUDSTONE FROM LEFT TO RIGHT IN THE PHOTO.

B. DETAIL OF BEDSET IN OUTCROP SHOWING CONSTITUENT BEDS, LAMINAESETS, AND LAMINAE. BEDDING SURFACES ARE PLANAR AND PARALLEL. EACH BED IS NORMALLY GRADED AND COMPRISES A HOMOGENEOUS LAMINASET OVERLAIN BY A LAMINASET COMPOSED OF EVEN, PLANAR LAMINAE. BASED ON THESE CRITERIA THE BEDS ARE TURBIDITES. TURBIDITES BOUNDED BY PLANAR PARALLEL BEDDING SURFACES ARE THE MOST COMMON CONSTITUENTS OF DELTA-FRONT DEPOSITS.

C. GAMMA-RAY WELL LOG FROM A WELL CORED AND LOGGED THROUGH THE PANTHER-TONGUE DELTA. THE WELL IS LOCATED 1.75 MILES NORTH NORTHEAST OF THE LEFT END OF THE OUTCROP SHOWN IN (A) ABOVE. INDIVIDUAL SANDSTONE BEDSETS ARE SHOWN ON THE WELL LOG.

D. A PORTION OF THE CORE CUT THROUGH THE PANTHER TONGUE DELTA. THE CORE IS FROM THE EXXON PRODUCTION RESEARCH COMPANY PRICE RIVER COAL NO. 2 (CSW SEC.11-T13S-R9E) DRILLED IN HARDSCRABBLE CANYON NORTHWEST OF HELPER, UTAH. THE CORE SHOWS A BEDSET COMPOSED OF BEDS, LAMINASETS, AND LAMINAE. EACH BED IS NORMALLY GRADED AND LIKE (B) ABOVE IS A TURBIDITE. BEDDING SURFACES ARE EVEN AND PARALLEL.

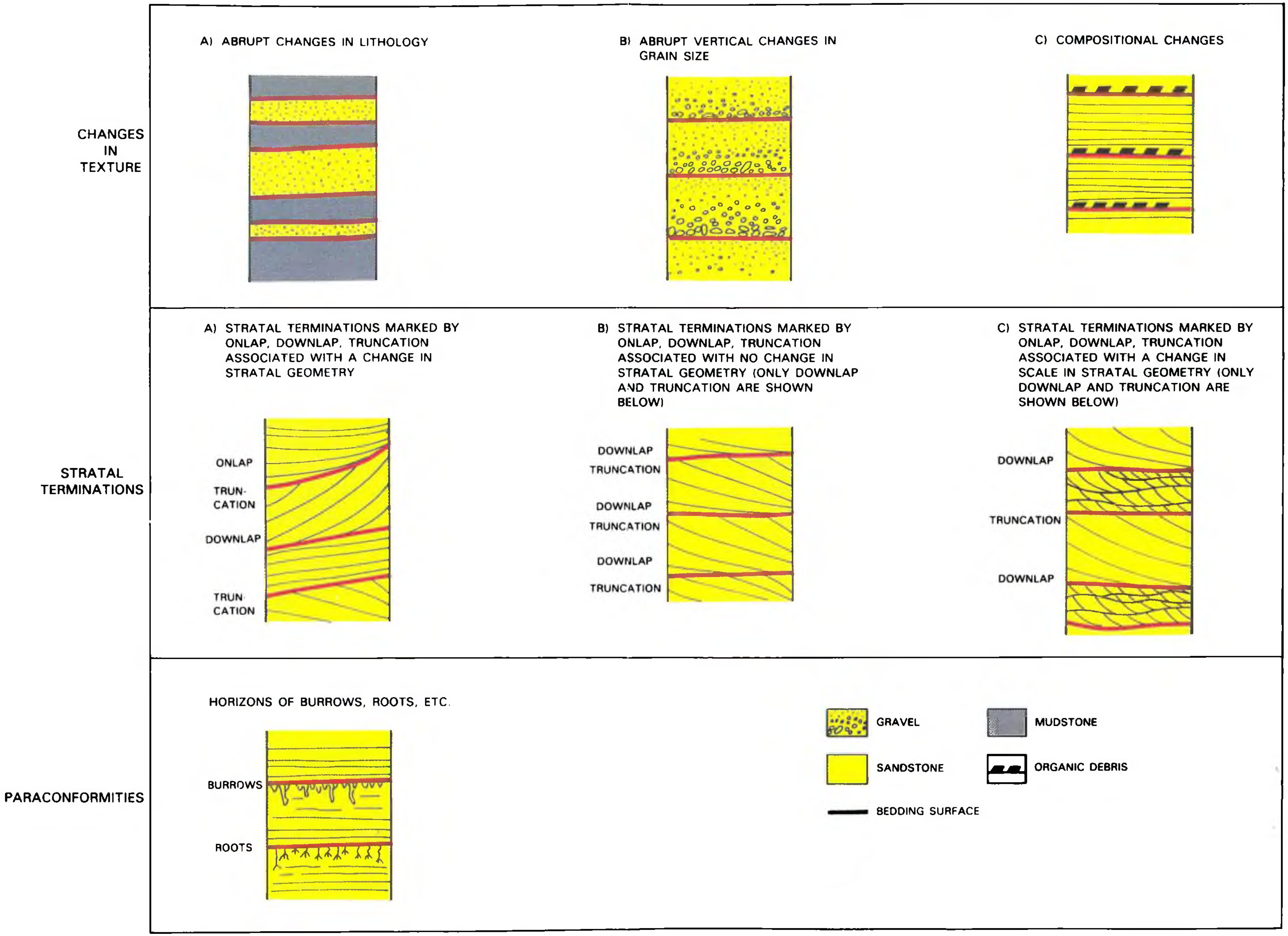


Figure 2—Criteria for picking bedding surfaces.

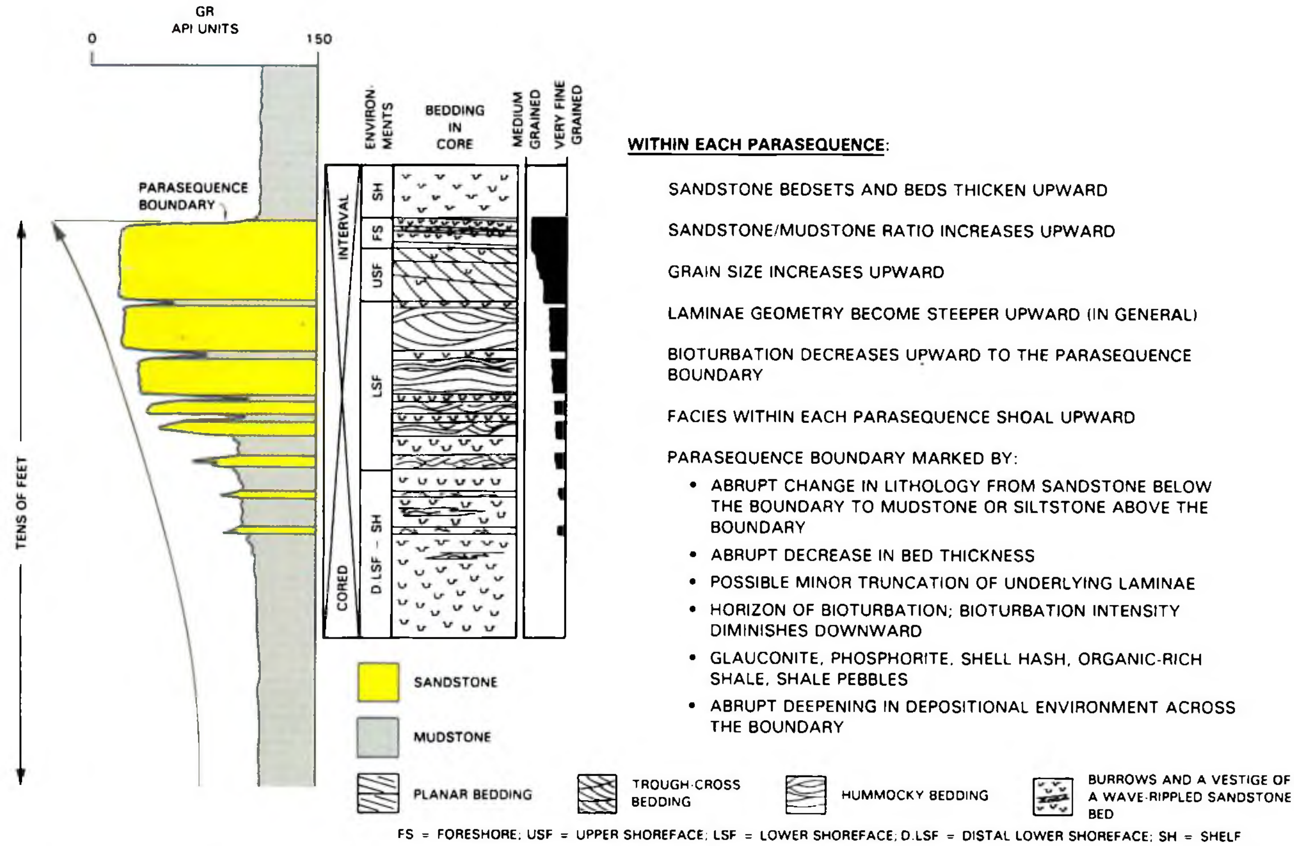


Figure 3A—Stratal characteristics of an upward-coarsening parasequence. This type of parasequence is interpreted to form in a beach environment on a sandy, wave- or fluvial-dominated shoreline.

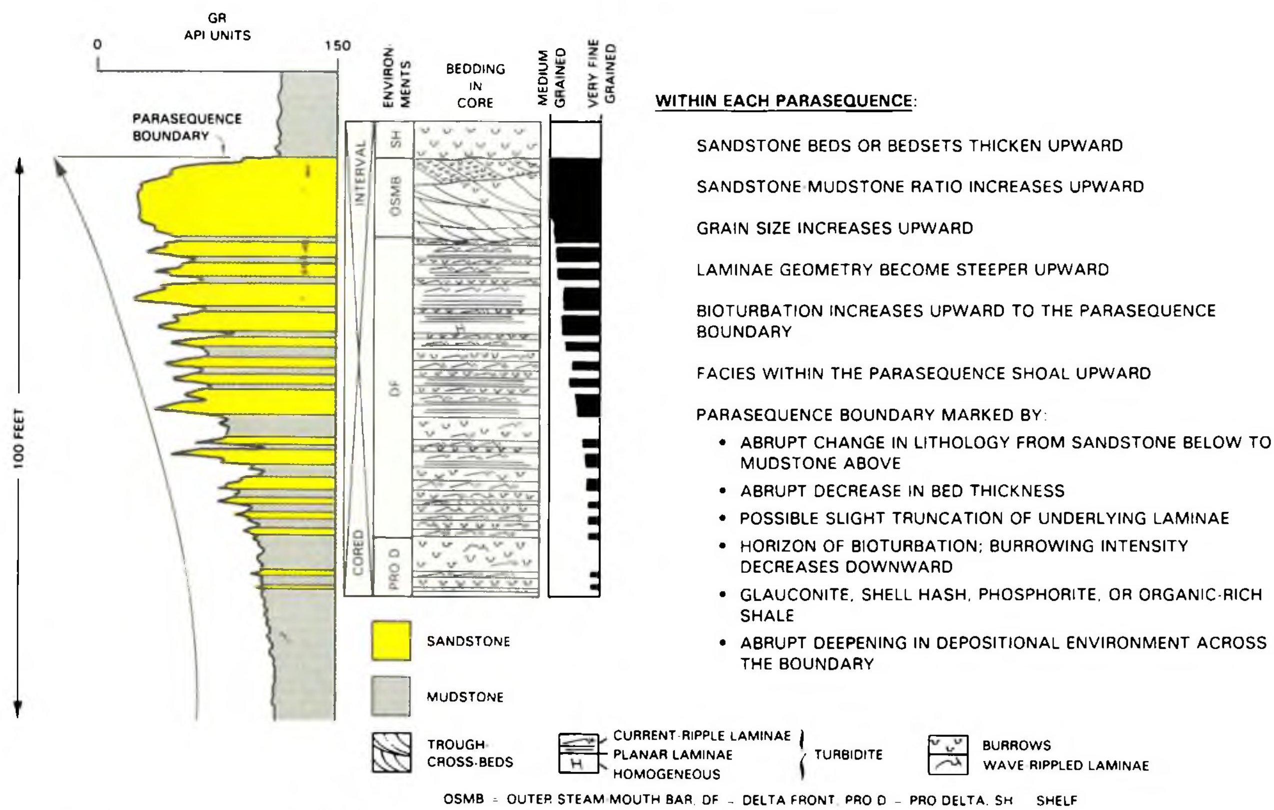


Figure 3B—Stratal characteristics of an upward-coarsening parasequence. This type of parasequence is interpreted to form in a deltaic environment on a sandy, fluvial- or wave-dominated shoreline.

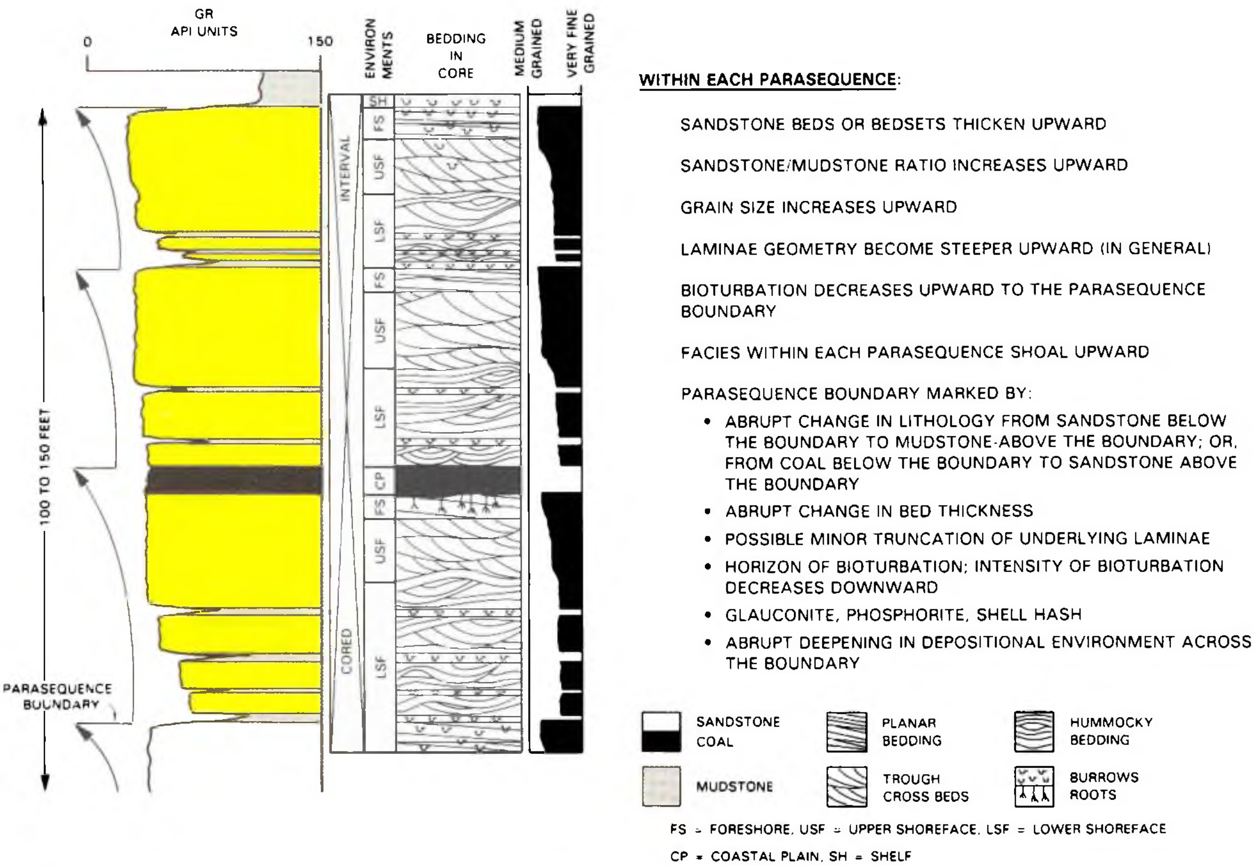


Figure 3C—Stratal characteristics of stacked upward-coarsening parasequences. These parasequences are interpreted to form in a beach environment on a sandy, wave- or fluvial-dominated shoreline where the rate of deposition equals the rate of accommodation.

are assumed to be rare in the rock record. Most "deepening-upward" facies associations probably are produced by a backstepping set of parasequences called a retrogradational parasequence set. In some environments, where siliciclastic deposition is condensed or the water depth is too great, lithologic variation may not be sufficient to recognize parasequences easily. In these sections, strata appear to deepen upward gradually, although careful observation may reveal subtle evidence of flooding surfaces marking parasequence boundaries.

The sources of sediment for the bedsets within parasequences are the river mouths at the shoreline. The parasequences fill the basin from the margin toward the center (Frazier, 1974) through the basinward migration of the shoreline accomplished by parasequence progradation. Sediment generally does not bypass the inner shelf to be deposited as parasequences on the middle to outer shelf except during times of relative falls in sea level. Exceptions to this will be found in physiographic settings similar to the modern coast of South Africa (Flemming, 1981), where strong ocean currents sweep sediment off the front of a delta and across the shelf in sand waves.

Parasequence Boundary

A parasequence boundary is a marine-flooding surface and its correlative surfaces. It is a planar surface of local to basinal extent and exhibits only minor topographic relief over large areas. A marine-flooding surface sharply separates deeper-water rocks, such as shelf mudstones, above, from shallower-water rocks, such as lower-shoreface sandstones, below. The flooding surface is characterized by minor submarine erosion and nondeposition, with a minor hiatus indicated. The amount of submarine erosion associated with a marine-flooding surface varies, but probably ranges from a few inches to tens of feet, with several feet being most common.

Few transgressive-lag deposits have been observed in cores or outcrops on marine-flooding surfaces which are not coincident with sequence boundaries. A *transgressive lag*, as we use it here, is defined as a sedimentary deposit, commonly less than 2 ft (0.61 m) thick, of relatively coarse-grained material composed of shells, shell fragments, clay rip-up clasts, calcareous nodules, siliciclastic gravel or pebbles. This material derives from underlying strata by shoreface erosion during a marine transgression, and is concentrated as

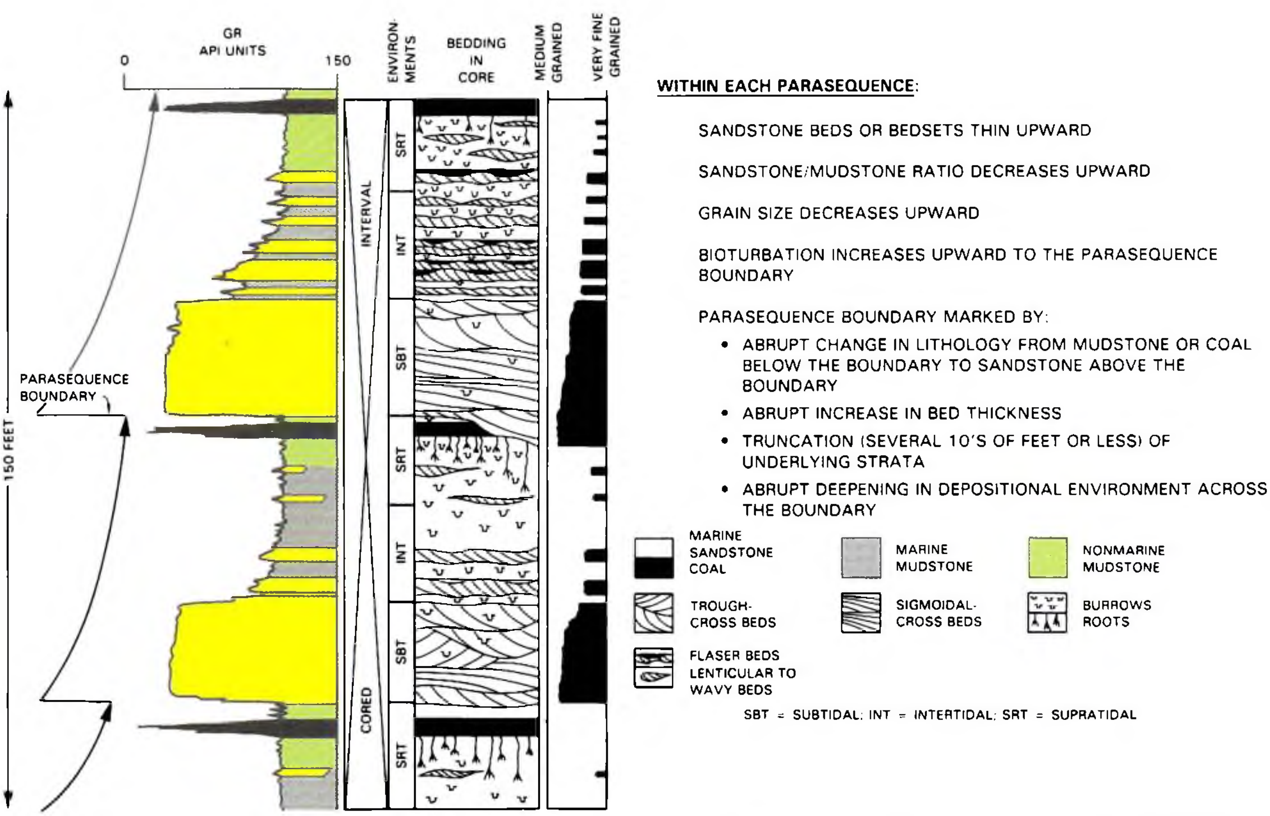


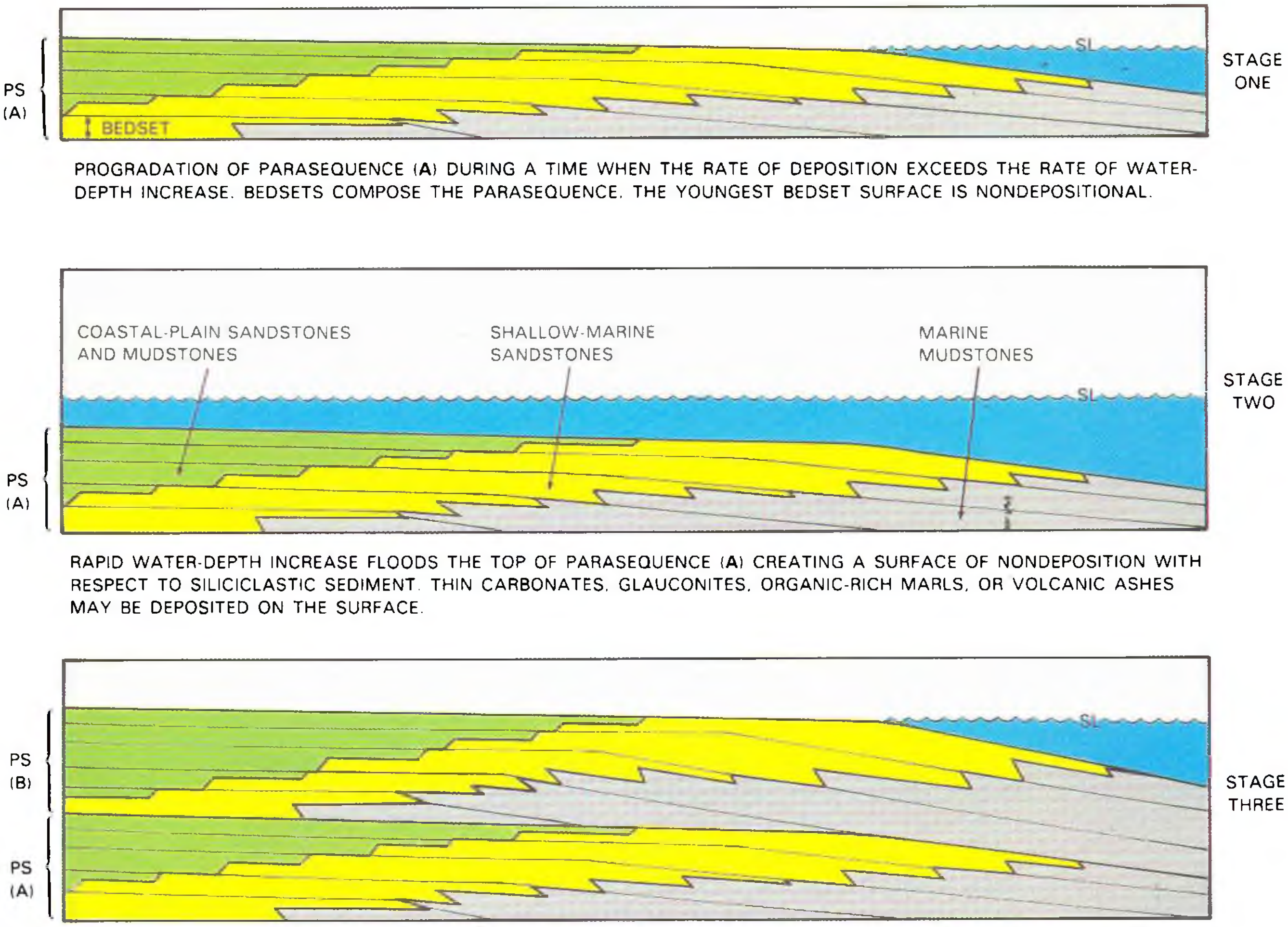
Figure 3D—Stratal characteristics of two upward-fining parasequences. These types of parasequences are interpreted to form in a tidal flat to subtidal environment on a muddy, tide-dominated shoreline.

a discrete bed on top of the transgressed surface, commonly on the inner to outer shelf. The majority of marine-flooding surfaces observed in cores or outcrops are marked by a sharp boundary free of concentrations of sedimentary particles such as the ones listed above. However, when a transgressive lag does occur on a marine-flooding surface, the lag commonly is clearly derived from the underlying strata; as, for example, in the case of a thin bed of siliciclastic pebbles at the top of a pebbly sandstone. More commonly, lag deposits are found on marine-flooding surfaces that are coincident with sequence boundaries. In this case, the lag does not have such a clear affinity to the underlying deposit. Four types of these lags are below; only the first type is a transgressive lag.

The first type of lag consists of discrete, irregularly shaped calcareous nodules up to 1 in. (2.54 cm) in diameter. The lag rests on a marine-flooding surface interpreted to be coincident with a sequence boundary at the base of an incised valley or in an interfluvial area. This lag derives from calcretes or discrete caliche nodules formed within a soil horizon during subaerial exposure of the sequence boundary. Subsequent transgression removes the relatively easily eroded soil and concentrates the nodules as lag on the transgressed surface. These nodules commonly are the only indication that a soil horizon existed, unless isolated remnants of the soil horizon can be found preserved in low areas on the transgressed shelf.

The second type of lag is caused by intense burrowing and wave or current reworking of the parasequence up to 5 ft (1.5 m) below the marine-flooding surface, which winnows out finer particles and concentrates coarser grains. This reworking is gradational down into the underlying strata; there is no surface separating the reworked deposits from the rest of the parasequence. This reworking is interpreted to form in response to storms and normal infaunal colonization subsequent to transgression, but before a significant amount of fine-grained sediment progrades over the flooding surface. In some places, the bioturbation and submarine exposure may lead to the formation of a firm ground. Commonly, this lag forms on a marine-flooding surface that is coincident with a sequence boundary, although this coincidence is not necessary for the lag to develop.

The third type of lag commonly recognized on marine-flooding surfaces occurs following a sea-level rise, but before an appreciable amount of finer-grained siliciclastic sediment can prograde over the shelf, allowing organic or inorganic carbonates to accumulate on the marine-flooding surface. Organic carbon-



PROGRADATION OF PARASEQUENCE (A) DURING A TIME WHEN THE RATE OF DEPOSITION EXCEEDS THE RATE OF WATER-DEPTH INCREASE. BEDSETS COMPOSE THE PARASEQUENCE. THE YOUNGEST BEDSET SURFACE IS NONDEPOSITIONAL.

RAPID WATER-DEPTH INCREASE FLOODS THE TOP OF PARASEQUENCE (A) CREATING A SURFACE OF NONDEPOSITION WITH RESPECT TO SILICICLASTIC SEDIMENT. THIN CARBONATES, GLAUCONITES, ORGANIC-RICH MARLS, OR VOLCANIC ASHES MAY BE DEPOSITED ON THE SURFACE.

PROGRADATION OF PARASEQUENCE (B) DURING A TIME WHEN THE RATE OF DEPOSITION EXCEEDS THE RATE OF WATER-DEPTH INCREASE. BEDSETS IN PARASEQUENCE (B) DOWNLAP ONTO THE BOUNDARY OF PARASEQUENCE (A). THERE IS AN ABRUPT DEEPENING OF FACIES ACROSS THE BOUNDARY OF PARASEQUENCE (A).

Figure 4—Progressive development of a parasequence boundary.

ates, in the form of shell beds, create widespread, tabular bodies up to 6 ft (1.8 m) thick that rest on the transgressive surface (Kidwell, 1989). Although these shell beds are winnowed and reworked by storms, they represent populations indigenous to the shelf; the shell beds do not derive from erosion of the underlying parasequence. In Miocene strata along the Calvert Cliffs of Maryland, these types of shell beds are interpreted to lie on marine-flooding surfaces coincident with sequence boundaries (Kidwell, 1989). Inorganic carbonates, in the form of oolites or pisolites, can form banks and bars on the marine-flooding surface, especially where it is coincident with a sequence boundary away from an incised valley. These grain types accumulate during a slow sea-level rise when the outer part of the shelf has been covered with shallow water following a sea-level lowstand. Wave agitation is sufficient to form the oolites or pisolites, and siliciclastic influx is minimal. Ultimately, when the continuing sea-level rise places the carbonate grains below the reach of wave agitation, the shoals cease to grow and may be reworked locally and distributed over the shelf by storms.

The fourth, and probably most common, type of lag is a channel lag lying on a sequence boundary at the base of an incised valley. This lag accumulates during a sea-level fall. The basal-channel lag consists of a variety of grain types but is most commonly composed of rounded chert, quartz, or quartzite pebbles ranging in thickness from thin lenses only one pebble thick to beds several feet thick. Because this lag forms during a sea-level fall, it is explained more fully in the "Incised Valley" discussion in the section on "Sequence Boundary Characteristics."

The marine-flooding surface has a correlative surface in the coastal plain and a correlative surface on the shelf. The correlative surface in the coastal plain is not marked by significant subaerial erosion, stream rejuvenation, downward shift in coastal onlap, or onlap of overlying strata; it may be marked by local erosion due to fluvial processes and local evidence of subaerial exposure such as soil or root horizons normally found in coastal-plain deposits. The correlative surface on the shelf is a conformable surface with no significant hiatus indicated and is marked by thin pelagic or hemipelagic deposits. These deposits include thin carbonates, organic-rich mudstones, glauconites, and volcanic ashes indicating terrigenous-sediment starvation. Strata across correlative surfaces usually do not indicate a change in water depth; commonly the correlative surfaces in the coastal plain or on the shelf can be identified only by correlating updip or downdip from a marine-flooding surface. In even deeper-water environments, such as the slope or basin floor, parasequence boundaries may also be unrecognizable.

The characteristics of parasequence boundaries suggest that they form in response to an abrupt increase in water depth that is sufficiently rapid to overcome deposition. The stages of parasequence-boundary formation are simplistically illustrated in Figure 4.

In two special cases, shown in Figure 5, parasequences may be bounded either above or below by sequence boundaries. In the first case (Figure 5, Example 1), a sequence boundary truncates a parasequence in the underlying transgressive systems tract and erodes into lower-shoreface sandstones (well A) and marine mudstones (well B). Subsequent deposition of a lowstand-shoreline parasequence on top of the sequence boundary results in (1) a younger parasequence bounded above by a marine-flooding surface and below by a sequence boundary, and (2) an older parasequence bounded below by a marine-flooding surface and above by an erosional sequence boundary. The correct parasequence interpretation in Example 1, based on recognition of the sequence boundary, is contrasted in Figure 5 with the incorrect parasequence interpretation that results if the sequence boundary is not identified.

In the second case (Figure 5, Example 2) the sequence boundary in well 2, expressed as a surface of subaerial exposure, coincides with a marine-flooding surface. This juxtaposition of surfaces results in a parasequence bounded above by a sequence boundary and below by a marine-flooding surface. There are three coincident surfaces at the top of the youngest parasequence in well 2: (1) the marine-flooding surface originally bounding the parasequence, probably formed at the end of the highstand, (2) the sequence boundary, expressed as a subaerial exposure surface, and (3) the last marine-flooding surface formed during the sea-level rise that terminated the lowstand.

Parasequence boundaries, within a framework of regional sequence boundaries, are the best surfaces to use for *local* correlation of time and facies from logs and cores, and as surfaces on which paleogeographic maps can be made, for several reasons. (1) Parasequence boundaries are easily recognizable surfaces that separate older beds from younger beds. (2) The boundaries form rapidly (similar observations have been made by other authors, notably Wilson, 1975; and Goodwin and Anderson, 1985), probably within hundreds of years to thousands of years, and approximate time markers useful for chronostratigraphy (Sears et al., 1941; Krumbein and Sloss, 1963; Wilson, 1975; Goodwin and Anderson, 1985). (3) Parasequence boundaries bound genetically related assemblages of facies, providing an essential framework for facies interpretation and correlation on well-log cross sections within the sequence. (4) Finally, they commonly are areally extensive enough for local subsurface correlation within a basin. However, parasequence boundaries usually cannot be easily correlated regionally with widely spaced well control. For this reason, and because parasequence distribution is very sensitive to sediment supply, parasequence boundaries usually are not good surfaces for regional correlation of time and facies.

Vertical Facies Relationships in Parasequences

Well-exposed outcrops in the Blackhawk Formation of east-central Utah were studied to document the vertical and lateral facies relationships in parasequences as a guide for subsurface correlation. These exposures also have been studied by Spieker (1949), Young (1955), Balsely and Horne (1980), Kamola and Howard (1985), and Swift et al. (1987). To relate outcrop observations of parasequences to subsurface expression, three wells were drilled on the outcrop by Exxon Production Research Company in 1982. Each well was cored and logged continuously with a suite of conventional electric- and nuclear-logging tools. The vertical facies relationships of parasequences from the Late Cretaceous age (Campanian) Blackhawk Formation are shown in Figure 6 in outcrop, core, and well log, the latter from one of the nearby 1982 wells. Each parasequence on the log is marked by an upward decrease in gamma-ray response, indicating an upward increase in the sandstone/mudstone ratio within the parasequence and generally an upward increase in the sandstone bed or bedset thickness. This vertical pattern of upward coarsening and thickening reflects parasequence progradation.

Each parasequence boundary in Figure 6 is marked by a blue line on the well log. The parasequence from interval A (160 to 218 ft, or 49 to 66 m) begins at the base with interbedded mudstones and burrowed, hummocky-bedded sandstones deposited in the lower shoreface of a beach. The upper part of the cored interval consists of trough and tabular cross-bedded sandstones and planar-laminated sandstones deposited, respectively, in the upper shoreface and fore-

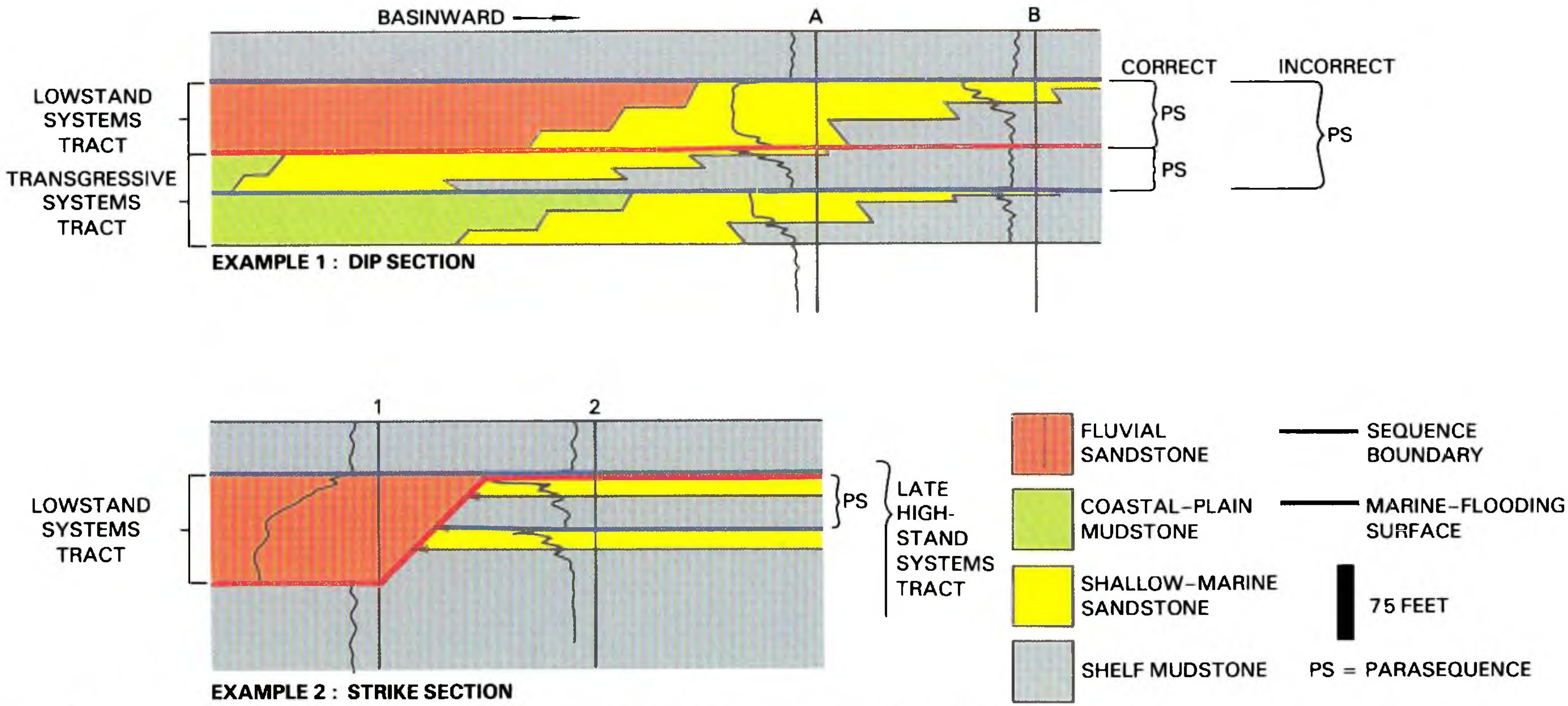


Figure 5—Two examples illustrating situations where parasequences are bounded above or below by sequence boundaries.
Example 1—If a sequence boundary occurs between two marine-flooding surfaces, as illustrated in well B, the parasequences are defined from sequence boundary to flooding surface (illustrated as "correct"), and not from flooding surface to flooding surface (illustrated as "incorrect").
Example 2—A marine-flooding surface bounding the youngest parasequence in the highstand systems tract is often coincident with a sequence boundary. In well 2 the parasequence boundary is coincident with a sequence boundary.

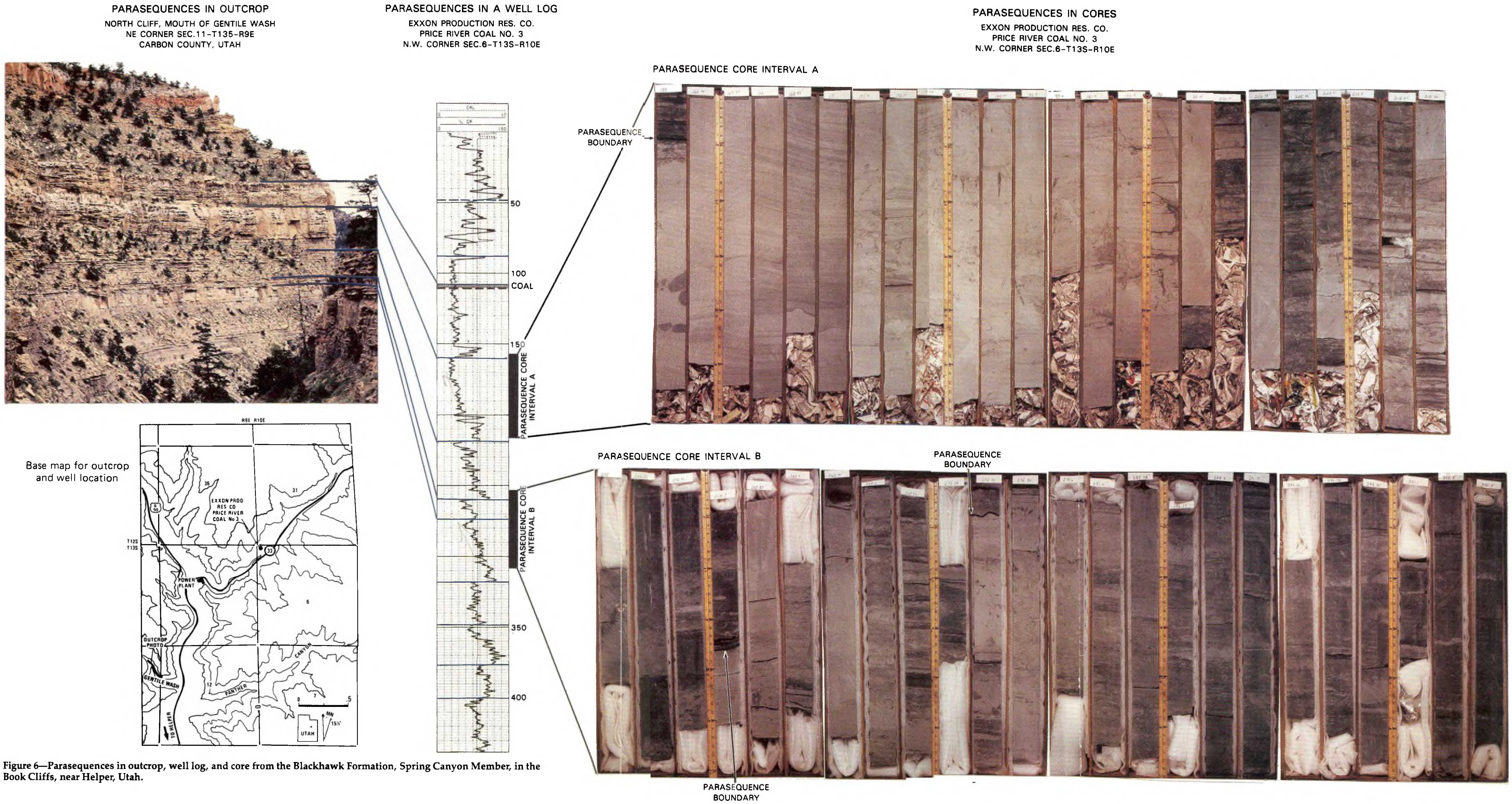


Figure 6—Parasequences in outcrop, well log, and core from the Blackhawk Formation, Spring Canyon Member, in the Book Cliffs, near Helper, Utah.

shore of the beach. The parasequence boundary occurs at 158.5 ft (48 m) near the top of the last core box. The boundary is marked by deeper-water, black, shelf mudstones lying sharply on burrow-churned, low-angle to planar-laminated sandstone beds with no intervening transgressive lag. In outcrop, this parasequence boundary can be traced approximately 15 mi (24 km) along depositional dip. This core was cut near the youngest, most basinward position of the foreshore in the parasequence.

Interval B (308 to 255 ft, or 94 to 78 m) contains two parasequences (Figure 6). The lower parasequence begins at the base of the core with burrowed, black mudstones and partially burrowed-churned, wave-rippled sandstones deposited on a shelf. This facies is overlain by burrowed, hummocky-bedded sandstones interbedded with thin, black mudstones deposited in the lower shoreface of a beach. The boundary for the lower parasequence is a burrowed surface at 274 ft (84 m) defined by black mudstones lying abruptly on hummocky-bedded sandstones. The upper parasequence in interval B begins at the base with burrowed, black, shelf mudstones and thin wave-rippled sandstones; like the lower parasequence, it is capped with burrowed, hummocky-bedded sandstones deposited in the lower shoreface of a beach. The parasequence boundary for the upper parasequence occurs at 260 ft (79 m). The boundary is a distinct surface defined by black mudstones in sharp contact with underlying burrowed and hummocky-bedded sandstones. In outcrop, these parasequence boundaries can be traced at least 12 mi (19 km) along depositional dip. As in interval A, both parasequence boundaries in interval B are devoid of transgressive lags in the core and outcrop.

The parasequence boundaries in intervals A and B are marine-flooding surfaces interpreted to result from an abrupt increase in water depth. This deepening is indicated by the facies contrasts across the parasequence boundaries. Vertical-facies associations within parasequences in intervals A and B do not exhibit any significant discontinuities and are interpreted to result from normal-shoreline progradation. Additional well-log responses of parasequences and parasequence boundaries in different regions and formations are shown in Figure 7.

Lateral Facies Relationships in Parasequences

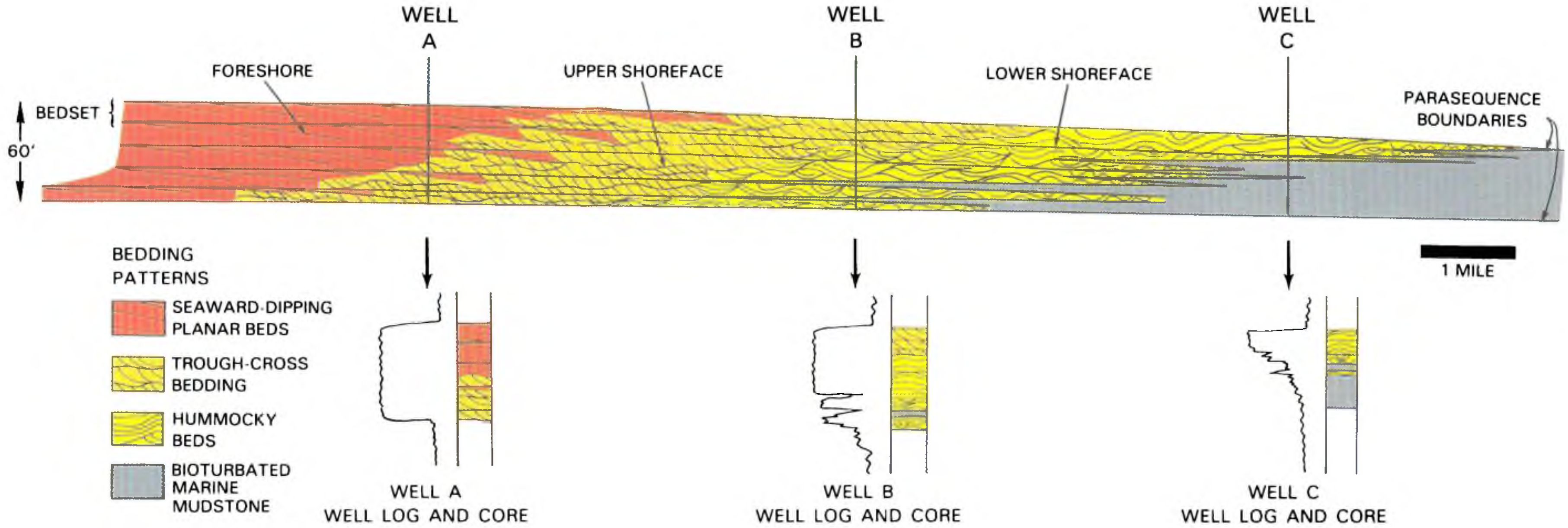
The lateral facies relationships, predicted rock types observed in cores, and well-log responses for a single parasequence interpreted to have been deposited in a beach environment are shown in Figure 8. Bedset surfaces are the throughgoing master surfaces that define the primary stratification within the parasequence. The facies changes within each bedset occur bed by bed. Because the types of facies changes occurring in each bedset within the parasequence are similar and there are no significant chronostratigraphic breaks between bedsets, a parasequence is considered to be a genetically related succession of beds and bedsets. Within a single bed in a beach parasequence (Figure 8), gently seaward-dipping, planar, parallel laminae of the foreshore change geometry basinward into more steeply dipping foreset laminae within trough-cross beds of the upper shoreface. These foreshore and upper-shoreface rocks compose the potential hydrocarbon reservoir in the parasequence. The trough-cross beds grade seaward bed-by-bed into hummocky beds of the lower shoreface. Finally, the same bedset deposited in the lower shoreface can be traced seaward to a point where the sandstone bedset thins to a few inches and may be so churned by burrowing organisms that its boundaries become indistinct. Campbell (1979) has documented similar lateral facies changes in the beach deposits of the Gallup Sandstone.

In the landward direction, foreshore and upper-shoreface bedsets within a parasequence either abruptly change facies into washover fans, which in turn change facies into coastal-plain mudstones and thin sandstones, or are truncated by tidal inlets. Because of progradation, the entire vertical succession of strata composing the parasequence is rarely complete at any point in the parasequence, as shown by the schematic well-log and core profiles in Figure 8.

Parasequences terminate in a landward direction by onlap onto a sequence boundary, by local fluvial-channel erosion in the updip coastal or alluvial plain, or by widespread fluvial incision associated with a sequence boundary. Parasequences lose their identity basinward by thinning, shaling out, and downlapping accompanied by stratal thinning onto an older parasequence, parasequence set, or sequence boundary. Shoreline parasequences often can be correlated on well-log cross sections for tens of miles into the basin before the parasequence boundaries become unrecognizable as flooding surfaces.

Interpretation of Depositional Mechanisms

Shallow-marine parasequences form when the rate of sedimentation in deltaic, beach, or tidal-flat environments is greater than the rate of accommodation along the coastline. Accommodation is defined as the new space available for sedimentation and is interpreted to be a function of eustasy and subsidence (Jervey, 1988; Posamentier et al., 1988). Parasequence boundaries are interpreted to form when the rate of sediment supply at the shoreline is less than the rate of accommodation. Under these conditions, the shoreline normally retreats rapidly and very little marine sediment is preserved in the stratigraphic record; commonly a marine-flooding surface is the only indication



STRATAL CHARACTERISTICS

1. BOUNDED BY FLAT, PARALLEL, SLIGHTLY SEAWARD-DIPPING PARASEQUENCE BOUNDARIES OF NONDEPOSITION OR SLIGHT EROSION.

2. INTERNALLY COMPOSED OF SEAWARD-DIPPING, IMBRICATE BEDSETS BOUNDED BY BEDSET SURFACES.

3. BEDDING PATTERN CHANGES SEAWARD IN A PREDICTABLE ORDER WITHIN EACH BEDSET.

4. VERTICAL SUCCESSION OF STRATA IS USUALLY INCOMPLETE AT ANY ONE POINT IN THE PARASEQUENCE.

5. THE PARASEQUENCE TERMINATES UPDIP BY
   - ONLAP ONTO A SEQUENCE BOUNDARY
   - LOCAL EROSION BY FLUVIAL CHANNELS OR BY WIDESPREAD FLUVIAL INCISION.

6. THE PARASEQUENCE TERMINATES DOWNDIP BY THINNING, SHALING OUT, AND DOWNLAPPING ONTO AN UNDERLYING PARASEQUENCE, PARASEQUENCE SET, OR SEQUENCE BOUNDARY.

Figure 8—Lateral facies relationships and postulated core and well-log responses for a beach parasequence.

that the rate of accommodation exceeded the rate of sediment supply.

Three different mechanisms can generate parasequence boundaries. One well-documented mechanism is the relatively rapid increase in water depth caused by compaction of prodelta mudstones in a delta lobe following distributary-channel avulsion (Frazier, 1967). The drowning of the lobe produces an abrupt, planar, slightly erosional surface, commonly with little or no preserved transgressive lag lying above it (Elliott, 1974). The resulting parasequence boundary has a lateral extent equivalent to the areal extent of the lobe itself. Frazier and Osanik (1967) showed that the three youngest lobes in the Holocene San Bernard delta in southeastern Louisiana have areal extents ranging from 300 to 3000 mi^2 (777 to 7770 km^2). The rates for lobe progradation range from 800 to 1400 years. Because the surfaces bounding each of these lobes are extensive areally and formed rapidly, they provide local time lines for chronostratigraphic and lithostratigraphic analysis over relatively large areas in the subsurface.

A second mechanism for the formation of a parasequence boundary is a rapid relative rise in sea level caused by subsidence along tectonically active faults. Earthquakes such as the 1964 earthquake in Alaska (Plafker, 1965) or the 1960 earthquake in Chile (Plafker and Savage, 1970) produced nearly instantaneous, maximum coastal subsidence of 6.5 and 9 ft (2 and 3 m), respectively. Plafker and Savage (1970) document a zone of subsidence 600 mi (963 km) long and 70 mi (112 km) wide along the Chilean coastline. Along low-lying shorelines, such subsidence could drown large areas of coastal deposits rapidly, thereby producing a parasequence boundary. Short-term increases in the rate of subsidence on the order of a few thousand years near coastal salt domes or growth faults also could produce local relative rises in sea level sufficient to drown coastal deposits and produce parasequence boundaries.

A third mechanism for parasequence boundary formation is eustasy. The relationship of eustasy and subsidence to parasequence and sequence deposition is presented in Figure 39 and is discussed later, in "Interpretations of Depositional Mechanisms" within the "Sequence" section.

PARASEQUENCE SET

Definition

A *parasequence set* is a succession of genetically related parasequences forming a distinctive stacking pattern bounded by major marine-flooding surfaces and their correlative surfaces. Parasequence set characteristics are summarized in Table 1.

Parasequence Set Boundary

Like parasequence boundaries, parasequence set boundaries are marine-flooding surfaces and their correlative surfaces. Figure 9 shows a parasequence set boundary with hummocky-bedded and burrowed, lower-shoreface sandstones lying in sharp contact on coastal-plain coals. Parasequence set boundaries (1) separate distinctive parasequence-stacking patterns, (2) may coincide with sequence boundaries, and (3) may be downlap surfaces and boundaries of systems tracts.

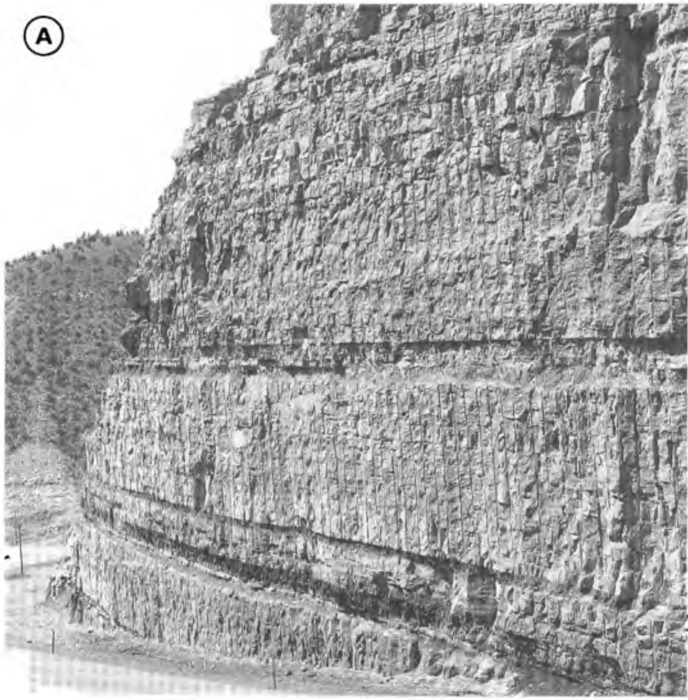
Types of Parasequence Sets

Stacking patterns of parasequences within parasequence sets are progradational, retrogradational, or aggradational (Van Wagoner, 1985), depending on the ratio of depositional rates to accommodation rates. Figure 10 schematically illustrates these stacking patterns and their well-log responses. In a *progradational parasequence set*, successively younger parasequences are deposited farther basinward; overall, the rate of deposition is greater than the rate of accommodation. In a *retrogradational parasequence set*, successively younger parasequences are deposited farther landward, in a backstepping pattern; overall, the rate of deposition is less than the rate of accommodation. Although each parasequence in a retrogradational parasequence set progrades, the parasequence set deepens upward in a "transgressive pattern." We use the term "retrogradation" in the dictionary sense (Gary et al., 1972) to mean "the backward (landward) movement or retreat of a shoreline or coastline." As Gary et al. (1972) pointed out, retrogradation is the antonym of progradation. In an *aggradational parasequence set*, successively younger parasequences are deposited above one another with no significant lateral shifts; overall, the rate of accommodation approximates the rate of deposition.

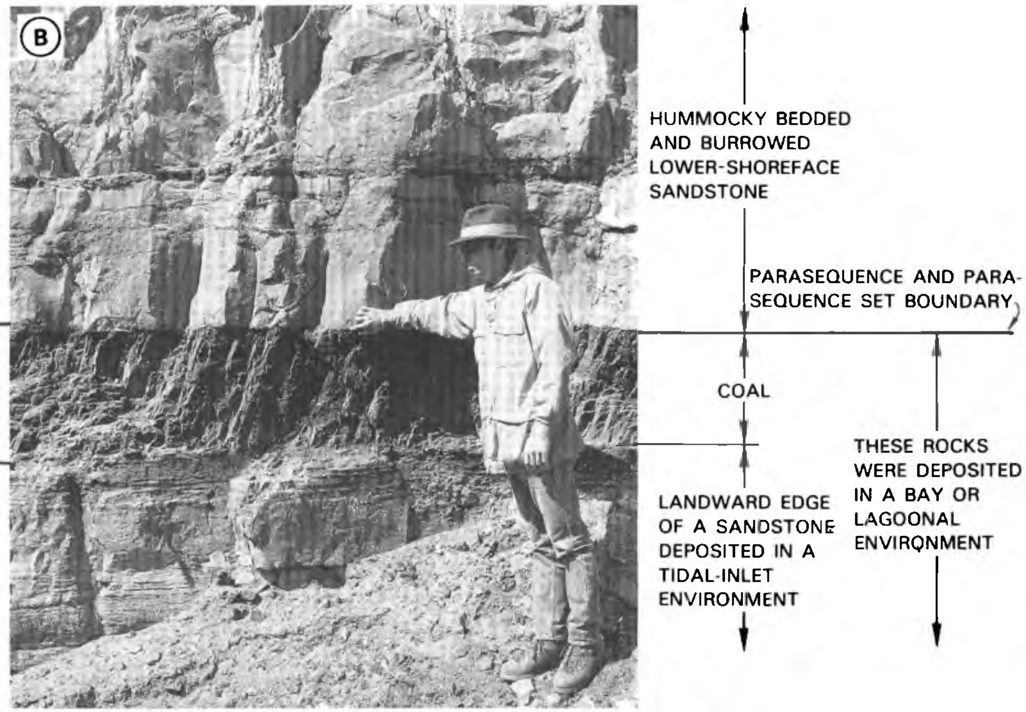
Vertical Facies Relationships in Parasequence Sets

Parasequence sets can be identified from a single well log. In a progradational parasequence set (Figure 11), successively younger parasequences contain sandstone with greater depositional porosities and higher percentages of rocks deposited in shallow-marine to coastal-plain environments than underlying parasequences. The youngest parasequence in the well may consist entirely of rocks that were deposited in a coastal-plain environment. In addition, younger parasequences tend to be thicker than older parasequences in the set.

In a retrogradational parasequence set (Figure 11), successively younger parasequences contain more shale or mudstone and higher percentages of rocks deposited in deeper-water marine environments, such as lower shoreface, delta front, or shelf, than



A. ROADCUT ALONG HIGHWAY 6, 3.3 MILES NORTH OF HELPER, UTAH. THE YOUNGEST COAL CAPS THE SPRING CANYON MEMBER AND IS OVERLAIN BY THE ABERDEEN MEMBER, BOTH OF THE BLACKHAWK FORMATION. THESE ROCKS ARE UPPER CRETACEOUS (CAMPANIAN) IN AGE. THE PARASEQUENCE SET BOUNDARY OCCURS ON TOP OF THE COAL (SEE PHOTO B FOR DETAILS).



B. THE FINGER OF THE GEOLOGIST IS ON A SURFACE THAT IS BOTH A PARASEQUENCE AND A PARASEQUENCE SET BOUNDARY. THE BOUNDARY IS SHARPLY DEFINED BY COALS BELOW AND BIOTURBATED, VERY FINE-GRAINED, HUMMOCKY-BEDDED SANDSTONES ABOVE. THE BOUNDARY IS REMARKABLY FLAT WITH NO EVIDENCE OF TRUNCATION, EROSIONAL RELIEF, OR TRANSGRESSIVE LAG. THIS SURFACE IS TYPICAL OF PARASEQUENCE AND PARASEQUENCE SET BOUNDARIES. THE BOUNDARY IS INTERPRETED TO BE A FLOODING SURFACE. THE INTERPRETATIONS OF THE DEPOSITIONAL ENVIRONMENTS ARE SHOWN TO THE RIGHT OF PHOTO B.

Figure 9—Outcrop expression of a parasequence set boundary.

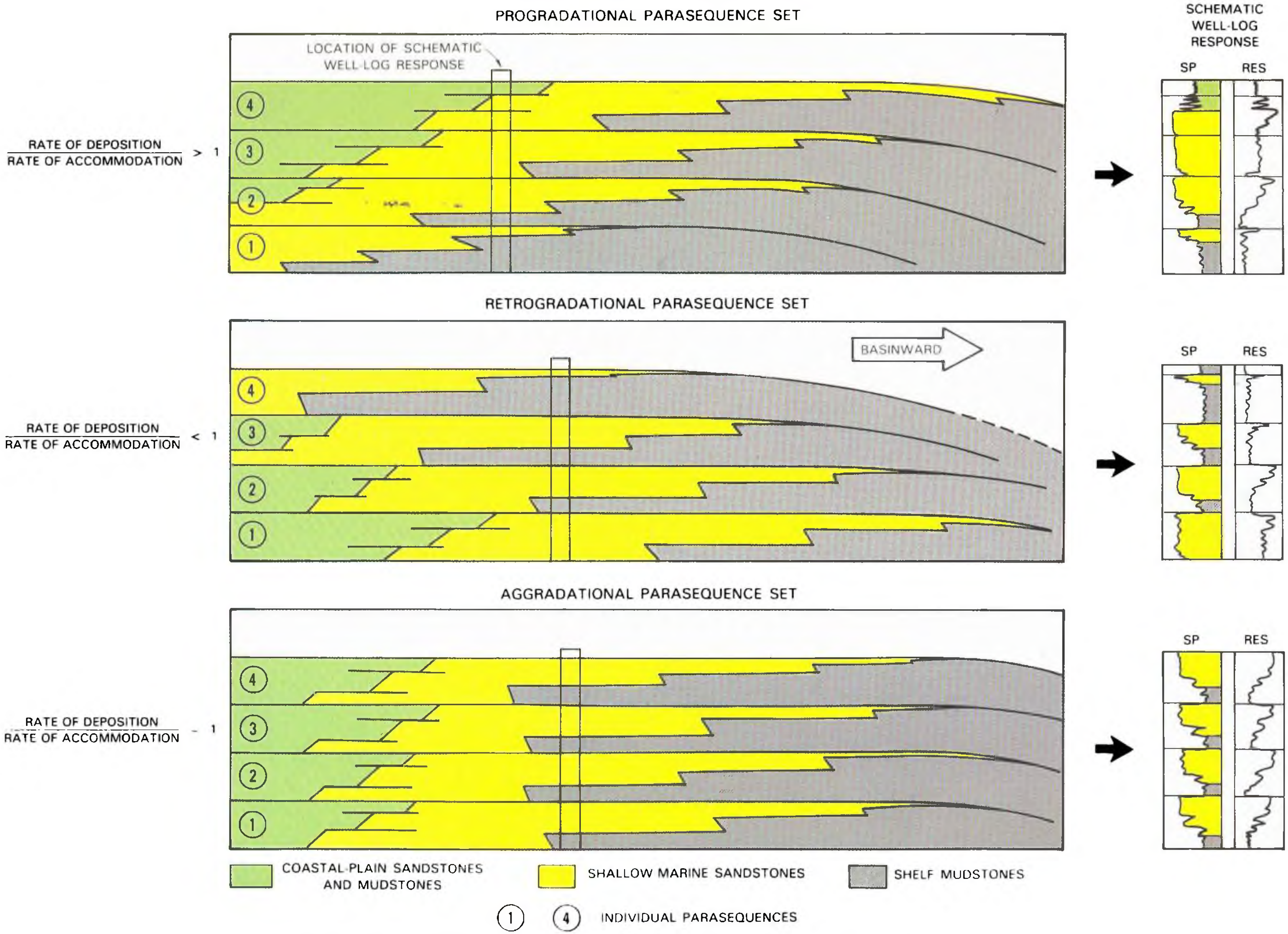


Figure 10—Parasequence-stacking patterns in parasequence sets; cross-section and well-log expression.

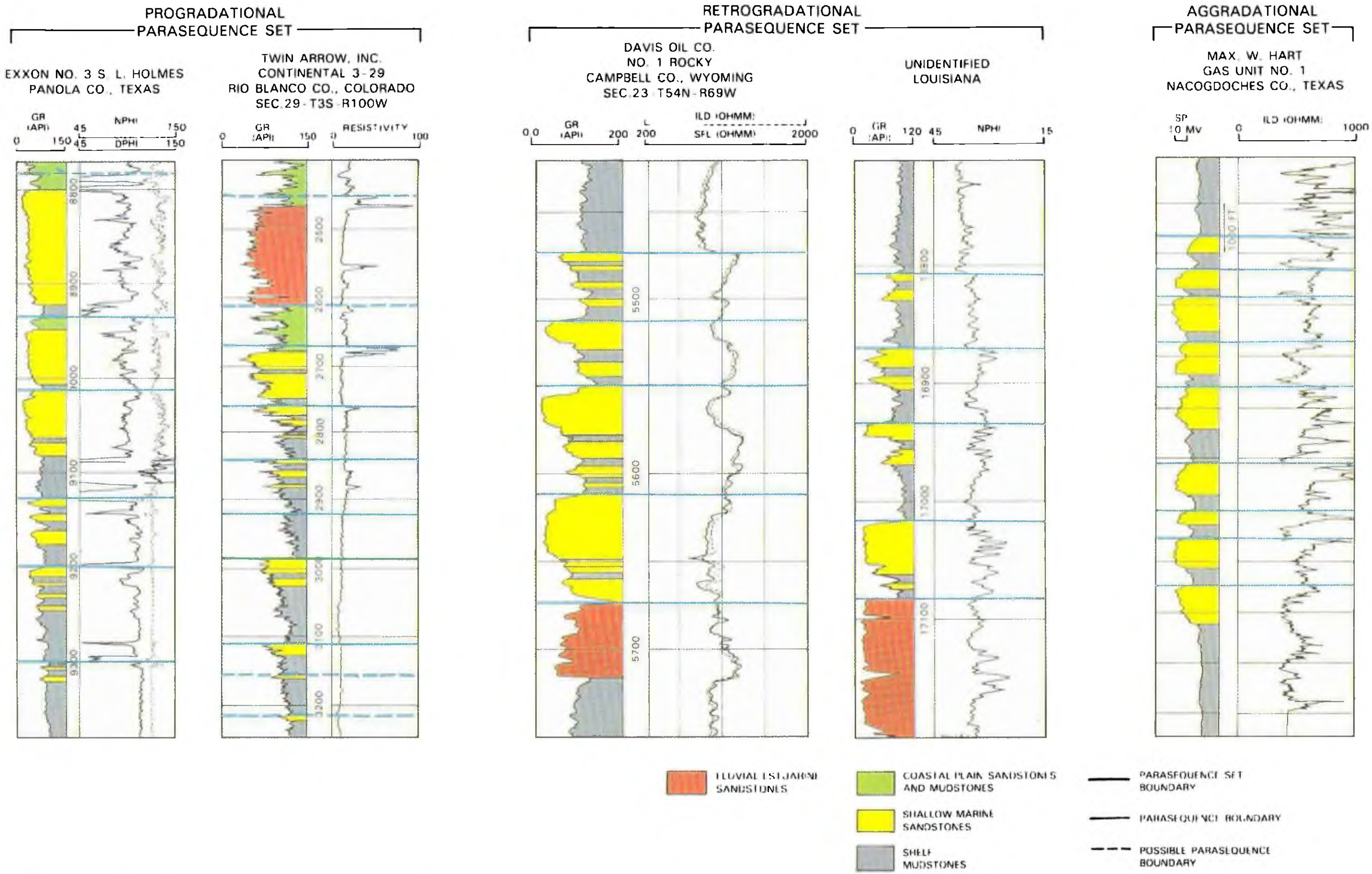


Figure 11—Well-log response of parasequence sets.

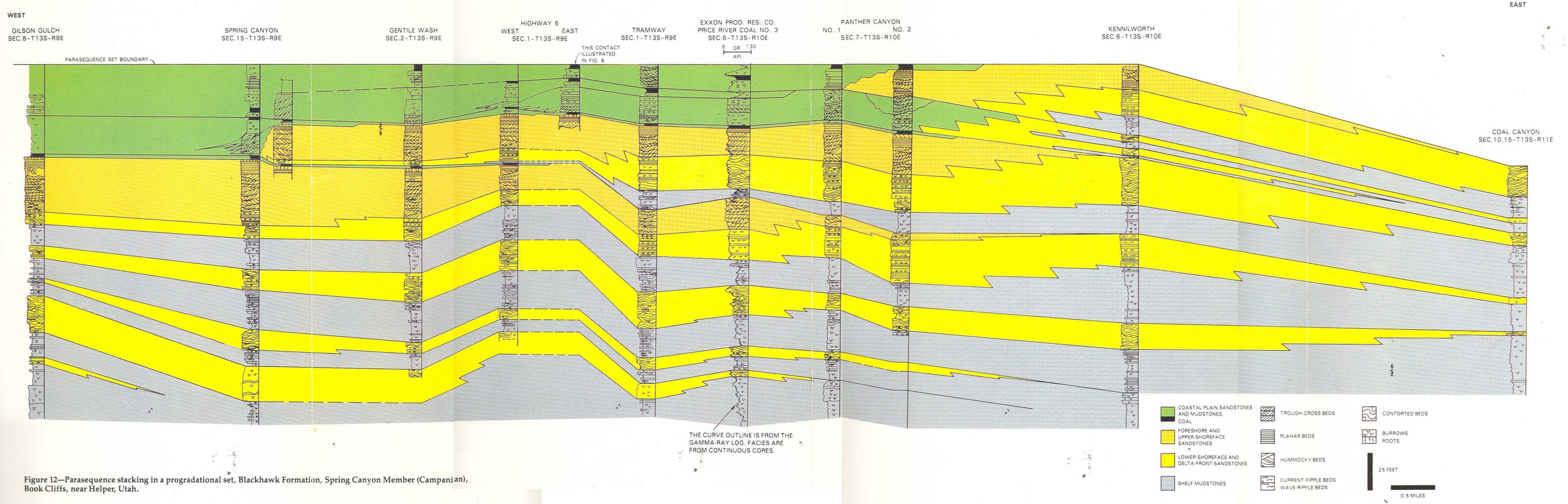


Figure 12—Parasequence stacking in a progradational set, Blackhawk Formation, Spring Canyon Member (Campanian), Book Cliffs, near Helper, Utah.

Figure 13—Rapid updip facies changes within a beach parasequence. This example is from a progradational parasequence set, Book Cliffs, near Helper, Utah. Figure 12 is a cross section through this progradational parasequence set. This updip facies change occurs at the Spring Canyon location on the cross section (Figure 12).

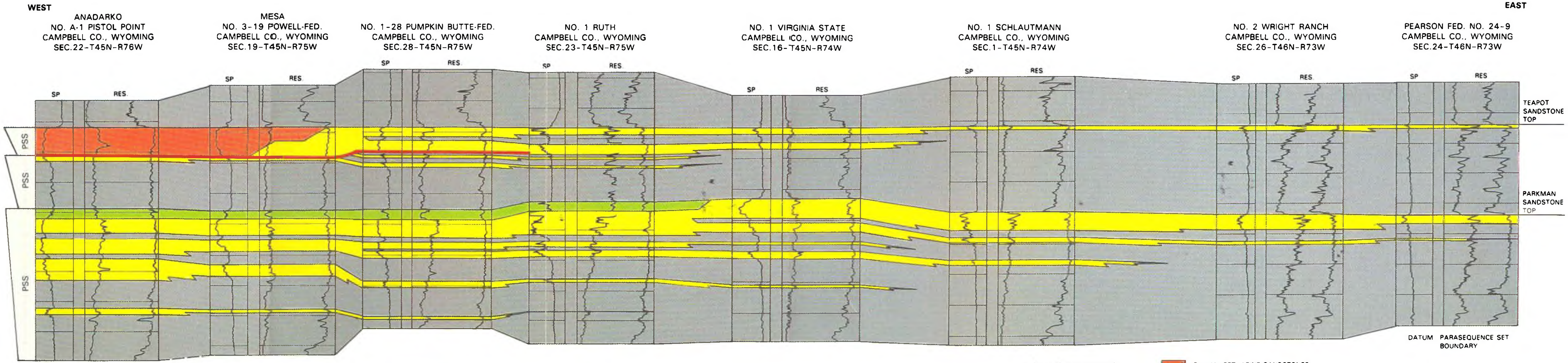


Figure 14—Parasequence stacking in progradational parasequence sets. Three parasequence sets are shown; one parasequence set is in the Parkman Sandstone, Mesaverde Formation (Campanian), two parasequence sets are in the Teapot Sandstone (Campanian). The Teapot Sandstone in the three western wells rests on an unconformity (sequence boundary). These wells are from the Powder River basin, Wyoming.

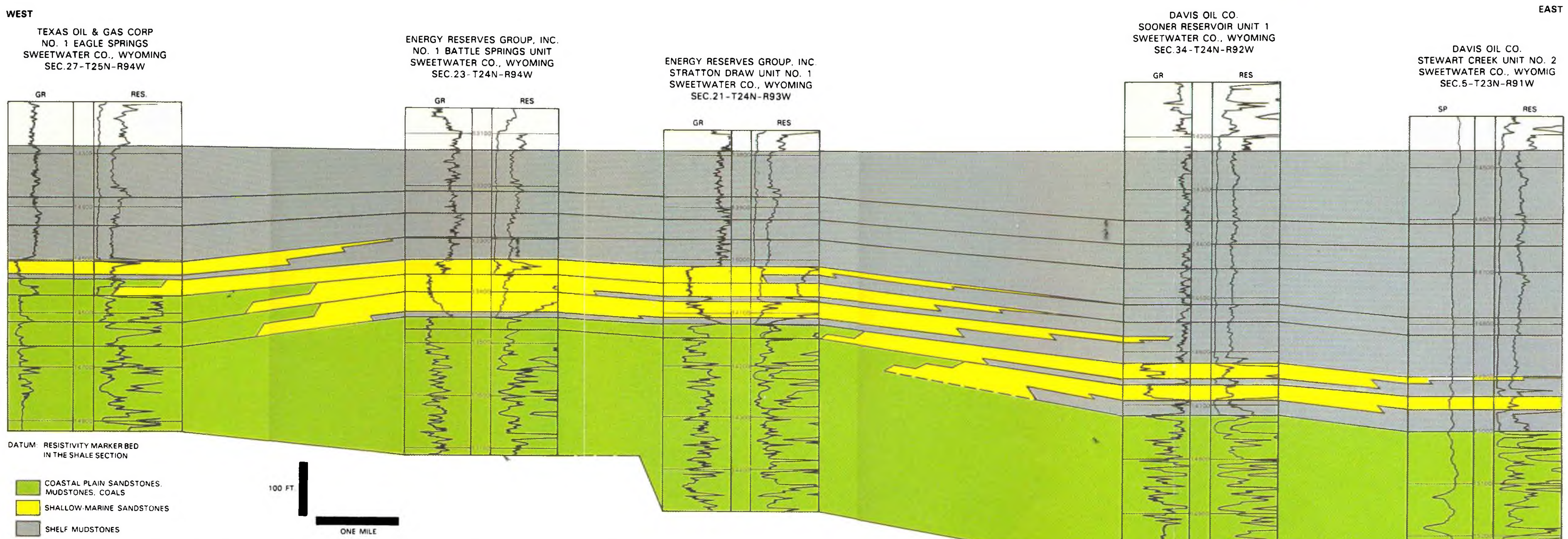


Figure 15—Parasequence stacking in a retrogradational parasequence set, Almond Sandstone, Late Cretaceous age, Washakie basin, Wyoming.

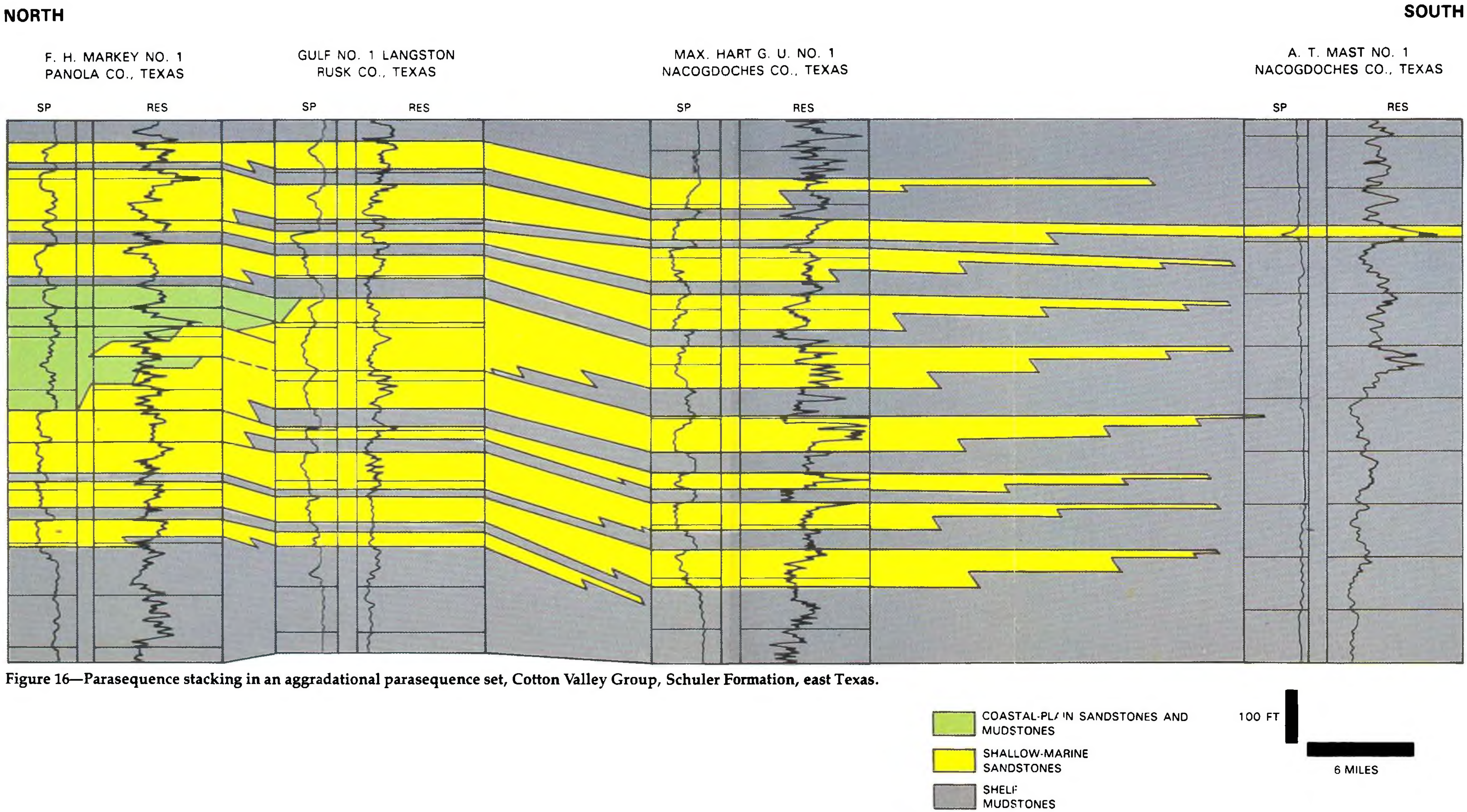


Figure 16—Parasequence stacking in an aggradational parasequence set, Cotton Valley Group, Schuler Formation, east Texas.

underlying parasequences. The youngest parasequence in the set commonly is composed entirely of rocks deposited on the shelf. In addition, younger parasequences tend to be thinner than older parasequences in the set.

In an aggradational parasequence set (Figure 11) the facies, thicknesses, and sandstone to mudstone ratios do not change significantly.

Lateral Facies Relationships in Parasequence Sets

The vertical expressions of different kinds of parasequence sets in single well logs (Figure 11) also have characteristic lateral expressions on cross sections. These subsurface and outcrop cross-section expressions are illustrated with four examples, shown in Figures 12 through 16.

In the first example, the dip-oriented distribution of parasequences in a progradational parasequence set from the Late Cretaceous age (Campanian) Blackhawk Formation exposed in the Book Cliffs, north of Price, Utah, is shown in Figure 12. A gamma-ray curve from the Exxon Production Research Company Price River Coal No. 3 well is included on the cross section. Facies data plotted on the gamma-ray curve are from continuous 3-in. (7.62-cm) cores recovered from that well (Figure 6). Successively younger parasequences step farther basinward, and produce the well-log patterns shown in Figure 11 for a progradational parasequence set. The updip pinchouts of porous marine sandstones into nonporous, coastal-plain mudstones also step basinward in successively younger parasequences. The highest depositional porosities in each parasequence are preserved just seaward of the pinchout of marine rocks into coastal-plain deposits. These pinchouts are very abrupt commonly occurring laterally over a distance of less than 100 ft (30 m). Such pinchouts can lead to confusion in well-log correlations because of the abrupt change in log shape between two closely spaced wells. One of the updip pinchouts within a parasequence is shown in Figure 13. In this example, foreshore and upper-shoreface sandstones are truncated updip by a landward-dipping erosional surface, interpreted to be cut by a migrating tidal inlet. A wedge-shaped sandstone body with a maximum thickness of 15 ft (4.6 m), consisting of imbricate, landward-dipping bedsets, and interpreted as a flood-tidal delta, rests on the erosional surface. The pinchout (Figure 13) occurs on the cross section (Figure 12) between Gentile Wash (Sec. 2, T13S, R9E) and Spring Canyon (Sec. 15, T12S, R9E).

In the second example, the dip-oriented distribution of parasequences in three progradational parasequence sets from the subsurface Parkman and Teapot sandstones of the Late Cretaceous age (Campanian) Mesaverde Formation, Powder River basin, Wyoming, is shown in Figure 14. In the Parkman parasequence set, successively younger parasequences step father basinward to the east, producing the well-log pattern characteristic of a progradational parasequence set (Figure 11). Only the marine-flooding surfaces on top of each parasequence are carried on the cross section; their seaward correlative surfaces are not indicated. The parasequence set is terminated by an abrupt increase in water depth that flooded across the top of the parasequence set and superimposed deeper-water marine mudstones on top of shallow-marine sandstones. Well-log correlation indicated that the top of the parasequence set is a planar surface. Each parasequence within the set is bounded by a minor marine-flooding surface; the parasequence set is bounded by the major marine-flooding surface that terminates the underlying stacking pattern.

The Teapot Sandstone (Figure 14) is composed of two progradational parasequence sets. The lower parasequence set is terminated by a sequence boundary marked by truncation of the underlying parasequences and a slight basinward shift in facies. A second progradational parasequence set, composed of two parasequences, rests on top of the sequence boundary. A marine-flooding surface separates deeper-water mudstones above the parasequence set boundary from shallow-marine sandstones below the boundary.

In the third example, the dip-oriented distribution of parasequences in a retrogradational parasequence set from the Late Cretaceous age (Campanian) Almond Formation and Ericson Sandstone, Mesaverde Group, Washakie basin, Wyoming, is shown in Figure 15. Successively younger parasequences step farther landward, producing the well-log pattern characteristic of a retrogradational parasequence set (Figure 11). Sandstones deposited in nearshore, shallow-marine environments compose the bulk of the middle part of the cross section, where porosities are best developed. Updip pinchouts of porous marine sandstones into nonporous, coastal-plain mudstones step landward with time.

In the fourth example, the dip-oriented distribution of parasequences in an aggradational parasequence set from the Late Jurassic or Early Cretaceous age, Cotton Valley Group, East Texas basin, is shown in Figure 16. Parasequences stack vertically with little or no lateral shift in facies, producing the well-log pattern characteristic of an aggradational parasequence set (Figure 11). A significant vertical thickness of porous sandstones may develop where this stacking occurs. In an updip position, or just seaward of the updip pinchouts of marine sandstones into coastal-plain mudstones, the porous sandstones may stack, with little or no intervening nonporous sandstone or mudstone, to form a potentially thick reservoir facies with good vertical continuity. In an intermediate position, the porous sandstones will be separated by shelf mudstones or thin beds of nonporous sandstones.

Correlation Concepts

Parasequence and parasequence set correlations commonly yield results that differ significantly from those obtained by conventional lithostratigraphic correlations that rely on formations, or formation "tops" of sandstone or mudstone intervals. To illustrate some of these differences, schematic cross sections through a progradational parasequence set and a retrogradational parasequence set are compared with typical lithostratigraphic correlations (Figures 17 and 18).

The progradational parasequence set cross section in Figure 17 was constructed using the parasequence set boundary as a datum. The shallow-marine and coastal-plain rocks of each younger parasequence step upward and basinward. The shallow-marine sandstones are potential reservoirs. Many are isolated above and below in mudstones, ensuring poor vertical communication and possibly separate oil-water contacts. Because of amalgamation of shoreline sandstones, some of the potential reservoirs have good vertical communication near the updip pinchouts of marine rocks into coastal-plain rocks.

The lithostratigraphic cross section in Figure 17 was constructed using the tops of the shallow-marine sandstones as a datum because this boundary (1) commonly is the site of coal deposition providing a good log marker, (2) is the most conspicuous boundary on the SP or gamma-ray log, and (3) provides a similar resistivity response on each log inasmuch as the facies, porosities, and fluids in each massive, shallow-marine sandstone are similar. If this datum is selected, as is commonly done, and the lithofacies are correlated by connecting the sandstone tops, the continuity of the reservoir is exaggerated, genetically different sandstones are linked together, and potential shallow-marine sandstone reservoirs are interpreted to change facies updip into marine shales and mudstones.

The retrogradational parasequence set cross section in Figure 18 was constructed using a parasequence set boundary as a datum. This boundary can be traced basinward into a diagnostic resistivity marker bed in the shale. The marine rocks in successively younger parasequences step landward or backstep. Each parasequence progrades and each shallow-marine sandstone changes facies updip into coastal-plain rocks. The shallow-marine sandstone reservoirs are isolated above and below in marine mudstones and commonly have separate oil-water contacts.

The lithostratigraphic cross section in Figure 18 was constructed using the top of the youngest, significant shallow-marine sandstone in each well as a datum. This horizon is a distinct lithologic break. It has a similar appearance in all of the wells and is easy to identify on the logs because it commonly is marked by an abrupt resistivity change. Correlating the logs using this surface leads to an interpretation of a continuous, relatively thin, shallow-marine sandstone. The continuity is exaggerated, and potential reservoir sandstones are incorrectly linked into the same sandstone body with an interpreted common oil-water contact. When production data suggest that there are at least two oil-water contacts in this reservoir, the geologist commonly adds a fault to explain the discrepancy between production data and the stratigraphic interpretation. Benthonic fauna usually are preserved in the shales just above the sandstone. Using the first occurrence of the benthonic foraminifera as a correlation tool results in the same correlation arrived at by using the sandstone tops, because these organisms are facies controlled.

SEQUENCE

Definitions

Sequence: A relatively conformable succession of genetically related strata bounded by unconformities or their correlative conformities (Mitchum, 1977). Parasequences and parasequence sets are the stratal building blocks of the sequence. Sequence characteristics are summarized in Table 1.

Unconformity: A surface separating younger from older strata along which there is evidence of subaerial-erosional truncation and, in some areas, correlative submarine erosion, or subaerial exposure, with a significant hiatus indicated (Van Wagoner et al., 1988). This definition restricts the usage of unconformity to subaerial surfaces and their correlative submarine erosional surfaces and is somewhat more restrictive than the definition of unconformity used by Mitchum (1977). Local, contemporaneous erosion and deposition associated with geological processes such as point-bar development or aeolian-dune migration are excluded from the definition of unconformity used in this book.

Conformity: A surface separating younger from older strata along which there is no evidence of erosion (neither subaerial nor submarine) or nondeposition, and along which no significant hiatus is indicated. It includes surfaces onto which there is very slow deposition or low rates of sediment accumulation, with long periods of geologic time being represented by very thin deposits.

Sequences can be subdivided into systems tracts (Van Wagoner et al., 1988; Posamentier et al., 1988) based on objective criteria including types of bounding surfaces, parasequence set distribution, and position within the sequence. Systems tracts also can be characterized by geometry and facies associations. *Systems tracts* are defined as a linkage of contemporaneous depositional systems (Brown and Fisher, 1977);

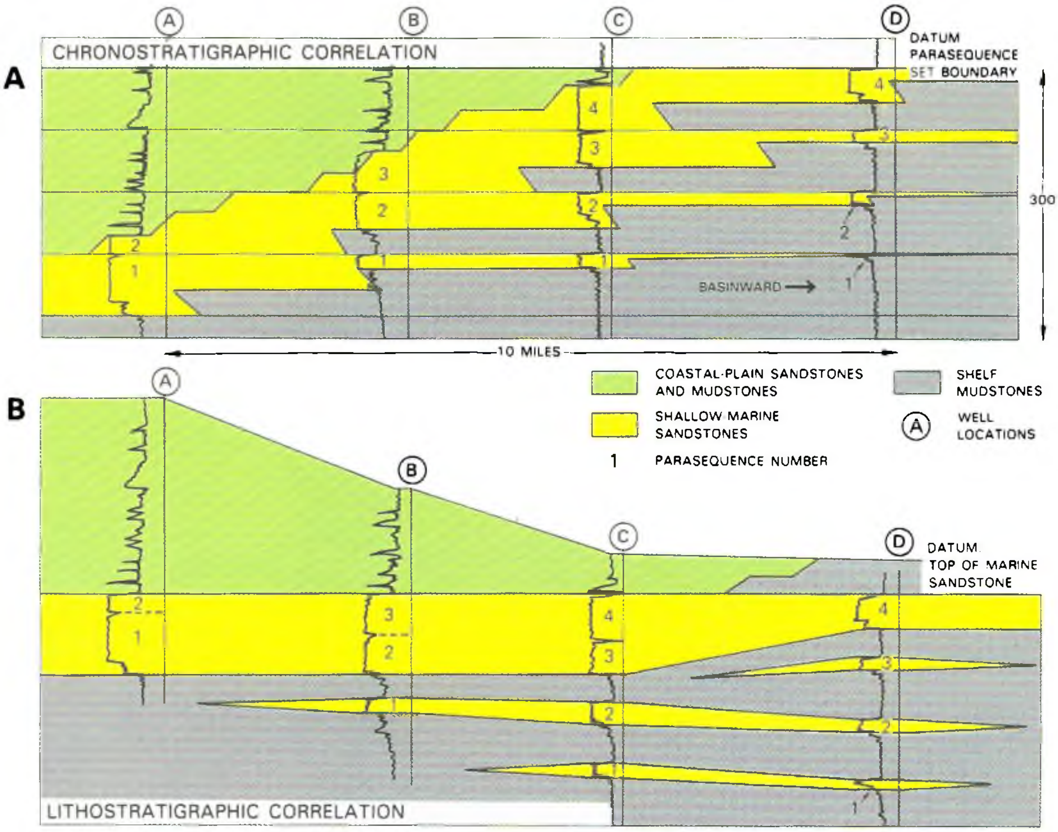


Figure 17—Comparison of (A) chronostratigraphic correlation and (B) lithostratigraphic correlation styles: progradational parasequence set.

depositional systems are defined as three-dimensional assemblages of lithofacies (Fisher and McGowen, 1967). Four systems tracts are defined in following sections: lowstand, shelf-margin, transgressive, and highstand systems tracts. *Lowstand* and *highstand* are descriptive terms that refer to position within the sequence; when referring to systems tracts these terms do not indicate a period of time or position on a eustatic or relative cycle of sea level. Type-1 and -2 sequences are recognized in the rock record (Van Wagoner et al., 1988); they are defined and identified on the basis of (1) arrangement of strata into systems tracts between the sequence boundaries and (2) types of bounding unconformities. *Type-1 sequences* are composed of lowstand, transgressive, and highstand systems tracts bounded beneath by type-1 unconformities and their correlative conformities. *Type-2 sequences* are composed of shelf-margin, transgressive, and highstand systems tracts bounded beneath by type-2 unconformities and their correlative conformities.

A type-1 sequence is interpreted to form when the rate of eustatic fall exceeds the rate of subsidence at the depositional-shoreline break, producing a relative fall in sea level at that position (Jervey, 1988; Posamentier et al., 1988). The *depositional-shoreline break* is the position on the shelf landward of which the depositional surface is at or near base level, usually sea level, and seaward of which the depositional surface is below sea level (Van Wagoner et al., 1988). This position coincides approximately with the seaward end of the stream-mouth bar in a delta or the upper shoreface in a beach environment. In previous publications the depositional-shoreline break has been referred to as the "shelf edge" (Vail and Todd, 1981; Vail et al., 1984).

The systems tract distribution within a sequence is determined, in part, by the relationship between the depositional-shoreline break and the shelf break. We define the *shelf break* as the physiographic province in a basin defined by a change in dip from the shelf (dipping less than 1:1000 landward of the shelf break) to the slope (dipping more than 1:40 seaward of the shelf break) (Heezen et al., 1959). During today's highstand of sea level, the shelf break ranges in water depth from 120 to 600 ft (37 to 183 m). In many basins, the depositional-shoreline break, at the time of a relative fall in sea level, is 100 mi (160 km) or more landward of the shelf break. In other basins, if the highstand systems tract has prograded to the shelf break, the depositional-shoreline break, at the time of the relative

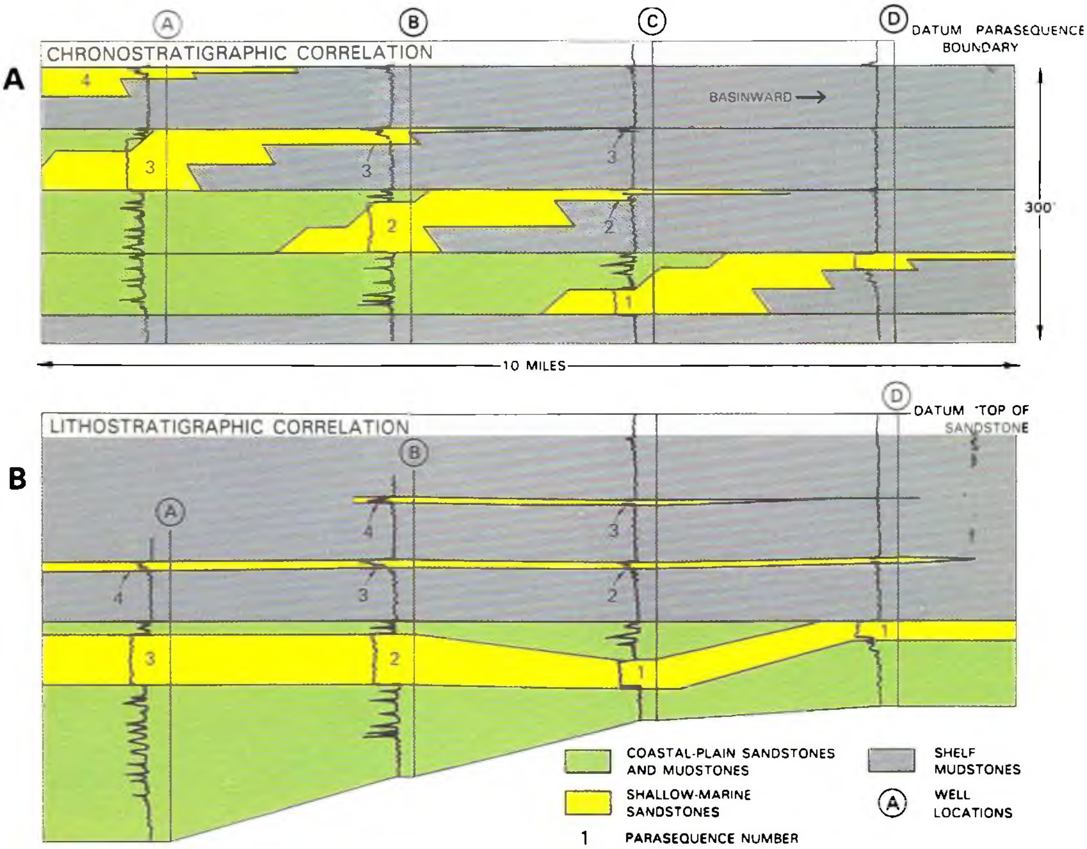


Figure 18—Comparison of (A) chronostratigraphic and (B) lithostratigraphic correlation styles: retrogradational parasequence set.

fall in sea level, may be at the shelf break.

A type-2 sequence boundary is interpreted to form when the rate of eustatic fall is slightly less than or equal to the rate of basin subsidence at the existing depositional-shoreline break at the time of the eustatic fall (Jervey, 1988; Posamentier et al., 1988). This means that there is no relative fall in sea level at the depositional-shoreline break for the type-2 sequence boundary.

The following discussion describes the stratal patterns of type-1 and type-2 sequences, and the characteristics of the sequence boundaries.

Stratal Patterns in Type-1 Sequences

Basin geometry fundamentally influences the stratal patterns in type-1 sequences; type-1 sequences deposited in basins with a shelf break have a different lowstand configuration from type-1 sequences deposited in basins with a ramp margin. Type-1 sequences and their constituent systems tracts deposited in basins with shelf-break margins and ramp margins are compared below.

Shelf-Break Margin

The distribution of parasequences and parasequence sets within the lowstand, transgressive, and highstand systems tracts of an idealized type-1 sequence is illustrated in Figure 19. This type of sequence is deposited in a basin with

(1) a well-defined shelf, slope, and basin-floor topography;
(2) shelf dips less than 0.5°, slope dips of 3 to 6°, with 10° dips along submarine canyon walls (Shepard, 1973);
(3) a relatively abrupt shelf break separating low-angle shelf deposits from much more steeply dipping slope deposits (Heezen et al., 1959);
(4) a relatively abrupt transition from shallow water into much deeper water;
(5) oblique clinoform patterns;
(6) incision in response to sea-level fall below the depositional-shoreline break if submarine canyons form; and
(7) probable deposition of basin-floor submarine fans and slope fans.

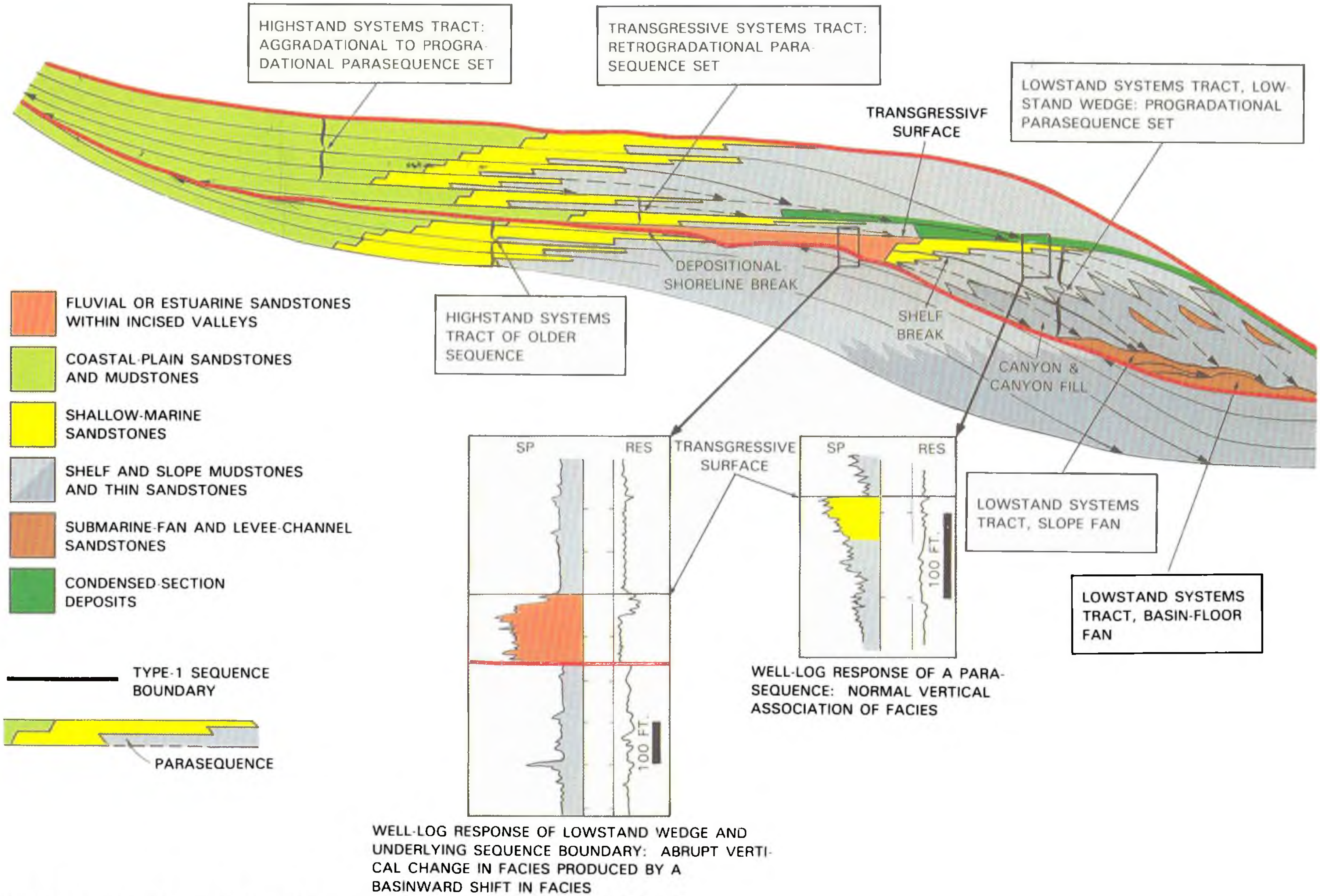


Figure 19—Stratal patterns in a type-1 sequence deposited in a basin with a shelf break.

In addition to being deposited in a basin with a shelf break, the following additional conditions must exist:

(1) sufficiently large fluvial systems to cut canyons and deliver sediment to the basin;
(2) enough accommodation for the parasequence sets to be preserved; and
(3) a relative fall in sea level of a rate and magnitude sufficient to deposit the lowstand systems tract at or just beyond the shelf break.

The component parasequence sets discussed below are the ones most commonly encountered in each systems tract. Variations in the rates of sediment supply and relative sea-level change along a basin margin can result in the simultaneous deposition of different parasequence sets in different places on the shelf. For this reason, boundaries between systems tracts may vary in time from place to place on the shelf within the same sequence. Fundamental stratal components of the systems tracts in an ideal sequence (Figure 19) are discussed below. Although a submarine fan is included in Figure 19, this inclusion is not meant to convey a particular thickness for the sequence, especially on the shelf. As mentioned previously in this book, sequences are defined by the component strata and types of bounding surfaces, not by thickness or time for formation. For example, an unconformity-bounded stratal unit, composed of systems tracts with no internal unconformities (as defined in this book) is a sequence. This sequence may be tens of feet thick and detectable only on well logs or in cores and outcrops, or it may be hundreds of feet thick and easily resolvable on seismic lines.

Lowstand Systems Tract

The lowstand systems tract consists of a basin-floor fan, a slope fan, and a lowstand wedge. Typically, the *basin-floor fan* is dominantly sand, consisting of Tab, Tac and truncated Ta Bouma sequences. It appears to be similar to the type I and type II fans of Mutti (1985). The basin-floor fan may be deposited at the mouth of a canyon, although it may occur widely separated from the canyon mouth, or a canyon may not be evident. It has no age-equivalent rocks on the slope or shelf. *Slope fans* are made up of turbidite-leveed channel and overbank deposits. They overlie the basin-floor fan and are downlapped by the overlying lowstand wedge (Vail, 1987). The slope fan appears to be similar to the type III fan of Mutti (1985). The *lowstand wedge* is composed of one or more progradational parasequence sets making up a wedge that is restricted seaward of the shelf break and that onlaps the slope of the preceding sequence. The proximal part of the wedge consists of incised-valley fills and their associated lowstand-shoreline deposits on the shelf or upper slope. The distal part of the wedge is composed of a thick, mostly shale-prone, wedge-shaped unit that downlaps onto the slope fan.

As strictly defined, parasequences are difficult to recognize in basin-floor and slope-fan environments because there are no criteria in these units with which to recognize shallowing upward. Fan lobes in these units characterized by upward thinning and fining bedsets or by upward thickening and coarsening bedsets may represent parasequences.

Incised valleys are entrenched fluvial systems extending their channels basinward and eroding into underlying strata in response to a relative fall in sea level. On the shelf, the incised valleys are bounded below by the sequence boundary and above by the first major marine-flooding surface, called the transgressive surface. The well log at the left in Figure 19 shows a common well-log pattern through an incised-valley fill. The blocky well-log pattern, interpreted from the log shape as a braided stream, lies in sharp contact with shelf mudstones. This abnormal vertical association of depositional environments is called a basinward shift in facies; it forms in response to a relative fall in sea level. A *basinward shift in facies* occurs when shallow-marine to nonmarine strata, deposited above a sequence boundary, lie directly on much deeper strata, such as middle- to outer-shelf mudstones and thin sandstones below the sequence boundary, with no intervening rocks deposited in intermediate depositional environments. The basinward shift in facies is a result either of erosion of the intervening gradational facies, or of nondeposition because of the rapid shift of environments. Differentiation of basinward shifts in facies from distributary channels is discussed in the section entitled "Sequence Boundary Characteristics." The well-log response through a single parasequence at the top of the lowstand wedge is illustrated in Figure 19.

Regional stratigraphic analyses, such as those documented in this book, suggest that a proportionately large number of the reservoirs in siliciclastic sequences occur within the lowstand systems tract.

Transgressive Systems Tract

The transgressive systems tract is bounded below by the transgressive surface and above by the downlap surface or maximum-flooding surface. Parasequences within the transgressive systems tract backstep in a retrogradational parasequence set. The systems tract progressively deepens upward as successively younger parasequences step farther landward. The downlap surface, coincident with the upper boundary of the youngest parasequence in the transgressive systems tract, is the surface onto which the clinoform toes of the overlying highstand systems tract may merge and become very thin. It is during the time of the transgressive to early highstand systems tracts that this condensed section is deposited.

The *condensed section* (Loutit et al., 1988) is a facies consisting of thin hemipelagic or pelagic sediments deposited as the parasequences step landward and as the shelf is starved of terrigenous sediment. The greatest diversity and abundance of fauna within the sequence are found in this terrigenous-starved interval. Deposition within the condensed section is continuous although the section commonly is thin, accumulates at very slow rates, and encompasses a great deal of time.

Condensed sections are most extensive at the time of maximum regional transgression of the shoreline (Loutit et al., 1988). These characteristics of condensed sections have two important implications for stratigraphic analysis. First, if the sampling of outcrop, core, or cuttings for biostratigraphic-age determination is not selective, the condensed section can be missed. If the condensed section is missed, there may be an apparent major time gap in the biostratigraphic record, prompting paleontologists to infer a major unconformity where deposition really was continuous. Second, the condensed section commonly contains more abundant, diverse, deep-water fauna than do rocks above or below. Few or no fauna are recovered from the largely fluvial, estuarine, or shallow-marine sandstones of the transgressive or lowstand systems tracts. If fauna from successive condensed sections are sampled through several sequences in a well, and no attention is paid to interpretations of depositional environments from well-log or seismic data in the same interval, a continuous, deep-water environment may be interpreted for the sampled interval. This interpretation misses the important sequence boundaries along which fluvial or shallow-marine, reservoir-quality sandstones might have been introduced farther into the basin. Furthermore, the sandstones might be interpreted erroneously as having been deposited in deep water.

Highstand Systems Tract

The highstand systems tract is bounded below by the downlap surface and above by the next sequence boundary. The early highstand commonly consists of an aggradational parasequence set; the late highstand is composed of one or more progradational parasequence sets. The ideal highstand is illustrated in Figure 19. In many siliciclastic sequences the highstand systems tract is significantly truncated by the overlying sequence boundary and, if preserved, is thin and shale prone.

Ramp Margin

In contrast to Figure 19, the type-1 sequence in Figure 20A was deposited in a basin with a ramp margin. Deposition on a ramp margin is characterized by

(1) uniform, low-angle dips of less than 1°, with most dips less than 0.5°;
(2) shingled to sigmoidal clinoforms (Mitchum et al., 1977);
(3) no abrupt breaks in gradient separating relatively low dips from much steeper dips;
(4) no abrupt changes in water depth from shallow water to much deeper water;
(5) incision to, but not below, the lowstand shoreline in response to a relative fall in sea level; and
(6) deposition of lowstand deltas and other shoreline sandstones in response to the sea-level fall (basin-floor submarine fans and slope fans unlikely to be deposited on the ramp margin).

Cretaceous strata in the interior foreland basin of the western United States and Canada contain examples of this type of sequence. Asquith (1970) showed well-defined examples of sigmoidal to shingled clinoforms with present dips of 0.5° or less in the Washakie, Big Horn, and Powder River basins in Wyoming.

Although the transgressive and highstand systems tracts in Figure 19 and Figure 20A are similar, the lowstand systems tracts in these two figures differ. Thick, shale-prone lowstand wedges, slope fans, and basin-floor fans are unlikely to form in the lowstand systems tract because the depositional dips on ramps are relatively low and uniform. Instead, the lowstand systems tract in a ramp margin typically consists of narrow to broad incised valleys, usually filled with tide-dominated deltaic deposits and age-equivalent, updip fluvial strata. Low-angle clinoforms, such as those documented by Asquith (1970), commonly are found on ramp margins within the transgressive or highstand systems tracts. Delta-front turbidites, such as those documented from the Panther Tongue delta (Figure 1), are common in this type of basin and may be mistaken for submarine fans.

Two end-members of type-1 sequence deposition are represented by Figures 19 and 20A. In the first end member (Figure 19), the relative fall in sea level is sufficient to move the lowstand shoreline beyond the depositional-shoreline break to the shelf break, resulting in probable canyon and submarine-fan formation. In the second end member (Figure 20A), either the relative fall in sea level moves the lowstand shoreline beyond the depositional-shoreline break but not to the shelf break, or no shelf break exists in the basin because the margin is a ramp, resulting in a lowstand systems tract consisting of a relatively thin wedge with no canyon or submarine-fan formation.

Stratal Patterns in Type-2 Sequences

The distributions of parasequence sets and systems tracts in a type-2 sequence are illustrated in Figure 20B. The lowest system tract in the type-2 sequence is the shelf-margin systems tract (Posamentier et al., 1988). It can be deposited anywhere on the shelf and consists of one or more weakly progradational to aggradational

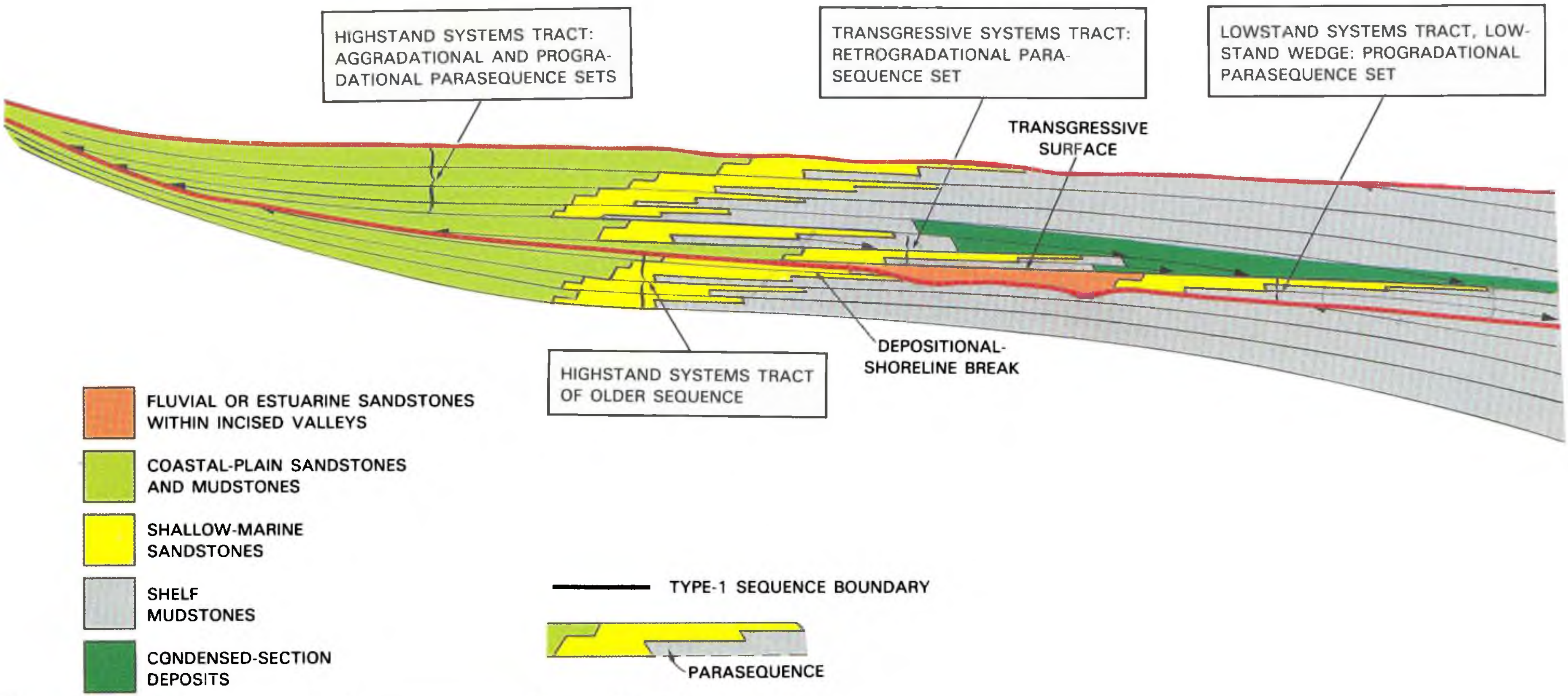


Figure 20A—Stratal patterns in a type-1 sequence deposited in a basin with a ramp margin.

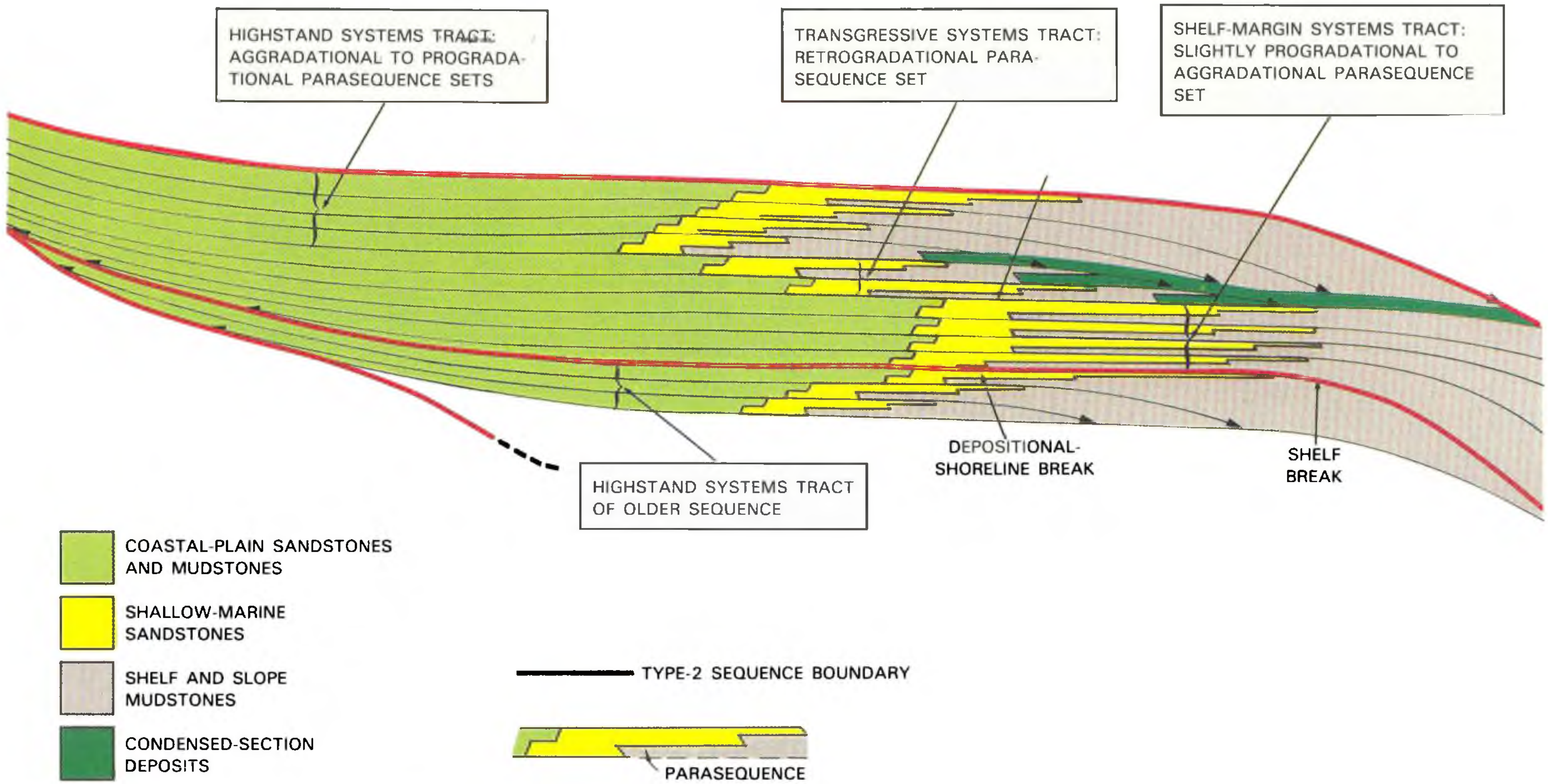


Figure 20B—Stratal patterns in a type-2 sequence.

parasequence sets composed of shallow-marine parasequences with updip coastal-plain deposits. The base of the shelf-margin systems tract is the type-2 sequence boundary, and the top is the first significant flooding surface on the shelf. The transgressive and highstand systems tracts in type-2 and type-1 sequences are similar.

Type-2 sequences (Figure 20B) and type-1 sequences deposited on a ramp (Figure 20A) superficially resemble each other; both lack fans and canyons, and both of their initial systems tracts (shelf-margin systems tract in the type-2 sequence and lowstand systems tract in the type-1 sequence) are deposited on the shelf. However, unlike the type-1 sequences deposited on ramps, there is no relative fall in sea level at the depositional-shoreline break for the type-2 sequence. Consequently, type-2 sequences do not have incised valleys and they lack the significant erosional truncation that results from stream rejuvenation and a basinward shift in facies.

Sequence Boundary Characteristics

A sequence boundary is an unconformity and its correlative conformity; it is a laterally continuous, widespread surface covering at least an entire basin and seems to occur synchronously in many basins around the world (Vail et al., 1977; Vail and Todd, 1981; Vail et al., 1984; Haq et al., 1988). A sequence boundary separates all of the strata below the boundary from all of the strata above the boundary (Mitchum, 1977) and has chronostratigraphic significance. Correlation of sequence boundaries on well-log cross sections provides a high-resolution chronostratigraphic framework for facies analysis. If sufficient well control is available, not only does this framework equal or surpass other tools in chronostratigraphic resolution, but, if necessary, the framework can be developed from the well-log data base. The following discussion of sequence boundaries is divided into three parts: recognition criteria, incised-valley attributes and examples, and correlation pitfalls.

Recognition Criteria

The criteria that identify the unconformable part of sequence boundaries in a *single* well log, core, or outcrop include a basinward shift in facies for a type-1 sequence boundary and a vertical change in parasequence stacking patterns for a type-1 or a type-2 sequence boundary. As an example of the latter criterion, consider the case of three parasequence sets arranged in vertical order from the oldest to the youngest: retrogradational, progradational (or aggradational), followed by retrogradational. In this case, there is commonly a sequence boundary at the top or the base of the progradational (or aggradational) parasequence set.

On a well-log or outcrop *cross section* the recognition criteria for the unconformable part of a type-2 sequence boundary include onlap of overlying strata, a downward shift in coastal onlap, and subaerial exposure with minor subaerial truncation, all landward of the depositional-shoreline break within the updip, coastal-plain part of the sequence where correlation is less precise. For this reason, these criteria are particularly difficult to recognize in well-log or outcrop cross sections. Type-2 sequence boundaries are most readily defined by the changes in parasequence stacking patterns described above. Based on this criterion, type-2 sequence boundaries in siliciclastic strata appear to be rare in most basins.

On a well-log or outcrop cross section the recognition criteria for the unconformable part of a type-1 sequence boundary include the following:

- Subaerial-erosional truncation, a laterally correlative subaerial-exposure surface marked by soil or root horizons, and laterally correlative-submarine erosion, especially in the deep-water slope environment must be present.
- Onlap of overlying strata either onto the margins of incised valleys or coastal onlap must exist.
- A downward shift in coastal onlap (Vail et al., 1977); however, this commonly cannot be demonstrated on well-log cross sections because much of the coastal onlap occurs in the updip, fluvial part of the sequence where accurate well-log correlation is difficult, and therefore, the criterion of a basinward shift in facies must be used.
- To confirm that erosional truncation and a basinward shift in facies marks a sequence boundary and not a local-distributary channel, one or more of these criteria must be demonstrated over a regionally significant area.

The unconformable part of a type-1 sequence boundary can be traced seaward into a conformable surface on the shelf or slope, commonly occurring at or near the base of a marine parasequence. Based on the criteria listed above, applied to the stratigraphic analysis of many basins around the world, type-1 sequence boundaries appear to predominate in siliciclastic strata.

Not all of the recognition criteria presented above occur everywhere along a particular type-1 sequence boundary in a basin. A type-1 sequence boundary has different physical expressions depending on where it is observed and on the variations along a basin margin in rates of sediment supply and sea-level change.

On the slope, seaward of the shelf break or in deeper-water environments, the most pronounced attributes of a type-1 sequence boundary are truncation and onlap. The distribution of these recognition criteria is controlled by the distribution of submarine canyons, slope failure, contour-current erosion set up by lowstand conditions, and the deposition of the basin-floor and slope fans.

On the shelf, the most pronounced attributes of a type-1 sequence boundary are truncation, a basinward shift in facies, and subaerial exposure. The distribution of these properties of type-1 sequence boundaries is controlled primarily by the distribution of incised valleys and the lithology of the strata that fill these valleys.

Incised Valleys

Incised valleys range in width from less than several miles to many tens of miles. They range in depth from tens to several hundreds of feet. Incised valleys form and fill in two phases. The first phase consists of erosion, sediment bypass through the eroded valleys, and deposition at the lowstand shoreline in response to a relative fall in sea level. The second phase consists of deposition within the valleys in response to a relative rise in sea level, generally during the late lowstand or transgressive systems tracts.

Because incised valleys form in these two temporally distinct phases, the fill may consist of a wide variety of rock types deposited in a variety of environments. Depositional environments and associated rock types within the upper reaches of the incised valleys include estuarine and braided-stream sandstones, fluvial sandstones showing evidence of significant tidal modification, or coastal-plain sandstones, mudstones or coals. These deposits, which lie above the sequence boundary, commonly rest directly on the middle- to outer-shelf mudstones and thin sandstones that lie below the boundary, with intervening rocks either eroded or not deposited in intermediate-depositional environments. As discussed above, this abnormal vertical association of facies marks a basinward shift in facies. Incised valleys also can be filled with marine mudstones if the rate of deposition of coarse-grained sediment is low relative to the rate of sea-level rise at the end of the lowstand.

Depositional environments and associated rock types within the lower reaches of the incised valleys vary and include lowstand-delta and tidal-flat sandstones and mudstones and beach and estuarine sandstones. Commonly, these shallow-marine strata, in the case of beaches or deltas, form one or more progradational parasequence sets. If tide-dominated deltas, consisting of tidal bars and tidal shoals within an estuary, form in the lower reaches of an incised valley there may be no deposition of sand-prone lowstand-shoreline facies across the shelf until the transgressive systems tract is deposited. Landward, these tide-dominated strata change facies into coarse-grained braided-stream deposits.

Adjacent to incised valleys, the erosional surface passes into a correlative subaerial-exposure surface marked by soils or rooted horizons. Three examples of incised valleys marking type-1 sequence boundaries, exhibiting the characteristics described above, are discussed in the following paragraphs.

The first example is a relatively narrow incised valley in the Muddy Sandstone, illustrated with a well-log cross section in Figure 21 showing the Clareton field in the eastern Powder River basin, Wyoming. The valley is approximately 6 mi (9.6 km) wide, 40 mi (64 km) long, and erodes 40 ft (18 m) into the underlying shelf mudstones of the Skull Creek Shale. The valley is filled with fine- to medium-grained sandstone and mudstone interpreted to have been deposited in a fluvial to estuarine environment. The fluvial to estuarine sandstones lying directly on the shelf mudstones represent a basinward shift in facies and, along with the truncation, sharply mark the sequence boundary. The incised-valley fill is encased laterally in the shelf mudstones; delta-front or lower-shoreface sandstones do not occur below the incised valley or adjacent to it. Shallow-marine parasequences, in a retrogradational parasequence set, overlie the fluvial or estuarine incised-valley fill. The sequence boundary defined by the incised-valley erosion can be correlated throughout the Powder River and Denver basins in Wyoming and Colorado (Weimer, 1983, 1984, 1988).

The second example illustrates three middle Miocene incised valleys from south-central Louisiana shown on a well-log cross section (Figure 22). This cross section is a small part of a regional study of middle Miocene sequence stratigraphy in south Louisiana. Approximately 700 well logs, six cores, and numerous biostratigraphic analyses were used to interpret the stratigraphy.

Depositional environments of the Miocene strata were interpreted from well-log shapes, core descriptions calibrated to well-log response, map patterns, and paleowater depth from biostratigraphy. Based on foraminifer-age dates, sequence 2 was placed at the top of the *Cibicides opima* biozone, sequence 1 at the base of the *Bigenerina humblei* biozone. These zonations suggest that sequence boundary 1 corresponds to the 13.8-Ma sequence boundary (L.C. Menconi, personal communication, 1988) on the Exxon sea-level curve (Haq et al., 1988) and that sequence boundaries 2 and 3 are not on the Exxon chart.

Broad, sheet-like geometries, well-developed truncation, and a basinward shift in facies are associated with these incised valleys (Figure 22). The sandstones within the incised valleys are interpreted as fluvial, possibly braided stream to estuarine in origin. In core, the blocky sandstones are medium- to coarse-grained, have sharp, erosional bases often overlain by the coarsest grains in the blocky sandstone, are nearly completely composed of trough-cross beds, and generally are composed of smaller 2- to 10-ft- (0.6- to 3 m-) thick fining-upward units. Based on biostratigraphic data, the marine mudstones below the sequence boundaries were deposited in inner- to middle-neritic water depths. Thin sandstones within or at the top of

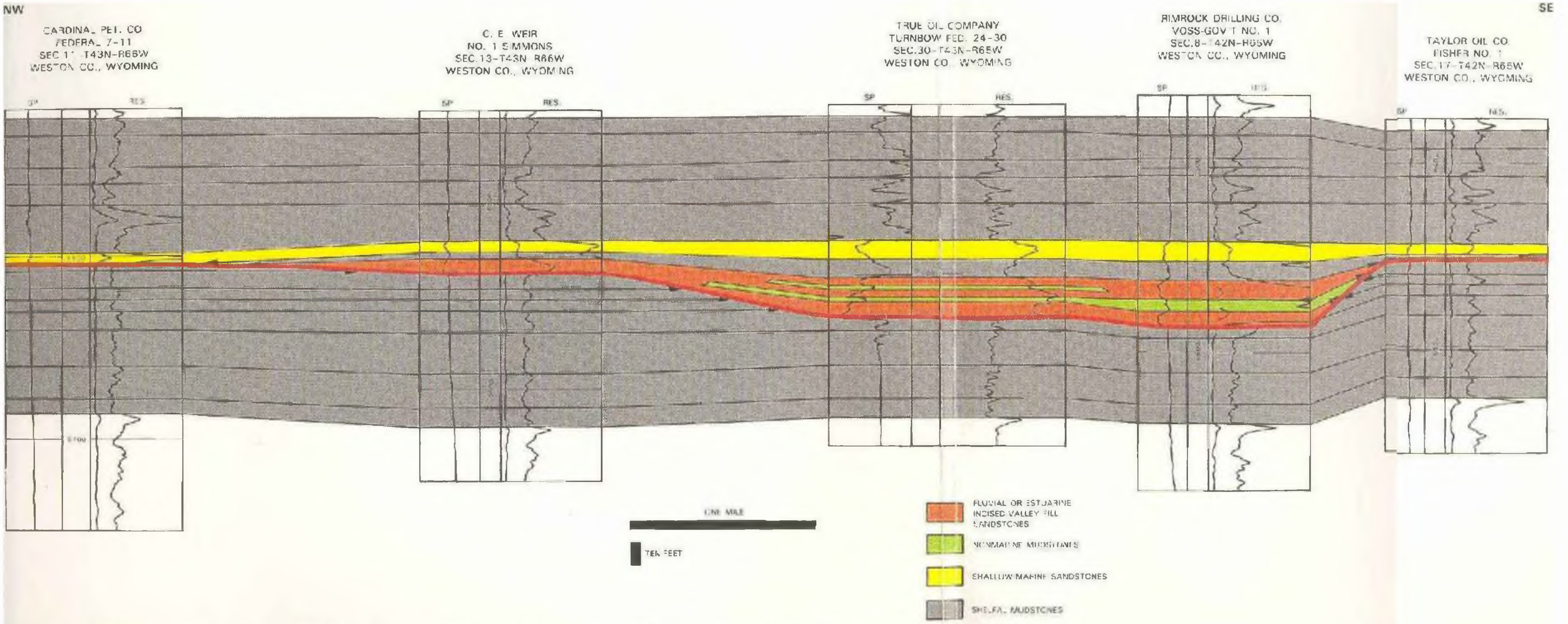


Figure 21—An incised valley in the Powder River basin. Albian-aged Muddy Sandstone within the incised valley erodes into the Skull Creek Shale. The sequence boundary at the base of the Muddy Sandstone is marked by regional-subaerial erosion or exposure, a downward shift in facies, and onlap.

Figure 22—Well-log cross section of the middle Miocene of south-central Louisiana illustrating three sequence boundaries overlain by incised-valley-fill fluvial or estuarine sandstones and mudstones. Beneath incised valleys, sequence boundaries are marked by a basinward shift in facies and truncation; away from the valleys the sequence boundaries are coincident with the marine-flooding surfaces at the tops of distal marine parasequences. This cross section is landward of the middle Miocene shelf break.

the mudstones but below the sequence boundaries are interpreted to be distal delta front or lower shoreface, based on well-log shape, thickness, areal distribution, and association with open-marine mudstones. This vertical juxtaposition of very shallow-marine to nonmarine strata directly above open-marine rocks marks the basinward shift in facies. These incised valleys extend many tens of miles updip and cut across underlying depositional environments, which range from fluvial to outer shelf.

Correlation of similar resistivity-curve patterns from well to well in the marine mudstones beneath the incised valleys provides an accurate and detailed chronostratigraphic framework, derived independently from biostratigraphic or radiometric control, for recognition of stratal terminations such as truncation. Where incised valleys are not present, for example along sequence boundary 2, the sequence boundary coincides with a marine-flooding surface. The sequence boundary in this interfluvial area may be marked by soil or root horizons if these products of subaerial exposure have not been removed by the subsequent sea-level rise.

Figure 23 is a paleogeographic map showing the distribution of the incised-valley fill and interfluves for the middle Miocene sequence boundary 2. The map illustrates the distribution of sediments on the depositional surface just before the first major marine-flooding event inundated the shelf, terminating lowstand deposition, and creating the transgressive surface. Contours show the gross incised-valley fill thickness; the major incised-valley axes are highlighted with heavy lines. The location of the cross section in Figure 22 is marked on the map. The incised-valley fill in the central and eastern part of the map is a broad sheet of sandstone at least 40 mi (64 km) wide, 25 mi (40 km) long, and locally, is up to 240 ft (73 m) thick, but averages about 150 ft (46 m) thick over the area. These dimensions of the valley fill are a minimum because the southern and eastern limits of the incised valley are outside the study area. Truncation and a basinward shift in facies can be observed everywhere below the incised valley in Figure 23.

This incised-valley sheet sandstone represents either a single, large incised Miocene river similar in dimensions to the modern Mississippi (Fisk, 1944) or a number of smaller rivers that coalesced during the sea-level fall. In the latter case, tributaries forming as sea level fell would erode progressively into interfluvial areas, allowing separated rivers to coalesce into a single, large alluvial valley. Tributary development in response to the sea-level fall would have begun first in the earliest exposed or most northerly strata (Figure 23), allowing ample time, in this case, for the separate valleys to coalesce. In the latest exposed, or most southerly strata, tributaries barely would have begun incising before the sea-level fall ended and the valleys were initially flooded. This condition may explain why the incised valleys in the central and eastern part of the map have more interfluves at their southern or downdip ends. Incised-valley sheet sandstones, commonly bifurcating to the south, are a typical reservoir pattern in Tertiary strata along the Texas and Louisiana Gulf Coast.

The incised-valley fill in the western part of the map is a relatively narrow sandstone 1 to 5 mi (1.6 to 8 km) wide, at least 40 mi (64 km) long, and up to 270 ft (82 m) thick. Except for thickness, these dimensions are comparable to the dimensions of the Muddy Sandstone incised valleys in the Powder River basin, Wyoming (Figure 21). These relatively narrow incised valleys probably formed when a single small- to moderate-sized river entrenched during a sea-level fall.

Not all type-1 sequence boundaries marked by erosional truncation associated with incised valleys exhibit a basinward shift in facies. The third example of a type-1 sequence boundary (Figure 24) illustrates truncation along one side of an interpreted incised valley at the 80-Ma sequence boundary (Haq et al., 1988) on the top of the Gammon Ferruginous Member of the Pierre Shale in the Powder River basin, eastern Wyoming. The incised valley is filled with siltstones, marine mudstones, and bentonites. In the cross section (Figure 24), 300 ft (92 m) of strata within the Gammon are truncated where the sequence boundary at the base of the incised valley cuts down to the southeast. Above the sequence boundary the Ardmore bentonite, interbedded marine mudstones, and the Sharon Springs Member of the Pierre Shale (another marine mudstone), onlap to the northwest. The Ardmore bentonite and lower half of the Sharon Springs are within the *Baculites obtusus* ammonite biozone (Gill and Cobban, 1966). Shallow-marine to fluvial sandstones recording a basinward shift in facies have not been observed directly above the sequence boundary. This pattern of truncation below the sequence boundary and onlap of marine mudstones above has been observed regionally within the Powder River basin.

The regional truncation below the 80-Ma sequence boundary, interpreted to have been formed by regional paleovalleys, is shown on Figure 25. In contrast to Figure 24, Figure 25 shows both sides of the major incised valley. This map illustrates the subcrop thickness in the Powder River basin from the 80-Ma sequence boundary to an underlying resistivity marker coincident with a sequence boundary at the base of the Sussex sandstone. Small, open circles indicate the distribution of well logs used to make the map. North–south erosional axes in Figure 25, indicated by heavy lines, are interpreted to be incised-valley axes cut during the 80-Ma sea-level fall (Haq et al., 1988). These axes suggest a dendritic drainage pattern; this regional pattern is unlikely to be produced by submarine erosion on this ramp margin. A rapid

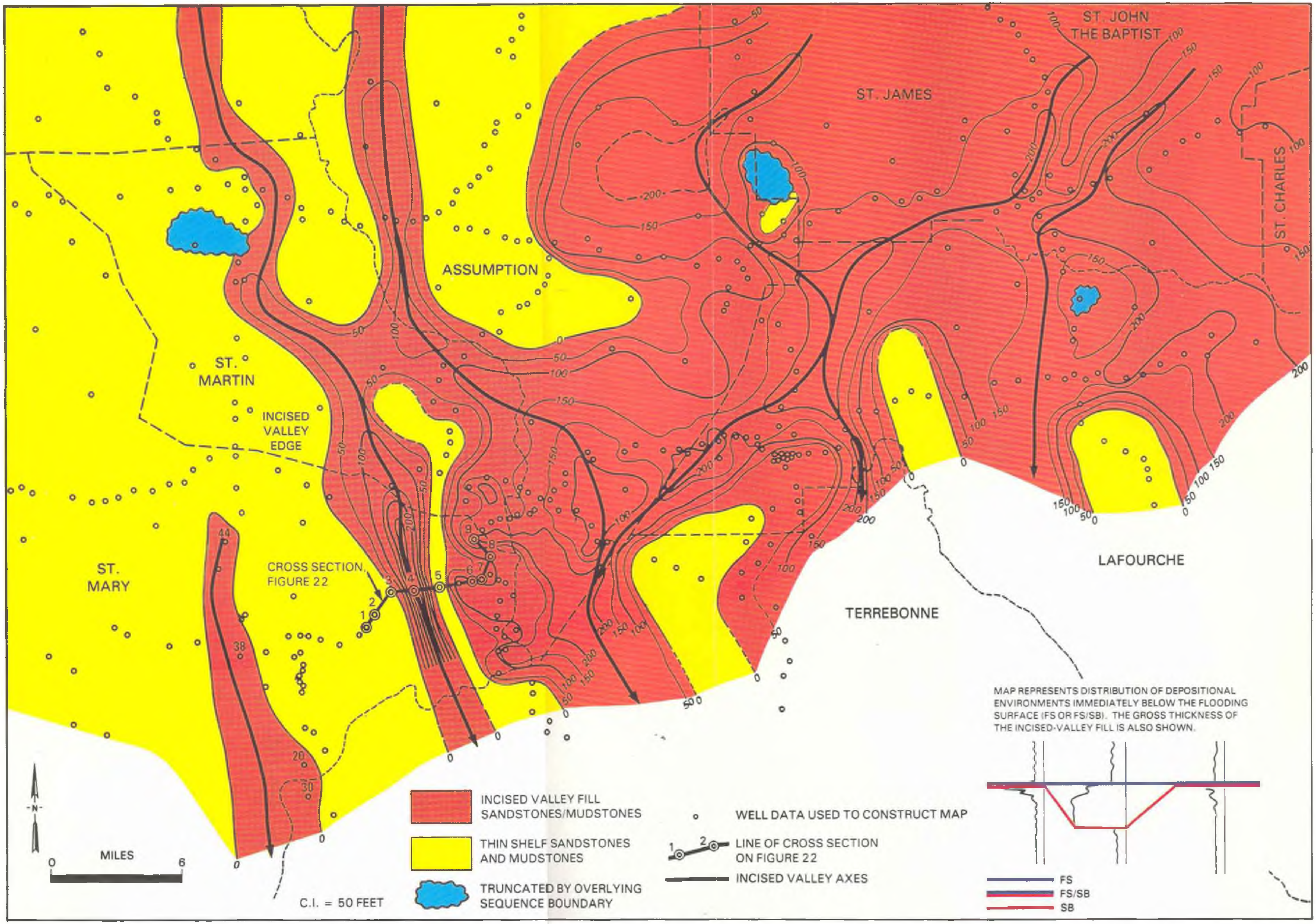


Figure 23—Paleogeographic map of the middle Miocene-aged sequence 2 from Figure 22 showing the distribution of the lowstand incised-valley fill below the transgressive surface in south-central Louisiana. Contours show the incised-valley-fill thickness; the major incised-valley axes are highlighted with heavy lines. The incised-valley fill is sheet-like in the eastern area of the map. This pattern is common in Tertiary incised valleys in the Gulf Coast and probably forms when several river systems coalesce during sea-level lowstand. The incised-valley fill is ribbon-like in the western area of the map, probably reflecting incision of a single fluvial system during sea-level lowstand. This pattern is developed in basins with small or widely spaced fluvial systems. The location of the cross section in Figure 22 is indicated on the map.

Figure 24—Truncation below and onlap onto the 80-Ma sequence boundary (Haq et al., 1988) on the east side of the Powder River basin, Wyoming. This regional-erosional surface, in this area, has been interpreted as a shelf break by Asquith (1970).

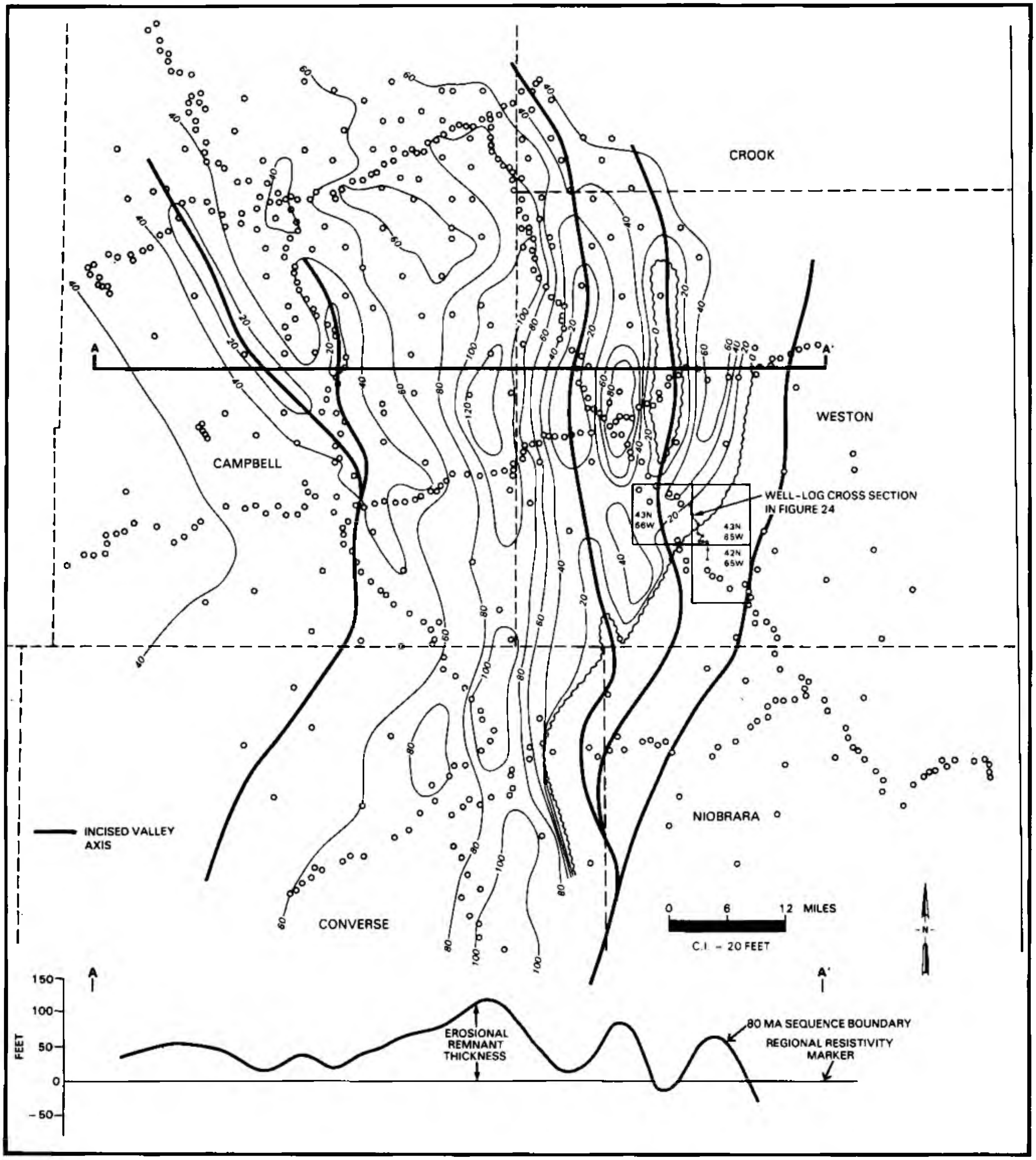


Figure 25—Subcrop map showing the thickness of the erosional remnant between the 80-Ma sequence boundary and an underlying regionally correlative resistivity marker. The resistivity marker is interpreted to have been nearly horizontal at the time of deposition. Cross-section A-A ' represents the relief of the erosional remnant. The surface represented by this map is interpreted to have been incised during the 80-Ma sea-level lowstand. Axes of interpreted incised valleys are shown on the map and coincide with the low areas on the A-A ' cross section. Following incision, a rapid relative rise in sea level drowned the valleys, which were subsequently filled with onlapping bentonites and prograding, downlapping marine mudstones.

relative rise in sea level following lowstand incision drowned the incised systems and, coupled with a probable low influx of coarser-grained sediment, prevented significant coarse-grained siliciclastic infill. Following the sea-level rise, bentonites and marine siliciclastic mudstones and shales, in shelf-perched clinoforms of subsequent transgressive and highstand systems tracts, filled the incised topography.

The 80-Ma sequence boundary has been recognized as a subaerial-erosion surface in other places in the western United States. In western Nebraska, DeGraw (1975) mapped extensive truncation at the top of the Niobrara just below the Ardmore bentonite. A map of the incision on the top of the Niobrara shows a complex fluvial-drainage system characterized by a north-south-trending trellis-drainage pattern (DeGraw, 1975). Basal Pierre siltstones and mudstones filling the incised topography are interpreted to be nonmarine (DeGraw, 1975). Another south- to southeast-trending paleodrainage system at the top of the Niobrara has been observed in central and eastern North and South Dakota (Shurr and Reiskind, 1984). This unconformity also occurs at the base of the *Baculites obtusus* ammonite zone (Shurr and Reiskind, 1984), establishing it as the 80-Ma sequence boundary.

Interpretation of a widespread sequence boundary at the top of the Niobrara or the base of the Pierre has implications for the sedimentary history of the Cretaceous seaway in the western United States. Figure 24 coincides closely with a portion of the cross section through eastern Wyoming presented in Asquith (1970, his figure 12). Asquith interpreted the surface at the top of the Gammon Ferruginous and unnamed members of the Pierre Shale as a depositional surface defining a shelf, shelf-break, slope, and basin-floor topography. Most of the clinoforms on Asquith's cross section have very low present dips, ranging from 18′ to 43′. These low-angle clinoforms or offlaps, coupled with the interpretation that the most steeply dipping surface on Asquith's figure 12 is erosional, not depositional, suggests that this is a ramp margin (see discussion of ramp margin stratal geometries in the section on "Stratal Patterns in Type-1 Sequences"). Furthermore, if the Cretaceous seaway from the eastern Powder River basin to eastern South Dakota were subaerially exposed, it is probable that the sea retreated from a large part of the North American craton at or about 80 Ma.

Finally, the paleodrainage patterns in this example are to the south, not to the east away from the highlands, as is considered normal in this foreland basin (Mallory, 1972). A fall in sea level in foreland basins, such as the Cretaceous basin of western North America, may result in realignment of fluvial axes with major, basinwide tectonic elements such as loci of regional subsidence parallel to thrust sheets. If so, lowstand-fluvial systems may flow perpendicular to the thrust sheets for a short distance into the basin, and then turn 90° to the north or south to flow parallel to axes of regional subsidence for long distances. During times of sea-level highstands in the foreland basin most fluvial systems flow perpendicular to the thrust sheets because this drainage orientation represents, in most cases, the shortest path to the sea. Shoreline parasequences prograde to the east and have north-south-oriented depositional strikes in highstand or transgressive systems tracts.

These three examples (Figures 21 through 25) of incised valleys on well-log cross sections and maps show that the physical expression of type-1 sequence boundaries in siliciclastic strata on a shelf or ramp can vary depending on incised-valley size, distribution, and fill. These aspects of incised valleys are, in turn, controlled by the dimensions, rate of sediment supply, and distribution of the rivers existing in the basin at the time of the sea-level fall. The variations in type-1 sequence boundary expressions and the relationships of these variations to incised valleys and their precursor fluvial systems are shown in Figure 26. In this figure, three different incised-valley types are illustrated—a relatively narrow, sandstone-filled valley like Figure 21; a relatively wide, sandstone-filled valley like Figure 22; and a shale-filled valley like Figure 24—each incising across subaerially exposed ramp or shelf strata deposited landward of the shelf break.

Each valley represents a different original fluvial type and each is associated with a different sequence-boundary expression. For example, a type-1 sequence boundary in a basin or portion of a basin with widely spaced rivers of moderate discharge and a moderate rate of relative sea-level rise will be marked by local truncation and a basinward shift in facies below relatively narrow, sandstone-filled incised valleys. Soil or root horizons in interfluvial areas, if not removed by the subsequent sea-level rise, will be widespread. The sequence boundary might only be recognized in a well log, core, or outcrop if they intersected the incised-valley fill. The position of the sequence boundary in the other well logs in the data base would have to be established by correlation from the wells that penetrated the valleys.

A type-1 sequence boundary in a basin or portion of a basin with numerous, closely spaced rivers or one large river with significant discharge and a low to moderate rate of relative sea-level rise will be marked by regional truncation beneath an extensive fluvial- or estuarine-sheet sandstone and a widely distributed basinward shift in facies. Because of extensive regional truncation, interfluve areas will only be locally preserved and soil horizons, commonly developed on interfluves will be rare. The sequence boundary will be recognized in most of the well logs, cores, and outcrops in the data base.

A type-1 sequence boundary in a basin or a portion of a basin with rivers carrying little or no bed load and a moderate to rapid rate of relative sea-level rise will be marked by truncation and widespread soil or root horizons or equivalent evidence of subaerial exposure, if preserved, but not by a basinward shift in facies. The sequence boundary would not be recognized in an individual well log and probably not recognized in cores. However, correlation demonstrating truncation of resistivity markers on well-log cross sections or seismic lines would readily reveal the incised valley and sequence boundary.

Finally, a type-1 sequence boundary in a basin or a portion of a basin with no rivers will be marked only by widespread evidence of subaerial exposure, if this evidence is not removed by the subsequent sea-level rise. A thin transgressive lag of calcareous nodules lying on the flooded sequence boundary is commonly the only indication that a soil horizon existed on the sequence boundary before the sea-level rise. This lag is discussed in more detail in the section "Parasequence Boundary Characteristics" and more briefly discussed at the end of this section. Significant erosion and a basinward shift in facies will not be associated with the sequence boundary in this case. The sequence boundary will probably not be recognized in a well log in the absence of core, and might be only recognized in the well if it were correlated from another area where it was more clearly expressed.

In Figure 26, different expressions of the type-1 sequence boundary on the shelf or ramp are labelled SB1 where they are beneath sandstone-filled incised valleys; SB2 where they are beneath shale-filled incised valleys; and SB3 to show where the sequence boundary is conformable on the shelf or ramp seaward of the lowstand shoreline. Marine-flooding surfaces marking parasequence boundaries are labelled FS, and subaerially exposed interfluves marking the sequence boundary away from the incised valleys coincident with the flooding surface are labelled FS/SB. Depositional environments, stratal terminations, and other diagnostic criteria associated with type-1 sequence boundaries in siliciclastic strata on a shelf or ramp are summarized in the table in Figure 26.

In addition to the criteria listed in the table in Figure 26, sequence boundaries can be marked by various types of lag deposits. These lags include:

(1) transgressive lags of calcarous nodules deposited on marine-flooding surfaces that are coincident with sequence boundaries (FS/SB) or on sequence boundaries within incised valleys. The calcareous nodules are derived by shoreface erosion from soil horizons formed during the subaerial exposure of the sequence boundary.

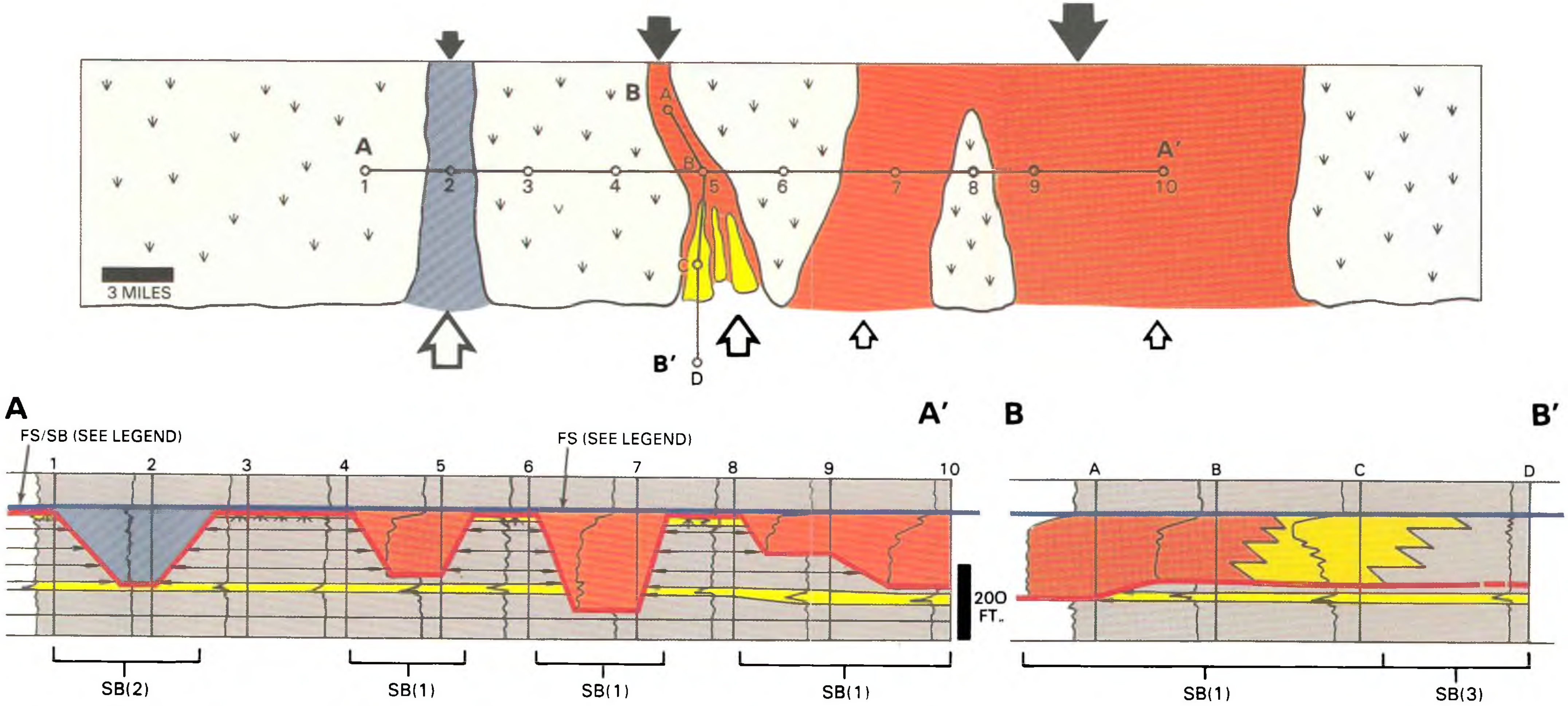
(2) organic or inorganic carbonates deposited on marine-flooding surfaces that are coincident with sequence boundaries.

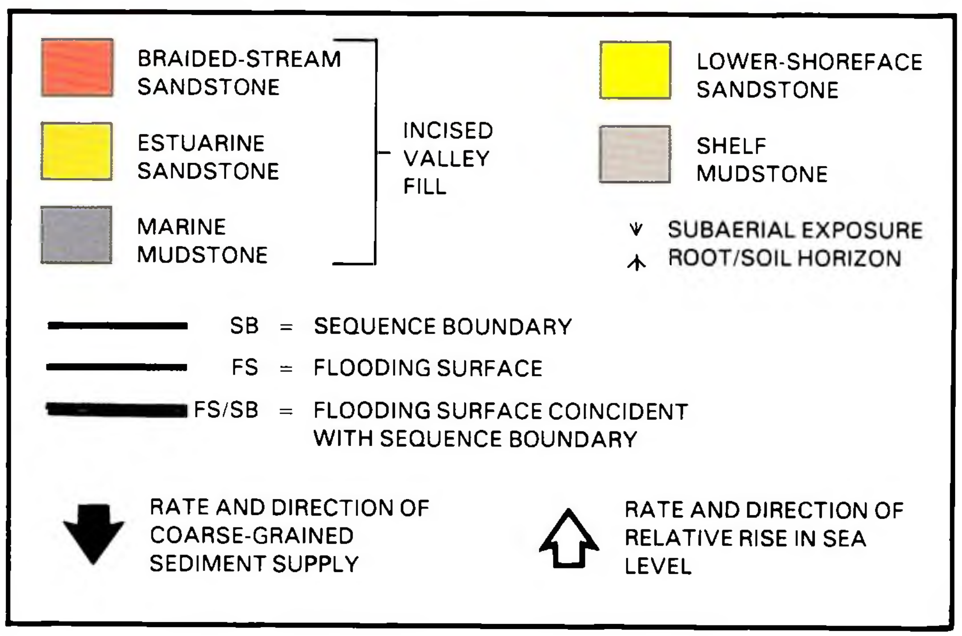
(3) basal-channel lags deposited on sequence boundaries within incised valleys.

The first two types of lags are discussed in the section "Parasequence Boundary." The third type of lag forms during sea-level fall as the shelf is eroded by fluvial channels forming the incised valleys. During incision, finer-grained shelf sediments are flushed through the valley system. Coarser-grained particles eroded from the shelf strata are concentrated as a basal lag as much as several feet thick on the sequence boundary in the valley. The lag derived from the shelf strata commonly consists of a wide variety of grain types including intertidal and open-marine shells, shark teeth, glauconite, phosphorite pebbles, shale rip-up clasts, and bones. The lag commonly shows evidence of subaerial exposure.

Basal-channel lags also may be derived from more proximal sources. These lags commonly consist of coarse grains of chert and quartz, well-rounded quartz and quartzite pebbles, and sandstone and shale rip-up clasts. It is common to find quartz and quartzite pebbles ranging in thickness from thin beds, only one pebble thick, to beds 1 or 2 ft (0.3 or 0.6 m) thick. Thin pebble beds may be deposited in the axes of incised valleys or at the edges of incised valleys, almost on valley interfluves. Commonly, basal-channel lags within valley axes consist of a mixture of particles derived from the incised shelf and more proximal sources. If the incised valley erodes into inner-shelf parasequences and the valley is filled with marine mudstones, or fine-grained estuarine or lower-shoreface strata, the basal-channel lag could be interpreted as transgressive lag with no apparent evidence of a relative fall in sea level. If the incised valley erodes into middle- or outer-shelf mudstones and the valley subsequently is filled with cross-bedded estuarine sandstones, the basal-channel lag could be interpreted as a transgressive lag overlain by a shelf-ridge sandstone.

In Figure 26, the sequence boundary between incised valleys (labelled FS/SB) is a soil or root horizon lying on a shallow-marine parasequence. This parasequence may be deposited during either the highstand systems tract of the previous sequence or the early part of the lowstand systems tract to which the incised valleys belong in Figure 26. The latter case probably occurs frequently in the rock record, forming in the following way. In the early stages of the relative fall in sea level, fluvial systems incise and move progressively seaward across the shelf as the shelf is exposed. Sediment eroded from the underlying highstand strata by the incised valleys is deposited seaward of and adjacent to the valley mouths, forming thin delta and beach parasequences. As the sea-level fall continues and incised valleys erode farther across the shelf, (1) new beach and delta parasequences are deposited farther out on the shelf at the mouths of incised valleys, (2) previously deposited parasequences are eroded in





		SB(1)	SB(2)	SB(3)	FS/SB	FS
DEPOSITIONAL ENVIRONMENTS AND ROCK TYPES	ABOVE THE SURFACE	COMMONLY COARSE-GRAINED FLUVIAL SUCH AS BRAIDED-STREAM TO ESTUARINE. LOWER-SHOREFACE OR DELTAIC PARASEQUENCES MAY BE PRESENT WITHIN THE UPPER PORTION OF THE FILL.	SHELF MUDSTONES OFTEN CONTAINING THIN WAVE AND CURRENT RIPPLES, BENTONITES.	DISTAL PORTIONS OF MARINE PARASEQUENCES, SHELF MUDSTONES.	SHELF MUDSTONES TO SHALLOW-MARINE PARASEQUENCES.	SHELF MUDSTONES TO SHALLOW-MARINE PARASEQUENCES. IN THE LATTER, HUMMOCKY BEDS ABOVE THE BOUNDARY COMMONLY REST SHARPLY ON TIDAL BEDS BELOW THE BOUNDARY.
	BELOW THE SURFACE	VARIABLE, RANGING FROM ALLUVIAL SANDSTONES AND MUDSTONES UPDIP TO SHELF MUDSTONES DOWNDIP		USUALLY SHELF MUDSTONES	VARIABLE, RANGING FROM ALLUVIAL SANDSTONES AND MUDSTONES UPDIP TO SHELF MUDSTONES DOWNDIP	VARIABLE, RANGING FROM FLUVIAL AND ESTUARINE SANDSTONES OF THE INCISED VALLEY UPDIP TO SHELF MUDSTONES DOWNDIP
STRATAL TERMINATIONS	ABOVE THE SURFACE	*ONLAP* ONTO THE VALLEY WALLS, *ONLAP* UP THE VALLEY AXIS (COASTAL ONLAP).	*ONLAP* ONTO THE VALLEY WALLS, *ONLAP* UP THE VALLEY AXIS, *DOWNLAP* DOWN THE VALLEY AXIS.	DOWNLAP INTO THE BASIN, ALTHOUGH THIS DOWNLAP MAY OCCUR OVER MANY 10's OF MILES.	STRATA CONFORMABLE WITH THE SURFACE OR DOWNLAP ONTO THE SURFACE	
	BELOW THE SURFACE	TRUNCATION	TRUNCATION	SURFACE IS CONFORMABLE WITH UNDERLYING STRATA.	SURFACE IS PARALLEL TO SUBPARALLEL TO UNDERLYING STRATA ALTHOUGH THERE MAY HAVE BEEN MINOR EROSION OCCURRING DURING THE FLOODING, PRODUCING A PLANAR SURFACE WITH LITTLE OR NO REGIONAL RELIEF.	
DIAGNOSTIC CRITERIA		• POSSIBLE ROOT OR SOIL HORIZON ON VALLEY FLOOR • SHARP SURFACE MARKING A MAJOR INCREASE IN GRAIN SIZE • POSSIBLE LAG BED RESTING ON SURFACE. LAG COMMONLY CONSISTS OF CLAY RIP-UP CLASTS, SHELLS, SHARKS TEETH, BONES, AND PHOSPHORITE PEBBLES	• POSSIBLE ROOT OR SOIL HORIZON ON VALLEY FLOOR • POSSIBLE LAG BED RESTING ON SURFACE; LAG WOULD BE SIMILAR IN COMPOSITION TO LAG DESCRIBED IN SB(1) CATEGORY • STRATA ABOVE THE SURFACE COULD BE FINER GRAINED THAN STRATA BELOW THE SURFACE	OFTEN COINCIDENT WITH A REGIONALLY PERSISTENT RESISTIVITY MARKER	*SOIL OR ROOT HORIZONS DEVELOPED ON THIS SURFACE.* REWORKING DURING THE MARINE FLOODING COULD OBSCURE OR REMOVE EVIDENCE OF THESE HORIZONS. THIN CARBONATE GRAINSTONES OR SHELL BEDS COMMON ON THE SURFACE. ORGANIC ENRICHMENT OF MUDSTONES ABOVE THE SURFACE POSSIBLE.	*SOIL OR ROOT HORIZONS ABSENT.* THIN CARBONATE GRAINSTONES COMMON ON THE SURFACE. ORGANIC ENRICHMENT OF MUDSTONES ABOVE THE SURFACE POSSIBLE.

Figure 26—Variations in type-1 sequence boundary expressions on the shelf, illustrated in map and cross-section view. Depositional environments, stratal terminations, and other diagnostic criteria of type-1 sequence boundaries and marine-flooding surfaces are summarized on the chart.

front of incised valleys or are partially to totally preserved and "stranded" on the shelf at the edges of, or adjacent to, the incised valleys, and (3) the "stranded" parasequences are overridden by the subaerial-exposure surface of the sequence boundary.

These "stranded" lowstand parasequences represent early lowstand systems tract deposition on the shelf or ramp. In basins with a shelf break, these parasequences could predate submarine-fan deposition before the sea-level fall reaches the shelf edge. Although they form during the early part of the sea-level fall, they are overlain by a regionally extensive unconformity marked by subaerial exposure and truncation labelled on Figure 26 as SB1, SB2, SB3, and FS/SB. Although it does not record the time of the initial sea-level fall over its entire extent, this unconformity is the sequence boundary because (1) it separates all of the rocks below from the rocks above; (2) although all points on the surface do not represent the same duration of time, one instant of time is common to all points when the sea-level fall ends and the unconformity is completely formed; (3) it is readily identified over most of its extent; (4) it is the surface that controls the distribution of overlying strata in the lowstand systems tract on the shelf; and (5) it forms relatively quickly, probably in less than 10,000 years.

The "stranded" lowstand parasequences below the sequence boundary commonly have the following stratal characteristics:

(1) they typically are deltaic or beach parasequences, but commonly consist of sharp-based, lower-shoreface sandstones;
(2) they have no significant updip coastal-plain equivalents, and there is no sediment accommodation updip because of the sea-level fall;
(3) they rest, commonly abruptly, on open-marine strata, although their bases cannot be interpreted as a basinward shift in facies;
(4) they rest on a conformable surface, and each parasequence gradually shoals upward;
(5) they are overlain by the unconformable part of the sequence boundary marked either by minor truncation or subaerial exposure; and
(6) they generally are thin because of reduced accommodation on the shelf; their thicknesses typically do not exceed tens of feet; and they also may vary in thickness due to a varying amount of truncation below the overlying sequence boundary.

Paleovalley distribution on the shelf is often controlled by tectonic features such as basement-involved faults, thrusts, and growth faults. Structural lows caused by salt withdrawal also control valley distribution. In many cases, the paleovalleys deposited in low areas controlled by tectonics or salt are incised and can properly be called incised valleys. In other cases, especially when the topography created by the tectonics or salt is not subdued, the paleovalleys have little or no truncation at their bases. When little or no truncation exists, the sequence boundary is still marked by a basinward shift in facies at the base of the paleovalley fill, but the paleovalley cannot properly be described as incised.

Correlation Pitfalls

To interpret type-1 sequence boundaries correctly in well logs, cores, or outcrops, it is critical to distinguish between incised valleys and local channels, such as distributary channels, in constructing an accurate chronostratigraphic framework. In the examples presented in Figures 21 through 25, we interpreted the vertical association of facies on the cross sections as incised valleys and not distributary channels or other local channels because the valleys are too wide to be distributary channels, the strata at the edges of the incised valleys are distal-marine sandstones and shelf mudstones, not delta-front or stream-mouth bar deposits, and valley fills occur along certain surfaces, i.e., sequence boundaries, that are widespread in the basin and not confined to one deltaic lobe. Criteria for the differentiation of incised valleys from distributary channels in a single well log and on a well-log cross section or in an outcrop are explained more fully in the following paragraphs.

Incised-valley interpretation is more difficult in a single well log than on a cross section because distributary channels, eroding deeply into underlying deltaic deposits, can juxtapose relatively coarse-grained strata directly on prodelta mudstones thereby mimicking a basinward shift in facies. However, where a distributary channel of a given delta lobe cuts into but not through the prodelta mudstones of the same lobe, the thickness of the distributary-channel fill cannot be much greater than the paleowater depth of the eroded mudstones. For example, if prodelta mudstones were deposited in 100 ft (30 m) of water, the fill of the distributary channel eroding into them must be nearly 100 ft (30 m) thick. This is not necessarily the case with incised valleys. Because incised valleys erode in response to a relative fall in sea level, the paleowater depth of the eroded mudstones beneath the sequence boundary is commonly much greater than the thickness of the valley fill. For example, shelf mudstones deposited in 300 ft (92 m) of water can be truncated by an incised valley only 30 ft (9 m) thick or less. As important as this relationship is, it is not always possible to determine accurately the paleowater depth of the strata imaged on a well log. Cores, cuttings, or an outcrop, if available, may provide enough data to interpret the paleowater depth.

Another important distinction between distributary channels and incised valleys that may be recognized in a core or outcrop is that the sequence boundary at the base of an incised valley commonly shows evidence of a hiatus between the times of erosion and deposition. Root zones, soils, or burrowed horizons can form on the valley floor during sea-level lowstand but before the valley is flooded and filled with sediment (Weimer, 1983). A distributary channel is always full of fresh water, or if discharge is low, salt water. It is unlikely that evidence of significant subaerial exposure will occur on a distributary-channel floor.

On a well-log cross section or in a relatively continuous outcrop, differentiation between incised valleys and distributary channels depends on an analysis of channel width and lateral-facies relationships. Distributary channels are relatively narrow. The distributary channels of the modern Mississippi River range from 500 to 5500 ft (153 to 1673 m) wide. Incised valleys are commonly several miles wide (Figures 21, 22, and 23) to many tens of miles wide (Figure 23). These widths can be identified on cross sections or in outcrops, and if possible, should be mapped regionally. Furthermore, widespread incised-valley erosion occurs along a single stratigraphic surface. Deltaic distributary channels usually stack to form multiple horizons.

It is critical to analyze the facies encasing the channel in order to distinguish between distributary channels and incised valleys. Distributary channels are encased in delta-plain or stream-mouth bar deposits (Figure 27). Even when the distributary channel of a given lobe erodes through the prodelta of that lobe into an underlying parasequence, most of the distributary-channel fill is laterally encased in stream-mouth bar deposits. Distributary channels can only step seaward if they have a subaqueous, shallow-water delta platform across which they can migrate. By their nature, distributary channels cannot be encased regionally in deeper-water deposits. For much of their length, incised valleys commonly are encased in middle- to outer-neritic mudstones because they incise during a relative fall in sea level.

Sequences in Outcrop and Subsurface

Examples of type-1 sequences and sequence boundaries, component parasequence sets, systems tracts, and facies associations in outcrops and well-log cross sections are illustrated in Figures 28 through 33. Each sequence in these examples is bounded by unconformities or their correlative conformities and contains lowstand, transgressive, and highstand systems tracts

The first example is from Cretaceous (Campanian) outcrops of the Grassy and Desert members of the Blackhawk Formation and the Castlegate, Buck Tongue, and Sego members of the Price River Formation exposed in the Book Cliffs between Green River, Utah and the Utah-Colorado border (Young, 1955; Hale and Van De Graaff, 1964; Van De Graaff, 1970; Gill and Hail, 1975; and Pfaff, 1985). Eight sequences are exposed in the cliffs, as follows: one sequence in the upper part of the Grassy and lower Desert members, one sequence within the upper Desert Member, another one within the Castlegate and Buck Tongue members, two sequences within the lower part of the Sego Member, and three sequences within the upper part of the Sego Member. These sequences were deposited on a ramp margin.

Figures 28 and 29 show the well-log response through the stratigraphic interval containing the sequences, parasequence sets, systems tracts, and parasequences in the Tenneco Rattlesnake State 2-12 (Figure 28) and the Exxon Production Research Co. (EPR) Sego Canyon no. 2 (Figure 29). The systems tracts are identified in the well logs using parasequence-stacking patterns and facies interpretations from outcrops and cores. The Tenneco well is 11 mi (18 km) north of the Desert and Castlegate outcrops, nearly on depositional strike with these strata; the Exxon well is 2 mi (3.2 km) north of the outcrops. The Sego, Buck Tongue, Castlegate, and Desert members were cored continuously in this well.

A measured section through the Sego, Buck Tongue, Castlegate, and Desert members at Thompson Canyon is illustrated in Figure 30. The measured section documents the vertical-facies associations and the sequence stratigraphy of these units.

A simplified outcrop cross section of the sequences and systems tracts in the upper part of the Desert Member, and the Castlegate, Buck Tongue, and Sego members is illustrated in Figure 31. Photographs of these strata in the cliff face at the Crescent Flat location on the cross section are also shown. This cross section is based on 135 sections measured between Green River, Utah, and Hunter Canyon, Colorado, supplemented with numerous outcrop panoramas.

The cross section in Figure 31, oriented west-southwest to north-northeast, is close to a depositional dip section with respect to the Castlegate Member, but is close to a depositional strike section with respect to the Sego. This occurs because, in the area of the thickest Castlegate exposure, the cliffs change orientation from east–west between Green River and Sagers Canyon, to northeast–southwest from just east of Sagers Canyon to the Colorado-Utah border, where the Sego is best exposed. The depositional dip for the Castlegate is to the southeast; the depositional dip for the Sego Member is to the south and southwest.

A map locating the Tenneco Rattlesnake State 2-12 (Figure 28), the EPR Co. Sego Canyon no. 2 (Figure 29), the measured section at Thompson Canyon (Figure 30), and the outcrop cross section (Figure 31), is illustrated in Figure 32.

Three backstepping parasequences near the top of the Grassy Member form the transgressive systems tract (Figure 28) of sequence 1. The parasequences are overlain by the highstand systems tract (Figures 28 and 29), which consists of a progradational parasequence set within the lower part of the Desert Member. Sequence 2 begins with a sequence boundary at the base of the upper part of the Desert Member (Figures 30 and 31), and is marked in outcrop by truncation and a basinward shift in facies. These two attributes can be traced 30 mi (48 km) down depositional dip from Tuscher Canyon to Sagers Canyon (Figure 31). The sequence boundary is the base of a regional incised valley; the lowstand systems tract of sequence 2 within the valley is composed of braided-stream, point-bar, and estuarine sandstones and mudstones arranged in an aggradational parasequence set. Along the outcrop, especially between Hatch Mesa and Coal Canyon, the incised valley cuts deeply into the underlying strata, resulting in juxtaposition of coal-bearing, coastal-plain rocks above the sequence boundary directly on dark gray, shelf mudstones.

The transgressive systems tract of sequence 2 (Figures 30 and 31) is best developed at Crescent Flat East and eastward to Sagers Canyon, where most of this systems tract changes facies to shelf sandstones and mudstones. Over this area, the transgressive systems tract consists of two lower-shoreface parasequences within a retrogradational parasequence set lying in sharp contact with a well-developed coalbed. The highstand systems tract is developed locally between Thompson and Sagers canyons (Figure 31), where it consists of beach parasequences stacked in a progradational parasequence set. Much of the transgressive and highstand systems tracts of sequence 2 is truncated by the boundary of sequence 3.

Sequence 3 (Figures 28 through 31) includes the Castlegate and Buck Tongue members of the Price River Formation. Sequence-boundary 3 is marked in outcrop by a basinward shift in facies and truncation (Figure 31) that can be traced at least 40 mi (64 km) down depositional dip from Tuscher Canyon to between Sagers and Cottonwood canyons (Figure 32).

The lowstand systems tract of sequence 3 is composed of fluvial sandstones and mudstones, coals, and estuarine sandstones and mudstones within the Castlegate Sandstone. The fluvial and estuarine rocks fill broad, coalesced, incised valleys that extensively dissect the underlying systems tracts of sequence 2. Sequence-boundary 3 incises progressively more deeply into the underlying Desert Member (Figure 31) in a landward and westward direction. Near Woodside Canyon, northwest of the town of Green River (Figure 32), the Desert Member, including all of sequence 2 and most of sequence 1, is absent because of this incised-valley truncation (Young, 1955). Alternately, sequence-boundary 3 incises progressively less deeply into the underlying Desert Member (Figure 31) in a seaward direction to the east. Between Sagers and Cottonwood canyons (Figure 32), the unconformable part of the sequence boundary merges with the top of the youngest lower-shoreface parasequence on the shelf and becomes a conformable surface (Figure 31). At West Salt Creek Canyon, a ferruginous, phosphatic oolite at the conformable surface lies on shelf mudstones and distal lower-shoreface hummocky beds, attesting to the substantial shallowing that must have occurred along this sequence boundary.

The transgressive systems tract in sequence 3 is composed of two parasequences in a retrogradational parasequence set and is best expressed in the EPR Co. Sego Canyon no. 2 (Figure 29). The highest organic-rich mudstones, with total organic-carbon values of 10%, lie on the transgressive surface at the top of the Castlegate, which is within the lowest part of the deepening-upward transgressive systems tract.

The highstand systems tract of sequence 3 (Figure 30) is composed of one complete and one incomplete beach parasequence within a progradational parasequence set (Figures 28 through 30).

Sequences 4 through 8 (Figures 29 and 30) occur within the Sego Member of the Price River Formation. Each sequence boundary is marked by regional-erosional truncation associated with incised valleys and a basinward shift in facies. Based on clinoform directions, orientations of channel-cross sections, and 408 paleocurrent directions measured on sigmoidal- and trough-cross beds, Sego incised valleys are oriented north-south and northeast-southwest with paleoflow to the south and southwest.

Sequence-boundary 8 is a major regional-erosional surface at the top of the lowstand-estuarine sandstone in sequence 7. This regional-erosional surface has as much as 100 ft (30 m) of relief locally and is overlain by fluvial sandstones, mudstones, and coals everywhere in the area studied. Some of the thickest coals in western Colorado and eastern Utah are in the lowstand systems tract of this sequence. In the area studied, sequence 8 is composed entirely of nonmarine strata.

Sequences 4 through 7 have similar systems tracts, facies associations, and sequence-boundary expressions; these attributes are summarized in the following description of sequence 4 (Figure 30). Sequence-boundary 4 is marked by truncation and a basinward shift in facies at the base of a regionally extensive incised valley approximately 15 mi (24 km) wide; incised valley edges can be seen clearly in outcrop. Shelf mudstones or wave-rippled siltstones and interbedded mudstones below the sequence boundary are overlain by upper-fine- to medium-grained, well-sorted sandstones above the sequence boundary. In places, a channel lag of clay clasts, shell and bone fragments, and phosphorite pebbles occurs at the base of the incised-valley sandstones. More commonly a lag of red clay clasts rests on the sequence boundary. The sandstones are composed of sigmoidal cross bedsets (Mutti et al., 1984, 1985) up to 3 ft (1 m) thick, with

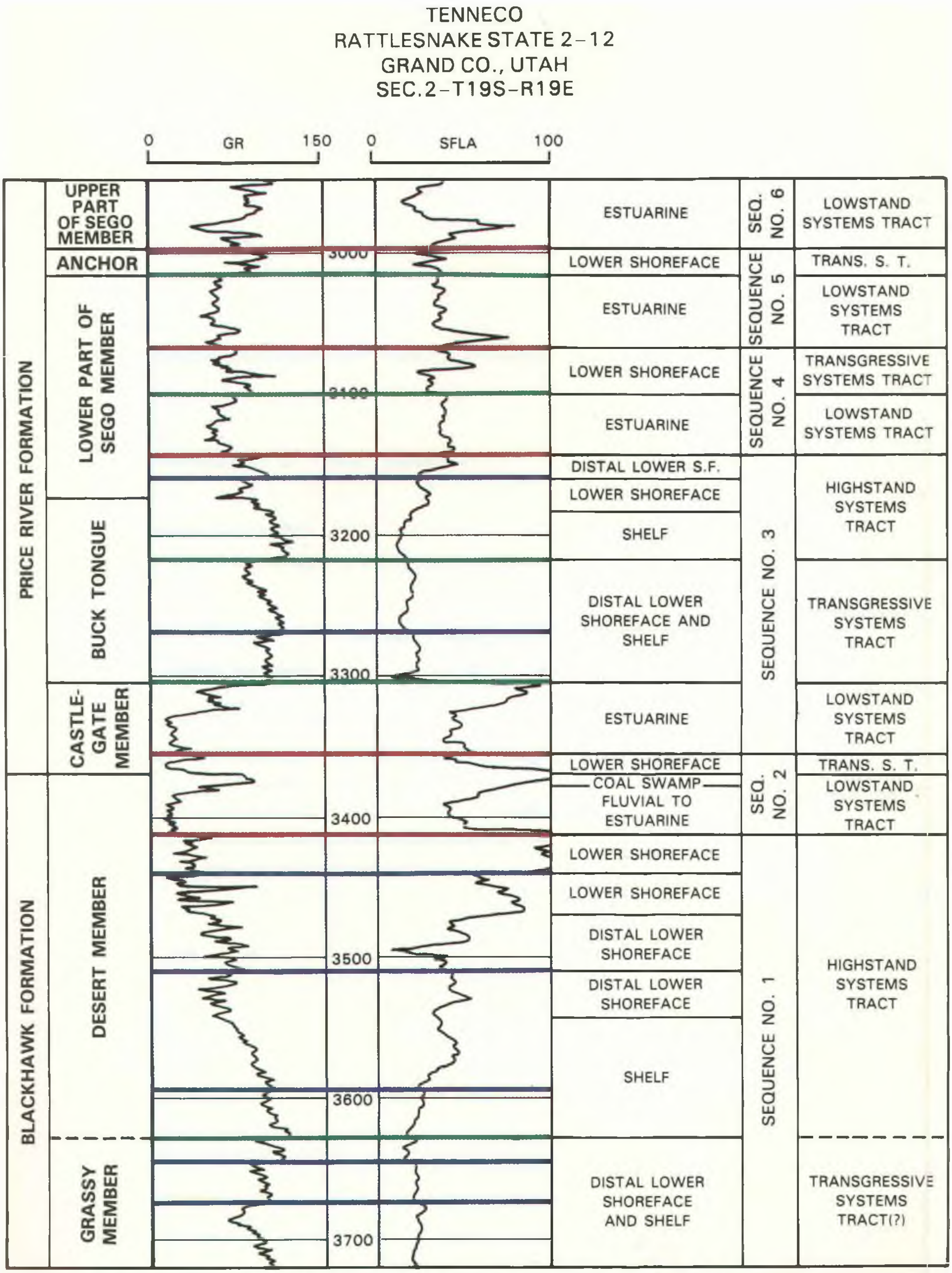


Figure 28—Well-log expression of sequence stratigraphy in the Tenneco Rattlesnake State 2-12, Book Cliffs, Utah.

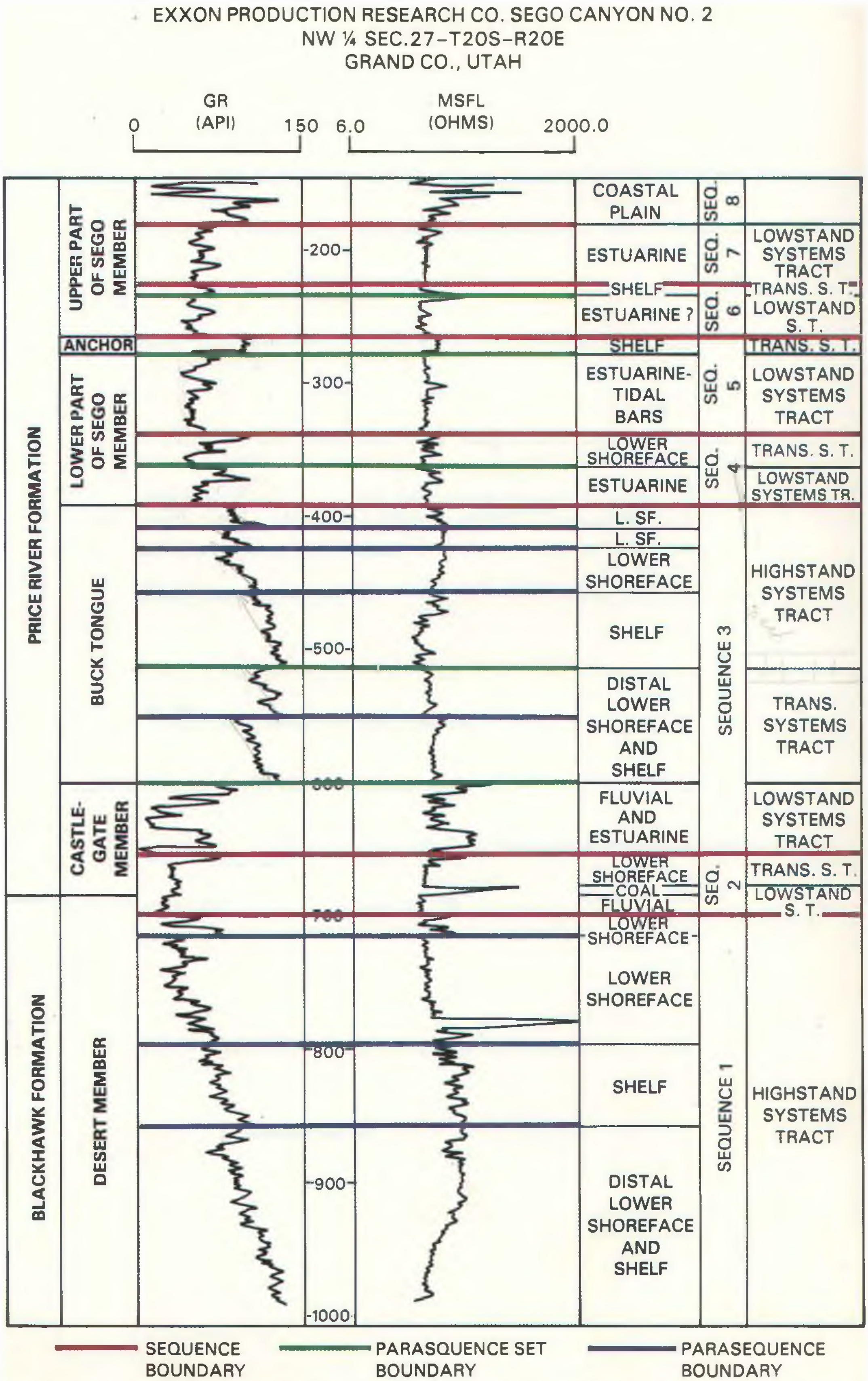


Figure 29—Well-log expression of sequence stratigraphy in the Exxon Production Research Co. Sego Canyon no. 2, Book Cliffs, Utah.

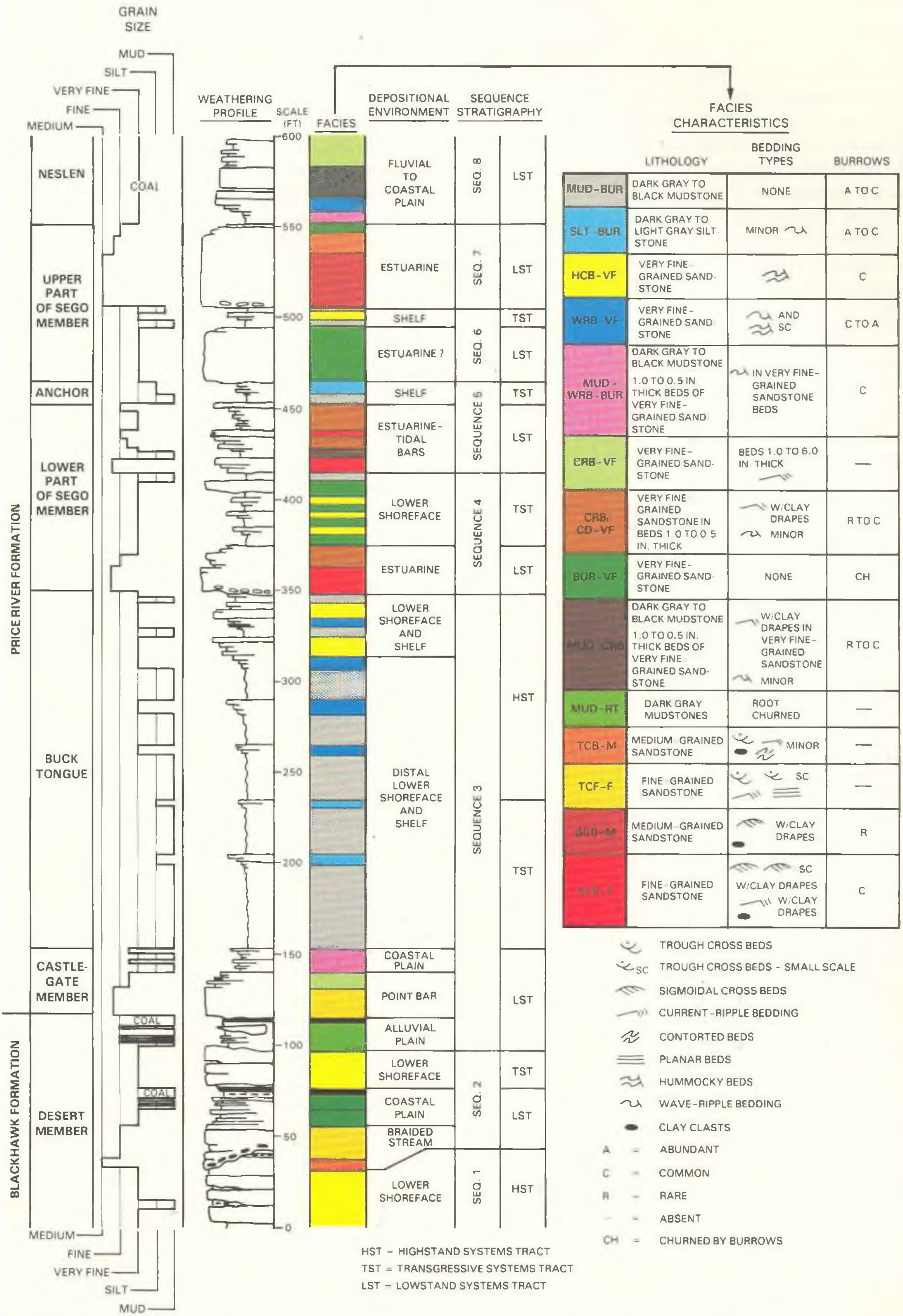


Figure 30—Measured section showing sequence stratigraphy and facies of Desert, Castlegate, Buck Tongue, and Sego members in Thompson and Sego canyons, Book Cliffs, Utah.

clay drapes on each foreset lamina. Foreset toes are extremely tangential and interbedded with clay drapes, small clay clasts, and current ripples. Reactivation surfaces are common within the cross bedsets; the upper bounding surfaces of the cross bedsets typically are convex upward. Tidal bundles can be recognized locally in the sandstones. Burrowing is minor within the sandstones, but when present is generally that of *Ophiomorpha* or *Thalassinoides*.

In some places, the sandstones lying on the sequence boundary have a more gradational base. In these places, the vertical succession begins with the above described lag, which is overlain by thin, upper-fine-grained, current-rippled sandstones with abundant interbedded clay drapes and minor zones of bioturbation. The sandstone beds gradually thicken upward, and there is a progressive decline in the amount of clay drapes; small-scale sigmoidal cross beds, up to 6 in. (15 cm) thick, and current ripples predominate. The upper part of this prograding unit is composed of the large-scale sigmoidal cross beds (described above) with minor current-ripple deposition.

These sandstones, lying on the sequence boundary, are interpreted to be tidal bars and shoals (Mutti et al., 1985) within a tide-dominated delta prograding into an estuary created by the flooding of the incised valley. These estuarine sandstones represent the lowstand systems tract of sequence 4. A sharp, planar surface separates the lowstand sandstones below from a 2- to 8-ft- (0.6- to 2.4-m-) thick interval of very fine-grained, hummocky-bedded sandstones above. The hummocky-bedded sandstones are overlain either by a parasequence or a sequence boundary. In the case of the parasequence boundary, marine mudstones and thin, wave-rippled, very fine-grained sandstones, locally up to 20 ft (6 m) thick, lie directly on the hummocky-bedded sandstones recording an increase in water depth. The top of the mudstones and thin sandstones is truncated by the next sequence boundary. The hummocky-bedded sandstones and overlying shelf mudstones and thin sandstones are interpreted as backstepping parasequences in the transgressive systems tract. If a highstand systems tract is present in the sequence, it is very thin and fine grained.

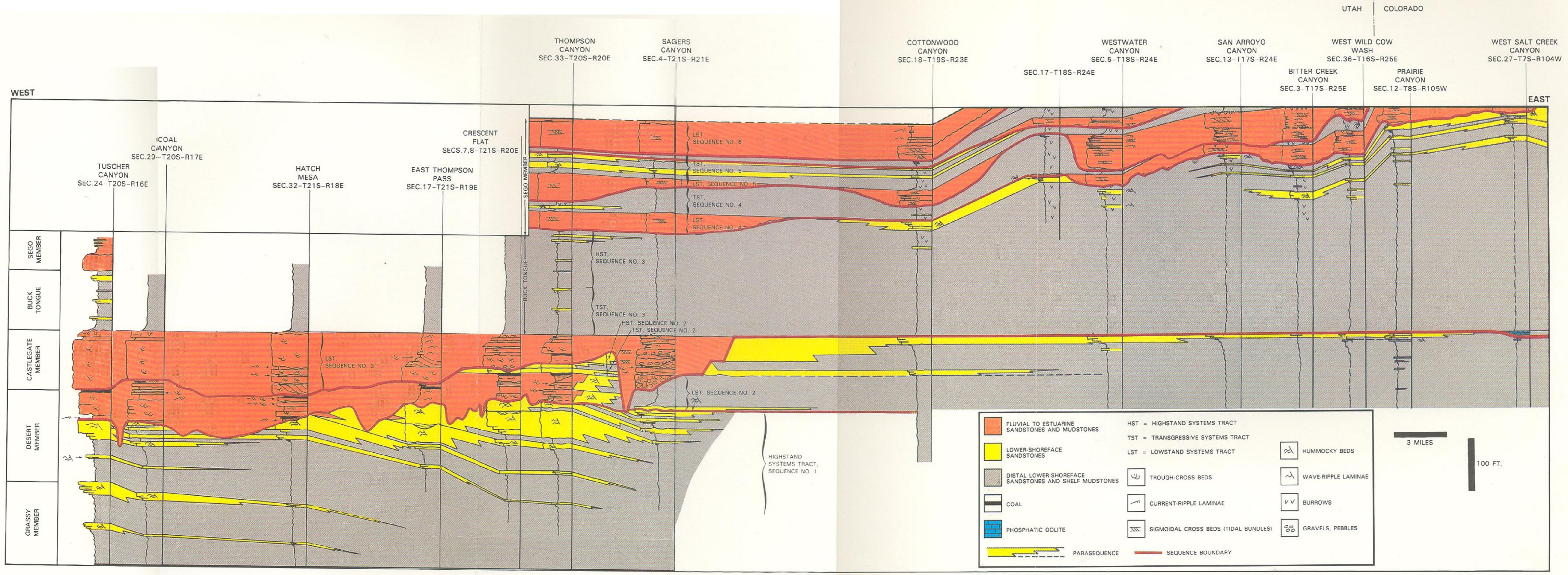
In the case of the sequence boundary, medium-grained estuarine sandstones erode into the lower-shoreface deposits and record a relative fall in sea level. It is important to note that the estuarine sandstones grade basinward into thinner, current-rippled sandstones and interbedded mudstones that in turn grade into shelf mudstones. Lower-shoreface, hummocky-bedded sandstones are not lateral-facies equivalents of the estuarine strata. In a landward direction, the estuarine deposits become sandier and coarser grained, eventually grading transitionally into coarse-grained, braided-stream sandstones and conglomerates.

Tide-dominated deltas were deposited within the partially flooded incised valley of sequence 4 during the early stages of a sea-level rise, presumably because the linear, relatively narrow embayment focused tidal currents. The incised valley gradually filled with tide-dominated deposits, while there was no deposition on the subaerially exposed shelf adjacent to the incised valley at this time. As sea level continued to rise, most of the shelf flooded. This flooding finally terminated tidal deposition within the incised valley, and created conditions for deposition of sheet-like, wave-dominated deposits over the entire shelf. The sharp contact separating the estuarine sandstones from the overlying lower-shoreface sandstones records this flooding. The progradation direction of the wave-dominated shoreline deposits of the transgressive systems tract appears to be oriented nearly parallel to the longitudinal axes of the incised valleys.

As the previous examples (Figures 28 through 30) show, the lithostratigraphic subdivision of these Cretaceous rocks does not always correspond to the chronostratigraphic or sequence subdivision (Figures 28 and 29). For example, the sequence boundary within the Desert Member separates the lower part of the Desert, interpreted as a highstand systems tract for sequence 1, from the upper part of the Desert, interpreted as a lowstand systems tract for sequence 2, with a potentially large, intervening stratigraphic gap. The sequence boundaries record the fundamental breaks in deposition; at each sequence boundary the "slate is wiped clean" and a new depositional record begins. Lithostratigraphic subdivisions commonly miss these fundamental boundaries, making it difficult to construct accurately a chronostratigraphic and regional-facies framework. Once the sequence-stratigraphic subdivision is made, the lithostratigraphic terminology is often so confusing that it needs to be modified substantially or abandoned.

The second example of sequences is a well-log cross section through middle Miocene strata of onshore Louisiana. The cross section is illustrated in Figure 33. These sequences are typical of much of the Tertiary rocks in the Gulf Coast basin. Five sequences can be recognized on this cross section (Figure 33). Each sequence boundary is marked by erosional truncation and a basinward shift in facies. The sequence boundaries have been mapped by means of these criteria, using nine other regional cross sections constructed from 700 well logs in addition to the cross section illustrated here; the sequences can be recognized over an area of at least 5600 mi^2 (14,500 km^2) in central and southern Louisiana. The systems tracts, parasequence sets, and facies within the five sequences are similar; sequence 1 typifies the distribution of these stratal components.

SIMPLIFIED CROSS SECTION FROM TUSCHER CANYON TO THE COLORADO-UTAH BORDER ILLUSTRATING SEQUENCES, COMPONENT PARASEQUENCE SETS, AND FACIES IN THE UPPER BLACKHAWK FORMATION (DESERT MEMBER) AND THE LOWER PRICE RIVER FORMATION (CASTLEGATE AND SEGO MEMBERS), CAMPANIAN IN AGE.



TWO PANORAMAS OF THE NORTH WALL OF THE CANYON WEST OF BLAZE CANYON (SEC.7–T21S–R20E). THE LOCATION OF THESE PANORAMA IS CALLED CRESCENT FLAT ON THE CROSS SECTION.

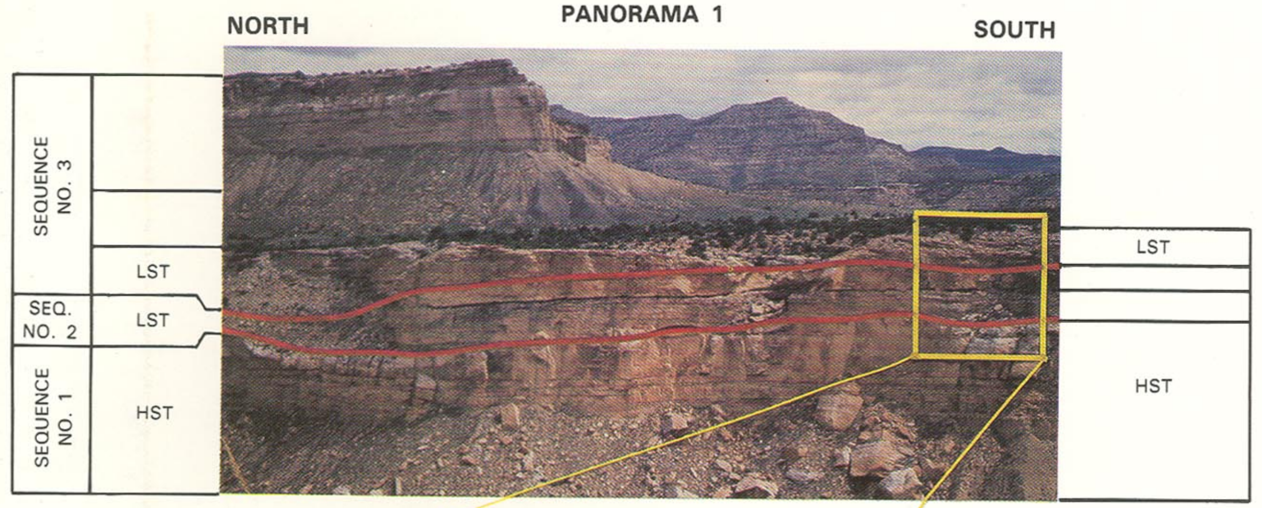


Figure 31—Sequences in the Castlegate, Desert, and Sego members outcropping in the Book Cliffs, Grand County, Utah.

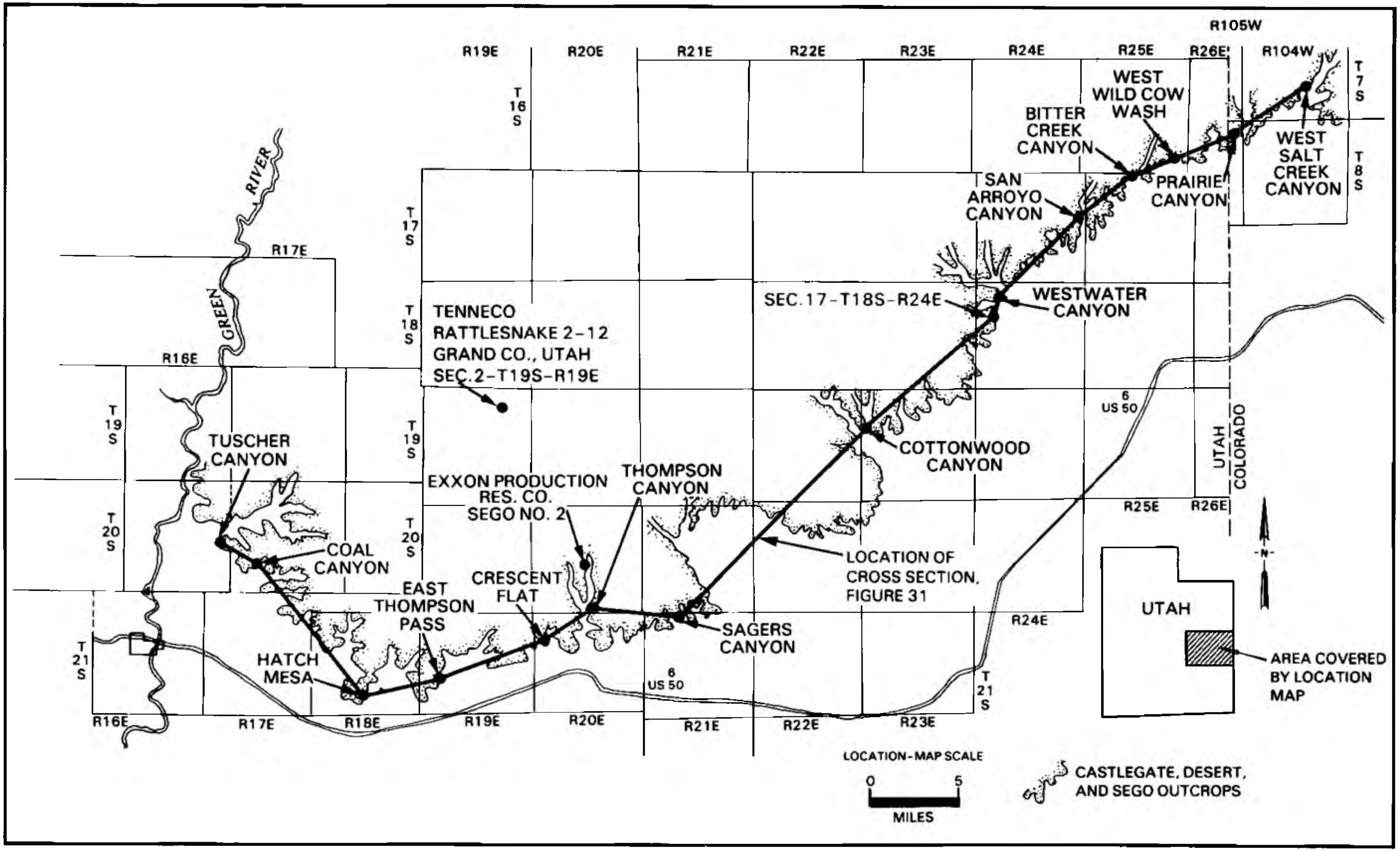


Figure 32—Map showing the location of the Desert, Castlegate, and Sego outcrops in eastern Utah, the Tenneco Rattlesnake State 2-12 (Figure 28), the EPR Co. Sego Canyon no. 2 (Figure 29), and the sequence cross section illustrated in Figure 31.

The lowstand systems tract of sequence 1 (Figure 33) consists of sandstones up to 250 ft (76 m) thick, characterized by a blocky to upward-fining SP well-log pattern. The sequence boundary at the base of the sandstones is a regional-erosional surface with local-erosional relief as great as 200 ft (61 m). The depositional environment of the sandstone is interpreted to have been fluvial or estuarine, filling a broad, incised-valley complex, based on log response and widely spaced core control. Maps constructed using the additional nine regional cross sections in the area show that the incised-valley complex is approximately 75 mi (120 km) wide. The depositional environment of the mudstones and thin sandstones below the sequence boundary is interpreted to have been middle to outer shelf, based on biostratigraphy and well-log responses. No intermediate water-depth deposits occur between the lowstand, incised-valley-fill sandstones and the underlying shelf mudstones of the previous sequence. Incised valleys of similar-aged sequences from Louisiana are illustrated in Figures 22 and 23.

The transgressive systems tract of sequence 1 (Figure 33) is composed of thin backstepping parasequences in a retrogradational parasequence set. A condensed section has not been identified in this systems tract. Only mudstones and very thin sandstones are preserved in the highstand systems tract. The coarser-grained part of the highstand systems tract apparently was truncated by the next sequence boundary. Erosion of the highstand systems tract by the overlying sequence boundary is common in many Tertiary sequences in the Gulf Coast basin. This pattern of systems tract distribution in sequence 1 is repeated in the other four sequences on the cross section.

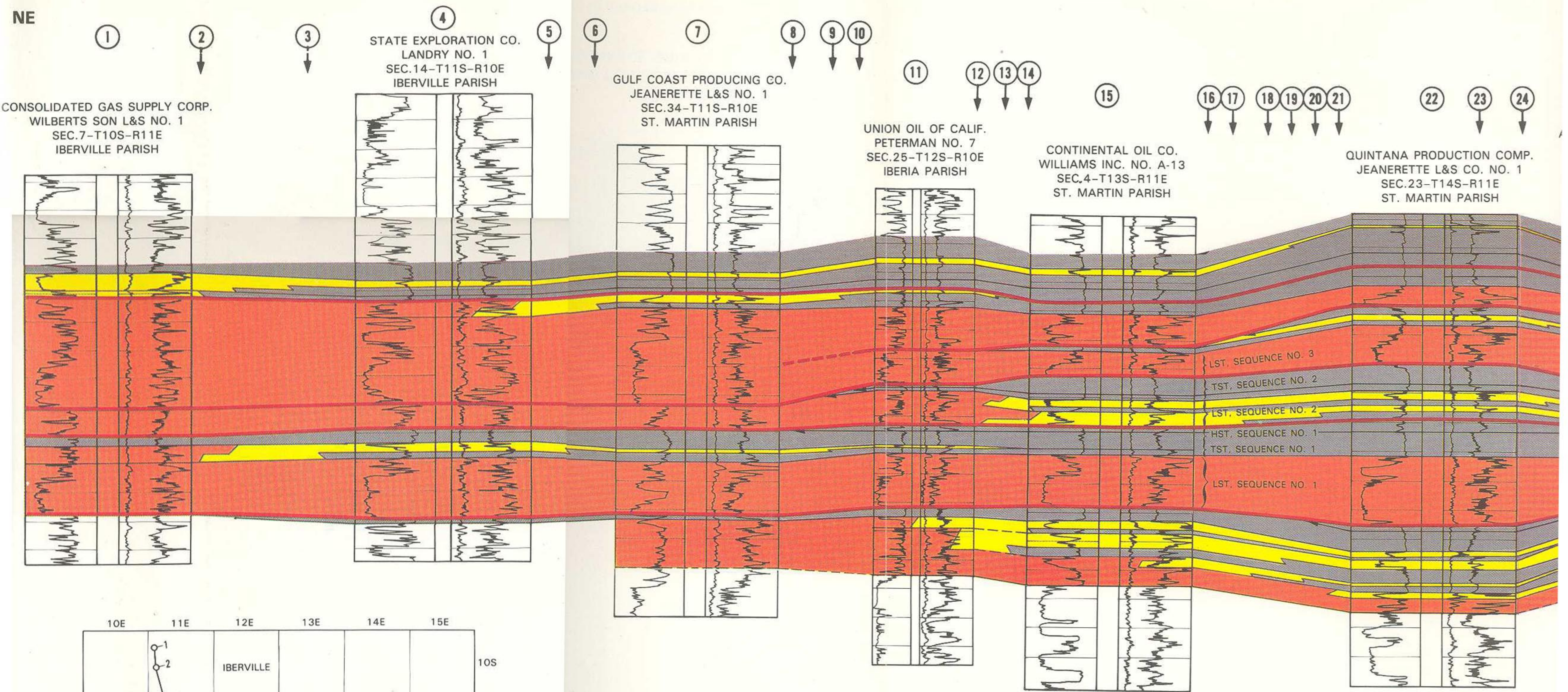
The mudstone in the transgressive and highstand systems tracts is within the *Cibicides opima* shale. Based on the fauna in this shale, the lower sequence boundary on Figure 33 is dated as 15.5 Ma (L.C. Menconi, personal communication, 1989) and appears on the Exxon global-cycle chart of Haq et al. (1988). The youngest sequence in Figure 33 is within the *Bigenerina humblei* biozone and corresponds to the Hollywood sandstone, an informal regional mapping unit within this biozone, suggesting an age date of 14.7 Ma for sequence boundary 5 (L.C. Menconi, personal communication, 1989). Based on these age dates, each of the five sequences in Figure 33 is interpreted to have been deposited during sea-level cycles lasting 100,000 to 200,000 years. These frequencies may be even higher if one assumes a significant hiatus on the third-order boundary representing basin-floor and slope-fan deposition. A model for the development of these high-frequency sequences and their implications for the interpretation of eustasy as a driving mechanism for sequence development are the topics of the next section.

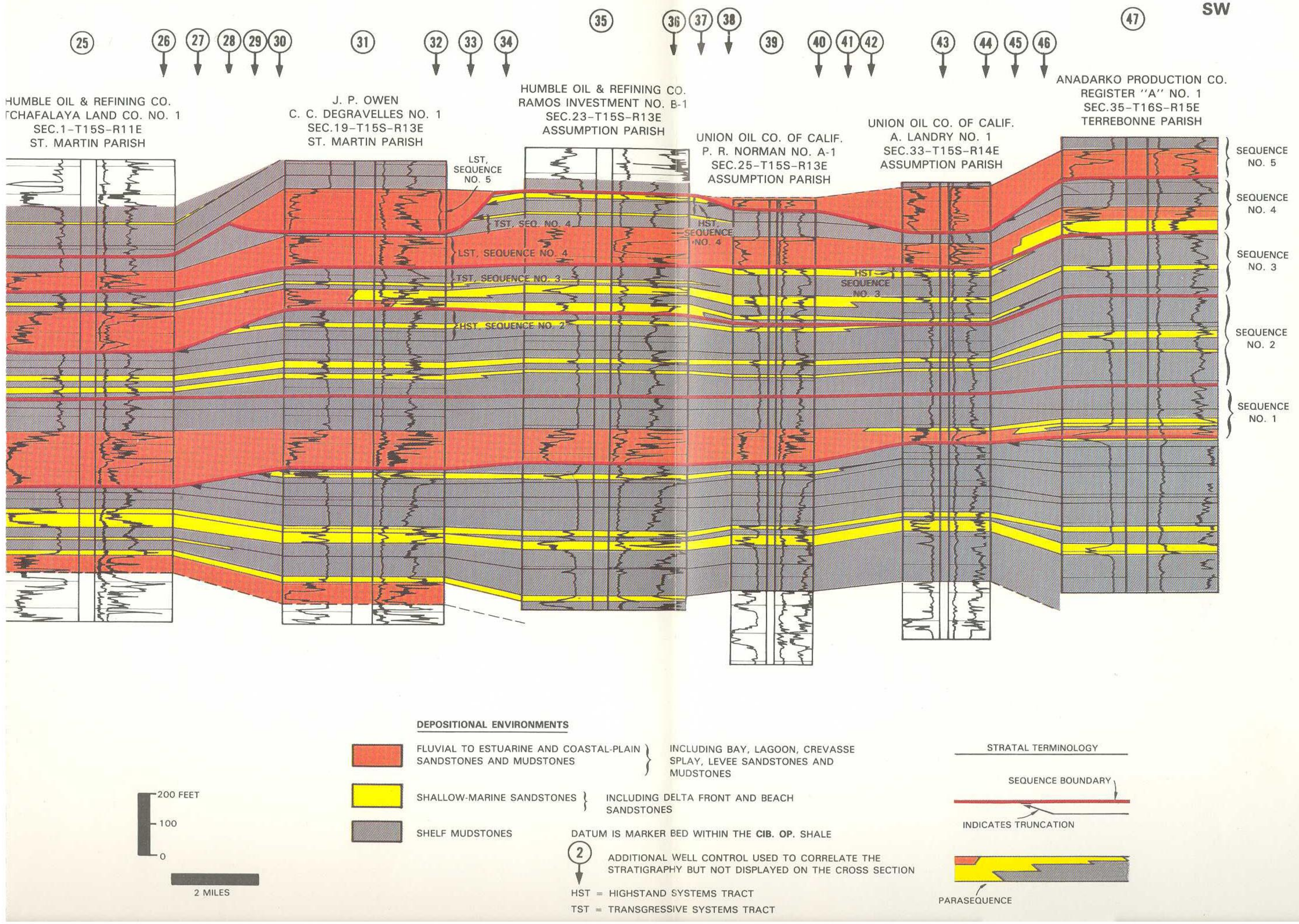
Interpretation of Depositional Mechanisms and Sequence Frequency

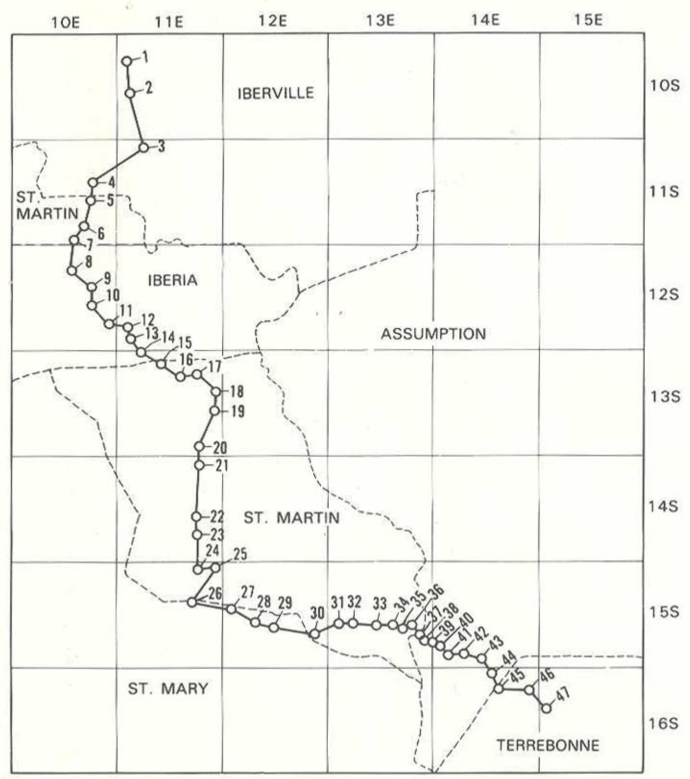
Sequences and their boundaries are interpreted to form in response to cycles of relative fall and rise of sea level. Jervey (1988) and Posamentier et al. (1988) presented an analysis of the interaction between eustasy (see figure 7, Posamentier and Vail, 1988) and basin subsidence that is interpreted to form sequence boundaries.

The interpreted relationship of stratal patterns to accommodation for a type-1 sequence with no significant incised-valley-fill deposition is shown in the block diagrams of Posamentier and Vail (1988, their figures 1 to 6). A variation of this idealized sequence, based on observations made in the Tertiary strata of the Gulf of Mexico, is shown in block diagrams in Figures 34 to 38 in this book. These block diagrams illustrate the successive evolution, over a period of 120,000 years, of a sequence similar to the sequences in Figure 33, with well-defined incised valleys and erosional truncation of the highstand systems tract. As the block diagrams illustrate, fluvial deposits within incised valleys are commonly coarse-grained, low-sinuosity channels reflecting slow rates of accommodation. Transgressive and early highstand-fluvial deposits are commonly finer-grained, high sinuosity channels and associated overbank strata reflecting high rates of accommodation. These two different fluvial-architectural patterns can be used as a guide to interpret sequences in totally nonmarine sections (Shanley and McCabe, 1989). A eustatic curve in the corner of each block diagram is color-coded to indicate the interpreted relationship of the systems tracts to eustasy. This eustatic curve is a graphic representation of the eustatic cycle of Jervey (1988), although at a higher frequency. Outcrop photographs illustrate the stratal characteristics of the facies that occur typically in each systems tract.

Parasequences and their boundaries also can be interpreted as responses to cycles of relative fall and rise of sea level. Sea-level cycles are classified by Vail et al. (1977) according to the duration of the cycle: third-order cycles, defined from fall to fall, have durations of 1 to 5 million years, fourth-order cycles have durations of hundreds of thousands of years. Following Vail et al. (1977) we assign to fifth-order cycles durations of tens of thousands of years. The relationship between this hierarchy of eustatic cycles, subsidence, and the deposition of sequences and parasequences is illustrated in Figure 39. In this figure, a third-order eustatic cycle (approximately one million years) is added to fourth-order cycles (approximately 120,000 years), and fifth-order cycles (approximately 50,000 years) to form a composite eustatic curve. Adding a total subsi-







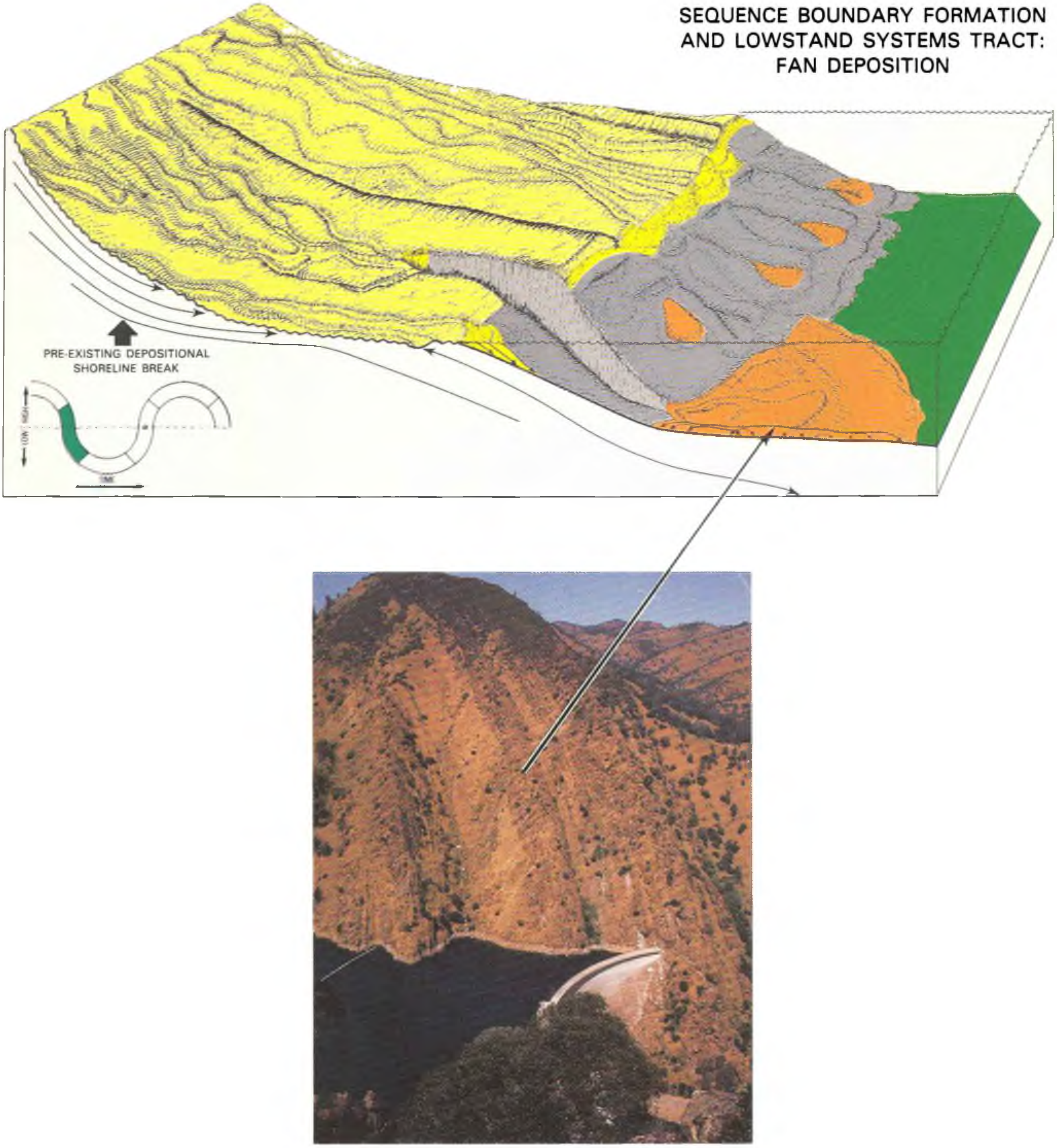
WELL-LOG LOCATION MAP
CENTRAL LOUISIANA

NO.	OPERATOR	WELL NAME	LOCATION	PARISH
1		ON CROSS SECTION		
2	WRATHER	SCHWING NO. 1	SEC.19–T10S–R11E	IBERVILLE
3	VAUGHN-AMARILLO-MCALESTER	WILBERT NO. 1	SEC.5–T11S–R11E	IBERVILLE
4		ON CROSS SECTION		
5	AMERADA-HESS ETHYL CO.	WILLIAMS NO. 3	SEC.23–T11S–R10E	ST. MARTIN
6	COASTAL STATES GAS CORP.	WILLIAMS NO. 1	SEC.35–T11S–R10E	ST. MARTIN
7		ON CROSS SECTION		
8	SAMEDAN	JEANERETTE NO. 1	SEC.10–T12S–R10E	IBERIA
9	CALIFORNIA	LEBOURGEOIS NO. 1	SEC.14–T12S–R10E	IBERIA
10	DIAMOND SHAMROCK	JEANERETTE NO. 1	SEC.23–T12S–R10E	IBERIA
11		ON CROSS SECTION		
12	UNION	PETERMAN NO. 10	SEC.30–T12S–R11E	IBERIA
13	UNION	PETERMAN NO. 1	SEC.31–T12S–R11E	IBERIA
14	CONTINENTAL	PETERMAN NO. 2	SEC.5–T13S–R11E	IBERIA
15		ON CROSS SECTION		
16	CONTINENTAL	JEANERETTE NO. B-3	SEC.10–T13S–R11E	ST. MARTIN
17	UNION	JEANERETTE NO. C-1	SEC.11–T13S–R11E	ST. MARTIN
18	UNION	JEANERETTE NO. 3	SEC.13–T13S–R11E	ST. MARTIN
19	SUN OIL CO.	SCHWING NO. 1	SEC.24–T13S–R11E	ST. MARTIN
20	CHEVRON	WILLIAMS NO. 2	SEC.35–T13S–R11E	ST. MARTIN
21	CALIFORNIA	JEANERETTE NO. 3	SEC.2–T14S–R11E	ST. MARTIN
22		ON CROSS SECTION		

NO.	OPERATOR	WELL NAME	LOCATION	PARISH
23	HUMBLE	GOODRICH NO. L-1	SEC.26–T14S–R11E	ST. MARTIN
24	HUMBLE	GOODRICH NO. C-3	SEC.2–T15S–R11E	ST. MARTIN
25		ON CROSS SECTION		
26	UNION	ST. LSE 1684 NO. 1	LAKE–T15S–R11E	–
27	CHEVRON	ST. LSE 4740 NO. 1	LAKE–T15S–R12E	ST. MARY
28	TENNESSEE	POLASKY NO. 1	SEC.36–T15S–R12E	ST. MARY
29	PAN AMERICAN	PUGH NO. 1	SEC.2–T15S–R12E	ST. MARY
30	HUNT TRUST	ST. LSE 5211 NO. 1	SEC.25–T15S–R12E	ST. MARTIN
31		ON CROSS SECTION		
32	QUINTANA	DEGRAVELLES NO. 1	SEC.20–T15S–R13E	ST. MARTIN
33	CALIFORNIA	ST. LSE 2810 NO. 2	SEC.21–T15S–R13E	ST. MARTIN
34	CALIFORNIA	ST. LSE 2162 NO. 1	SEC.22–T15S–R13E	ST. MARTIN
35		ON CROSS SECTION		
36	OCCIDENTAL	SHINN NO. 1	SEC.23–T15S–R13E	ST. MARTIN
37	HUMBLE	NORMAN NO. B-2	SEC.25–T15S–R13E	ASSUMPTION
38	HUMBLE	NORMAN NO. 1	SEC.25–T15S–R13E	ASSUMPTION
39		ON CROSS SECTION		
40	UNION	WILLIAMS NO. 9	SEC.30–T15S–R14E	ASSUMPTION
41	SUPERIOR	FRYMIRE NO. 1	SEC.31–T15S–R14E	ASSUMPTION
42	UNION	PENNISON NO. 1	SEC.32–T15S–R14E	ASSUMPTION
43		ON CROSS SECTION		
44	QUINTANA	LUTCHER MOORE NO. 1	SEC.64–T16S–R14E	ASSUMPTION
45	QUINTANA	MECOM NO. 1	SEC.71–T16S–R14E	ASSUMPTION
46	CALIFORNIA	MELANCON NO. 1	SEC.5–T16S–R14E	TERREBONNE
47		ON CROSS SECTION		

LIST OF WELL LOGS USED TO CONSTRUCT THE CROSS SECTION

Figure 33—Sequences from the middle Miocene of central Louisiana. The *Cibicides opima* shale composes most of sequence 2. The lower unconformity of sequence 1 is the 15.5-Ma sequence boundary on the cycle chart of Haq et al., 1987.

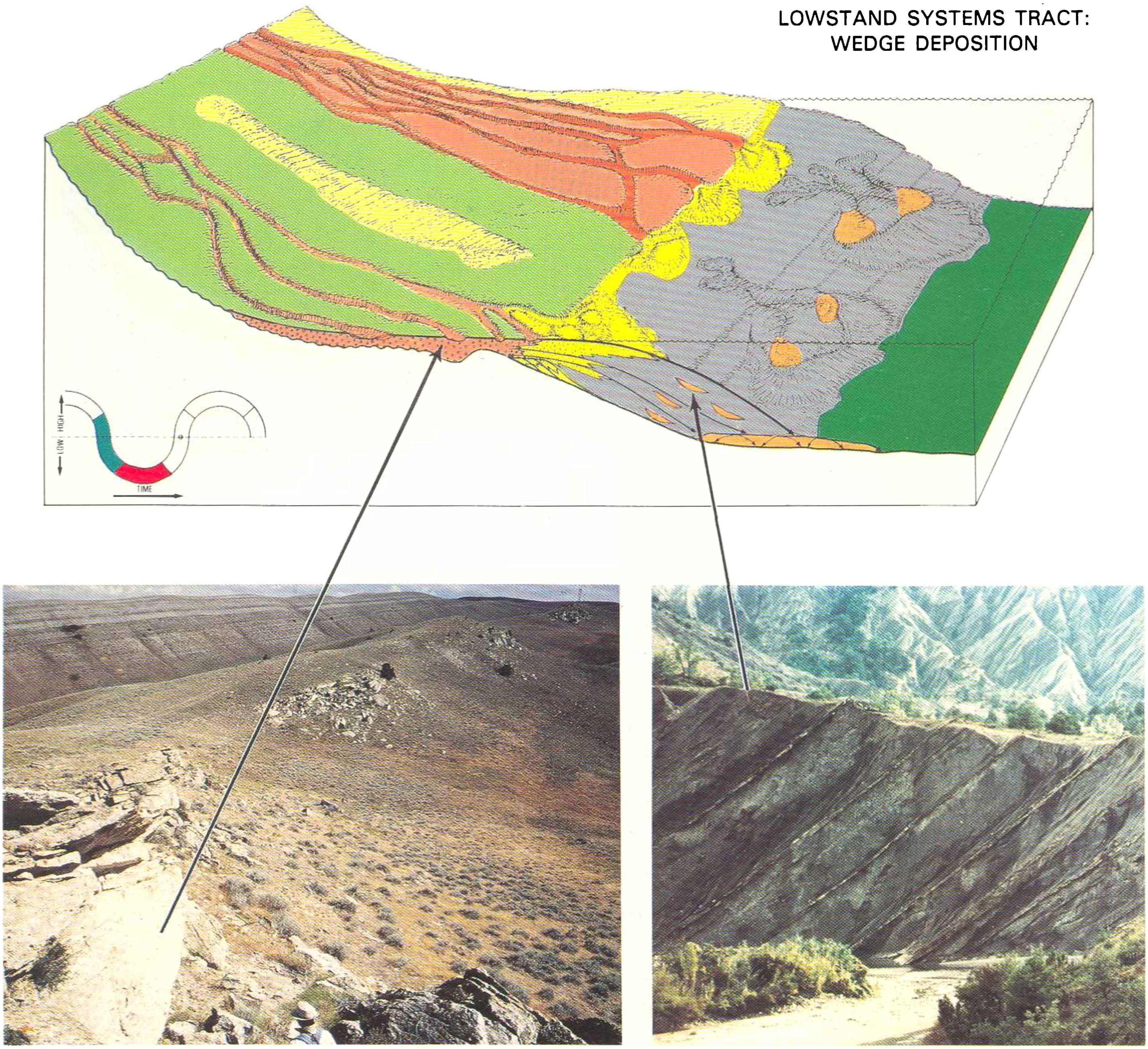


- RATE OF EUSTATIC FALL EXCEEDS RATE OF SUBSIDENCE
- SEA LEVEL FALLS TO SHELF BREAK, SHELF IS EXPOSED, INCISED; CANYON CUT
- SLOPE-PERCHED DELTAS AND SUBMARINE FANS ARE DEPOSITED

PHOTOGRAPH

SUBMARINE-FAN SANDSTONES; VENADO MEMBER, CORTINA FORMATION, TURONIAN, MONTICELLO DAM, CALIFORNIA

Figure 34—Sequence evolution: 1. Rapid relative fall of sea level.



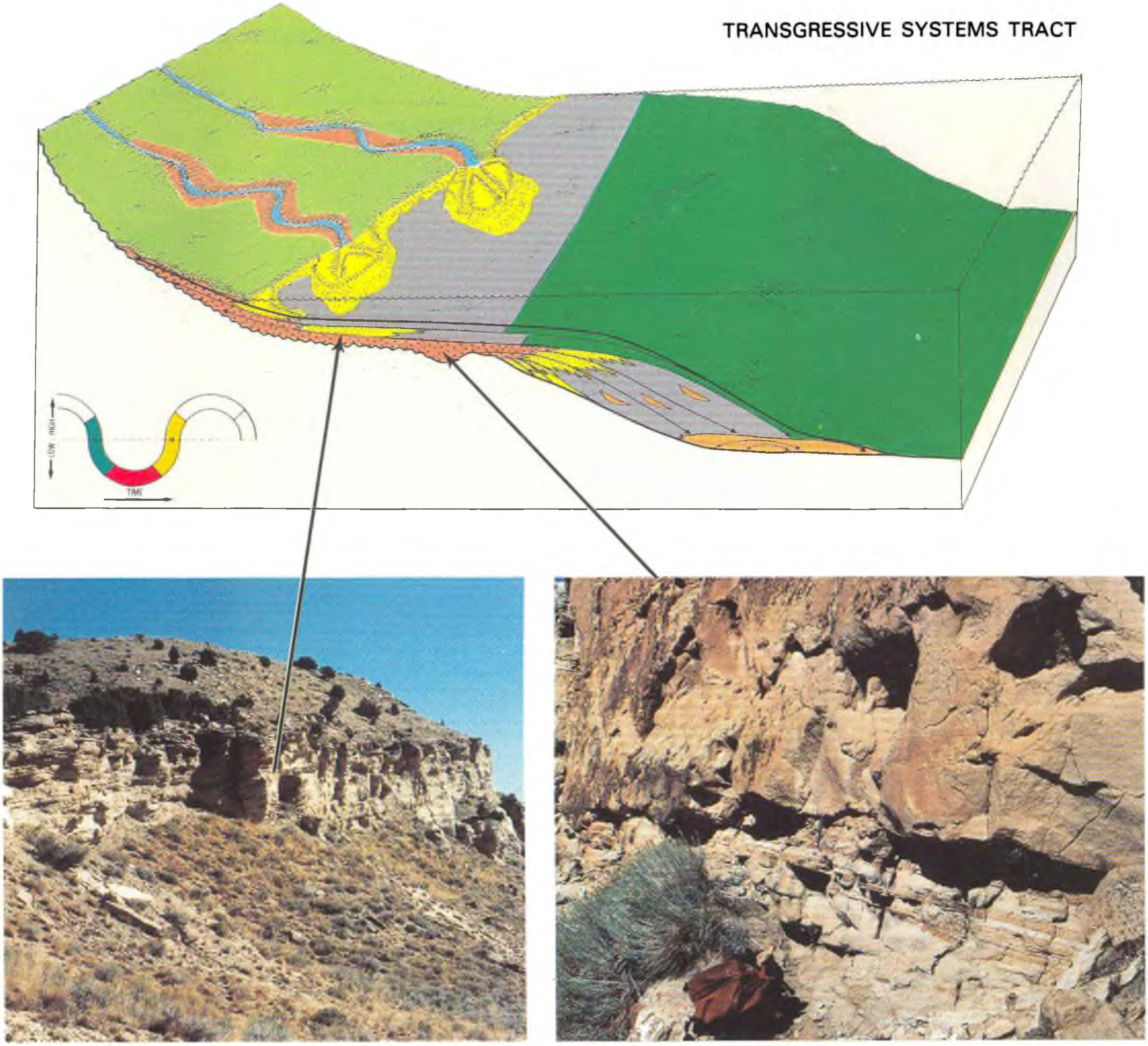
- RATE OF EUSTATIC FALL DECREASES, REACHES A STILLSTAND, AND RISES SLOWLY
- SUBMARINE-FAN DEPOSITION CEASES
- COARSE-GRAINED, BRAIDED-STREAM OR ESTUARINE SANDSTONES AGGRADE WITHIN THE FLUVIAL SYSTEMS OFTEN FILLING INCISED VALLEYS IN RESPONSE TO THE SEA-LEVEL RISE
- FINE-GRAINED TURBIDITES DEPOSITED ON THE SLOPE FORM A SHALE-PRONE WEDGE WITH THIN TURBIDITE SANDSTONE BEDS THAT DOWNLAP ON TOP OF THE FAN

PHOTOGRAPHS

LEFT: ESTUARINE INCISED-VALLEY FILL SANDSTONES; MUDDY SANDSTONE, WIND RIVER BASIN, WYOMING

RIGHT: LOWSTAND-WEDGE TURBIDITE SANDSTONES AND MUDSTONES, SPAIN

Figure 35—Sequence evolution: 2. Slow relative fall, stillstand, and slow relative rise of sea level.



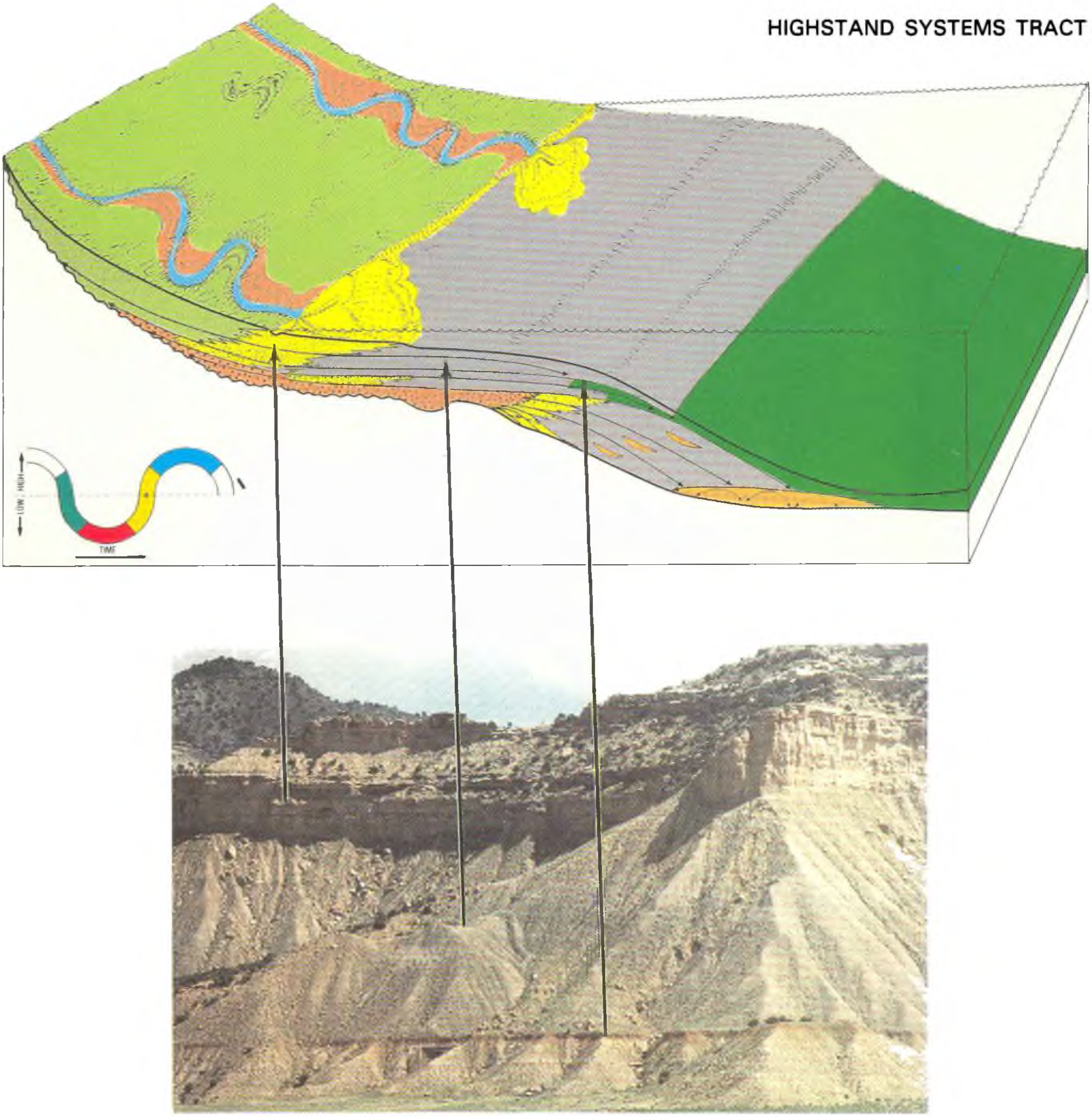
- RATE OF EUSTATIC RISE IS AT A MAXIMUM
- DURING BRIEF SLOWDOWNS IN RATE OF RISE PARASEQUENCES PROGRADE BUT OVERALL STACK IN BACKSTEPPING PATTERN
- ORGANIC-RICH FACIES (CONDENSED SECTION) MOVES UP ONTO THE SHELF
- FLUVIAL SYSTEMS TYPICALLY SHIFT FROM A BRAIDED TO MEANDERING PATTERN

PHOTOGRAPHS

LEFT: RETROGRADATIONAL PARASEQUENCE SET, TRANSGRESSIVE SYSTEMS TRACT; TOP TEAPOT SANDSTONE, BIG HORN BASIN, WYOMING

RIGHT: BRAIDED-STREAM INCISED VALLEY-FILL SANDSTONE; TEAPOT SANDSTONE, BIG HORN BASIN, WYOMING

Figure 36—Sequence evolution: 3. Rapid rise of sea level.

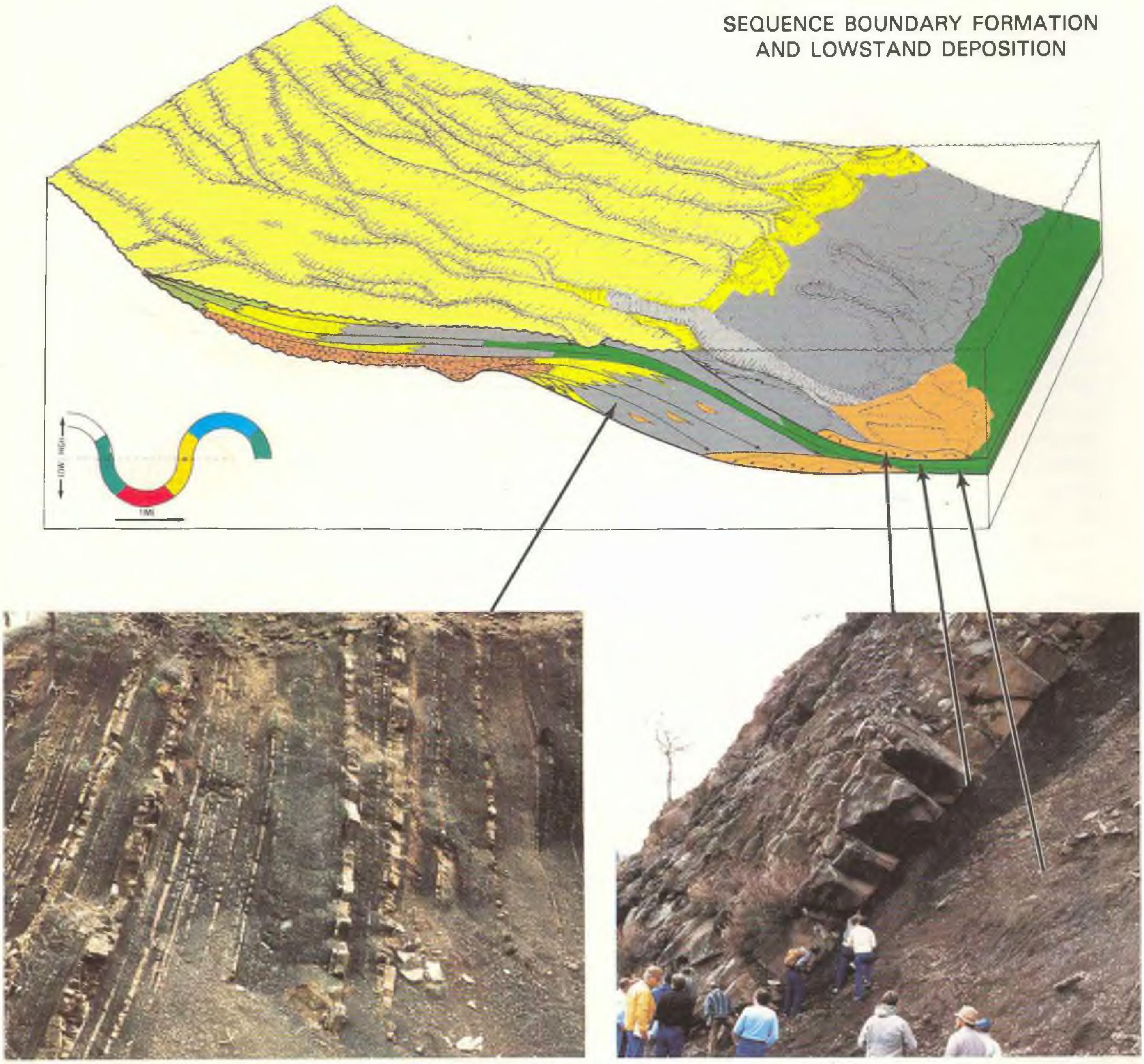


- RATE OF EUSTATIC RISE IS AT A MINIMUM AND IN THE LATE HIGHSTAND, FALLS SLOWLY
- RATES OF DEPOSITION GREATER THAN THE RATES OF SEA-LEVEL RISE, PARASEQUENCES BUILD BASINWARD IN AGGRADATIONAL TO PROGRADATIONAL PARASEQUENCE SETS OF THE HIGHSTAND SYSTEMS TRACT
- PARASEQUENCES DOWNLAP ONTO THE CONDENSED SECTION

PHOTOGRAPH

CONDENSED SECTION (PHOSPHATIC OOLITES) AND PROGRADATIONAL PARASEQUENCE SET, HIGHSTAND SYSTEMS TRACT; CASTLEGATE, BUCK TONGUE, AND SEGO MEMBERS, PRICE RIVER FORMATION, BOOK CLIFFS, DOUGLAS CREEK ARCH, COLORADO

Figure 37—Sequence evolution: 4. Slow relative rise, stillstand, and slow relative fall of sea level.



- RATE OF EUSTATIC FALL EXCEEDS RATE OF SUBSIDENCE
- SEA LEVEL FALLS TO SHELF BREAK, SHELF IS EXPOSED, INCISED; CANYONS CUT
- SLOPE-PERCHED DELTAS AND SUBMARINE FANS ARE DEPOSITED

PHOTOGRAPHS

LEFT: LOWSTAND-WEDGE TURBIDITE SANDSTONES AND MUDSTONES; BOXER FORMATION, SACRAMENTO VALLEY, CALIFORNIA

RIGHT: SUBMARINE-FAN SANDSTONES AND CONGLOMERATES (WITH A SEQUENCE BOUNDARY AT THEIR BASE) RESTING ON LOWSTAND-WEDGE MUDSTONES; SALT CREEK CONGLOMERATE ON THE LADOGA SHALE, SACRAMENTO VALLEY, CALIFORNIA

Figure 38—Sequence evolution: 5. Rapid relative fall of sea level.

dence rate of 0.5 ft/1000 years (15 cm/1000 years) to the composite eustatic curve gives a curve of the relative change in sea level, assumed to be defined at the depositional-shoreline break. The linear-subsidence curve on Figure 39 is drawn as an ascending, rather than descending, line to indicate that the net effect of subsidence is a relative rise in sea level.

Two types of fourth-order cycles, designated cycle "A" and cycle "B," compose the relative change in sea-level curve (Figure 39). Fourth-order cycle "A" is defined from sea-level fall to sea-level fall. If we assume adequate sediment supply, this fourth-order cycle deposits a sequence bounded by subaerial unconformities. Fifth-order cycles superimposed on the fourth-order cycle form parasequences bounded by marine-flooding surfaces. A schematic outcrop or well-log profile of the strata deposited during fourth-order cycle "A" is illustrated in Figure 39. The dark-orange shading on the relative sea-level curve shows the ages and positions on the curve of strata that have a low-preservation potential because of incised-valley erosion; near incised valleys most of the highstand deposits will be truncated.

Fourth-order cycle "B" (Figure 39) is defined from rapid rise (transgression) to rapid rise. This fourth-order cycle deposits parasequences bounded by marine-flooding surfaces, if we assume that no differential subsidence occurs in the basin. A schematic outcrop or well-log profile of the strata deposited during this fourth-order cycle "B" is illustrated in Figure 39. However, if the rate of subsidence decreases landward of the depositional-shoreline break so that the rate of eustatic fall exceeds the rate of subsidence in this updip position and thereby produces a downward shift in coastal onlap in the coastal plain, cycle "B" may deposit a type-2 sequence.

In this schematic example (Figure 39), depending on the interaction between the rates of eustasy and subsidence, fourth-order cycles deposit sequences or parasequences; fifth-order cycles deposit parasequences or have no depositional expression. If the subsidence rate is increased well above 0.5 ft/1000 years (15 cm/1000 years) in this example, the third-order cycle will deposit a sequence, referred to as a third-order sequence; the fourth-order cycles will form parasequences that are the components of the third-order sequence. If the subsidence rate is decreased well below 0.5 ft/1000 years (15 cm/1000 years) in this example, the fourth-order cycles will deposit only sequences, referred to as fourth-order sequences, composed of fifth-order parasequences. In this situation, the fourth-order sequences stack to build a third-order unit, tentatively called a third-order *composite sequence*, composed of sequence sets (Van Wagoner and Mitchum, 1989) of fourth-order sequences. In our experience, this situation is typical of many siliciclastic sequences deposited in depocenters, at least since the Pennsylvanian. We have observed fourth-order sequences within sequence sets in Pennsylvanian strata of the western and central United States, Cretaceous strata of the western United States, and most of the Tertiary strata in the northern Gulf of Mexico.

It is worth repeating that, in this book, we define sequences and parasequences based on their physical characteristics and not on the frequency of the sea-level cycle that resulted in their deposition. Although parasequences and fourth-order sequences may, under certain circumstances, be produced by sea-level cycles of the same duration, we do not treat them as synonymous stratal units as some authors do (e.g., Wright, 1986).

Finally, the interpreted role of eustasy in sequence deposition can be evaluated by referring back to Figures 22, 23, and 33 illustrating type-1 sequences from the Miocene of Louisiana. Although from the Miocene, these sequences are typical of most Tertiary-aged sequences along the Gulf Coast. As previously mentioned, these sequence boundaries are regional-erosional surfaces with 100 to 200 ft (30 to 60 m) of truncation covering at least thousands of square miles. Fluvial to estuarine sandstones above these sequence boundaries lie abruptly on outer- to mid-shelf mudstones with no intermediate shallow-marine deposits. Typically these sequences occur with a frequency of 100,000 to 200,000 years (Figure 33). The erosional truncation and vertical-facies associations marking these boundaries were produced by a basinward shift in the shoreline of tens of miles, as determined from facies relationships on the cross sections.

These stratal characteristics of the Miocene sequence boundaries formed in response to a relative fall in sea level. Two mechanisms can produce the relative fall: regional-tectonic uplift, or eustasy. Although they do not result in a basinward shift or a relative fall of sea level, rapid-deltaic progradation and distributary-channel erosion are also considered.

The Tertiary structural style of the northern Gulf Coast basin is characterized by detached, down-to-the-basin normal faults and local-salt features. These structures are diagnostic of a passive-margin tectonic setting where no dynamic plate-tectonic processes occur. The Tertiary of the northern Gulf Coast basin contains no evidence of thermal- or compressional-tectonic events that could cause regional uplift (Murray, 1961; Rainwater, 1967), especially at the frequencies necessary to produce the observed Miocene sequence boundaries.

The interpreted fluvial and estuarine sandstones of the lowstand systems tract of sequences 1 to 3 in Figure 22 and sequence 1 in Figure 33 were deposited in incised valleys that appear to be tens of miles wide (Figure 23), based on data from nine regional well-log cross sections and 23 paleogeographic maps constructed in central Louisiana. The incised valleys cut

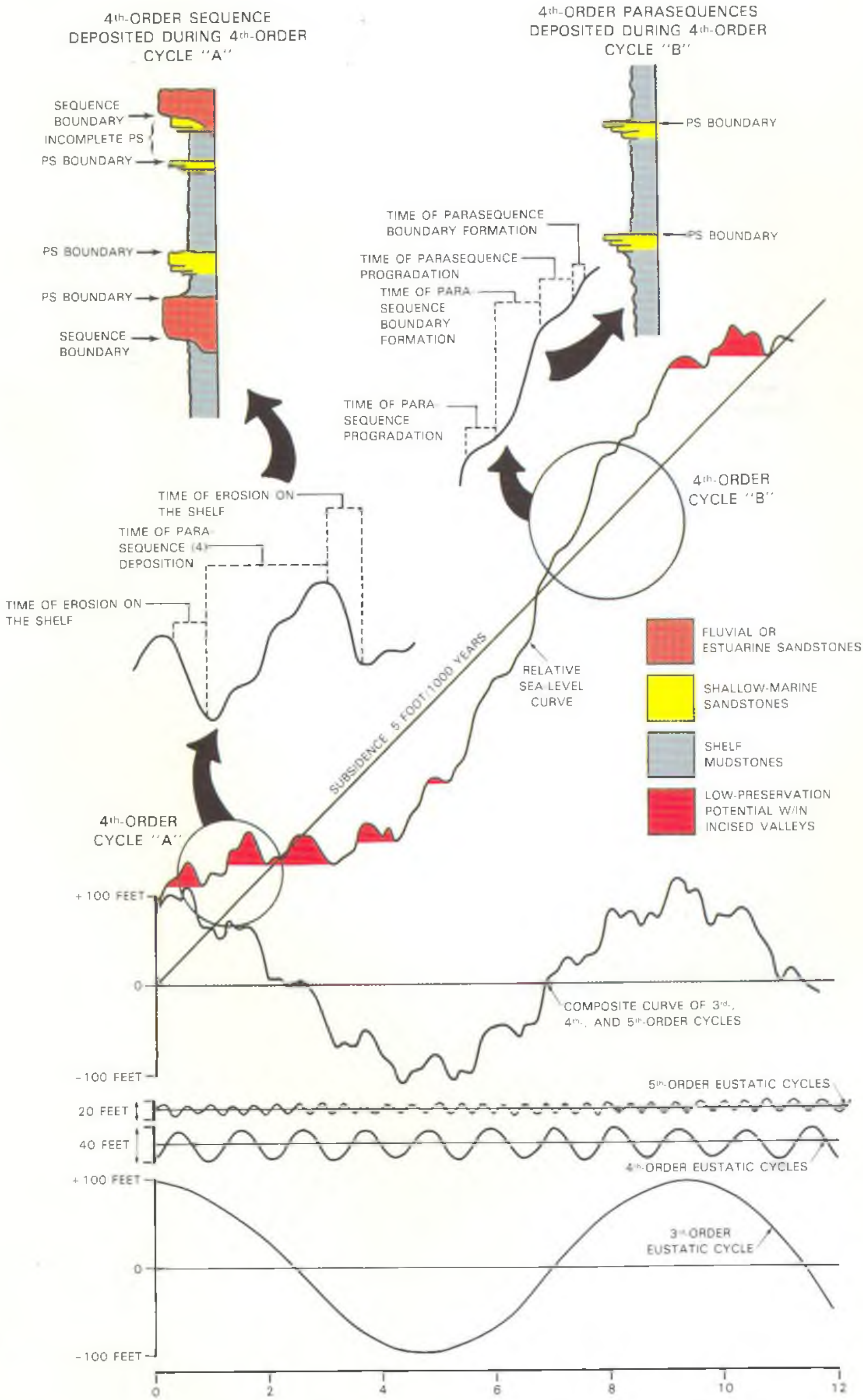
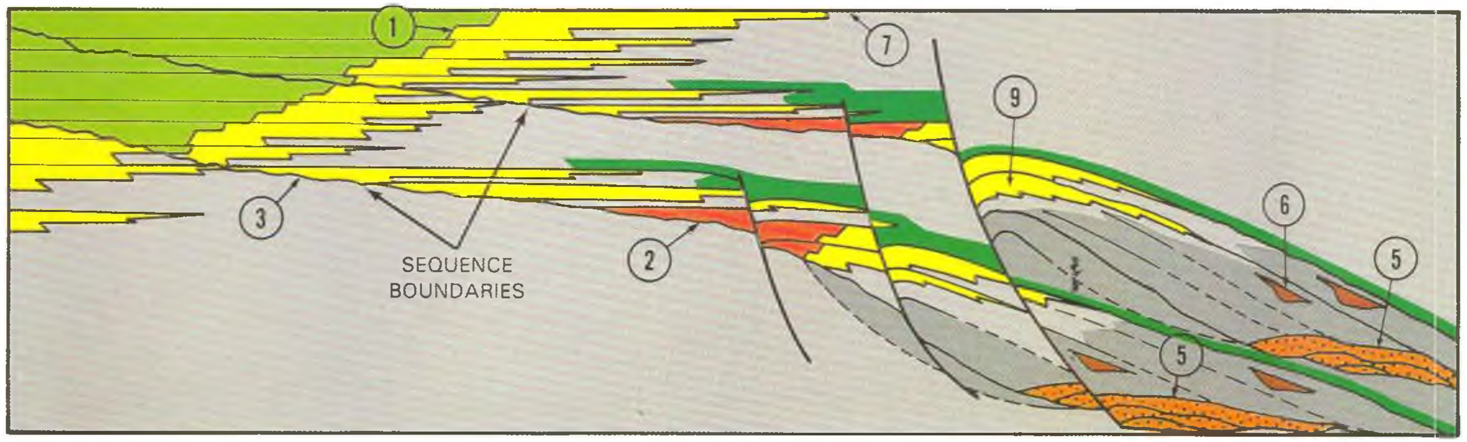


Figure 39—Interaction of eustasy and subsidence to produce parasequences and sequences.

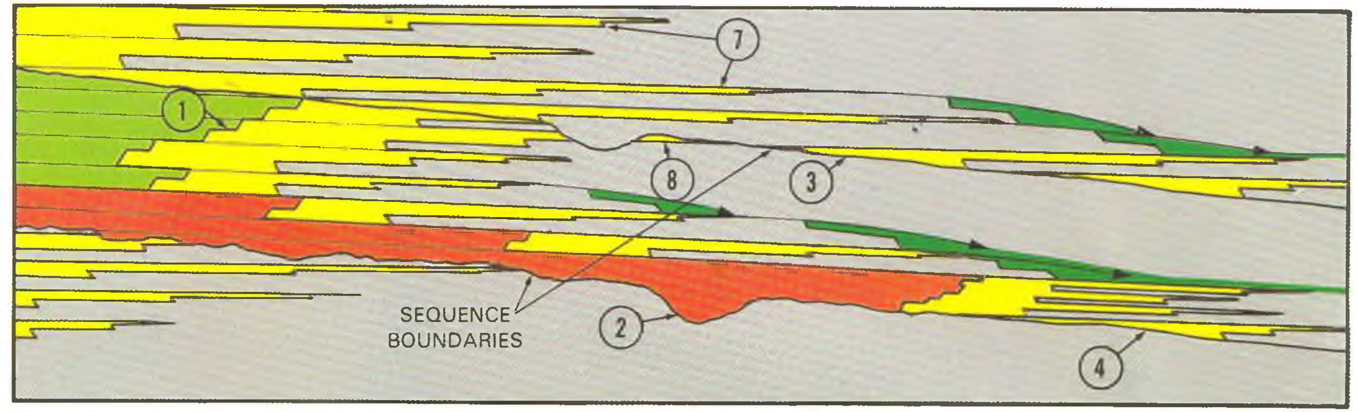
SHELF-EDGE TYPE MARGIN

FOUND IN: CONTINENTAL-MARGIN BASINS ON ATTENUATED CONTINENTAL TO OCEANIC CRUST



RAMP-TYPE MARGIN

FOUND IN:
- CRATONIC BASINS ON CONTINENTAL CRUST
- CONTINENTAL-MARGIN BASINS ON ATTENUATED CONTINENTAL CRUST
- LACUSTRINE BASINS ON CONTINENTAL OR ATTENUATED CONTINENTAL CRUST



NO.	PLAY TYPE	RESERVOIR-FACIES TYPE	POTENTIAL SEAL	EXAMPLES
1	UPDIP PINCH OUT	BEACH OR DELTAIC SANDSTONES	COASTAL-PLAIN MUDSTONES	FALL RIVER SANDSTONE, POWDER RIVER BASIN
2	INCISED VALLEY	BRAIDED-STREAM OR ESTUARINE SANDSTONES	SHELF MUDSTONES	YEGUA, MIOCENE; GULF OF MEXICO; MUDDY, POWDER RIVER BASIN
3	SHELF ONLAP	BEACH, DELTAIC, ESTUARINE, OR SUBTIDAL TO TIDAL-FLAT SANDSTONES	SHELF MUDSTONES	WOODBINE, TUSCALOOSA; GULF OF MEXICO
4	BASINALLY RESTRICTED ONLAP	DELTAIC SANDSTONES	SLOPE/BASIN MUDSTONES	
5	SUBMARINE FAN	SUBMARINE-FAN, TURBIDITE SANDSTONES	SLOPE/BASIN MUDSTONES	PLEISTOCENE, GULF OF MEXICO
6	LOWSTAND WEDGE	SMALL, AREALLY RESTRICTED FANS – COMPOSED OF THIN TURBIDITE SANDSTONES	SLOPE/BAIN MUDSTONES	YEGUA, GULF OF MEXICO
7	DOWNDIP PINCH OUT	DELTAIC, BEACH, OR SUBTIDAL SANDSTONES (NEED STRUCTURAL TILT)	SHELF MUDSTONES	PARKMAN SANDSTONE, SHANNON SANDSTONE, POWDER RIVER BASIN
8	TRUNCATION	BEACH OR DELTAIC SANDSTONES	SHELF MUDSTONES	WILCOX, GULF OF MEXICO; SUSSEX, POWDER RIVER BASIN
9	FAULT CLOSURE	1, 2, OR 3 ABOVE	SHELF MUDSTONES	PLIOCENE, PLEISTOCENE; GULF OF MEXICO

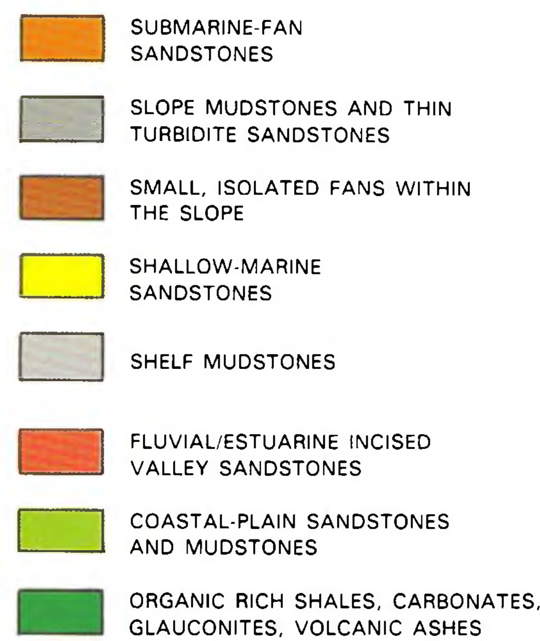


Figure 40—Play types along shelf-edge and ramp-type margins.

into, and are encased in, outer- to mid-shelf mudstones and thin, distal-marine sandstones. Delta-front and stream-mouth sandstones are absent, both lateral to and beneath, the blocky sandstones of the incised valleys—according to interpretation of well-log shapes and regional correlations. Furthermore, the fluvial or estuarine sandstones within the incised valleys do not change facies into shoreline deposits along depositional strike. In comparison, distributary channels of rivers like the modern Mississippi may erode into or through prodelta deposits, but are laterally encased in stream-mouth bar and delta-front sandstones. These lateral-facies relationships exist because a distributary channel builds seaward over the subaqueous-delta platform, even if deltaic progradation is extremely rapid. Sixteen deltas associated with the Mississippi River have been deposited in the last 7000 years and record extremely rapid progradation (Frazier, 1974). However, the preserved deltas and delta lobes show distributary-channel deposits encased in stream-mouth bar and delta-front deposits (Fisk 1961, Gould, 1970). The notable lack of subaqueous, sandy deltaic deposits beneath the sequence boundary or adjacent to the incised valley-fill sandstones in Figures 22 and 33 argues strongly against rapid-deltaic progradation associated with large rates of sediment supply as a mechanism for sequence-boundary formation.

If tectonic uplift and distributary-channel erosion associated with deltaic progradation are ruled out as viable mechanisms for the formation of sequence boundaries, then eustasy is the most likely mechanism to explain the stratal geometries observed in Figures 22 and 33. Pleistocene eustatic falls produced surfaces and facies associations (Fisk, 1944; Frazier, 1967, 1974; Suter and Berryhill, 1985; Suter et al., 1987; Boyd et al., 1988) identical to those seen in the Miocene of the Gulf Coast (Figure 33). Carbon-isotope curves provide evidence for Miocene eustatic changes (Renard, 1986).

The role that tectonism plays in forming or enhancing sequence boundaries is widely debated by stratigraphers. Pitman and Golovchenko (1983) stated that changes in sea level rapid enough to match the Exxon cycle chart (Haq et al., 1987, 1988) can be formed only by glacially induced sea-level fluctuations. Yet others (e.g., Thorne and Watts, 1984) have pointed out that large parts of the geologic column apparently lack evidence of glacial activity. Therefore, the formation of sequence boundaries has been attributed alternatively by many scientists to tectonism (Sloss, 1979, 1988; Bally, 1980, 1982; Watts, 1982; Thorne and Watts, 1984; Hallam, 1984; Parkinson and Summerhayes, 1985; Miall, 1986; Cloetingh, 1988; Hubbard, 1988; and others).

However, the type of tectonic events that would produce rapid, short-term fluctuations in sea level remains unclear, especially those tectonic events that would produce type-1 unconformities. Cloetingh (1988) has advanced the idea of rapid alternations in intraplate stresses, interacting with deflections of the lithosphere caused by sediment loading. Although Cloetingh did not define a frequency at which these tectonic events might occur, he suggested that this type of activity might occur episodically on time scales of "a few million years" to produce "apparent" sea-level changes of more than 327 ft (100 m) along the flanks of sedimentary basins. This mechanism, although not cyclic in nature, might be one explanation for some second-order cycles (9–10 m.y. frequency) on the Exxon cycle chart, but does not satisfactorily explain the higher-frequency third-order or fourth-order cyclicity.

Hubbard (1988), attributing major control of the formation of sequence boundaries to tectonic forces, discussed this point of view. He described two types of sequence boundaries within the Santos, Grand Banks, and Beaufort basins. One type (megasequence) appears to be caused by folding and/or faulting related to the onset of stages in the evolution of a given basin, such as rift onset, synrift faulting, and rift termination. These sequence boundaries represent tectonic episodes rather than true cyclic frequency, and average 49 m.y. in their occurrence. Sequence boundaries of the second type are unstructured, and separate transgressive and/or regressive wedges. They are interpreted to be the result of the interaction of the rates of change of basin subsidence and sediment input with that of long-term global, tectono-eustatic sea level. These sequence boundaries are probably noncyclic and have a modal frequency range of 10 to 15 m.y. Hubbard attempted to demonstrate that the surfaces are not synchronous between basins because each basin has a different history.

Members of the Exxon group have worked in all three basins that Hubbard described and have recognized those sequence boundaries he described. In addition, we described other boundaries that are less prominently developed, but that are important nevertheless in controlling sediment distribution and lithologies within the basin. These occur at the higher frequency expected from the Exxon cycle chart. We certainly agree that Hubbard's "megasequence" boundaries, occurring during onset of stages of basin evolution or other structural events, are *tectonically enhanced*, and become the most prominent and important surfaces in structural analysis of a basin. Similarly, unconformities bounding transgressive-regressive wedges are enhanced because the wedges commonly are produced by subordinate phases of basin subsidence. Because their enhancement is controlled by basinal tectonism, we would not expect the enhancement to extend beyond the limits of the individual basins.

However, the higher-frequency sequences, when dated as accurately as possible using biostratigraphy, appear to be synchronous between the basins. The presence of these sequences strongly suggests that the higher-frequency eustatic overprint is superposed on the lower-frequency or non-cyclic tectonic and sediment-supply controls. Hubbard's (1988) article is excellent for its description of sequence-stratigraphic, basin-analysis procedures, and for its use of tectonically enhanced sequences to describe and date basin development. However, we feel that there is a much stronger interrelationship than he recognized between eustasy and tectonism in controlling sediment type and distribution within the basin.

Although tectonism is the dominant control in determining the shape of the basin, the rate of sediment supply, and possibly even the longer-term, second-order arrangement of sequences, we believe that eustasy controls the timing and distribution of higher-frequency third- and fourth-order sequences.

EXPLORATION APPLICATION AND PLAY TYPES

The stratigraphic concepts we document in this book have broad application to exploration and production. The concepts provide techniques for chronostratigraphic correlation of well logs that result in (1) more accurate surfaces for mapping and facies correlation, and (2) higher-resolution chronostratigraphy for improved definition of plays, especially stratigraphic traps.

The concepts also provide techniques for lithostratigraphic correlation of well logs, thereby yielding (1) a more effective method for evaluating sandstone continuity and trend directions in reservoirs, superior to conventional correlation methods using sandstone or shale tops, (2) improved methods for predicting potential reservoir, source, and sealing facies away from the well, and (3) an alternative to exploration concepts such as offshore-bar reservoirs—resulting in more accurate trend prediction.

Finally, these concepts provide tools for looking at mature basins in fresh ways that result in (1) definition of new play types, opening up heavily drilled basins for new exploration, (2) improved ability to define and locate subtle, but potentially profitable, stratigraphic traps, (3) re-evaluation of producing fields to extend their lives and increase reserves, and (4) a more integrated stratigraphic framework for risking new plays. Figure 40 summarizes potential stratigraphic- and combination structural/stratigraphic-play types associated with the sequences and parasequences on two different basin margins: a margin with a shelf break, referred to in Figure 40 as a shelf-edge-type margin, and a ramp-type margin.

CONCLUSIONS

Sequence stratigraphy provides a powerful methodology for analyzing time and rock relationships in sedimentary strata. Fundamental to sequence stratigraphy is the recognition that sedimentary rocks are composed of a hierarchy of stratal units, from the smallest megascopic unit, the lamina, to the largest unit considered in this book, the sequence. With the exception of the lamina, each of these units is a genetically related succession of strata bounded by chronostratigraphically significant surfaces. Correlation of these bounding surfaces provides a high-resolution chronostratigraphic framework for facies analysis and prediction of rock types at a regional to reservoir scale.

Sequences are the fundamental stratal units of sequence stratigraphic analysis. A sequence boundary is a chronostratigraphically significant surface; it separates all of the rocks above the boundary from all of the rocks below. In most cases, the rocks above the boundary have no physical or temporal relationship to the rocks below. Although sequence boundaries do not form instantaneously, they probably form in from a few thousand to about ten thousand years, and so form very rapidly in geologic terms. For these reasons, recognition of sequence boundaries is critical for accurate facies interpretations and correlations.

A sequence boundary is a better surface for the regional correlation of time and facies than is a transgressive surface. This is true primarily because the timing of the formation of a sequence boundary is not affected by variations in sediment supply; conversely, the timing of the formation of a transgressive surface, at the top of a regressive unit, is controlled strongly by sediment supply. Temporal and spatial changes in the rate and distribution of sediment entering a basin are common. Furthermore, the sequence boundary is accompanied usually by regional erosion and onlap that control facies distribution. The transgressive surface is marked by slight erosion and no onlap.

Sequences are composed of parasequences and systems tracts. Parasequence boundaries are most useful for local correlation of time and facies within the chronostratigraphic framework of individual sequences. Parasequences stack to form aggradational, progradational, and retrogradational parasequence sets. Parasequence sets generally coincide with the systems tracts within the sequence in shallow-marine to nonmarine facies. They are less evident in deeper-water facies of the basin-floor and slope fans. Systems tracts provide a high degree of facies predictability away from the well bore or outcrop within the sequence. This predictability is especially important for analyzing reservoir, source, and seal facies within a basin or a field.

Three systems tracts are recognized in the ideal type-1 sequence: lowstand, transgressive, and highstand systems tracts. The lowstand systems tract is composed of a basin-floor fan, a slope fan, and a lowstand wedge. On the shelf the most conspicuous component of the lowstand wedge is the incised valley. A large proportion of hydrocarbons produced from siliciclastic rocks comes out of the lowstand systems tract. The transgressive systems tract is composed of backstepping parasequences, which can also contain hydrocarbon reserves. The transgressive systems tract can also be very thin, and its top can be a condensed section. The highstand systems tract is composed of aggradational to progradational parasequence sets. Typically, the highstand systems tract is truncated significantly by the overlying sequence boundary. Most type-1 sequences consist of a well-developed lowstand systems tract, a thin transgressive systems tract, and a shale-dominated and truncated highstand systems tract. Type-2 sequences are composed of shelf-margin, transgressive, and highstand systems tracts. In our experience, type-2 sequences are not common in siliciclastic strata.

Type-1 sequences occur with a high frequency from the Pleistocene back, at least, to the Pennsylvanian. High-frequency sequences are interpreted to form in response to sea-level cycles of 100,000 to 150,000 years. High-frequency sequences stack to form sequence sets, which, in turn, form composite sequences. Many of the third-order sea-level cycles on the Exxon Global Cycle Chart (Haq et al., 1988) may have resulted in the deposition of composite sequences composed of sequence sets. The recognition of composite sequences is critical for providing a regional framework for tying the stratigraphy of depocenters, where high-frequency sequences are best expressed, to time-equivalent areas of low-sediment supply. The recognition of high-frequency sequences is essential for developing accurate reconstructions of sea-level change through time and for developing a detailed picture of reservoir, source, and seal distribution within a stratigraphic unit. Finally, if our contention that high-frequency sequences are significant components of siliciclastic strata is correct, there is a fourth-order cyclicity superposed on the third-order cyclicity predicted by the Exxon Global Cycle Chart (Haq et al., 1988). This will have an impact on vertical-facies interpretation in siliciclastic strata because facies continuity may not exist across these high-frequency boundaries.

Recognition of the units in the stratal hierarchy, including sequences and parasequences, is based only on the physical relationships of the strata. These relationships are determined from core, outcrop, well-log, or seismic data. Application of sequence stratigraphy to stratigraphic analysis proceeds, in many basins, independently of inferred regional or global depositional mechanisms.

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