CHAPTER ONE

Historical Perspective

1.1 What is sequence stratigraphy?

Sequence stratigraphy is a subdiscipline of stratigraphy, the latter being defined broadly as 'the historical geology of stratified rocks'. There have been many definitions of sequence stratigraphy over the years, but perhaps the simplest, and that preferred by the authors, is 'the subdivision of sedimentary basin fills into genetic packages bounded by unconformities and their correlative conformities'. Sequence stratigraphy is used to provide a chronostratigraphic framework for the correlation and mapping of sedimentary facies and for stratigraphic prediction.

Several geological disciplines contribute to the sequence stratigraphic approach, including seismic stratigraphy, biostratigraphy, chronostratigraphy and sedimentology. These are discussed in more detail in forthcoming chapters. Note that lithostratigraphy is not considered to contribute usefully to sequence stratigraphy. Lithostratigraphy is the correlation of similar lithologies, which are commonly diachronous and have no time-significance (Fig. 1.1). Lithostratigraphic correlation is useful provided the sequence stratigraphic boundaries enveloping the interval of interest are constrained.

1.2 The evolution of sequence stratigraphy

Sequence stratigraphy is often regarded as a relatively new science, evolving in the 1970s from seismic stratigraphy. In fact sequence stratigraphy has its roots in the centuries-old controversies over the origin of cyclic sedimentation and eustatic versus tectonic controls on sea-level. Much of this early debate has been summarized recently in a set of historical geological papers edited by Dott in 1992 (1992a), entitled 'Eustasy: the Ups and Downs of a Major Geological Concept', and the interested reader is referred to this volume for more detail. Other historically important collections of sequence stratigraphic papers include American Association of Petroleum Geologists (AAPG) Memoir 26, published in 1977, and Society of Economic Paleontologists and Mineralogists (SEPM) Special Publication 42, published in 1988.

Sacred theories

The Deluge and the story of Noah is the most well-known of the earliest references to sea-level change. To the early investigators of sea-level change, the veracity of the Deluge

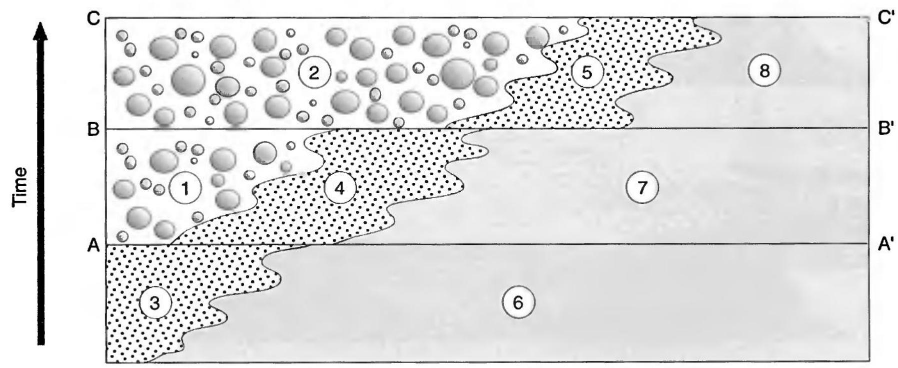


Fig. 1.1 The difference between sequence stratigraphy, which has a geological time significance, and lithostratigraphy, which correlates rocks of similar type. A lithostratigraphic correlation would correlate conglomerate units 1 and 2, sandstone units 3, 4 and 5 and mudstone units 6, 7 and 8. A sequence stratigraphic correlation would correlate time lines A–A′, B–B′ and C–C′

was not in question, but its origin was the subject of considerable debate by scientists and clergy alike. Perhaps the most popular of several theories were Burnet's *Sacred Theory of the Earth*, published in 1681, and the *Telliamed* of de Maillet, published in 1748 (and recently revisited by Carozzi in 1992). De Maillet proposed that following the formation of the Earth by the accretion of the ashes of burning suns over the cortex of an extinguished sun, a water envelope which developed around the planet gradually diminished in volume through time, and in so doing created the topography we see today. In effect, de Maillet interpreted sea-level changes on Earth as a 'single falling limb of a cosmic eustatic cycle' (Carozzi, 1992). This concept of a one-way sea-level fall was known as Neptunian theory. The erosion of primitive mountains by marine processes and the development of a series of offlapping sediment packages as implied by de Maillet and other Neptunists is illustrated schematically in Fig. 1.2.

The eighteenth century

The eighteenth century also saw the beginning of detailed stratigraphic analysis of rock units, and the recognition of unconformities as primary bounding surfaces. In 1788, Hutton first appreciated the significance of unconformities separating cycles of 'uplift, erosion and deposition', and unconformities were used by stratigraphers such as Sedgwick and Murchison in the following century to establish physical boundaries for geological periods (Sedgwick and Murchison, 1839). As the great stratigraphers continued with their practical approach, William Buckland (1823) proposed the concept of Diluvialism which was to

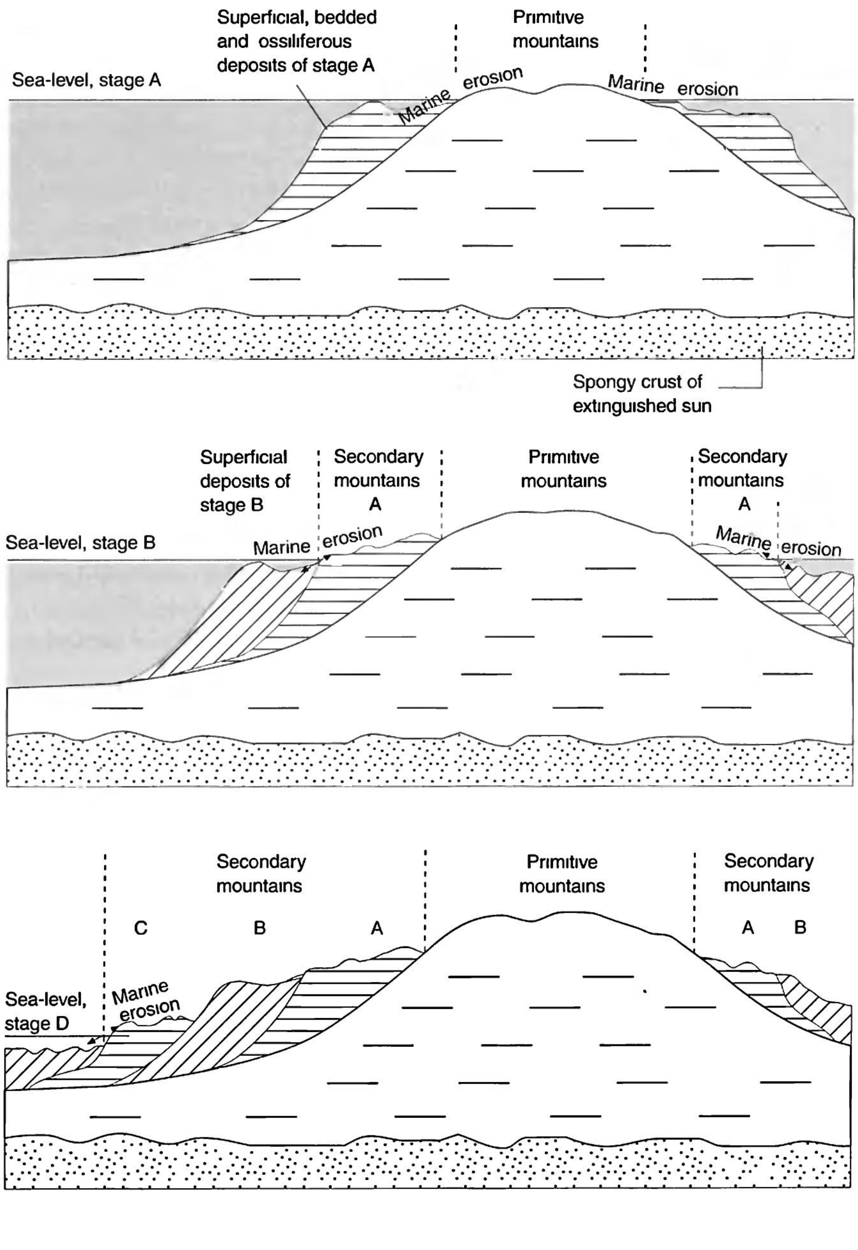


Fig. 1.2 Schematic illustrations of the development of sedimentary units by marine erosion and deposition during continuously falling sea-level, according to de Maillet (1748). After Carozzi (1992)

eclipse Neptunian theory. In diluvial theory, the geological products immediately preceding the flood were referred to as antediluvial, and those following the flood were referred to as post-diluvial or alluvial. However, the attraction of diluvial theory also soon waned as further geological evidence served to counter the simplistic notion of a single dramatic flooding event.

The nineteenth century

In the middle of the nineteenth century, the eustatic versus tectonic controls on sea-level change debate began in earnest with the glacial theories of Lyell and Agassiz. Lyell and others (including Celsius and Linnaeus) observed raised beaches along the coastline of Scandinavia and noted evidence of falling sea-levels from centuries-old marks on shoreline outcrops. Lyell concluded that the land was being slowly and differentially elevated (Lyell, 1835), a fact confirmed by Bravais in 1840 who had observed tilted beaches along fjords of the Scandinavian Arctic coast. At about the same time, Agassiz (1840) was developing his theories of glaciation, and MacLaren, on reviewing Agassiz's glacial theory in 1842, saw the potential of melting ice-caps as a major control on global sea-level. Unfortunately, neither Agassiz nor MacLaren received acceptance for their ideas for at least two more decades, until Croll (1864), in a forerunner of Milankovitch theory (1920), published the concept of orbitally forced glaciations.

The early twentieth century

By the late nineteenth century, glacial theory was thus able to explain eustatic sea-level change and isostatic uplift. However, it was to be several decades before glacial eustasy was resurrected as a control on sedimentary rhythmicity; other explanations of global eustasy took precedence, notably the work of Eduard Suess. Suess first coined the term eustasy in 1906, when he attributed the patterns of onlap and offlap of sedimentary units to global sea-level changes. Suess favoured a mechanism whereby sea-level was lowered by subsidence of the sea-floor, and raised by the displacement of seawater by oceanic sedimentation. He refused to believe the evidence for differential land uplift from Scandinavia, concluding that the Baltic was 'gradually emptying' (Suess, 1888). However, the majority of geologists in the early twentieth century still held the Lyellian view that the major control on sea-level at any point along the coast was the movement of the land. Despite the general lack of support for Suess' ideas, a number of American geologists began to develop concepts of global controls on unconformity development. Foremost amongst these was Chamberlin, who in 1898 and 1909 published his theory on the 'diastrophic control of stratigraphy by world-wide sea level changes'. Three diagrams from his first paper show this to be a precursor of modern sequence stratigraphic concepts (Fig. 1.3).

Chamberlin's ideas were developed by several American

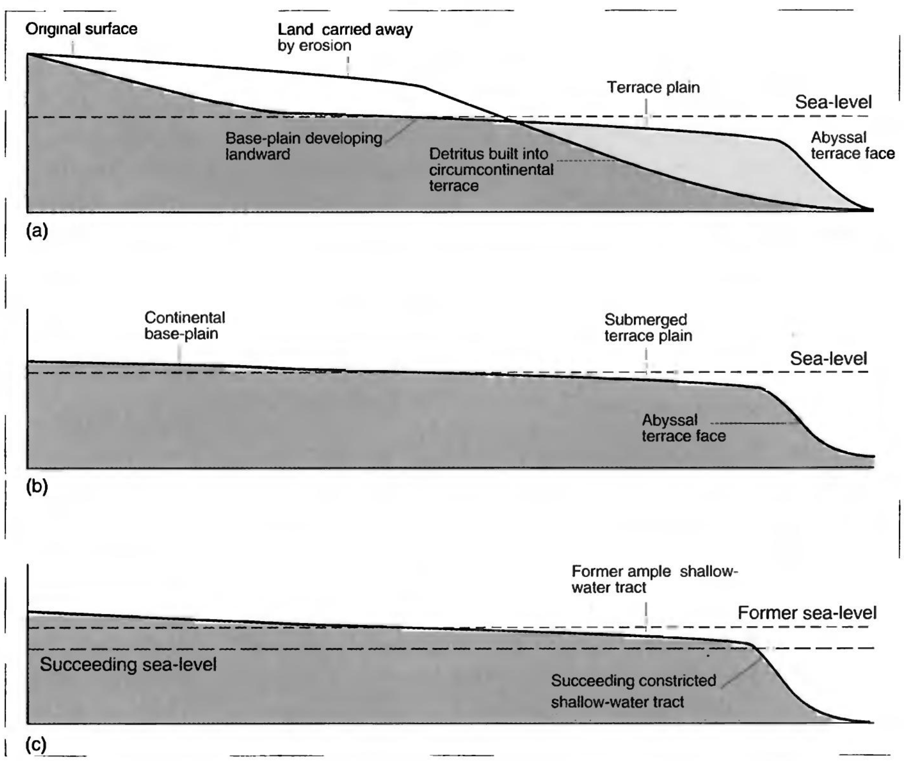


Fig. 1.3 Reproduction of figures from Chamberlin (1898) on diastrophism, unconformities and geological time divisions. From Dott (1992b)

geologists in the following decades, particularly in Palaeozoic systems of the Mid-west. Most notable amongst these were Ulrich and Schuchert, the latter using early palaeogeographic concepts and facies theory to re-create past environments bounded by global unconformities. However, the single most important publication from the 'eustatic' school was that of Grabau, a contemporary of Ulrich and Schuchert, whose 'pulsation theory' postulated rhythmic transgressions and regressions caused by changing heat flow from inside the Earth. The resulting 'pulse beat of the Earth', published in Grabau's 'The Rhythm of the Ages' in 1940, had a periodicity of about 30 million years and caused the development of global unconformities, which could be used to divide the stratigraphic record. Prior to Grabau's work, European geologists, notably Stille (1924) had begun to develop ideas about global unconformities caused by global tectonism with resulting eustatic effects, akin to modern low-order eustatic cycles.

On a smaller scale, sedimentary rhythms were being observed on a scale of metres in coal-bearing Carboniferous (Pennsylvanian) strata in Illinois and Kansas. In 1935, following further studies on Pleistocene glacio-eustatic changes, Wanless and Shepard proposed a control on the development of these Pennsylvanian 'cyclothems' by the accumulation and melting of Gondwana glaciers. This study and others like it resurrected the glacial-eustatic control on sedimentation developed by Croll many decades earlier.

The case for periodicity at a variety of scales in the stratigraphic record was thus becoming compelling. However, as with so many scientific bandwagons, it eventually ran out of steam. In a keynote address to the geological community in 1949, Gilluly argued for orogenesis as a continuous, rather than episodic process, and as a result the concept of rhythmicity of low order (tens of millions of years) gradually lost credibility. The Carboniferous cyclothems were then reinterpreted as autocyclic products, resulting from delta lobe switching and the internal reorganization of sedimentary systems. This latter point also emphasizes the ascendancy of process sedimentology in the early 1960s. Dott (1992a) amusingly points out that at that time many stratigraphers preferred to call themselves sedimentologists!

The late twentieth century

Ironically, Sloss, Krumbein and Dapples (1949) first outlined the concept of stratigraphic sequences at the same meeting that Gilluly proposed his ideas on orogenic continuum. Sloss, Krumbein and Dapples defined sequences as 'assemblages of strata and formations' bounded by prominent interregional unconformities. Despite the negative reaction to these ideas, Sloss (1963) published his major sequences correlateable across the North American Craton, the Indian Tribal names of which still appear as 'super sequences' on the Haq *et al.* (1987) chart. Sloss's ideas were developed further by his graduate students at Northwestern University, one of whom was Peter Vail. Also published at this time was Harry Wheeler's classic 1958 paper on time-stratigraphy which contains many of the concepts in use today, as well as an early attempt to introduce sequence stratigraphic terminology.

Seismic stratigraphy

The next major breakthrough in sequence stratigraphy was in the 1960s and 1970s, when the development of digitally recorded and processed multichannel seismic data made large scale two-dimensional images through basins available. Vail *et al.* (1977a) in AAPG Memoir 26 is perhaps the most referenced work on sequence stratigraphy to date. It summarizes work carried out by Vail and his co-workers, first in the Carter Oil Company and subsequently at the Exxon Production Research Corporation, through the 1960s and early 1970s (Vail and Wilbur, 1966; Mitchum *et al.*, 1976). This period of time marks a break where industry took the lead from academia in the development of sequence stratigraphy. Further papers on seismic sequence stratigraphy followed, and the ideas were gradually extended to incorporate both borehole and outcrop data (Vail *et al.*, 1984). In this work, eustatic sea-level was emphasized as the controlling mechanism for sequence development. In 1985 AAPG Memoir 39 appeared, in which Hubbard *et al.* proposed a tectonic mechanism for the subdivision of basin fill into 'megasequences', driven by changes in tectonic process. The tectonic versus eustatic debate was beginning afresh, although for many at this time seismic stratigraphy was synonymous with eustatic sea-level change, possibly because of its appeal as a global predictive tool for hydrocarbon exploration. In 1987, the Haq *et al.* global sea-level cycle chart was published. This is possibly the single most contentious of all the 'Exxon school' publications, chiefly because the supporting evidence for the curves has not been released. It remains unclear whether local corrections for tectonic uplift or subsidence have been applied, and the dating of unconformities to the accuracy implied by the chart has been challenged (Miall, 1991).

The sequence stratigraphy bandwagon rolls

Special Publication 42 of SEPM, *Sea Level Changes — an Integrated Approach*, was published in 1988 and introduced new concepts such as accommodation space and parasequences, and many of the concepts and principles described in Chapter 2 of this book. Special Publication 42 was important because it opened up the subject to a broader geological community beyond industrial seismic interpreters. In the late 1980s and in this decade, many sequence stratigraphic publications have appeared, some of which uncritically apply the tools and techniques, and some of which are strongly critical. Many question the validity of the interbasinal correlations upon which the Haq *et al.* (1987) curve is based, and others have questioned the validity of certain aspects of the sequence stratigraphic models presented in SEPM 42, such as Miall (1991) and Schlager (1992). Galloway (1989) presented an alternative model for the development of depositional units or 'genetic stratigraphic units' bounded by major flooding surfaces, rather than unconformities. Pitman (1978) has suggested that the origin of sequences and onlap patterns can be explained by variations in subsidence at continental margins, whereas Cloetingh (1988) and Kooi and Cloetingh (1991) proposed that relative sea-level changes and the formation of sequences of millions of years duration can be explained by intraplate stresses rather than eustatic sea-level changes.

The most recent developments in sequence stratigraphy have been in the area of high-resolution subseismic-scale sequence stratigraphy and computer modelling of sedimentary fill. Van Wagoner *et al.* (1990) led the way with the publication of a colourful text on high-resolution sequence stratigraphy from outcrops, logs and core. This stimulated excellent work in superbly exposed marine and marginal marine settings, such as the Jurassic of the Yorkshire Coast and the Cretaceous of the Western Interior Seaway, USA (see also Posamentier and Weimer, 1993, for review). High-resolution sequence stratigraphy also has been combined with work on metre-scale rhythmic successions, particularly bedded platform carbonates and mixed siliciclastic carbonate units (Hardie *et al.*, 1986; Goldhammer *et al.*, 1991). Milankovitch theory of orbital forcing has been revived by sequence stratigraphers in order to explain the origin of high-frequency subsequence-scale cycles. Computer modelling packages have been developed to analyse and replicate the sedimentary fill of basins, at scales from a few metres to entire basins. Basin-wide models include those developed by Royal Dutch/Shell, and published by Aigner *et al.* (1990), and the SEDPAK program developed at the University of South Carolina. Smaller scale cyclicity has been modelled by software such as Mr Sediment (Goldhammer *et al.*, 1989) and by Bosence and Waltham (1990).

The future

The future direction of sequence stratigraphy is difficult to predict, given the turbulent history of the sea-level change debate. At least in the short-term, carbonate systems require further case studies to demonstrate the importance (or otherwise) of controls other than sea-level change. Posamentier and Weimer (1993) have also emphasized the need for further work on the applicability of the concepts to non-marine and deep-marine settings, and further validation (or otherwise) of the sea-level cycle chart from outcrop and subsurface data. Schlager (1992) and others also argue for a more sedimentological approach to sequence stratigraphy, accounting for the autocyclicity of sedimentary processes within the sequence stratigraphic framework. At the very least we can expect considerable debate and further critiques of the subject. This level of activity and debate is all a far cry from the early 1960s when stratigraphy was unfashionable, before Peter Vail and others rescued the subject from its decline.

CHAPTER TWO

Concepts and Principles of Sequence Stratigraphy

2.1 Introduction

The stratigraphic signatures and stratal patterns in the sedimentary rock record are a result of the interaction of tectonics, eustasy and climate. Tectonics and eustasy control the amount of space available for sediment to accumulate (accommodation), and tectonics, eustasy and climate interact to control sediment supply and how much of the accommodation is filled. Autocyclic sedimentary processes control the detailed facies architecture as accommodation is filled. The purpose of this chapter is to introduce the principles that govern the creation, filling and destruction of accommodation. It then shows how these principles are used to divide the rock record into sequences and 'systems tracts', which describe the distribution of rocks in space and time.

The chapter uses siliciclastic systems to introduce the concepts and principles of sequence stratigraphy. Carbonate systems differ from clastic systems in their ability to produce sediment '*in situ*', and they respond in a different manner to accommodation changes. Carbonates are therefore discussed separately in Chapter 10.

2.1.1 Basin forming processes

Tectonism represents the primary control on the creation and destruction of accommodation. Without tectonic subsidence there is no sedimentary basin. It also influences the rate of sediment supply to basins. Tectonic subsidence results from two principle mechanisms, either extension or flexural loading of the lithosphere. Figure 2.1 illustrates theoretical tectonic subsidence rates in extensional, foreland and strike-slip basins. These curves in effect govern how much sediment can accumulate in the basin, modified by the effects of sediment loading, compaction and eustasy.

Extensional basins form in a variety of plate tectonic settings, but are most common on constructive plate margins. In extensional basins, tectonic subsidence rates vary systematically through time, with an initial period of very rapid subsidence caused by isostatic adjustment to lithosphere stretching, followed by a gradual (60–100 million years) and decreasing thermal subsidence phase as the asthenosphere cools. This systematic change in tectonic subsidence rate has a strong influence on the geometry of the basin-fill, such that it may be possible to divide the stratigraphy into pre-, post- and syn-rift phases (these phases have been termed *megasequences*; Hubbard, 1988). In the simple *syn-rift megasequence* model the sediments are deposited in the active fault-controlled depocentres of the evolving rift and can show roll-over and growth into the active faults. Differential subsidence across the extensional faults may exert a strong control on facies distributions. In the *post-rift megasequence*, any remaining rift-related topography is gradually buried beneath sediments that fill the subsiding basin and onlap the basin margin, creating the typical 'steers head' geometry (McKenzie,

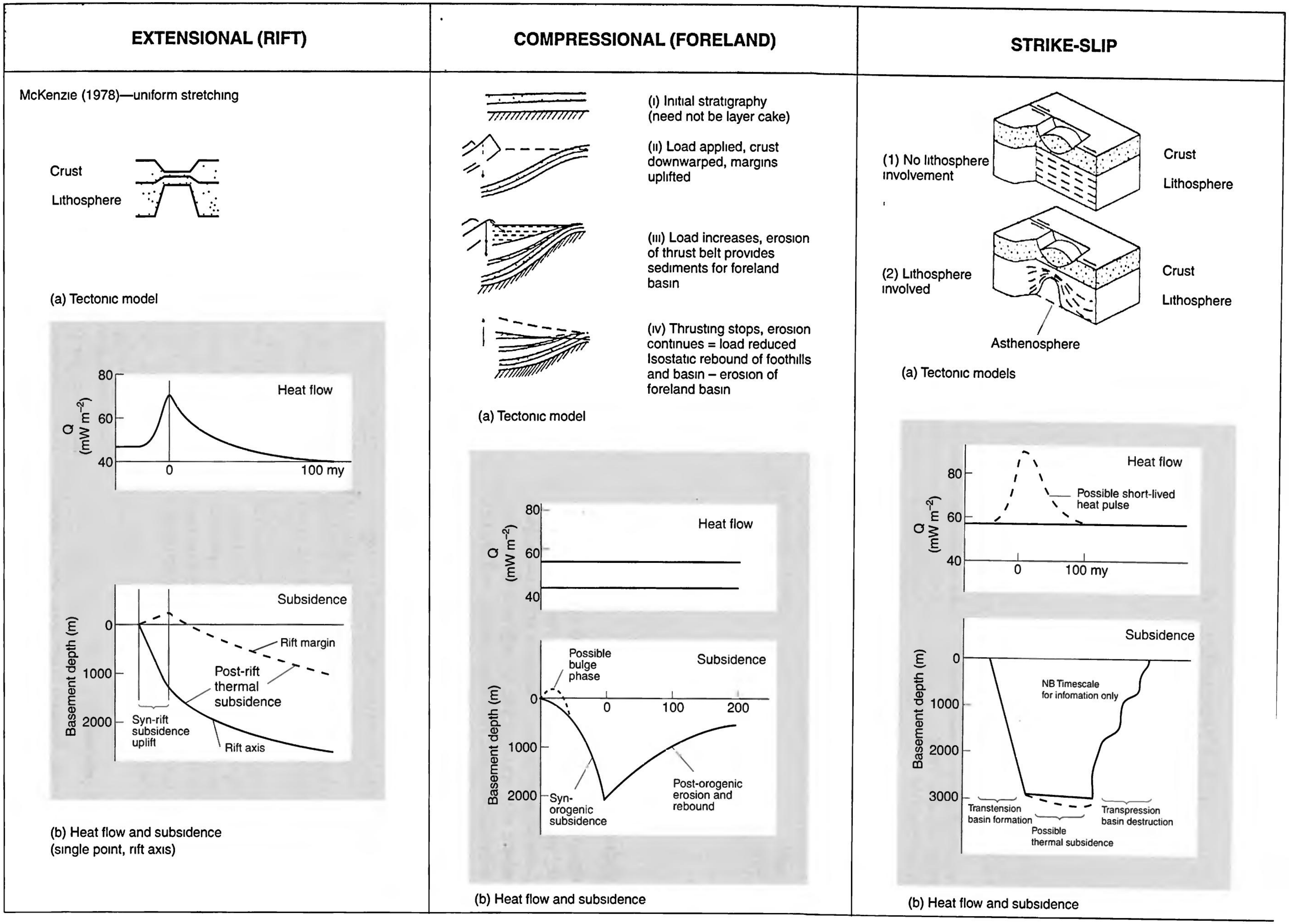


Fig. 2.1 Tectonic subsidence histories in rift, foreland and strike-slip basins

1978). The syn-rift and post-rift megasequences in a marine rift will contain sequences in which development is controlled by higher frequency changes in relative sea-level.

Foreland basins develop in response to loading of the lithosphere below thrust belts. The lithosphere bends in response to loading as the thrust sheets are emplaced, and creates a depression that is accentuated towards the load. The sedimentary fill to this foreland basin has a characteristic wedge shape, thickening towards the thrust front and forming a foreland basin megasequence. The width of the basin is proportional to the rigidity of the underlying lithosphere, and the depth is proportional to the size of the load. Foreland basins formed adjacent to growing mountain belts are characterized by large, and initially rapidly increasing, rates of sediment supply. Cessation of thrusting and continued erosion of the mountain belt leads to an eventual decrease in load, and many foreland basins become uplifted.

Strike-slip basins do not have a characteristic subsidence pattern, although in general, rates of subsidence (and uplift) are extremely rapid.

Tectonic subsidence curves provide a fundamental control on sediment accommodation, upon which higher frequency controls, such as eustasy, fault movement and diapirism, are superimposed. Figure 2.2 shows calculated tectonic subsidence curves for two real basins. In the Llanos Basin, Colombia, sediment supply has exceeded tectonic subsidence. The basin has remained full to base level, with excess sediment bypassed northwards to the sea. The subsidence curve shows slow subsidence through the late Cretaceous and early Tertiary, linked to thermal subsidence in a back-arc basin setting. Two distinct increases in subsidence rate occur in the mid–late Eocene and mid-Miocene, corresponding to two phases of mountain building in the Andes.

In the South Viking Graben example (Fig. 2.2), typical of a number of rifts, sedimentation has not always kept pace with true tectonic subsidence. This led to periods in the Cretaceous where water depths increased and sediment starvation occurred. In the Tertiary, uplift of the Scottish mainland and adjacent North Sea Basin resulted in increased sediment input to the basin (Milton *et al.*, 1990), which locally filled to base level. The remainder of the basin subsequently filled with sediment, resulting in the present-day shallow sea. Separation of the syn-rift and post-rift in this basin is difficult, because the transition occurred during a period of sediment starvation (Milton, 1993).

During periods of rapid basin subsidence, sequence boundaries generated by higher frequency eustatic sea-level falls will be obscured. In times of slow tectonic subsidence or basin uplift, sequence boundaries will be enhanced.

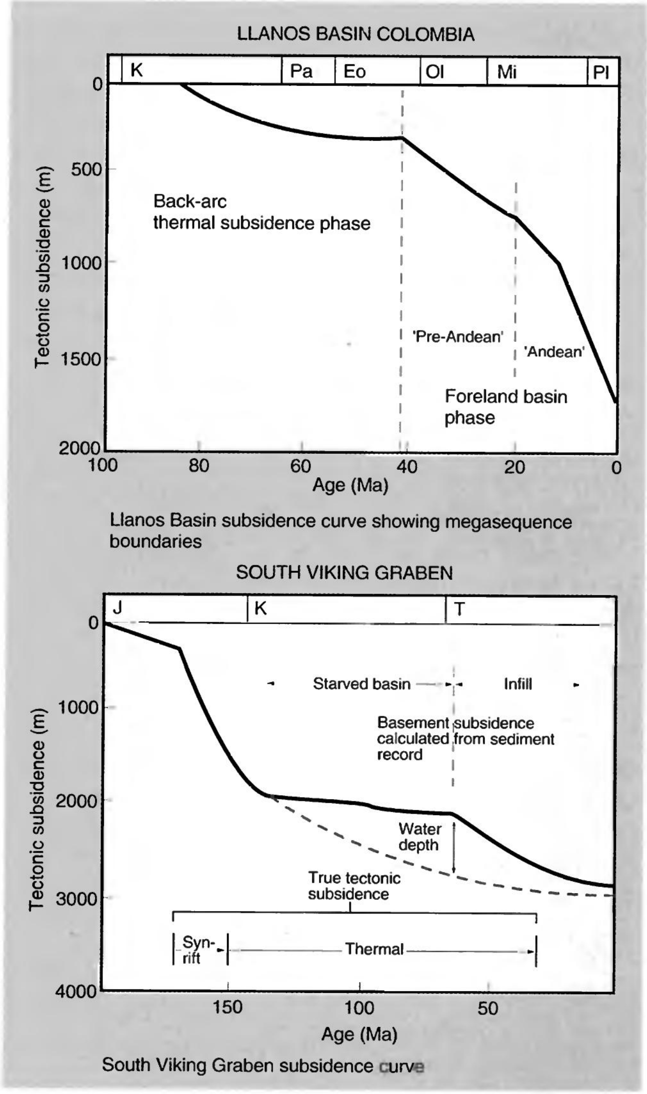


Fig. 2.2 Calculated tectonic subsidence histories for two sedimentary basins (reproduced by permission of BP Exploration Ltd)

2.1.2 Basin-margin concepts

Many of the concepts and principles of sequence stratigraphy are based on the observation from seismic data that prograding basin-margin systems often have a consistent depositional geometry (Fig. 2.3). *Topset* is a term used to describe the proximal portion of the basin-margin profile characterized by low gradients ($<0.1°$). Topsets effectively appear flat on seismic data and generally contain alluvial, deltaic and shallow-marine depositional systems.

The *shoreline* can be located at any point within the topset. It can coincide with the offlap break or may occur hundreds of kilometres landward. The proximal termination of the topset is usually termed the point of *coastal onlap*, referring to the up-dip limit of coastal-plain or paralic facies. *Clinoform* is used to describe the more steeply dipping portion of the basin-margin profile (commonly $>1°$) developed basinward of the topset. Clinoforms generally contain deeper water depositional systems characteristic of the slope. The slope of the clinoform generally

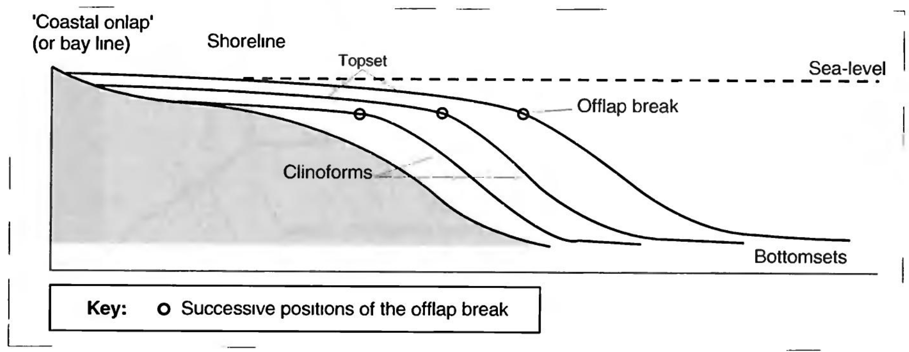


Fig. 2.3 Typical profile of a prograding basin-margin unit, comprising topsets and clinoforms separated by a break in slope; the offlap break. Bottomsets may also be present

can be resolved on seismic data. *Bottomset* is a term sometimes used to describe the portion of the basin-margin profile at the base of the clinoform characterized by low gradients and containing deep-water depositional systems.

The main break in slope in the depositional profile occurs between topset and clinoform and is called the *offlap break* (Vail *et al.*, 1991). The offlap break previously has been termed the *shelf edge* (Vail and Todd 1981; Vail *et al.*, 1984), leading to a confusion with the *shelf break*, i.e. the edge of the modern continental shelf, which is usually a relict feature rather than depositional feature. The term *depositional shoreline break* (Van Wagoner *et al.*, 1988) also has been used, but this implies that the main break in slope in a depositional profile coincides with the shoreline. The term offlap break is preferred here as it does not imply coincidence of the main break in slope with the shoreline.

The topset–clinoform profile results from the interplay between sediment supply and wave, storm and tidal energy in the basin. Sediment enters the proximal end of the profile through river systems and is distributed across the topset area by wave- and/or current-related processes. These may include fluvial currents, tidal currents, storm currents, etc. However, these topset transport processes are effective only at relatively shallow depths of up to a few tens of metres, and to move sediment into deeper water a slope must develop in order to allow sediment transportation by gravity processes. The clinoforms build to the angle needed to transport sediment at the required rate. Slope angle is strongly influenced by sediment calibre. Coarse-grained sediment, with a higher angle of rest, will build up steeper slopes than fine-grained sediment (Kenter, 1990). Also, carbonate systems generally can build steeper depositional slopes (up to 35°) than fine-grained clastic systems (0.5–3°) owing to their greater shear strength. Steeper slopes in clastic systems generally are either made of coarser grade material or are zones of erosion and sedimentary bypass.

The importance of the offlap break on the depositional systems is most apparent during relative sea-level fall (see 2.2.1). When relative sea-level fall exposes the offlap break, rivers commonly incise in order to re-equilibrate to lowered *base level*, with the result that the river becomes entrenched at its mouth (discussed in 2.4.3). The response of the depositional systems to this fall in relative sea-level depends on the nature of the basin margin (Fig. 2.4).

Shelf-break margins are those with well developed depositional clinoforms. Fluvial entrenchment during sea-level fall may result in focusing of the sediment load to discrete locations on the clinoform slope. Failure of the sediment mass has the capacity for forming large turbidity currents and submarine fan deposits. Shelf-break margins are typical of passive continental margins at times of slow rise of relative sea-level, when the delta systems can easily prograde to the shelf edge.

Ramp margins are characterized by relatively shallow water depths, where storms and current processes can operate over much of the area of deposition. Depositional angles are generally less than 1° and seismic clinoforms (if resolved) are shingled with a dip of around half a degree. The offlap break on a ramp margin is likely to be at the shoreline, where fluvial gradients pass into slightly steeper shelf or delta-front gradients. The response of the depositional systems in a ramp setting to relative sea-level change is therefore different from the shelf-break margin. In particular, deep-water turbidite deposition during lowstand may be absent, or of only minor significance. Depositional systems will, instead, be translated basinward without significant slope bypass or basinal deposition. Any turbidites found on a siliciclastic ramp margin are likely to be delta-front turbidites, rather than detached submarine fans (Van Wagoner *et al.*, 1990).

Many modern delta systems can be considered to form ramp margins, as generally they are shelf deltas prograding on to the drowned topsets of a previous shelf-break margin (Fig. 2.4). Frazier (1974) has shown that deposition on the continental shelf of the Gulf of Mexico is confined to the Mississippi delta, which is prograding into about 100 m of water. The rest of the shelf is effectively an area of non-deposition. The Mississippi delta presently forms a ramp margin, although very little extra progradation is needed for the delta to reach the shelf edge, and for the margin to become a shelf-break margin.

Rift margins characterize basins undergoing active crustal extension. Extensional faults have a strong influence on both palaeogeography and sediment influx rates. The spatial

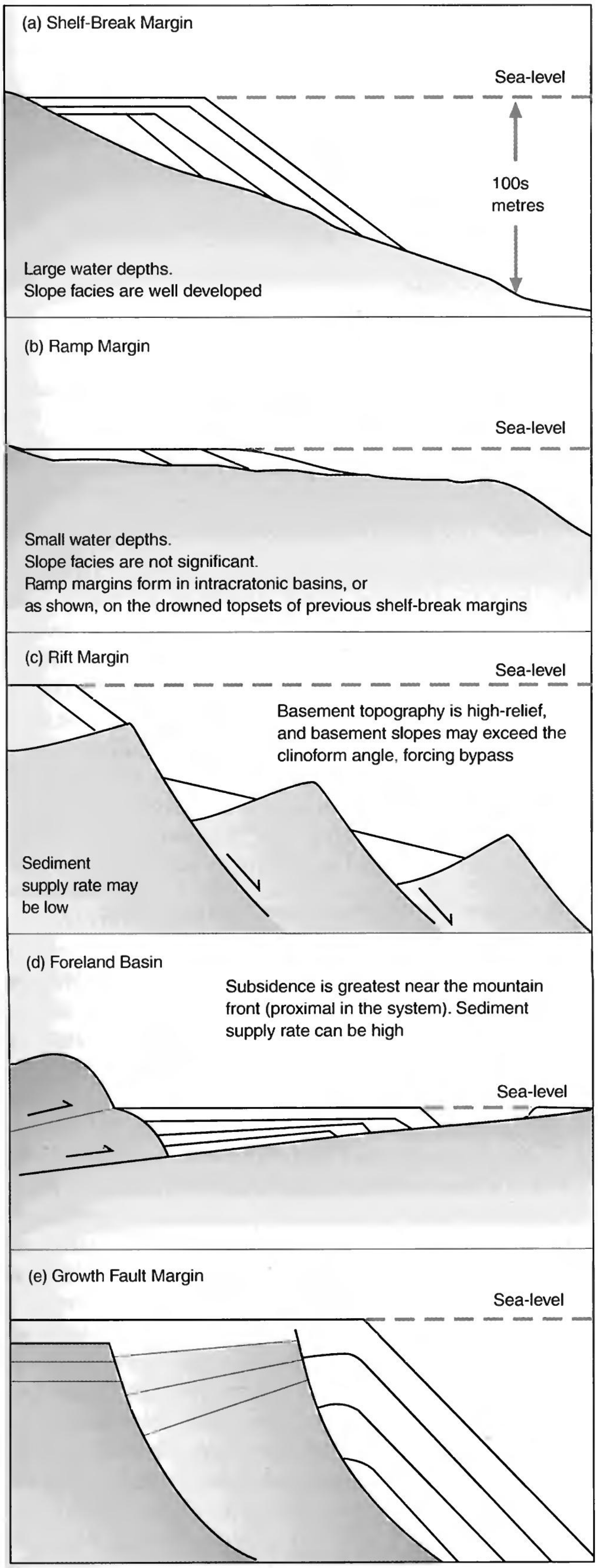


Fig. 2.4 Basin-margin types: (a) shelf-break margin; (b) ramp margin; (c) rift margin; (d) foreland basin, (e) growth-fault margin

distribution of sediment accommodation within the rift is controlled largely by tectonics. Subsidence rates generally will increase from the margins to the centre of the rift, although each individual fault block will have its own pattern of accommodation. The foot-wall crest will see the least subsidence and may experience uplift and erosion, whereas the hanging wall will experience progressively greater subsidence rates towards the controlling fault. The depositional systems that develop will depend on whether the rift is marine or continental. Transfer zones in the rift margin may control sediment entry points. Rift margins may be characterized by high topographic relief and relative sediment starvation, because sediment is bypassed towards the rift centre. Basin-margin systems may build out into deep water with long clinoform slopes and relatively minor topsets (Fig. 2.4). There is little potential for trapping coarse material in the topsets, and much may be bypassed to the basin.

Foreland-basin margins vary depending on whether sediment is being fed axially along the foreland basin or directly into the foreland basin from the thrust belt. In the latter case, the rate of tectonic subsidence increases towards the foreland thrust belt, i.e. the sediment source area. In other words, sediment accommodation may be relatively high in proximal areas compared with the basin centre. This has a marked affect on stratal geometries and may result in the aggradation of thick topset deposits, with little opportunity for seismic-scale clinoforms to develop (Posamentier and Allen, 1993).

Growth-fault margins are characterized by gravity driven syn-sedimentary extensional faults. The rate of subsidence is considerably greater on the hanging-wall side of the growth fault, resulting in an expanded sedimentary succession. The effect of the growth fault in the depositional systems developed will depend on whether the fault had a topographic expression on the sea-bed. At times when the hanging wall was a topographic low relative to the foot wall, facies differentiation occurs across the fault, with thick, deeper water clastic systems on the downthrown side. Growth-fault margins are discussed further in section 9.3.3.

2.2 Relative sea-level, tectonics and eustasy

2.2.1 Definitions of sea-level

In order to understand the controls on sequence development, it is first necessary to define what is meant by eustasy, relative sea-level and water depth (Fig. 2.5, from Jervey, 1988).

Global eustasy

Eustasy is measured between the sea-surface and a fixed datum, usually the centre of the Earth. Eustasy can vary by changing ocean-basin volume (e.g. by varying ocean-ridge volume) or by varying ocean-water volume (e.g. by glacio-

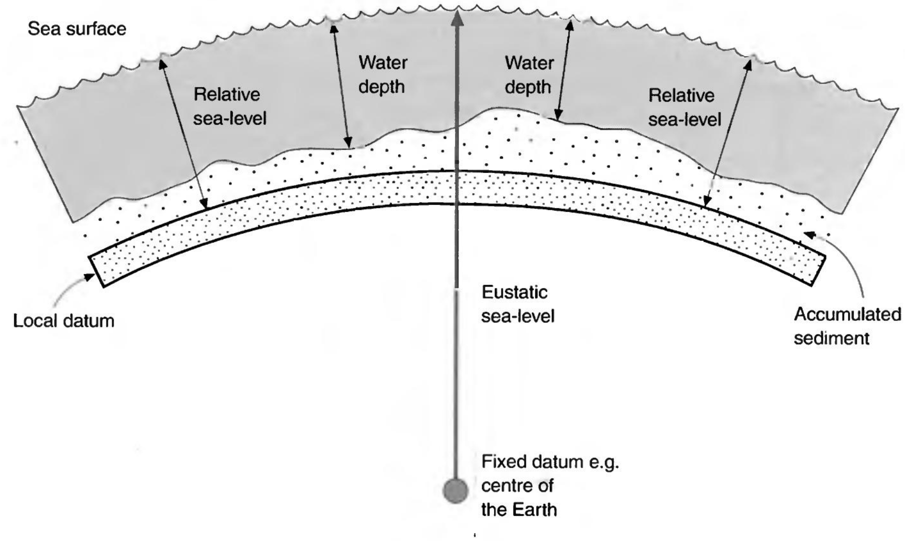


Fig. 2.5 Definitions of sea-level (after Jervey, 1988). 'Eustatic sea-level', base level for fluvial systems entering the ocean

eustasy). The interpretation of eustatic changes from the rock record is a complex and controversial topic, which will be discussed in detail in section 2.2.4. For the moment it is important only to emphasize that it can rise or fall, thus varying the base level for erosion on a global scale, where *base level* is defined as the level above which deposition is temporary and erosion occurs (see 2.2.2).

Relative sea-level

Relative sea-level is measured between the sea-surface and a local moving datum, such as basement or a surface within the sediment pile (Posamentier *et al.*, 1988). Tectonic subsidence or uplift of a basement datum, sediment compaction involving subsidence of a datum within the sediment pile, and vertical eustatic movements of the sea-surface all contribute to relative sea-level change. Relative sea-level 'rises' due to subsidence, compaction and/or eustatic sea-level rise, and 'falls' due to tectonic uplift and/or eustatic sea-level fall. Relative sea-level should not be confused with water depth, which is measured between the sea-surface and the sea-bed in any given geographic location at a point in time. The term *equilibrium point* is sometimes used to distinguish the point on a depositional profile where the rate of relative sea-level change is zero. The equilibrium point will separate, at any given time, the zone at the basin margin where relative sea-level is falling from the zone where relative sea-level is rising.

2.2.2 Accommodation

Eustasy and subsidence rate together control the amount of space available for sediment accumulation — this is conventionally termed accommodation. *Accommodation* is defined as the space available for sediment to accumulate at any point in time (Jervey, 1988). Accommodation is controlled by *base level* because in order for sediments to accumulate, there must be space available below base level. The base-level datum varies according to depositional setting (Fig. 2.6). In alluvial environments base level is controlled by the graded stream profile, which is graded to sea-level or lake-level at its distal end (Mackin, 1948; and Chapter 7). In deltaic and shoreline systems base level is effectively equivalent to sea-level. In shallow marine environments base level is ultimately also sea-level, although fairweather wave base can form a temporary base level in the form of a 'graded shelf profile'. Sediment supply fills the accommodation created and controls water depth:

$$\Delta\text{accommodation} = \Delta\text{eustasy} + \Delta\text{subsidence} + \Delta\text{compaction}$$

Sediment supply fills available accommodation. If the rate of sediment supply exceeds the rate of creation of accommodation at a given point, water depths will decrease:

$$\Delta\text{water depth} = \Delta\text{eustasy} + \Delta\text{subsidence} + \Delta\text{compaction} - \text{sediment deposited}$$

A series of cartoons in Fig. 2.7 illustrates the relationship between accommodation, relative sea-level and water depth in shoreline–shelf depositional systems. In these examples relative sea-level change and new sediment accommodation added are the same because base level is taken at the sea-surface.

In the discussion that follows we examine the relationship between relative sea-level and accommodation in shoreline–shelf depositional systems. Fluvial systems and the controls on the graded stream profile are discussed in Chapter 7, paralic systems are discussed in Chapter 8, submarine fans in Chapter 9 and carbonates in Chapter 10.

2.2.3 Accommodation through time

In order to understand how accommodation varies through time it is useful to consider how different rates of tectonic

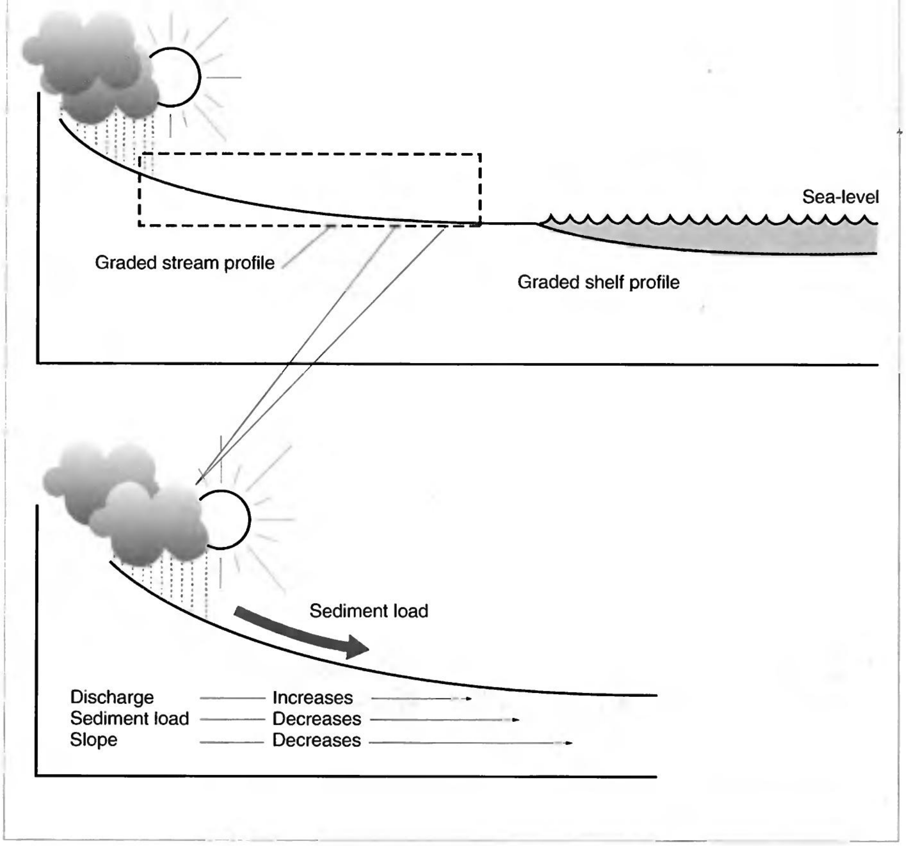


Fig. 2.6 Definitions of base level in fluvial, shoreline and shelf environments

subsidence and, in this case, the same sinusoidal eustatic sea-level curve combine to give different rates of addition and destruction of accommodation (rise and fall of relative sea-level) (Fig. 2.8; from Jervey, 1988).

In Fig. 2.8 subsidence is represented as a straight line, the gradient of which indicates the rate of subsidence at each point. The different gradients can be thought of as representing positions in a basin with increasing subsidence rates or changes in subsidence rate through time. Eustasy is represented by the same smooth curve in each case. The change in relative sea-level through time is found simply by the addition of the two curves. Relative sea-level is, in this case, equivalent to accommodation because the curves begin at zero water depth.

Where slow subsidence occurs, maximum accommodation is developed near the eustatic maximum. When eustasy falls to its original position, accommodation falls to a value representing that created only by subsidence. With increased rates of subsidence, the time of maximum accommodation is progressively later. Points in the basin where subsidence rates are very high experience no decrease in accommodation even though eustatic fall may be occurring. Note that the same curves could be produced theoretically by adding variable rates of tectonic subsidence and uplift to a flat eustatic curve.

2.2.4 Orders of cyclicity and global correlation

A depositional sequence represents a complete cycle of deposition bounded above and below by erosional unconformities. The sequence has a maximum duration, which is measured between the correlative conformities to the bounding unconformities. Thus, the duration of the sequence will be determined by the event controlling the creation and destruction of accommodation, i.e. tectonic subsidence and/or eustasy. Tectonic cycles of subsidence and uplift and eustatic cycles of rising and falling sea-level can operate over different time periods, and it is useful to classify sequences in terms of their order of duration, commonly termed first, second, third, fourth order, etc. (Fig. 2.9). A basin-fill can then be divided into a hierarchy of sequences, each representing the product of a particular order of tectonic or eustatic cycle.

In Fig. 2.9, from Duval *et al.* (1992), four orders of stratigraphic cycle are depicted. The continental encroachment cycle is defined by the very largest scale (> 50 million years) cycles of sedimentary onlap and offlap of the supercontinents. There are only two such cycles in the Phanerozoic, according to the Haq *et al.* (1987) sea-level curve. First-order continental encroachment cycles are considered to be controlled by tectono-eustasy, i.e. changes in ocean basin volume related to plate tectonic cycles (Pitman, 1978).

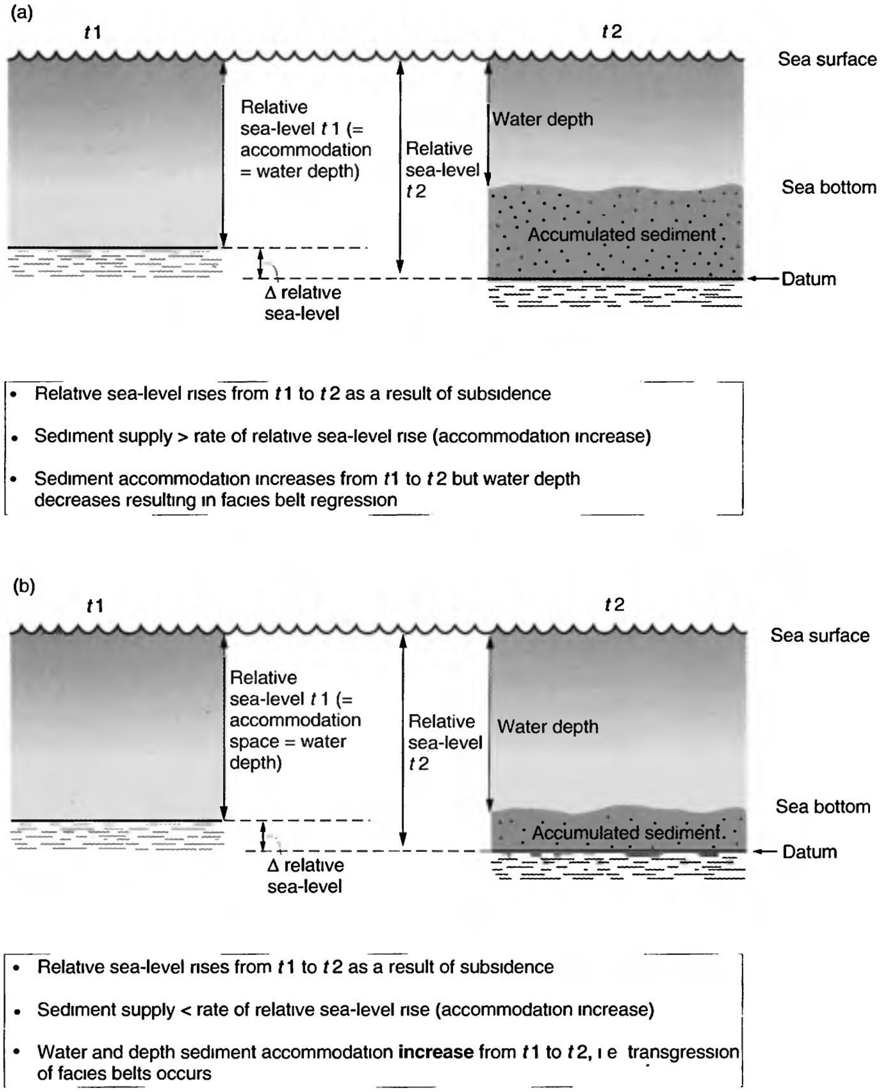


Fig. 2.7 This series of diagrams illustrates how eustatic sea-level rise/fall and subsidence/uplift can create/destroy accommodation. The rate at which sediment fills the space created controls water depth and whether *progradation* or *retrogradation* of facies belts is observed. In (a) relative sea-level rises and accommodation increases from time 1 to time 2 owing to subsidence, but the rate of sediment accumulation at this point is greater than the rate of relative sea-level rise and so water depth decreases from time 1 to time 2. In the depositional record, the interaction of these parameters would result in an overall regressive character to the vertical facies succession. (b) Relative sea-level rises and accommodation increases from time 1 to time 2 owing to subsidence, but the rate of sediment accumulation is less than the rate of relative sea-level rise and so water depths increase from time 1 to time 2. In the depositional record this may be apparent as a transgressive vertical facies succession. The same patterns would occur for a relative sea-level rise due to a eustatic rise of the same rate. (c) (*opposite*)

Second-order (3–50 million years) cycles are the building blocks of the first-order sequences and represent particular stages in the evolution of a basin. They may be caused by changes in the rate of tectonic subsidence in the basin or rate of uplift in the sediment source terrane.

Third-order (0.5–3 million years) sequence cycles are the foundation of sequence stratigraphy because they are often of a scale well-resolved by seismic data. They are identified by the recognition of individual cycles of accommodation creation and destruction. These cycles are considered by Vail *et al.* (1991) to be controlled by glacio-eustasy, although other tectonic mechanisms are possible (Cloetingh, 1988).

Composite sequence is a term sometimes used to describe second- or third-order sequences made up of higher order sequences (Mitchum and Van Wagoner 1991; and see 2.4.10).

Fourth-order (0.1–0.5 million years) 'parasequence' cycles represent individual shallowing upward facies cycles bounded by surfaces of abrupt deepening. These may be related in part to autocyclic processes within the sedimentary system.

The theory of eustatic control on deposition is a unifying stratigraphic concept that has attracted geologists for many generations (see Chapter 1 and papers in Dott, 1992a,b). If it were true that a global eustatic signature was overprinted on all stratigraphic successions, then it would be possible to date a stratigraphic section from the pattern of sequences and systems tracts, and to predict stratigraphy in unsampled areas from a knowledge of the global standard. A proposed global sea-level chart was first published by Vail *et al.* (1977a), and updated by Haq *et al.* (1987), based on measurements from basins around the globe. This chart is taken to support the theory that third-order relative sea-level variations are mostly eustatic in origin. Sceptics would argue that the chart is based on that theory rather than proof of it.

More discussion and controversy have been caused by

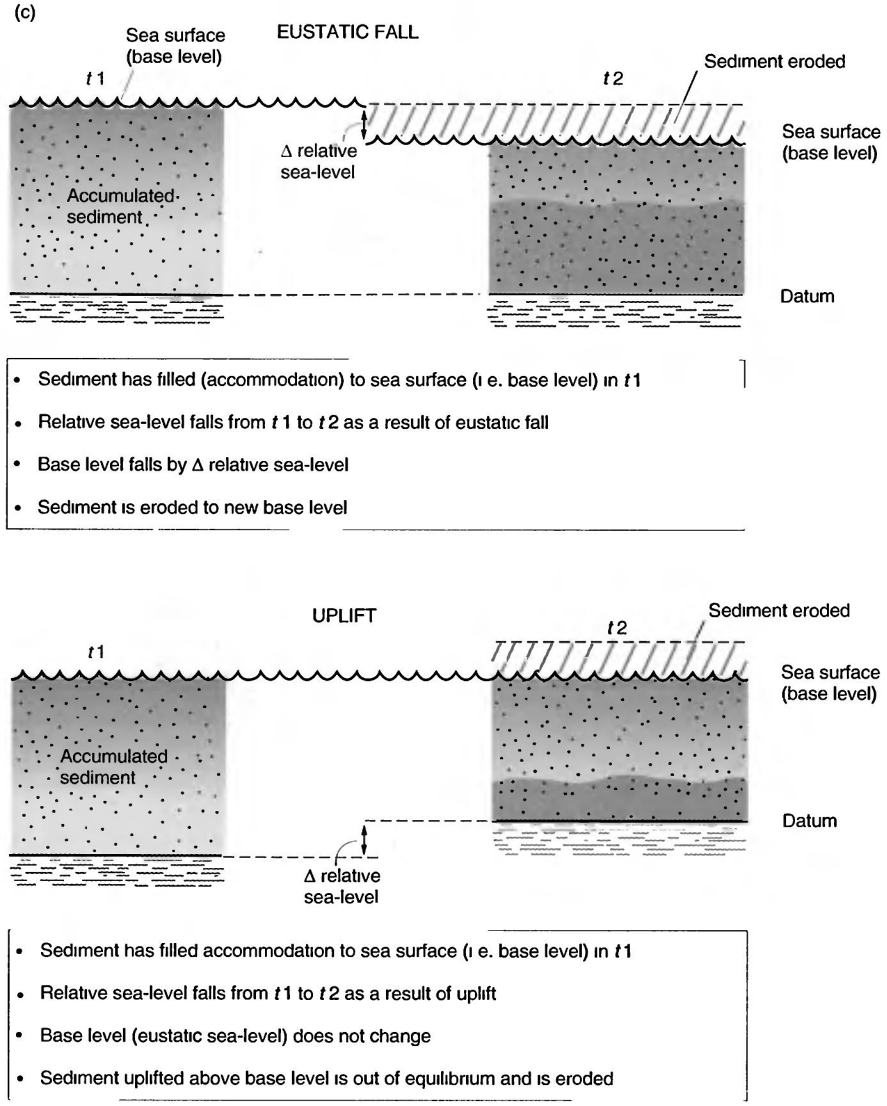


Fig. 2.7(c) Illustrates time when accommodation space is destroyed. This can occur by two mechanisms — tectonic uplift (and more locally salt or mud diapirism) and/or eustatic sea-level fall. The effect of relative sea-level fall is manifest at the basin margin (where accommodation space is limited) by *erosion*, and at the basin centre (where accommodation space is not limited) by *an increase in sediment supply*. The increase in sediment supply to the more basinal areas is due to both the erosion of previously deposited sediment and bypass of areas where accommodation space is now filled

the concept of a global eustatic signal than any other aspect of sequence stratigraphy. It is beyond the scope of this work to cover this discussion in detail and the following is a brief summary only.

There are a number of aspects of the Haq *et al.* (1987) chart that have excited comment:

1 The data to support the Haq chart have never been published fully, in particular the evidence for the global correlatability of the sequence boundaries. Miall (1986, 1992) has been a consistent critic of the Haq *et al.* curve, stating:

> The basic premise of the Exxon cycle chart, that there exists a globally correlatable suite of third order eustatic cycles, remains unproven...There are some specific cases where global synchroneity is suggested by detailed stratigraphic documentation (e.g. 4th- and 5th-order glacioeustatic cycles in the Neogene and, possibly the late Palaeozoic; 1st- and 2nd-order cycles related to changing rates of global sea floor spreading), but for the greater part of the Phanerozoic column no such proof is available. (Miall, 1991)

Miall also points out that it is arguable whether global biostratigraphic control is accurate enough to correlate third-order relative sea-levels unambiguously. Thus, for the moment, the global synchroneity of eustatic cycles has to remain to a degree an article of scientific faith rather than scientific fact.

2 The mechanism for generating third-order-scale eustatic sea-level changes is problematic in certain periods of geological time. Increases in the global ice volume during glaciations provide a mechanism for eustatic falls in sea-level in the late Cenozoic and late Palaeozoic, but no such mechanism exists for the (presumed) ice-free Cretaceous and Jurassic. Cloetingh *et al.* (1985) have proposed intraplate stress as a tectonic mechanism for generating third-order plate-wide relative sea-level cycles. Finally, it is not

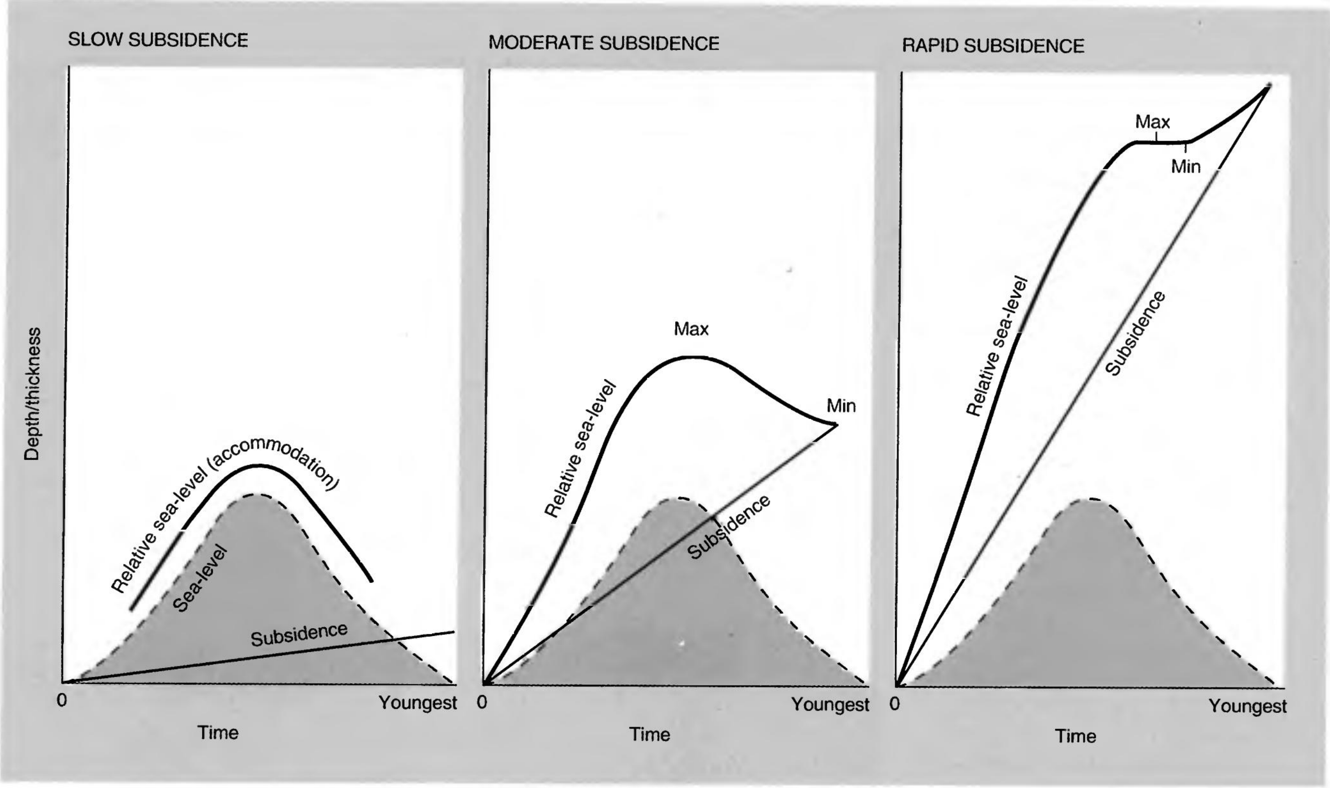


Fig. 2.8 Accommodation space through time (from Jervey, 1988)

accepted universally that the eustatic signal will be visible in all basins, and several stratigraphers believe that it may be obscured by a tectonic signal (e.g. Hubbard, 1988).

However, work is continuing to date basin-margin unconformities more accurately and to correlate these with the Neogene oxygen isotope record and hence directly to ice volume changes (e.g. Miller *et al.*, 1991, 1993). Also, various projects are underway to accurately date and correlate sequence boundaries on a regional scale in Europe, e.g. De Graciansky *et al.* (1993).

2.3 Sediment supply

The rate of sediment supply controls both how much and where accommodation is filled. The balance between sediment supply and relative sea-level rise controls whether facies belts prograde basinward or retrograde landward, and the calibre of sediment supplied to the basin has a strong influence on sedimentary facies. The first part of this section considers the principles controlling siliciclastic sediment supply to the basin margin and how sediment supply may vary through time. The second part considers how accommodation is filled in locations with high, moderate and low rates of sediment supply. The principles of carbonate sediment production and supply are discussed in Chapter 10.

2.3.1 Principles of clastic sediment supply

River transport is the principle means of transporting material from the continental interior to the depositional basin. The volume and grade of sediment delivered to the basin margin is a complex function of hinterland physiogeography, tectonics and climate. Studies of modern rivers show huge variations in the rate of sediment supplied to the continental margins (Fig. 2.10). Around 70% of the total load is supplied from only 10% of the land area, and just three rivers, the Ganges, the Brahmaputra and the Huang He (Yellow) supply 20% of the total fluvial load (Summerfield, 1991).

The amount of sediment supplied to the basin margin is a function of both the fluvial drainage basin area and the mechanical denudation (erosion) rate. Tectonism at both local and regional scale affects fluvial drainage basin shape, size and relief and also the geology of the provenance area, and will control the calibre of sediment eroded. The rate of fluvial denudation is a complex function of relief within the drainage area and climate. Climate influences not only the erosive power of the river by controlling discharge but also the erodibility of soils in the drainage basin, and the presence or absence of stabilizing vegetation. Present-day mechanical denudation rates vary from less than 1 mm per 1000 years in the St Lawrence river drainage basin to 640 mm per 1000 years in the Brahmaputra drainage basin.

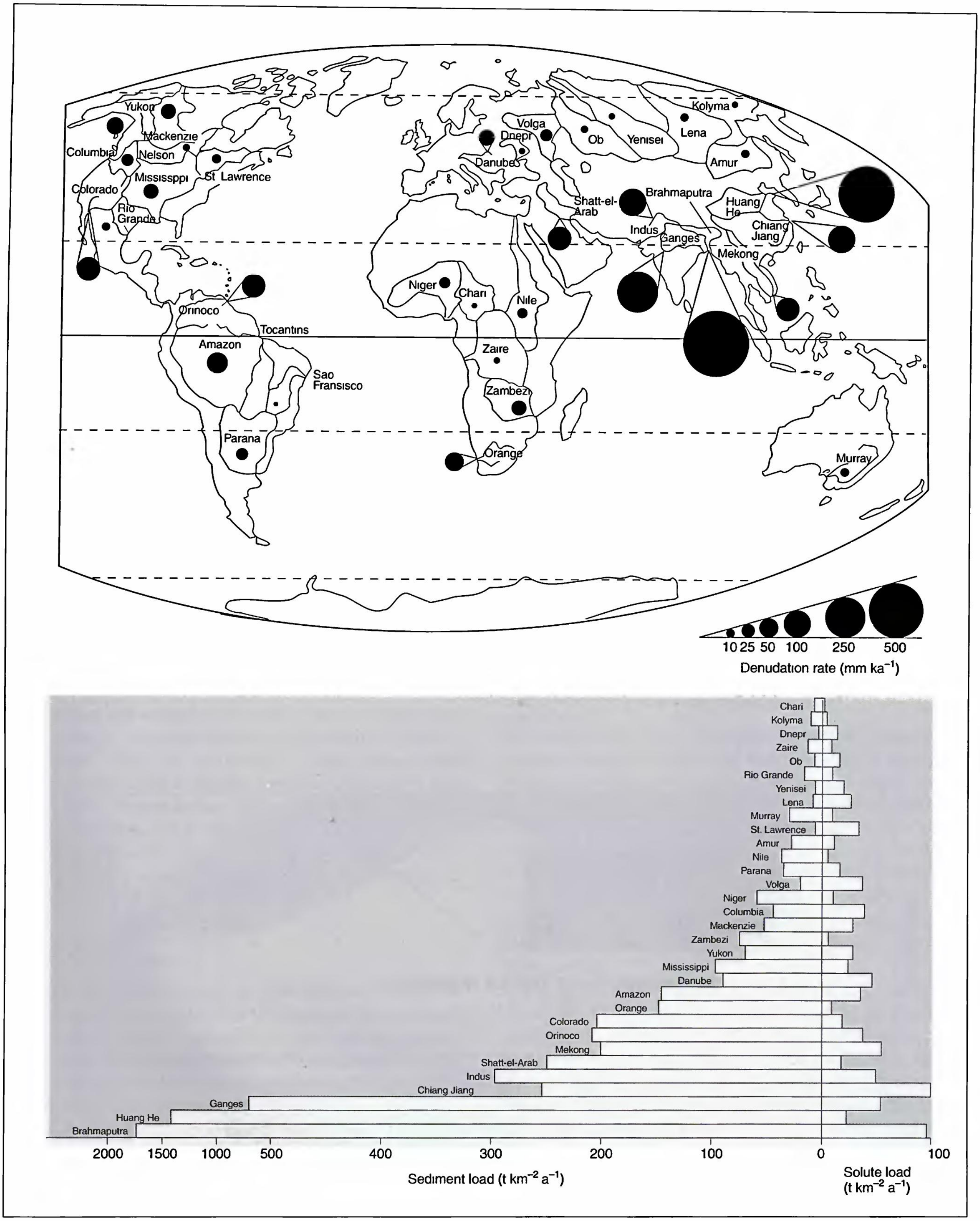


Fig. 2.10 (a) Denudation rates for the world's 35 largest drainage basins based on solid and solute data. Allowance has been made for the non-denudational component of solute loads. Source rock density is assumed to be 2700 kg m^{-3}. (b) Sediment and solute loads for the world's largest drainage basins. Both figures from Summerfield (1991)

A tributary of the Huang He river in China is an extreme example, with a denudation rate of 19 800 mm per 1000 years, because it drains over 3000 km^2 of loess-covered terrain in a semi-arid region of sparse vegetation (Summerfield, 1991).

It clearly would be unwise to think of sediment supply to the basin margin as being either spatially or temporally constant. Local sediment supply will depend on the proximity to a fluvial entry point to the basin margin. There also may be a linkage between glacio-eustatically controlled relative sea-level cycles and climate in the fluvial drainage basin (Blum, 1990). This will mean that sediment supply may vary through a sea-level cycle in a fashion characteristic of the drainage basin.

2.3.2 Filling of accommodation

The amount of sediment supplied to locations in the basin is a function of both the general rate of sediment supply to the basin and the proximity to sediment entry points to the basin. Figure 2.11 (from Jervey, 1988) considers the relationship of facies, relative sea-level and rates of sediment accumulation at three fixed points in the basin with identical relative sea-level curves but with differing rates of sediment supply. These could represent points along a continental margin at varying distances from a point source. Each model begins at time 0 with zero water depth, i.e. the shoreline is exactly located at the model location. For the purpose of illustration, Jervey distinguishes marine 'mud-prone' from coastal plain 'sand-prone' facies in Fig. 2.11. Sediment grain-size is clearly a function both of sediment supply 'type' as well as sediment supply 'rate'. The sediment supply rate is held constant through the relative sea-level cycle in this simple model.

At the location with low rates of sediment influx, accommodation always exceeds sediment accumulation, the coastline migrates landward and transgression ensues, with considerable water depths being developed. Mud-prone marine facies may be expected to accumulate at some distance from a coastline located marginward of the figured depositional site. The rate of accumulation in this case reflects rate of supply of sediment to this point in the basin.

With a moderate rate of sediment influx, the sea-floor can aggrade to sea-level (base level). The rate of increase of accommodation initially exceeds the ability of sediment supply to maintain the sediment surface at sea-level and a transgression ensues. During the transgression water depth increases at this location and marine shales are deposited. As the rate of relative sea-level rise diminishes, regression of the shoreline commences. Regression of the shoreline

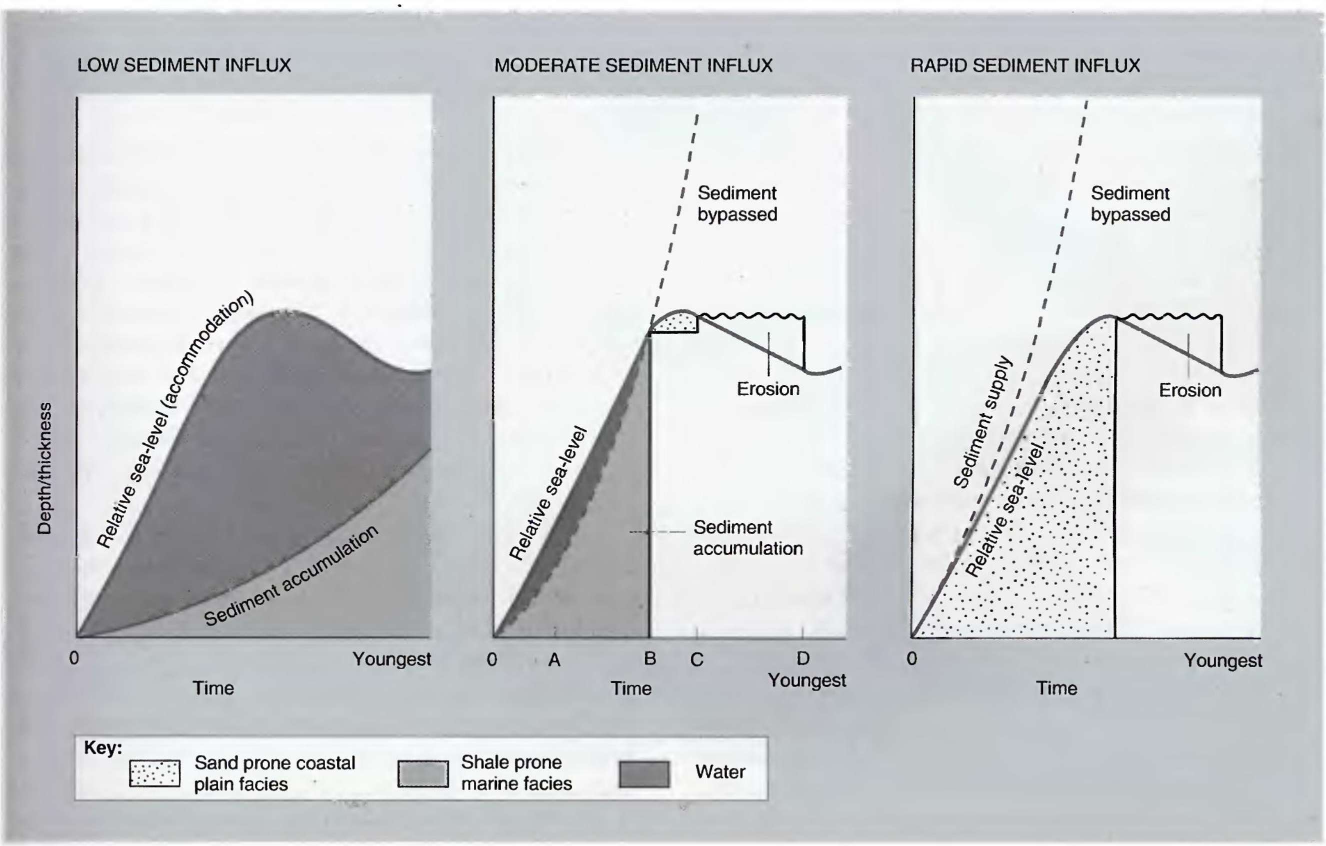


Fig. 2.11 Water depth and facies relationships with varying sediment supply rates (from Jervey, 1988)

continues until marine facies have aggraded to sea-level and the coastline again reaches the location shown. Thereafter, sediment supply rate exceeds the rate of creation of accommodation, the sediment surface is maintained at sea-level and coastal-plain facies accumulate. Excess sediment which cannot be accommodated in the coastal plain is transported basinwards. When accommodation decreases (relative sea-level falls) there is the potential for erosion of sediment deposited previously.

Where sediment influx is rapid, sediment supply rate always exceeds the rate of creation of accommodation and coastal/delta-plain sediments will accumulate. Regression of the shoreline will be continuous through the sea-level cycle. The rate of accumulation at this point in the basin is limited by the rate of accommodation increase. Erosion is likely when accommodation is removed during relative sea-level fall.

2.3.3 Basin architecture

In order to understand the behaviour of a topset/clinoform margin through time it is necessary to consider the balance between the rate of sediment supply and the rate of creation of *topset accommodation volume* (sometimes termed 'shelfal accommodation volume'). The rate of change of accommodation volume is a function of the magnitude of the sea-level rise multiplied by the topset area (Milton and Bertram, in press). If, during the same interval, the basin margin is supplied with a greater volume of sediment, then topset accommodation volume will be completely filled, and sediment will deposit on the clinoforms allowing the offlap break to prograde basinwards (Fig. 2.12).

Progradational geometries therefore occur when sediment supply exceeds the rate of creation of topset accommodation volume and facies belts migrate basinward. On seismic data progradation is expressed as *clinoforms* that show the basinward migration of the offlap break. *Regression* is a term that will be used here to refer specifically to basinward movement of the shoreline.

Aggradational geometries occur when sediment supply and rate of creation of topset accommodation volume are roughly balanced. Facies belts stack vertically and the offlap break does not migrate landward or basinward.

Retrogradational geometries occur when sediment supply is less than the rate of creation of topset accommodation volume. Facies belts migrate landward and the former depositional offlap break becomes a relict feature. *Transgression* is used here to refer specifically to the landward movement of the shoreline.

These phases of progradation, aggradation and retrogradation are not continuous but are made up of smaller (subseismic) scale progradational units called *parasequences* (see 2.5). Parasequences stack together in *parasequence sets* to make up the depositional geometries observable on seismic data.

The next section will show that the principles of cyclic changes in accommodation through time can be used to divide the sedimentary record into packages deposited during characteristic phases of the sea-level cycle.

2.4 Sequences and systems tracts

2.4.1 Sequences and sequence boundaries

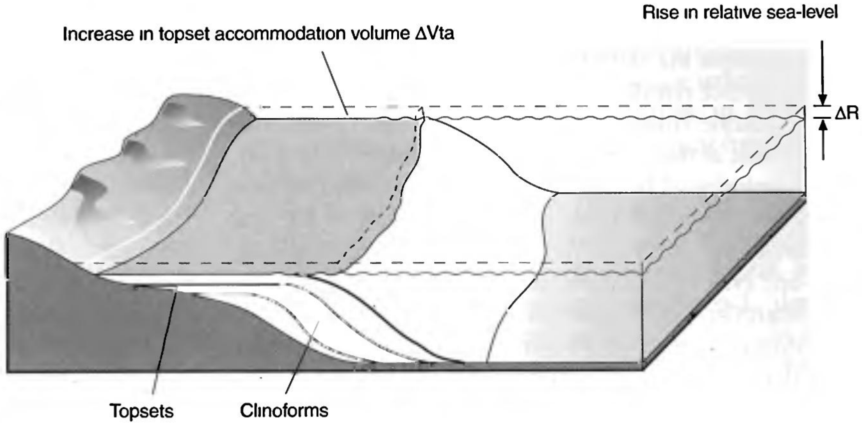
The term 'sequence', as applied in sequence stratigraphy, was defined originally by Mitchum *et al.* (1977a) as:

> A stratigraphic unit composed of a relatively conformable succession of genetically related strata bounded at its top and base by unconformities or their correlative conformities.

This generalized definition does not specify the scale or duration of the sequence, nor does it imply any particular mechanism for causing the unconformities. The term 'unconformity' in this definition was an initial cause of confusion, because the precise usage of the term can vary. Mitchum *et al.* (1977a) initially included marine hiatuses and condensed intervals in the term 'unconformity', but as models of cyclic deposition driven by relative sea-level variations developed, it became clear that basin-margin subaerial unconformities needed to be distinguished from basin-centre marine hiatuses. For the purpose of defining sequences, the term 'unconformity' is now restricted to a much narrower definition, namely 'a surface separating younger from older strata along which there is evidence of subaerial erosion and truncation (and in some areas correlative submarine erosion) and subaerial exposure and along which a significant hiatus is indicated' (Van Wagoner *et al.*, 1988).

Thus sequences are units bounded by significant subaerial erosion surfaces. Units bounded by marine condensed surfaces, surfaces of transgression, or marine onlap surfaces, are not sequences by this definition. It is interesting to note that the Exxon workers 'seriously considered using the term "synthem" instead of "sequence"' (Mitchum *et al.*, 1977a). In retrospect this might have avoided a lot of confusion with the sedimentological use of the term 'sequence', and with other cycle-defined 'sequences' (such as the genetic depositional sequences of Galloway, 1989). However 'synthem stratigraphy' does not have the same ring to it!

At first sight the definition quoted above seems simple enough. However, in practice, it is not so simple to apply. It is difficult to demonstrate non-marine exposure from a well log or from a seismic data set and tracing an erosion surface offshore into its 'correlative conformity' is often problematic. The term 'significant' in the clarification of Van Wagoner *et al.* (1988), quoted above, is not particularly helpful because it gives no indication of what magnitude of discontinuity is significant. *Composite sequences* (discussed in 2.4.10) are allowed to contain unconformities, provided these unconformities are of a higher 'order' than the ones which bound the sequence, and are therefore not significant.



The increment of topset accommodation volume ΔVta caused by a rise in relative sea-level ΔR is equal to the product of ΔR and the topset area

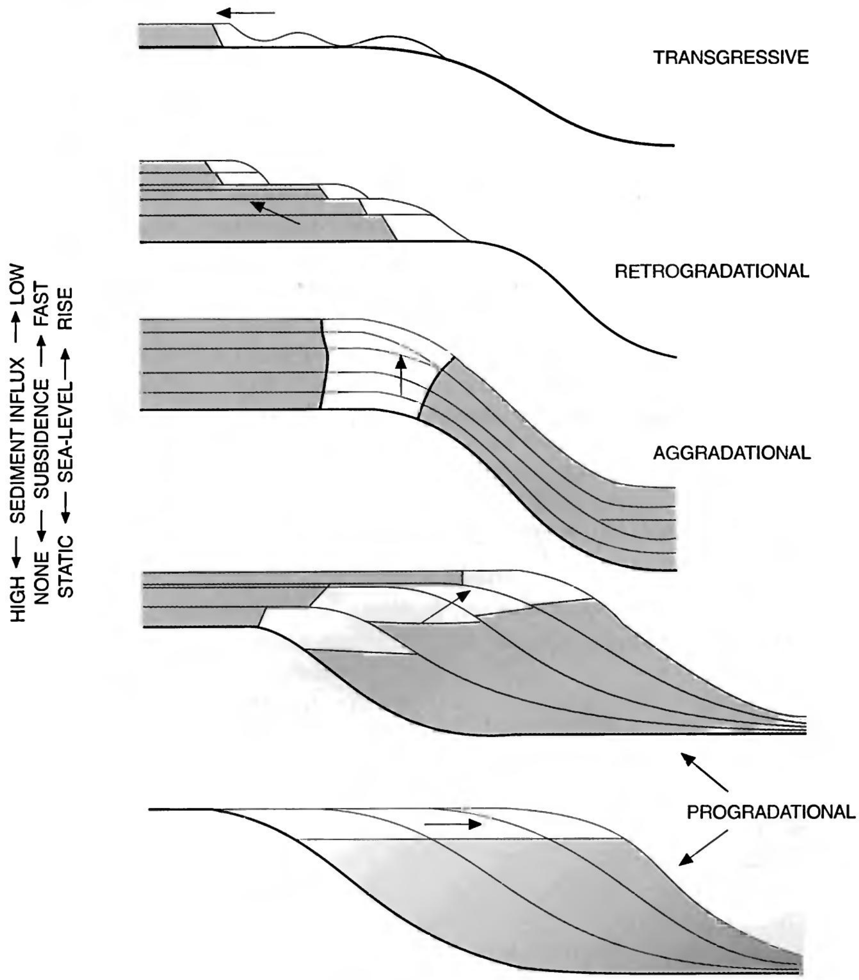


Fig. 2.12 Depositional architecture as a function of accommodation volume and sediment supply (after Galloway, 1989)

The stratigrapher must define 'significant' on the scale of the particular study.

Nevertheless the restricted principle is straightforward, and a sequence represents one cycle of deposition bounded by non-marine erosion, deposited during one 'significant' (on the scale of the study) cycle of fall and rise of base level. In the majority of basins, base level is controlled by sea-level, and a sequence is the product of a cycle of fall and rise of relative sea-level. An idealized sequence resulting from one cycle of base level change is shown in Fig. 2.13, after figures in Van Wagoner *et al.* (1988). This is a *type 1 sequence*, where the fall in relative sea-level is sufficiently large that the first topsets within the sequence onlap the clinoforms of the previous sequence, implying a fall in relative sea-level at the position of the offlap break. *Type 2 sequences* are described below in section 2.4.6.

According to Van Wagoner *et al.* (1988), a type 1 sequence boundary is characterized by subaerial exposure and concurrent subaerial erosion associated with stream rejuvenation, a basinward shift in facies, a downward shift in coastal onlap, and onlap of overlying strata. Coastal onlap is a term used to describe the onlap point on topset strata at the basin margin (see Chapter 3). As a result of the basinward shift in facies, non-marine or marginal marine rocks, such as braided-stream or estuarine sandstones, may directly overlie shallow-marine rocks, such as lower shoreface sandstones or shelf mudstones, across a sequence boundary with no intervening rocks deposited in intermediate depositional environments. This facies superposition is termed a *facies dislocation*. A type 1 sequence boundary is interpreted by Van Wagoner *et al.* (1988) to form when the rate of eustatic fall exceeds the rate of basin subsidence at the offlap break, producing a fall in relative sea-level at that position.

2.4.2 Systems tract definition

The idealized type 1 sequence shown on Fig. 2.13 is representative of a shelf-break margin. It can be seen to be comprised of a number of distinct depositional packages. It was observed in the early days of seismic stratigraphy that deposition in a basin was not uniform and continuous but occurred in a series of discrete 'packets' bounded by seismic reflection terminations (see Chapter 3). Workers in Exxon found that these packages generally were arranged in a predictable fashion in the majority of sequences they observed on seismic data. These packages are known as *systems tracts*.

This term systems tract was first defined by Brown and Fisher (1977) as a linkage of contemporaneous depositional systems, where a depositional system is a three-dimensional assemblage of lithofacies, genetically linked by active (modern) or inferred (ancient) processes and environments (after Fisher and McGowen, 1967).

A systems tract is therefore a three-dimensional unit of deposition, and the boundaries of a systems tract are depositional boundaries of onlap, downlap, etc. The seismic expression of a systems tract is a unit of conformable reflections bounded by surfaces of reflection termination ('seismic-stratigraphic units' of Brown and Fisher, 1977; 'seismic sequences' of Mitchum *et al.*, 1977a; referred to as 'seismic packages' in Chapter 5).

Systems tracts are recognized and defined by the nature of their boundaries and by their internal geometry. Within any one relative sea-level cycle, three main systems tracts that characterize different parts of the relative sea-level cycle are frequently developed (Fig. 2.13).

It is easy to become lost in the complexities of systems tract terminology, and it is always worth remembering the purpose of stratigraphic division into systems tracts. The systems tract represents the fundamental mapping unit for stratigraphic prediction, because it contains a set of depositional systems with consistent palaeogeography and depositional polarity, and for which a single palaeogeographic map can be drawn.

2.4.3 Lowstand systems tract

The basal (stratigraphically oldest) systems tract in a type 1 depositional sequence is called the *lowstand systems tract.*

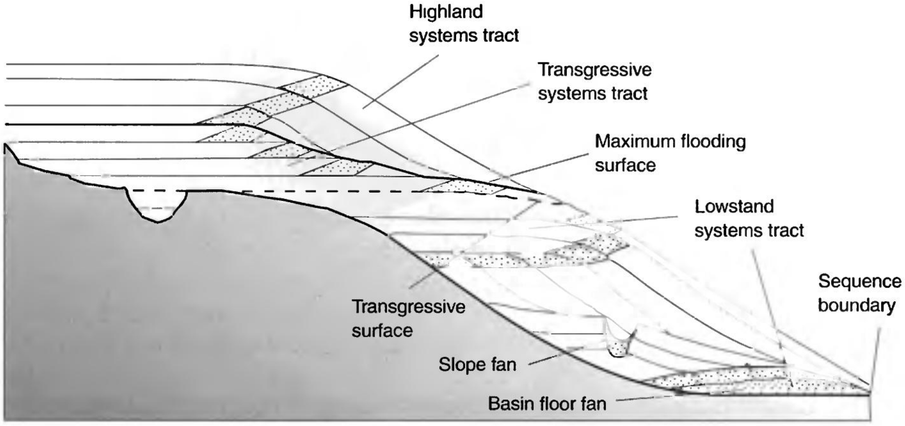


Fig. 2.13 Stratal geometries in a type 1 sequence on a shelf-break margin. Five separate sedimentary packages are shown, traditionally assigned to three systems tracts; lowstand, transgressive and highstand

The lowstand systems tract is deposited during an interval of relative sea-level fall at the offlap break, and subsequent slow relative sea-level rise.

Falling relative sea-level at the offlap break of a shelf-break margin will have an extreme effect on the river systems. Prior to the fall in relative sea-level, the rivers will have more-or-less maintained a graded river profile with an erosional upper portion and a depositional lower portion (alluvial plain and coastal plain). The rivers will have been free to avulse, responding to rises in relative sea-level over this lower portion. When relative sea-level falls at the offlap break, the river profile must adjust to the lowered base level (see Chapter 7). The river incises into the previously deposited topsets; the alluvial plain, coastal plain and/or shelf deposits of the previous sequence. These reworked sediments, and the fluvial load from the hinterland, are delivered directly on to the previous highstand clinoform slope. Because the river is not free to avulse, the sediment is focused towards the same point on the slope. This is an inherently unstable situation, and sedimentation processes are dominated by large-scale slope failure resulting in bypass of the slope and deposition of submarine fans in the basin. These processes continue to dominate the sedimentary record while relative sea-level is falling and the river system is forced to incise.

At the relative sea-level low point the river profile stabilizes again, and a prograding topset–clinoform system can then be established. The first topset of this system will onlap below the level of the previous offlap break. This is known as a *downward shift in coastal onlap below the level of the offlap break*, and is indicative of a type 1 sequence boundary. The rate of rise of relative sea-level is initially low, and together with the limited topset area of the prograding system, this results in a low rate of creation of topset accommodation (see Fig. 2.15). This will be outpaced by sediment supply, and so the system will prograde. However the accelerating rate of creation of accommodation volume eventually may outpace sediment supply, resulting in a change from progradation to aggradation and retrogradation, and the onset of the next (transgressive) systems tract.

The lowstand systems tract therefore consists of two parts; a unit of submarine fans deposited during falling relative sea-level, and a topset/clinoform system, initially progradational but becoming aggradational, deposited during a slow rise of relative sea-level. These can be treated as separate and distinct systems tracts, because the fans and the topset–clinoforms need never have been in depositional continuity. They are traditionally both placed in a single lowstand systems tract, on the basis that the boundary between the two may be gradational rather than distinct, with submarine fans forming much of the slope portion of the lowstand wedge (Posamentier and Vail, 1988).

Lowstand submarine fans

Frequently two distinct fan units can be recognized within the lowstand submarine fans; an initial *basin floor fan* unit, detached from the foot of the slope, and a subsequent *slope fan* unit, abutting the slope, occasionally referred to in older literature as 'slope front fill' (see Fig. 2.14). Van Wagoner *et al.* (1988) describe the basin-floor fan as being characterized by submarine fan deposits on the lower slope or basin floor. Fan formation is associated with the erosion of canyons into the slope and the incision of fluvial valleys into the shelf. Siliciclastic sediment bypasses the shelf and slope through the valleys and the canyons to feed the basin-floor fan. The base of the basin-floor fan (coincident with

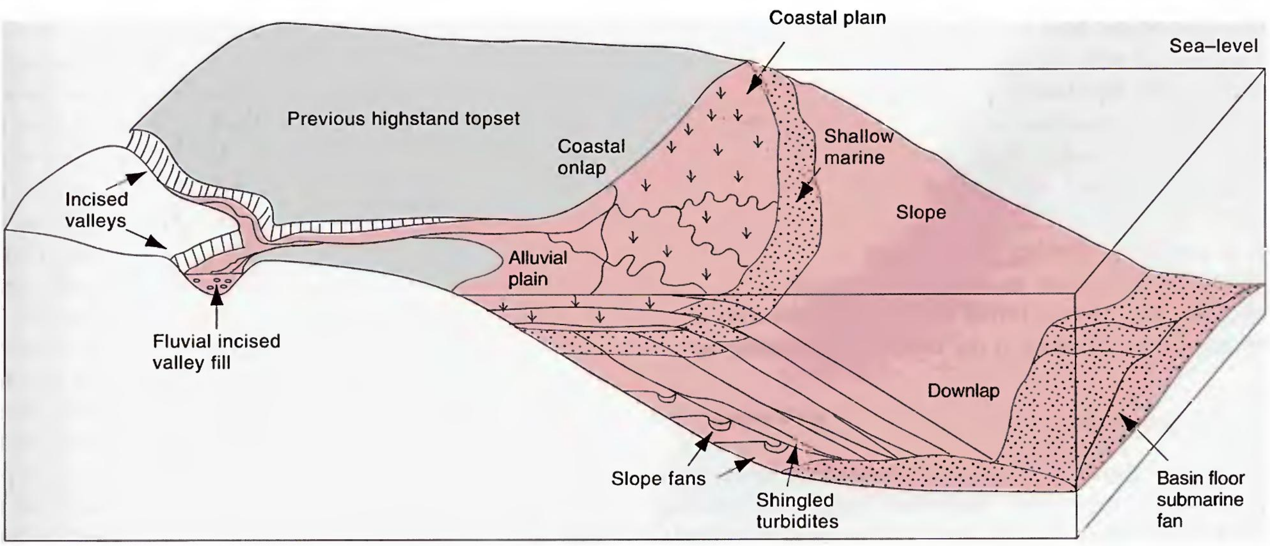


Fig. 2.14 Components of the lowstand systems tract on a shelf-break margin. These include a basin-floor fan and a slope fan, but the diagram also shows the active systems of the lowstand wedge; namely valley fill, alluvial and coastal plain topsets, a shallow marine belt and an active slope system, which in its early stages may contain shingled turbidites

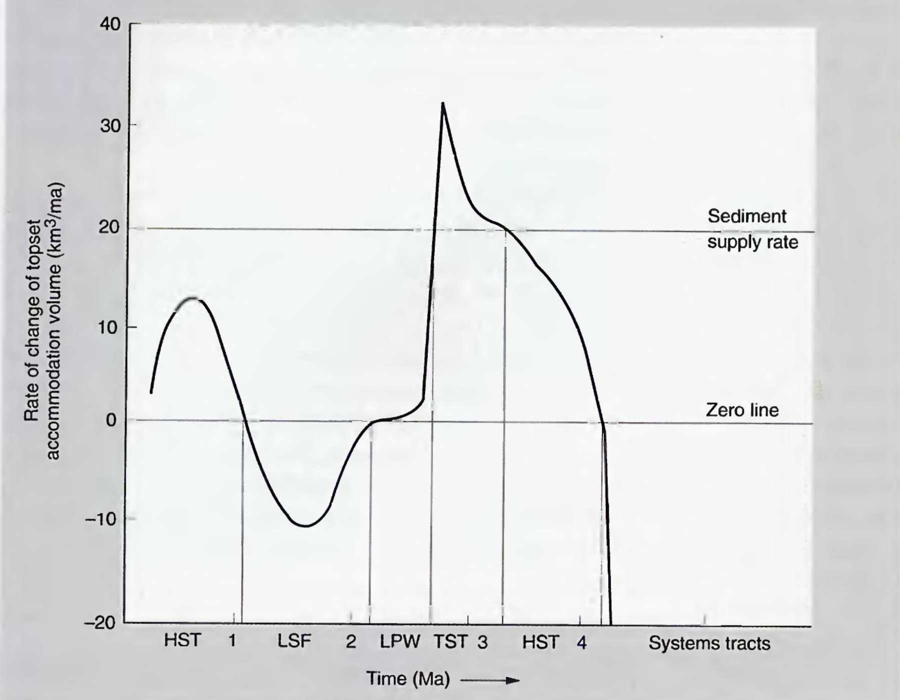


Fig. 2.15 The relationship between relative sea-level, topset accommodation volume, and systems tracts, in a simple numerical model with sinusoidal relative sea-level and constant sediment supply. The large change in topset accommodation volume between the lowstand and transgressive systems tracts is caused by flooding back over the highstand topsets. HST, highstand systems tract; LSF, lowstand submarine fan; LPW, lowstand prograding wedge; TST, transgressive systems tract

the base of the lowstand systems tract) is correlated with the type 1 sequence boundary and the top of the fan may be a downlap surface to the subsequent lowstand prograding wedge, if it prograded far enough, or may be a downlap surface for any overlying slope fans. Basin-floor fan deposition, canyon formation, and incised-valley erosion are interpreted to occur during a fall in relative sea-level over the entire topset area.

Slope fans are described by Van Wagoner *et al.* (1988) as characterized by turbidite and debris-flow deposition on the middle or the base of the slope. Slope-fan deposition can be coeval with the basin-floor fan or with the early portion of the lowstand wedge. The top of the slope fan may be a downlap surface for the middle and upper portions of the lowstand wedge. Slope fans are typically described as being composed of channel–levee complexes (see Chapter 9).

It is not clear whether two distinct fan units will be visible in all sequences in all basins, and the interpreter should beware of force-fitting a twofold subdivision on a submarine fan succession if the data do not warrant it.

Lowstand prograding wedge

As described above, the lowstand prograding wedge is a topset–clinoform system deposited during accelerating relative sea-level rise. It is separated from the overlying transgressive systems tract by a *maximum progradation surface*, marking a change in parasequence stacking geometry from progradational (in the lowstand wedge) to retrogradational (in the transgressive systems tract). Deposition of the lowstand prograding wedge is confined initially to the areas around the mouths of the incised rivers (Fig. 2.15). Little if any topset accommodation volume is created at this time, and the bulk of the sediment bypasses the topsets to be deposited on the clinoform slope. Slope instability and occasional fan deposition is likely to occur, and the bottom sets of the early lowstand prograding wedge may contain interbedded turbidites, which often have a characteristic 'shingled' seismic facies.

As relative sea-level begins to rise, the fluvial valleys incised into older topsets during falling relative sea-level begin to be back-filled in their lower reaches with fluvial or estuarine deposits, while topsets of the prograding wedge begin to be deposited. Accelerating relative sea-level rise results in a facies association indicative of increasing accommodation volume, such as an upwards increase in coals, overbank shales, lagoonal facies, tidal influence, etc., and a decrease in the connectivity of fluvial sandbodies. The transition to the overlying transgressive systems tract may be a gradational turnaround from progradation to retrogradation, or may, as in Fig. 2.15, be abrupt, as relative sea-level rises above the level of the offlap break of the previous sequence, resulting in a huge increase in topset accommodation volume. This boundary can be called the *maximum progradation surface*, the *transgressive surface*, or the *top lowstand surface*.

The lowstand prograding wedge is often sandier than the preceding highstand wedges, owing to the recycling of sands from the highstand topsets. In a predominantly muddy system, sandy lowstand wedges can be sealed against underlying shales of the highstand systems tract and overlying shales of the transgressive systems tract, thus forming stratigraphic traps.

2.4.4 Transgressive systems tract

The *transgressive systems tract* is the middle systems tract of both type 1 and type 2 sequences (Figs 2.13, 2.16 and 2.18). It is deposited during that part of a relative sea-level rise cycle when topset accommodation volume is increasing faster than the rate of sediment supply. It contains mostly topsets, with few associated clinoforms, and is entirely retrogradational. The active depositional systems are topset systems; alluvial, paralic, coastal plain and shelfal. Any deltas are shelf deltas. These systems may show evidence of an undersupply of sediment, and may be rich in coals, overbank deposits and lagoonal or lacustrine deposits. Drainage systems may be flooded to form estuaries. Wide shelf areas are characteristic of transgressive systems tracts, and tidal influence may be widespread. The transgressive systems tract passes distally into a condensed section characterized by extremely low rates of deposition and the development of condensed facies such as glauconitic, organic rich and/or phosphatic shales (Chapter 11), or pelagic carbonates.

The maximum rate of rise of relative sea-level occurs some time within the transgressive systems tract, and the end of the systems tract occurs when the rate of topset accommodation volume decreases to a point where it just matches sediment supply, and progradation begins again. This point is known as the *maximum flooding surface*.

Topsets of the transgressive systems tract tend to have a lower sand percentage than those of other systems tracts, because little of the mud-grade sediment bypasses the topsets. The transgressive systems tract can therefore often host sealing horizons to topset reservoirs, and sometimes also source beds (see Chapter 11). Posamentier and Allen (1993b) proposed a new component of transgressive systems tracts, which they termed the 'healing phase' component. They showed several examples of sediment wedges banked against the foot of highstand clinoforms, which they related to sediment reworked basinwards during transgression. An alternative view of these wedges could be that they may be the lowstand components of higher order sequences within a composite systems tract, or products of retrogressive slumping of the highstand slope.

The present-day depositional systems over much of the globe form a transgressive systems tract. Wide continental shelves are common (many of which are the flooded topsets of the last lowstand). Most of the major deltas are shelf deltas and most of the major fans are inactive. Estuaries and tidal seas are common around northwest Europe, whereas the eastern USA coast is dominated by retreating barrier coastlines and lagoons, with deep-sea sedimentation generally restricted to rare turbidites sourced from retrogressive slumping of the continental slope.

2.4.5 Highstand systems tract

The *highstand systems tract* is the youngest systems tract in either a type 1 or a type 2 sequence (Figs 2.13, 2.18). It

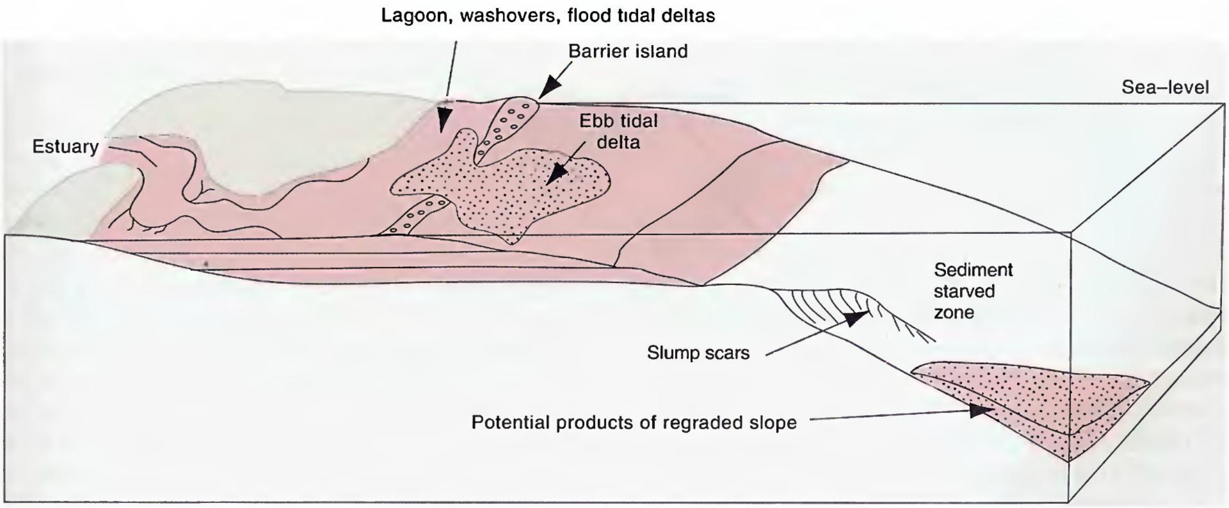


Fig. 2.16 Components of the transgressive systems tract. These are all topset systems, and here are shown to have significant tidal influence, due to the wide shelfal area of the drowned lowstand topsets. Deposition includes estuarine, lagoonal, barrier and tidal depositional systems, which pass seaward into a shelfal condensed zone

represents the progradational topset–clinoform system deposited after maximum transgression and before a sequence boundary, when the rate of creation of accommodation is less than the rate of sediment supply (Fig. 2.17). The highstand systems tract is characterized by a decelerating rate of relative sea-level rise through time, resulting in initial aggradational and later progradational architecture. Depositional systems may be similar initially to those in the transgressive systems tract, but the infill of shelf areas by progradation, and the decrease in the rate of relative sea-level rise, may lead to a decrease in tidal influence during a highstand systems tract, and a decrease in the amount of coal, and of overbank, lagoonal and lacustrine shales. Channel sandbodies will become more common and more connected.

Posamentier and Vail (1988) discuss various models which imply that the late highstand systems tract is characterized by significant fluvial deposition. They used the concept of the 'bay line', which they defined as the line to which stream profiles are graded and where fluvial processes are replaced by paralic and shelf processes. The bay line also represents the coastal onlap point during relative sea-level rise. Late in the highstand systems tract the bay line begins to migrate basinward as relative sea-level falls in the proximal part of the depositional profile, and Posamentier and Vail (1988) suggest significant alluvial accommodation will be generated. These models are an oversimplification and have been a source of considerable misunderstanding, e.g. Miall (1991), Shanley and McCabe (1994) and discussion in Chapters 7 and 8.

2.4.6 Type 2 sequence boundary and the shelf-margin systems tract

Relative sea-level may fall over the proximal area of the highstand topsets, without falling at the offlap break. A sequence boundary results, but not one characterized by fluvial incision or submarine fan deposition. The sequence boundary is recognized on seismic data by a downward shift in coastal onlap to a position landward of the offlap break, where topset reflections can be seen onlapping an older topset (Fig. 2.18). This is known as a type 2 sequence boundary, and the subsequent systems tract is known as a *shelf-margin systems tract.* It consists of prograding topsets and clinoforms, and is progradational initially but becomes aggradational upwards, passing eventually into a retrogradational transgressive systems tract. The shelf-margin systems tract may be very difficult to recognize in outcrop or on a well-log data base, and is differentiated from the underlying highstand systems tract by a subtle unconformity only, and possibly by a change in parasequence stacking pattern. It also could be recognized in a grid of wells, or a large area of outcrops, as the onlap of one parasequence on to another, and the merging of flooding surfaces (e.g. the merging of two coal seams).

The type 2 sequence boundary and shelf-margin systems tract are sometimes misused in the literature because of the difficulty in demonstrating a basinward shift in coastal onlap towards but not beyond the offlap break. Seismic resolution is often insufficient to resolve the subtle change in dip where topset onlaps topset. The change from progradation to aggradation, which is also considered characteristic of the type 2 boundary, is not definitive on its own because other factors such as decreasing sediment supply

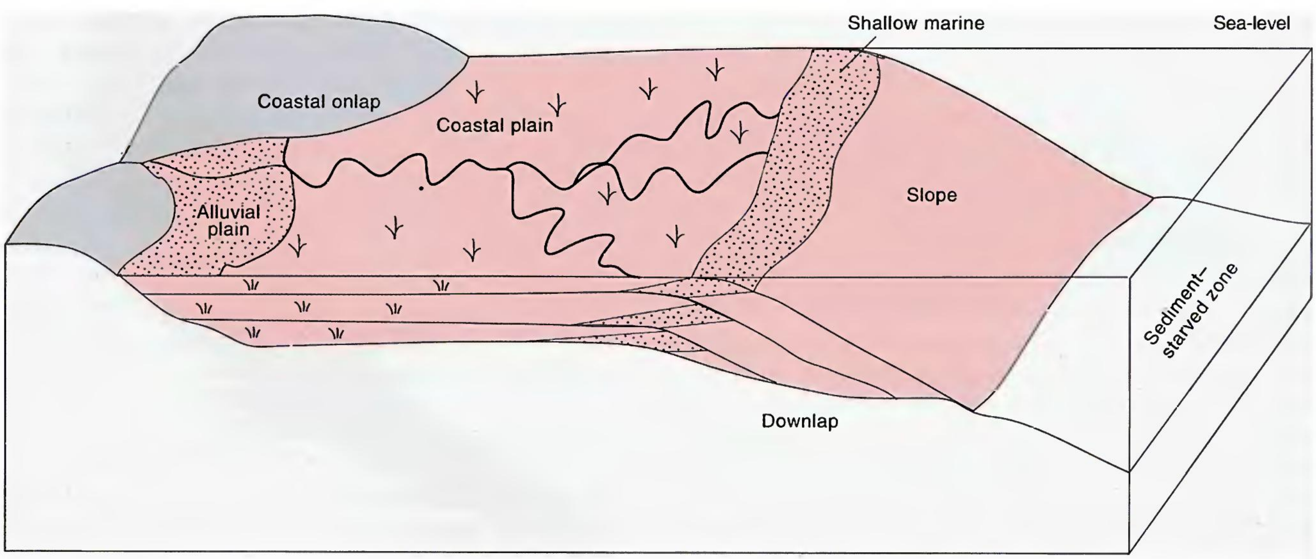


Fig. 2.17 Components of the highstand systems tract on a shelf-break margin. These include topset (alluvial, coastal), shallow marine, and slope systems

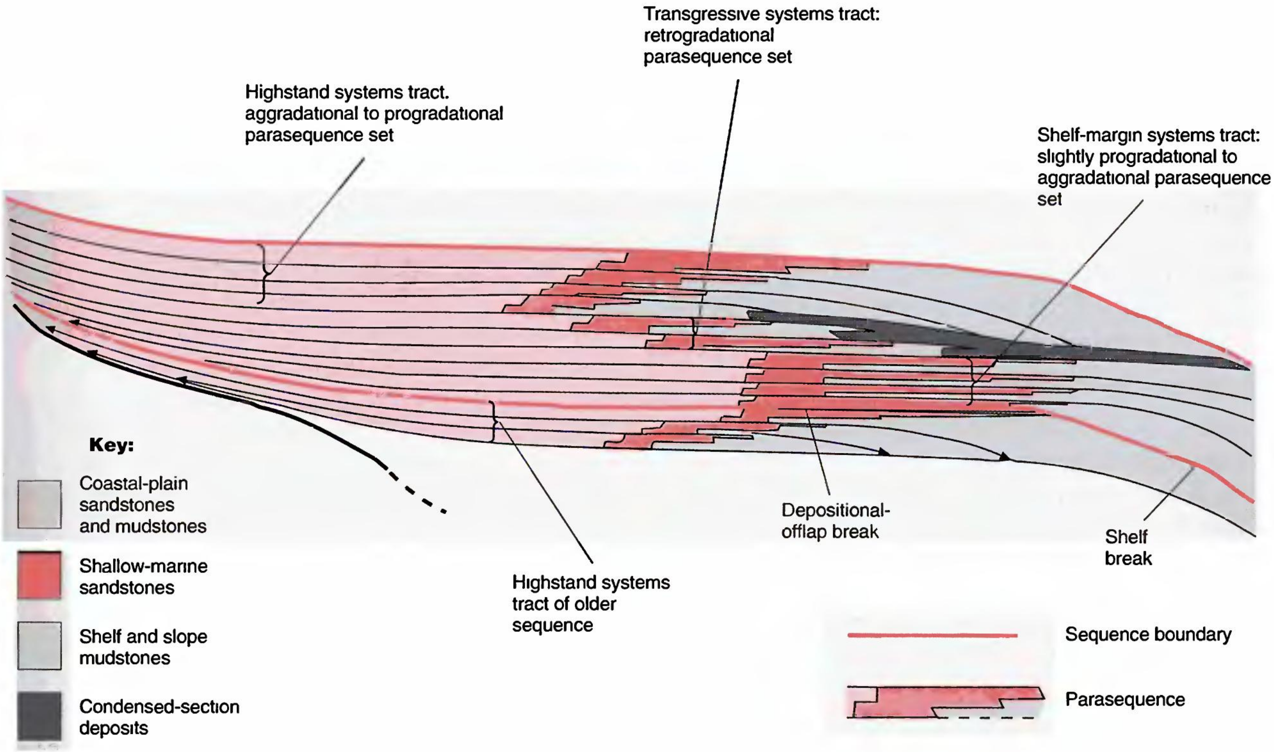


Fig. 2.18 A type 2 sequence. The type 2 sequence boundary is recognized from a downward shift in coastal onlap landward of the offlap break. This downward shift does not result in shelf bypass, fan deposition, or incision of the highstand topsets. The sequence boundary is overlain by a shelf-margin systems tract; a system tract of topsets with a predominantly aggradational stacking pattern. The rate of sea-level fall at the shoreline is equal to, or less than, the subsidence (from van Wagoner *et al.*, 1988)

rates could also affect such a pattern. In outcrop studies type 2 sequence boundaries are often used simply to distinguish minor sequence boundaries. Note also that a type 2 sequence boundary may pass laterally into a type 1 sequence boundary depending on the tectonic subsidence pattern in the basin.

2.4.7 Lowstand systems tracts on a ramp margin

The systems tracts described above are developed on a shelf-break margin, where the clinoform slope is steep enough and deep enough to allow large-scale failure and the formation of submarine fan systems. On a *ramp margin*, the lowstand systems tract was described by Van Wagoner *et al.* (1988) as consisting of a relatively thin lowstand wedge that may contain two parts (Fig. 2.19). The first part is characterized by stream incision and sediment bypass of the coastal plain. This is interpreted to occur during a relative fall in sea-level when the shoreline steps rapidly basinward until the relative fall stabilizes. The second part of the wedge is characterized by a slow relative rise in sea-level, the infilling of incised valleys, and continued shoreline progradation. This results in a lowstand wedge composed of incised valley-fill deposits up-dip and one or more progradational parasequence sets down-dip. The top of the lowstand wedge is the transgressive surface; the base of the lowstand wedge is the sequence boundary (discussed in detail in 8.3.4).

During falling relative sea-level on a ramp margin there is no bypass of sediment to the basin floor. Instead the sediment may be deposited as a set of downstepping prograding wedges, known as forced regressive wedges (Posamentier *et al.*, 1992). A number of these may be preserved between the highstand and lowstand prograding wedges. Posamentier (1993) referred to these wedges as *forced regressive wedge systems tracts*, which at the site of deposition are overlain by 'regressive subaerial surfaces of erosion', and underlain by 'regressive marine surfaces of erosion'. The latter are likely to pass landward into subaerial unconformities, and therefore both boundaries are strictly sequence boundaries. These forced regressive wedges are often sand-rich, and may form attractive stratigraphic traps where encased in shale. A number of examples of forced regressions, i.e. falls in relative sea-level on a ramp margin, are presented by Posamentier *et al.* (1992), and Posamentier and Chamberlain (1992) describe the detailed stratigraphy of a ramp-margin lowstand systems tract within the Viking Formation of Canada.

The transgressive and highstand systems tracts on a ramp margin are similar to those on a shelf-break margin,

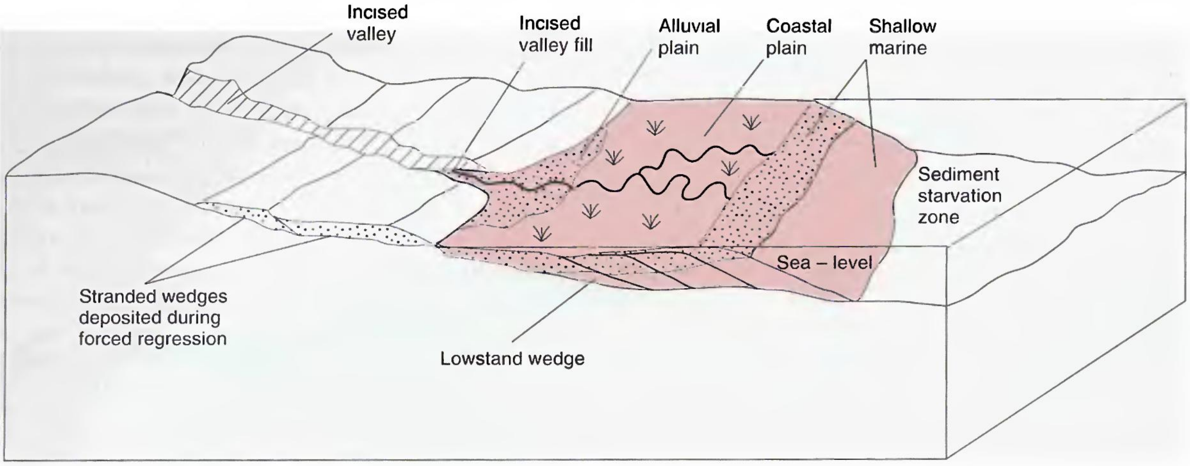


Fig. 2.19 Components of the lowstand systems tract on a ramp margin. The water depth is not great enough to allow development of a significant slope, and therefore turbidite systems are not developed during falling relative sea-level

although there is no significant clinoform component to the highstand systems tract.

2.4.8 Controls on systems tract boundaries

The actual time of initiation of a systems tract is interpreted by Van Wagoner *et al.* (1988) to be a function of the interaction between eustasy, sediment supply and tectonics. To this can be added 'topset area', which has a significant bearing on the development of the transgressive surface and the maximum flooding surface. Figure 2.15 shows the relationship between topset accommodation and systems tracts in a simple system of continuous subsidence and sinusoidally varying eustatic sea-level. The systems tract boundaries occur at the following points.

1 The type 1 sequence boundary (base of the lowstand systems tract) occurs when the rate of relative sea-level rise is zero and decreasing at the offlap break. The time at which this occurs is a function of eustasy and subsidence.

2 The boundary between the lowstand fans and the lowstand prograding wedge occurs when the rate of relative sea-level rise is zero and then increasing at the offlap break. The time at which this occurs is a function of eustasy and subsidence.

3 The boundary between the lowstand prograding wedge and the transgressive systems tract occurs when the rate of creation of topset accommodation volume equals, and is just about to exceed, the rate of sediment supply. Topset accommodation volume is the change in accommodation volume over the topset area during a rise in relative sea-level, and when this exceeds the sediment supplied, a transgression will occur. The timing of the boundary is a function of eustasy, subsidence, sediment supply, and topset area, and may occur when sea-level first floods back over the previous highstand topsets (as in Fig. 2.15).

4 The boundary between the transgressive systems tract and the highstand systems tract (the maximum flooding surface) occurs when the rate of creation of topset accommodation volume equals, and is just about to fall below, the rate of sediment supply. This is a function of eustasy, subsidence, sediment supply, and topset area.

It can be seen from the above that the timing of most of the systems tract boundaries is affected by very many factors. Those affected by fewest factors are the sequence boundary and the top of the lowstand fans.

The relative volumes of the systems tracts will be a function of their duration, and the sediment supply rate. There may be a linkage between supply rate and systems tract, for example in high latitudes where low sea-level in a glacial period may be associated with ice cover in the fluvial drainage basin. These factors, and factors such as the basin topography, can seriously distort the ideal sequence geometry shown in Fig. 2.15. It is very rare to find a seismic line that looks like this ideal model. This does not mean the model is wrong, merely that it should not be used as a template for interpretation.

2.4.9 Other possible systems tracts within a relative sea-level cycle

Van Wagoner *et al.* (1988) suggest that systems tracts should be defined objectively on the basis of the types of bounding surface, their position in a sequence (if this can be determined) and on their internal geometry. Two other theoretical systems tracts are not recognized in the original Exxon scheme. These are shown on Fig. 2.20, and described below.

The *midstand systems tract* (or *forced regressive systems tract* of Hunt and Tucker (1992); see 2.4.7) represents an entire sequence where at no time subsidence was sufficiently high to outpace sediment supply and allow transgression. This might be expected in basins with low or negative tectonic subsidence and/or high rates of sediment supply. Third-order scale midstand systems tracts on a shelf-break

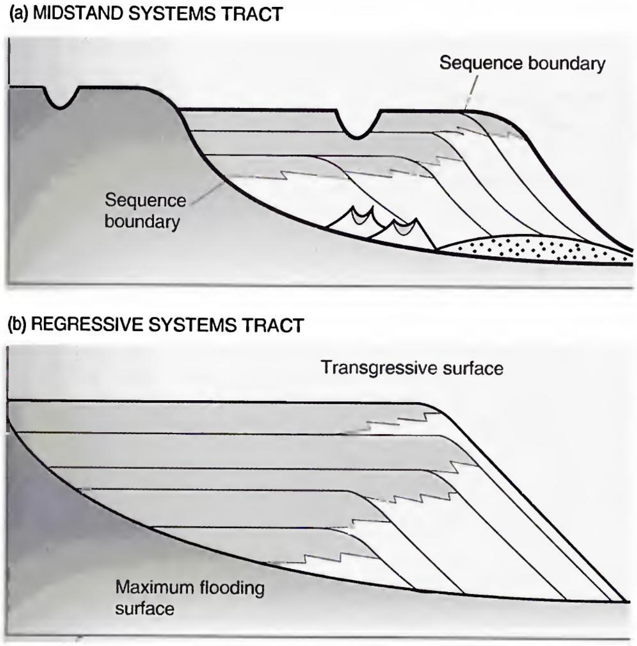


Fig. 2.20 Systems tracts not currently described in the Exxon scheme; (a) midstand systems tract; (b) regressive systems tract

margin have been described by Jones and Milton (1994) and Milton and Dyce (1995) from the Palaeogene of the North Sea at a time of basin-margin uplift. On a shelf-break margin (such as the North Sea Tertiary, or the Rhone delta) the midstand systems tract may comprise a unit of fans and a prograding wedge. On a ramp margin only prograding wedges will be developed.

The *regressive systems tract* (Fig. 2.20) is a theoretical systems tract formed between two rapid rises in relative sea-level separated by a slow rise (or by a pulse of increased sediment supply during continuous rate of rise). The systems tract is bounded below by a maximum flooding surface, and consists of a prograding wedge. The prograding wedge is bounded above by a maximum progradation surface. The internal geometry of the wedge is aggradational to progradational to aggradational again. Regressive systems tracts would be expected when eustatic cycles were superimposed on rapid background subsidence, so that not even type 2 sequence boundaries formed during the eustatic fall. Alternatively they may occur during a steady rise in relative sea-level with fluctuating sediment supply. Regressive systems tracts were predicted to occur in foreland basins by Posamentier and James (1993), although these authors termed them shelf-margin systems tracts, despite the absence of an underlying sequence boundary.

2.4.10 Composite (second and third order) sequences and systems tracts

Composite sequences were defined by Mitchum and Van Wagoner (1991) as 'successions of genetically related sequences in which the individual sequences stack into lowstand, transgressive and highstand sequence sets'. Figure 2.21 shows a composite sequence; bounded by two sequence boundaries but also containing four higher order sequence boundaries. Figure 2.22 shows the relative sea-level curve associated with Fig. 2.21, where two orders of cyclicity are apparent.

Most second- and many third-order sequences will contain higher order sequence boundaries, so it is important to state the order at which a sequence or systems tract is defined. For example, the highstand systems tract of a second-order composite sequence may, in reality, be a *highstand sequence set*, i.e. a stack of higher order sequences where the topset prograding parasequences may be dominant, but some higher order lowstand deposits may also exist. This has been demonstrated by Jones and Milton (1994), where all the systems tracts in a second-order sequence in the North Sea Tertiary contain third-order-scale lowstand fans. These fans particularly dominate the stratigraphy in the second-order lowstand systems tract. Finally, it is important to remember that systems tract boundaries in a composite sequence will be gradational, representing the interfingering of sequences and systems tracts of a higher order.

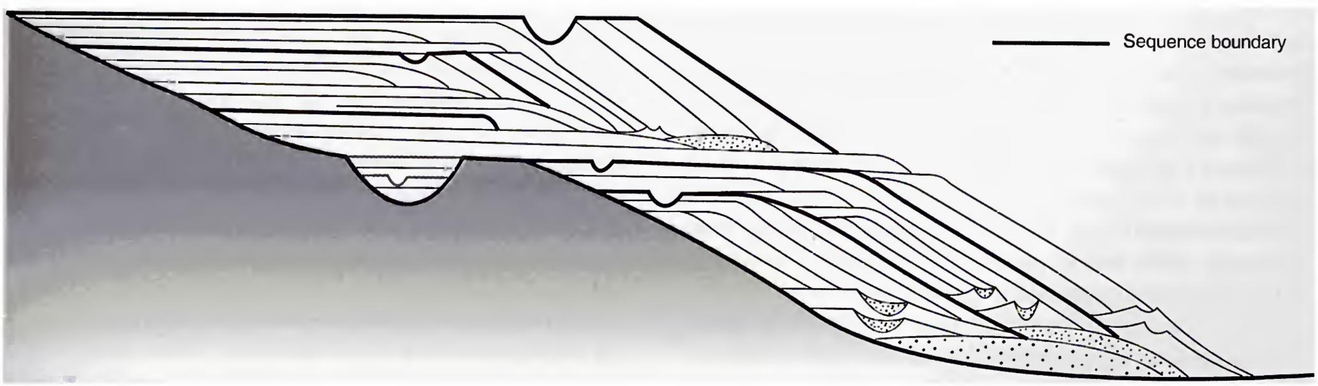


Fig. 2.21 A type 1 composite sequence, consisting of a stack of five higher order sequences. The high-frequency sequences form the building blocks of the composite sequence, and their nature is determined by their position in the composite sequence

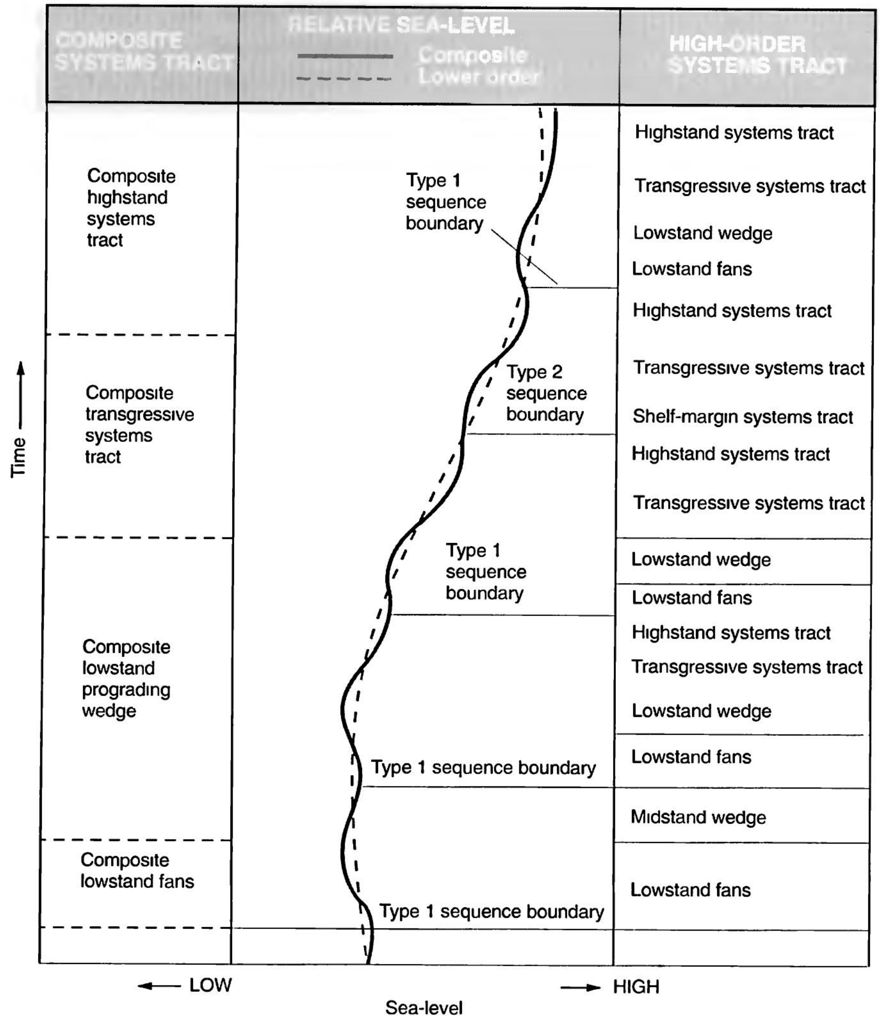


Fig. 2.22 The combination of low-order and high-order variations in relative sea-level implied in Fig. 2.21. The nature of the high-order sequences (their type, and the systems tracts within them) are determined by their position within the low-order sequence

2.4.11 Genetic stratigraphic sequences

Sequences, as discussed above, are cyclic stratigraphic units bounded by subaerial unconformities. However, because deposition is cyclic, the choice of the boundary is relatively arbitrary. Galloway (1989), after the work of Frazier (1974), suggested another means of subdividing the stratigraphy, using the maximum flooding surface as the cycle boundary. He therefore defined a *genetic stratigraphic sequence* as a package of sediments recording a significant episode of basin-margin outbuilding and basin filling, bounded by periods of widespread basin-margin flooding (Fig. 2.23).

With hindsight it is regrettable that he used the term 'sequence' here, instead of the 'depositional episode' of Frazier (1974), because this has led to a great deal of confusion. Some workers have used the Mitchum *et al.* (1977a) definition of the term, and some the Galloway (1989) definition. The sequence boundary, the maximum flooding surface and the maximum progradation surface are all valid correlation surfaces for dividing the stratigraphy. Each surface has its advantages and disadvantages as a primary stratigraphic boundary.

The sequence boundary can be recognized easily on seismic data by a downward shift in coastal onlap (as described in section 3.2.4). It indicates bypass and resedimentation into the basin, and is associated with the development of basinal reservoirs and hydrocarbon play systems. Recognition of the sequence boundary is therefore of great practical value in stratigraphic prediction for petroleum exploration. The timing of the sequence boundary is independent of sediment supply variations, so it is relatively isochronous. It is, however, difficult to see in a log and core data set, difficult to date precisely (occurring within proximal sediments potentially barren of fossils), and difficult to trace into the basin (except where associated with submarine fans).

The maximum flooding surface is also easily recognizable on seismic data, and can be identified easily on logs and in core. It is associated with topseal and often also with source-rock development. It may be represented by condensed marine facies, with a rich and easily dated fauna. It

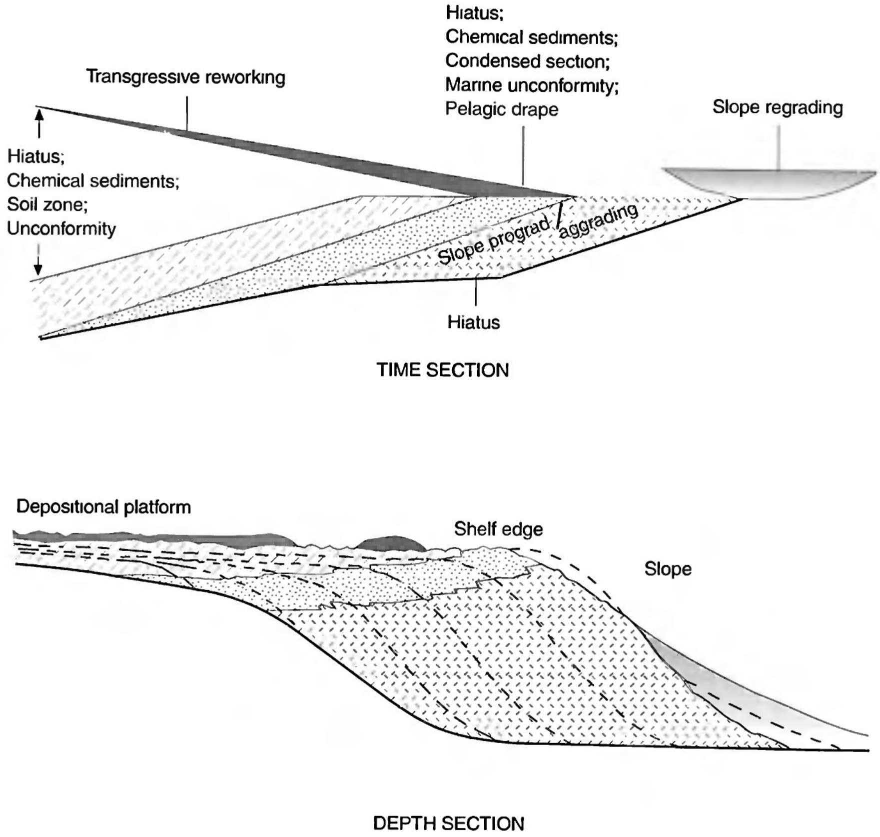


Fig. 2.23 A genetic stratigraphic sequence after Galloway (1989)

can be traced into the basin, where it correlates with a condensed interval, but may be difficult to trace into the proximal alluvial plain. Where the system is made up of several prograding lobes, it may be difficult to define precisely which lobe was most landward, and therefore where the maximum flooding surface lies. It is not associated with any particular reservoir development, and is seldom obvious in outcrop, where mudstones associated with the maximum flooding surface tend to be eroded or obscured.

The maximum progradation surface, or transgressive surface, also has been proposed by some workers as a natural surface for subdividing stratigraphy. This surface represents the furthest extent of topset reservoirs into the basin. It is recognized easily on seismic, outcrop, log and core data. It may be difficult to date precisely, and is difficult to correlate into proximal settings, and where the system is made up of several prograding lobes, it may be difficult to define precisely which lobe prograded the furthest and therefore where the maximum progradation surface lies.

The term sequence is usually now restricted to a unit bounded by subaerial unconformities. However, the most obvious surfaces in a basin are often the condensed intervals and maximum flooding surfaces (Loutit *et al.*, 1988). These can be used for a pragmatic initial subdivision of the stratigraphy into mapping units, such as in the Jurassic North Sea studies of Partington *et al.* (1993a). However, this subdivision into genetic stratigraphic 'sequences' *sensu* Galloway (1989) is not an end in itself. A complete understanding of palaeogeography and facies distribution is achieved only by subdivision into systems tracts, which requires identification of the sequence boundaries and transgressive surfaces as well as the maximum flooding surfaces.

2.5 High-resolution sequence stratigraphy and parasequences

2.5.1 Introduction

The concept of stratigraphic cycles driven by rises and falls in relative sea-level was developed using seismic data. This has a relatively coarse resolution of many tens to hundreds of metres, and seismic-based stratigraphy has been termed 'low-resolution sequence stratigraphy' by Posamentier and Weimer (1993). *High-resolution sequence stratigraphy* integrates observations at a log, core or outcrop scale (Chapter 4). These detailed data sets allow gross stratal geometries to be linked with the internal facies assemblages. In addition, high-resolution shallow seismic data over modern sedimentary systems has contributed much detail to the understanding of bedding geometries within sequences and systems tracts.

High-resolution sequence stratigraphy is used increasingly as a tool in hydrocarbon reservoir description (e.g. Posamentier and Chamberlain, 1992; Reynolds, 1994). The key introductory publication covering the theory and practice of high-resolution sequence stratigraphy is that of Van Wagoner *et al.* (1990).

2.5.2 Parasequences and their continental equivalents

Shallow marine sediments are commonly arranged into regular upward-coarsening units with an upward-shoaling facies succession, separated by much thinner units representing an upwards-deepening facies succession (Fig. 2.24, see also Fig. 8.4). Often the upward-deepening component is represented only by a hardground or omission surface marking a transition from shallower to significantly deeper water facies. In sequence stratigraphic terminology, these cycles are termed *parasequences*. Van Wagoner *et al.* (1990) defined parasequences as relatively conformable successions of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative surfaces. In special positions within the sequence, parasequences may be bounded either above or below by sequence boundaries.

The marine flooding surface in this definition is a surface separating younger from older strata across which there is evidence of an upward increase in water depth. This deepening is commonly accompanied by minor submarine erosion or non-deposition (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.

The recognition of parasequence boundaries, and their discrimination from sequence boundaries, was addressed by Van Wagoner *et al.* (1990), who suggest that shallow-marine parasequence boundaries are essentially flat, relatively condensed sections that represent abrupt deepening, and may be characterized by marine carbonate, phosphate or glauconite accumulations. The boundaries also mark abrupt changes in lithology and bed thickness, and are occasionally associated with lag deposits. Where lags do occur they are composed only of sediment reworked from below.

Parasequences form as a result of an oscillation in the balance between sediment supply and accommodation volume. Fluctuations in sediment supply due to autocyclic processes, such as avulsion and lobe switching, are probably the major control on parasequence formation. However, high-frequency variations in relative sea-level would produce high-order sequences that could look very similar to parasequences, especially if they were type 2 sequences.

Parasequences are defined by the marine flooding surface, and so cannot be defined in settings where changes in water depth are unrecorded. However, it is likely that parasequences have correlative equivalents in non-marine strata, such as fluvial avulsion cycles (although this has yet to be proved). Marine flooding surfaces could probably be correlated with coal beds on the coastal plain, and with widespread overbank mudstones and wet palaeosols on the alluvial plain. There are no criteria for recognizing para-

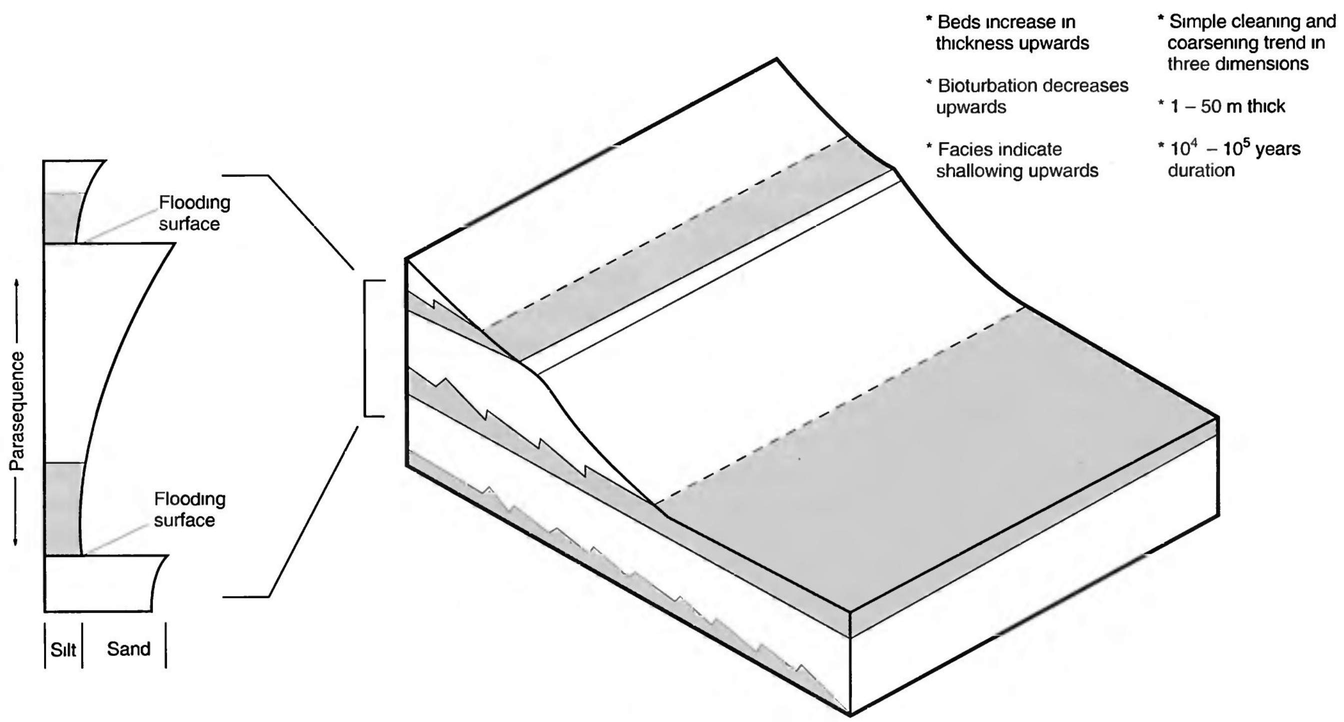


Fig. 2.24 An idealized parasequence. This represents a shallow-marine parasequence in a wave or storm-dominated setting, where upward coarsening can be related directly to upward shallowing

sequences in a deep marine setting. Parasequences will have correlative units on the clinoform slope representing the units of slope deposition fed by each delta lobe. Unless sediment is also bypassed to the basin floor, parasequences will have no equivalents in submarine fan facies. Mitchum and Van Wagoner (1991) speculated that individual fan lobes or leveed channels in the deep marine setting may reflect individual parasequences.

2.5.3 Parasequence sets

A parasequence set was defined by Van Wagoner *et al.* (1990) as a succession of genetically related parasequences forming a distinctive stacking pattern bounded by major marine-flooding surfaces and their correlative conformities (Fig. 2.25). In some cases one or both boundaries of a parasequence set will be a sequence boundary. Whilst parasequences may represent individual topset reflections within a systems tract on seismic data, parasequence sets often represent the entire topset component of that systems tract.

A stacking pattern refers here to the architecture of a vertical succession of parasequences. Progradational, retrogradational and aggradational stacking patterns can be recognized (Figs 2.25 and 2.26). In a progradational stacking pattern, the facies at the top of each parasequence become progressively more proximal higher in the succession. In a retrogradational stack the facies become more distal upwards, and in an aggradational stack the facies at the top of each parasequence is similar.

The topsets of the lowstand and highstand prograding wedges generally consist of a progradational parasequence set, whereas transgressive systems tracts consist entirely of a retrogradational parasequence set. The terms 'systems tract' and 'parasequence set' are not always synonymous (Posamentier and James, 1993), and in areas of high subsidence and sediment input, more than one parasequence set can exist in a systems tract. Parasequence sets are considered here as a class of depositional unit intermediate between parasequence and sequence, and the major marine flooding surfaces that bound parasequence sets may form subregional correlation markers.

2.5.4 Parasequence thickness trends

The thickness of a parasequence is controlled primarily by the water depth into which the shoreline progrades. This water depth represents the rise in relative sea-level since abandonment of the previous parasequence, and parasequence thickness is therefore a product of the rate of rise of relative sea-level and the periodicity of the parasequences.

If parasequence periodicity is relatively constant, then a slow rate of rise of relative sea-level results in thin parasequences, and a rapid rate of rise results in thick parasequences. Changes in the rate of relative sea-level rise should then be recognizable from trends in parasequence

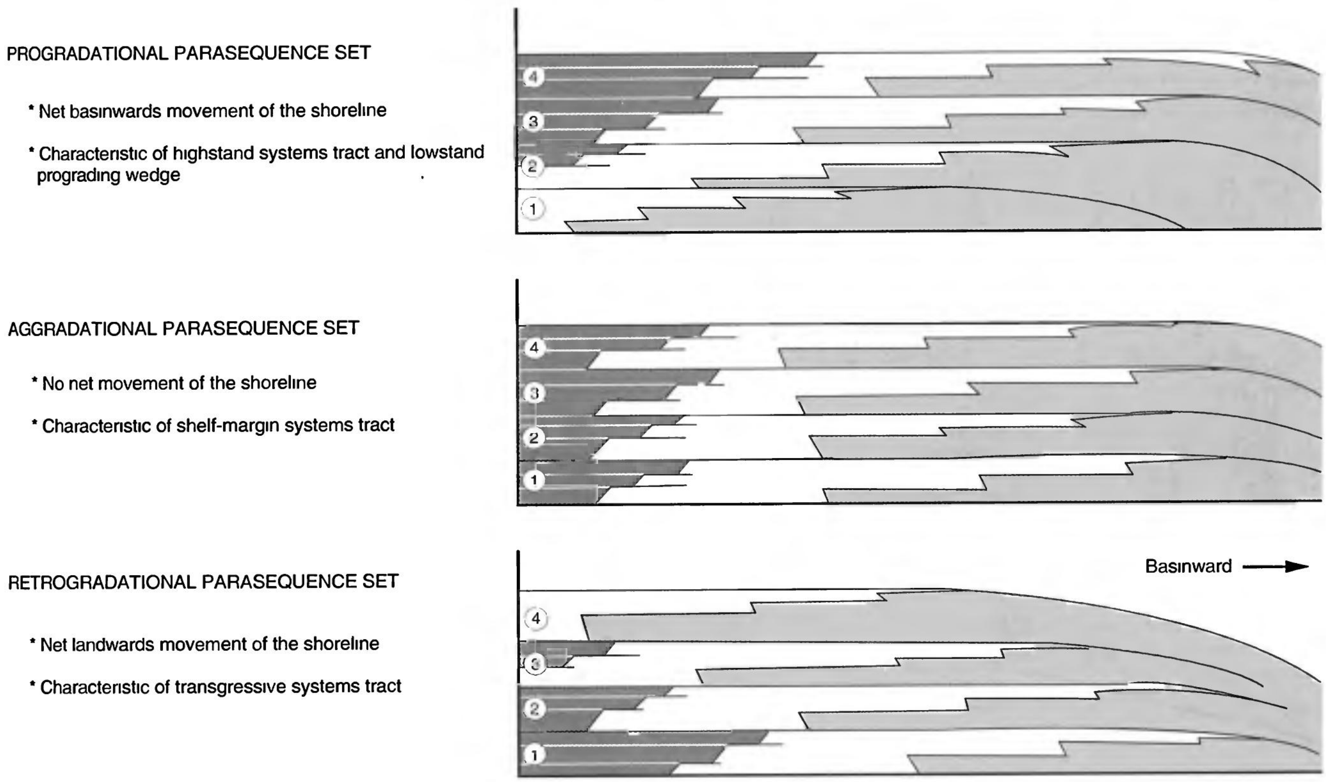


Fig. 2.25 Parasequence sets (after Van Wagoner *et al.*, 1988)

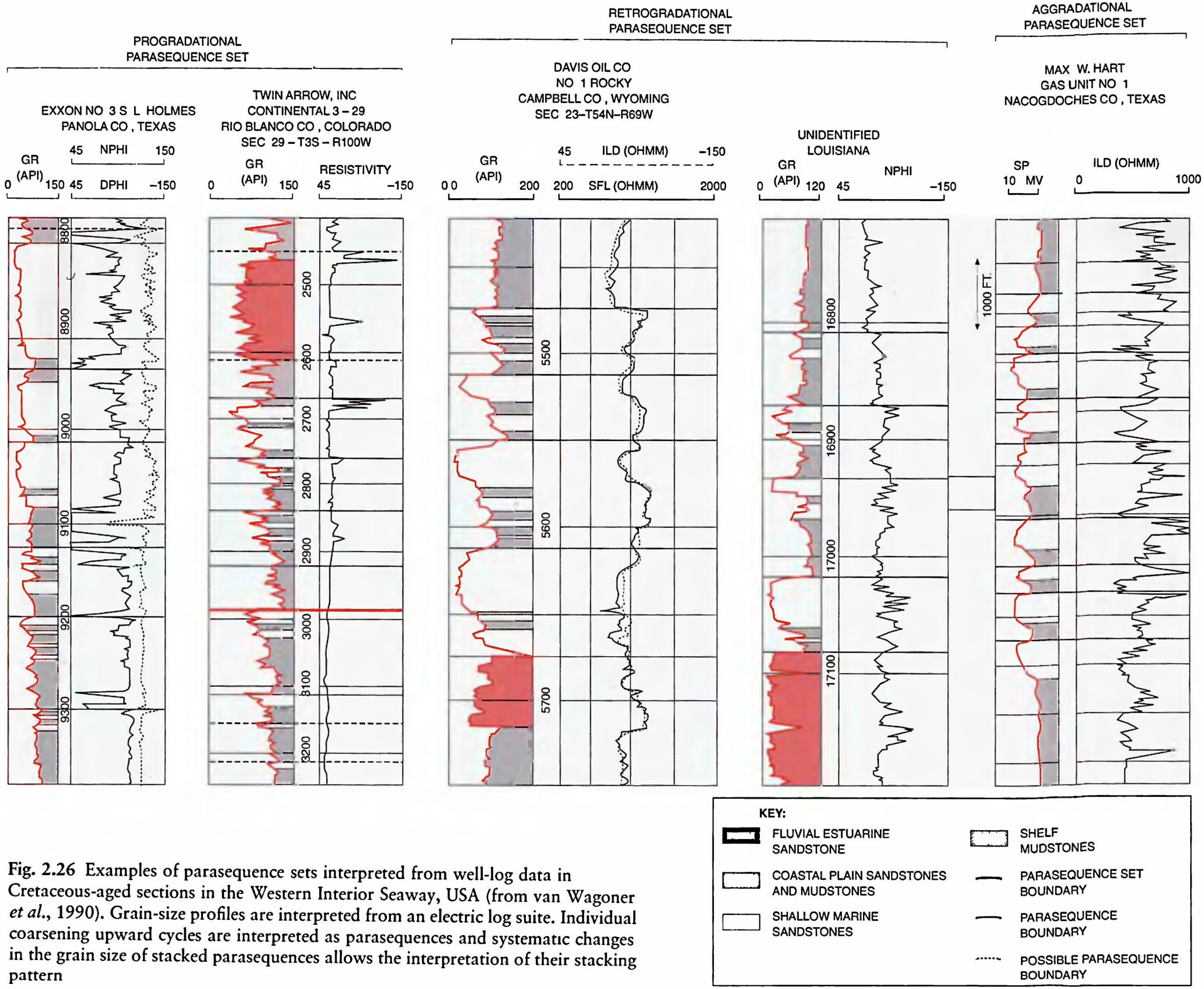


Fig. 2.26 Examples of parasequence sets interpreted from well-log data in Cretaceous-aged sections in the Western Interior Seaway, USA (from van Wagoner *et al.*, 1990). Grain-size profiles are interpreted from an electric log suite. Individual coarsening upward cycles are interpreted as parasequences and systematic changes in the grain size of stacked parasequences allows the interpretation of their stacking pattern

thickness. These ideas were documented by Posamentier *et al.* (1988), who suggested that a lowstand prograding wedge will be characterized by upwards-thickening parasequences (reflecting accelerating relative sea-level rise) whereas a highstand prograding wedge will be characterized by upward-thinning parasequences, due to decelerating relative sea-level rise.

This thickness analysis can be applied in limited circumstances. Retrogradational parasequence sets, for example, often show a thinning upward trend, due to basinward thinning of the individual parasequences. This is not related to decreasing rates of relative sea-level rise. Thickness trend analysis also assumes a constant parasequence frequency, which may not be a valid assumption in many cases.

2.5.5 Sequence boundaries

As described earlier, sequence boundaries can be recognized on seismic data from a downward shift in coastal onlap, implying a fall in relative sea-level, with exposure and erosion of the highstand topsets. In a core, well log or outcrop data set, the downward shift in coastal onlap is rarely evident. Direct evidence for exposure, erosion and forced regression must be sought instead (Fig. 2.27; see also section 8.3.1). A *facies dislocation* is a surface where rocks of a shallower facies rest directly on rocks of a significantly deeper facies. The trend of gradual shallowing predicted by Walther's law is thus 'dislocated'. This dislocation may be obvious, such as where a coal bed overlies an outer shelf mudstone, or it may subtle, such as an upper shoreface facies overlying lower shoreface with middle shoreface absent. In shallow marine settings the facies dislocation is often associated with an abrupt grain-size increase. A facies dislocation implies a fall in relative sea-level and the development of a subaerial unconformity, although this could be up-dip from where the dislocation is observed. It thus marks a sequence boundary or its correlative conformity. Facies dislocations commonly are developed over the more distal areas of the highstand topsets, and the highstand clinoforms.

Incised valleys are described by Van Wagoner *et al.* (1990) as entrenched fluvial systems that extend their channels basinward and erode into underlying strata in response to a fall in relative sea-level. On the shelf, lowstand deposits filling the incised valleys are bounded below by a sequence boundary and above by the transgressive surface. A facies dislocation may be present at the base of the incised valley, although a grid of wells, or an outcrop data set, may be needed to prove the existence of an incised valley.

Incised valleys are differentiated from distributary channels by being deeper and wider than the scale of an individual channel or channel belt. The valley incises below the level of the distributary mouth bars, and often contains a more proximal fluvial facies deposited as part of the aggradational late lowstand prograding wedge. Alternatively they may contain estuarine to marine facies deposited as part of the transgressive systems tract.

Between the incised valleys in the more proximal areas of the highstand topsets, sequence boundaries may be very hard to recognize. Any evidence of exposure, such as major palaeosols, reddening, or weathering will be fairly superficial and may be removed by subsequent transgressive erosion. These surfaces are known as E/T surfaces, for the superposition of erosion and transgression (Walker and Eyles, 1991). The only evidence for a sequence boundary may lie in the components of the transgressive lag, which could be significantly coarser grained than the underlying succession, or may contain material that is not derived solely from the underlying succession.

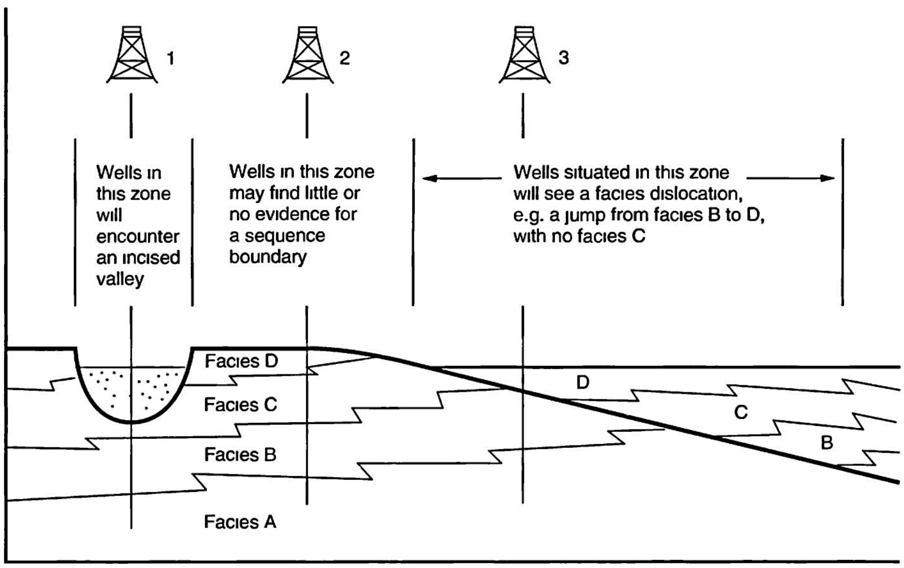


Fig. 2.27 Lines of evidence for the presence of a sequence boundary. These include a downward shift in coastal onlap (recognizable on seismic data and from correlation of a grid of wells), a facies dislocation recognized in well 3, and valley incision recognized in well 1. Well 2 may find little or no evidence for the sequence boundary

In rare cases a sequence boundary may be recognized by the truncation of underlying parasequences (e.g. fig. 24 in Van Wagoner *et al.*, 1990). Care must be taken, however, to ensure that the limits of the parasequences are erosional, not depositional.

2.5.6 Maximum flooding surfaces

In well log, core or outcrop data sets, maximum flooding surfaces are recognized as the boundary between a transgressive unit, or retrogradational parasequence set, and an overlying regressive unit, or progradational parasequence set (Fig. 2.28). In a proximal direction the maximum flooding surface may lie within an aggradational parasequence stack, and it passes into a shelfal and basinal condensed section in a distal direction. The condensed section may be represented by a distinctive log facies or lithofacies, such as a glauconitic horizon, chert band, limestone band or high-radioactivity, low-velocity shale. The distinctive character of maximum flooding surfaces, and the widespread development in the basin of the equivalent condensed interval, makes them the easiest of the sequence stratigraphic surfaces to identify (Loutit *et al.*, 1988). They are equivalent to the bounding hiatal surfaces that define the genetic stratigraphic units of Galloway (1989).

It should be noted that other condensed sections may form within a sequence, which are not equivalent to the maximum flooding surface; such as the boundary between the basin-floor fans and the slope fans, the boundary between the slope fans and the lowstand prograding wedge, and surfaces of major avulsion within a systems tract.

2.5.7 Ravinement surfaces

A *ravinement surface* is a surface of transgressive erosion. Swift (1968) described transgressive intervals in cratonic basins as commonly appearing to rest disconformably on underlying strata. The underlying strata sometimes may be pre-existing deposits of earlier cycles, but often are the marginal marine deposits of the contemporaneous cycle. The significance of such disconformities was first noted by Stamp (1921), who showed that the surf zone of a transgressing sea may bevel the marginal deposits of the coast being transgressed. Stamp called the resultant disconformity a ravinement.

The most widely accepted mechanism of landward beach and barrier migration is termed shoreface retreat, in which, as sea-level rises, sediment is eroded from the upper shoreface and emplaced in the lower-shoreface–offshore area as storm generated beds, or in the lagoon as a series of washover fans (Bruun, 1962; and Fig. 2.29). As the upper shoreface or breaker zone passes across the former barrier it erodes the lagoonal and washover facies deposited during the earlier stages of transgression and the lower shoreface facies therefore overlies a planar erosion surface. This erosion surface is referred to as the shoreface erosion plane, or the ravinement surface (Stamp, 1921).

The extent of erosion at the shoreface depends on the rate of rise of relative sea-level. In areas of rapid subsidence and/or relatively rapid sea-level rise, a comparatively complete transgressive unit may be preserved, whereas with slow subsidence and/or sea-level rise shoreface erosion is more pronounced and the transgressive unit is attenuated (Fischer, 1961).

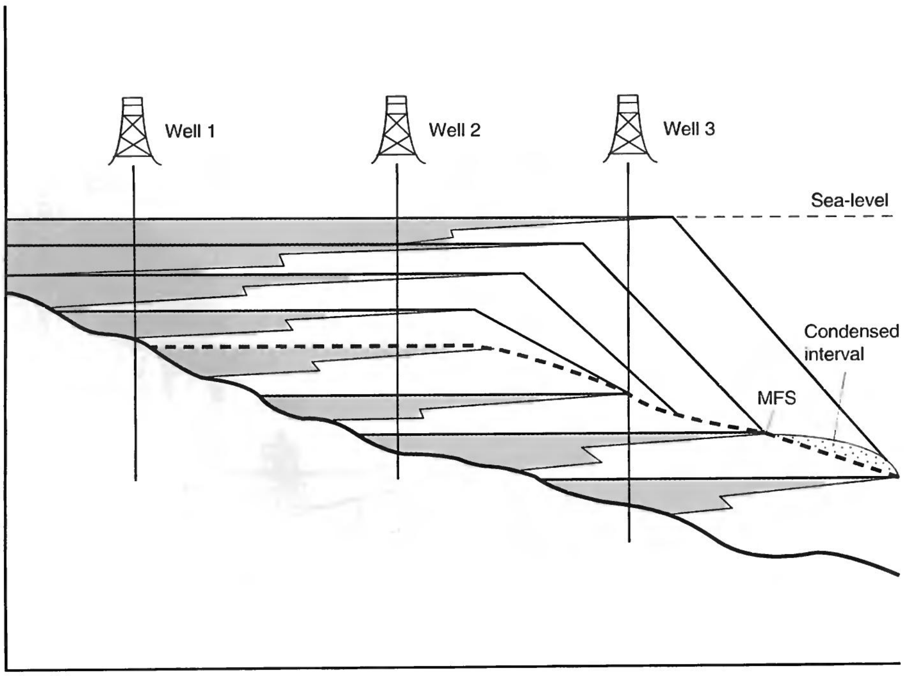


Fig. 2.28 Representation of a maximum flooding surface (MFS) in well data. In proximal wells the maximum flooding surface lies within aggradational topset facies and may be difficult to distinguish from a simple parasequence boundary. In distal wells the maximum flooding surface clearly overlies a retrogradation parasequence set

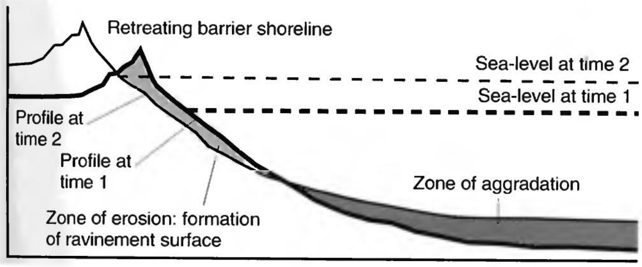


Fig. 2.29 Formation of a ravinement surface by transgressive erosion during shoreface retreat

The ravinement surface acts like a facies belt moving in parallel with other coastal facies belts during transgression. It results in a marine flooding surface; a rapid progression upwards from a supralittoral facies to a sublittoral facies with no intervening preservation of intermediate facies. This may form a striking surface in log, core or outcrop data, but is in detail diachronous and forms within a single episode of transgression. Ravinement surfaces therefore may bound every parasequence in a parasequence set. However, a more major ravinement surface may form at the transgressive surface (the boundary between the lowstand prograding wedge and the overlying transgressive systems tract). Other major ravinement surfaces may bound parasequence sets within a single systems tract.

2.5.8 Problems and pitfalls of high-resolution sequence stratigraphy

High-resolution sequence stratigraphy, performed on a subsurface data set, is not easy. The major problems can be summarized as follows:

1 Recognition of parasequences, and indeed of the depositional setting of the interval being studied, is tenuous without core control, good biostratigraphic control, or seismic indicators of basin setting (e.g. confirmation that it is a topset interval).

2 Correlation of parasequences may not be straightforward. One parasequence may look very much like another. Correlation will be easiest with closely spaced wells, or where some parasequences have a diagnostic log shape, or a marker lithology such as a prominent coal bed.

3 Recognition of sequence boundaries is not easy in the areas between the incised valleys and is not easy generally within a succession of parasequences.

4 Differentiating incised valleys from non-incised channel deposits can be very tricky. Much argument often focuses around the significance of anomalously large channel units. Van Wagoner *et al.* (1990) give several guidelines.

5 Systems tract boundaries can be recognized as surfaces where parasequence correlation lines terminate. Parasequences either onlap these surfaces, are truncated below them, or pinch out depositionally below them (in the case of the maximum flooding surface). It can be difficult to know which form of termination you are dealing with, and therefore the nature of the surface involved.

6 In an outcrop data set, high-resolution sequence stratigraphy is generally easier. There is abundant facies information, and surfaces can be traced laterally with relative ease. Differentiating incised from non-incised channels may still be difficult, although ideally the base of an incised valley could be traced laterally into an exposure/erosion surface. Outcrops generally are not continuous and correlating across gaps is perhaps the major problem. There will be no seismic data to allow the outcrop information to be placed in the context of stratal geometry, although sometimes large cliff sections give geometric information at a seismic scale (e.g. Bosellini, 1984, Fig. 4.5).