# CHAPTER 6 Deltas

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## 6.1 INTRODUCTION

Deltas are discrete shoreline protuberances formed where rivers enter oceans, semi-enclosed seas, lakes or lagoons and supply sediment more rapidly than it can be redistributed by basinal processes. Generally, deltas are served by well-defined drainage systems which culminate in a trunk stream and supply sediment to a restricted area of the shoreline, thus resulting in the formation of a depocentre. Less organized and immature drainage systems produce numerous closely spaced rivers which induce uniform progradation of the entire coastal plain rather than point-concentrated progradation and depocentres. At the river mouth, sediment-laden fluvial currents which have previously been confined between channel banks suddenly expand and decelerate on entering the receiving water body. As a result the sediment load is dispersed and deposited, with coarse-grained bedload accumulating near the river mouth, whilst finer grained, suspended sediment is transported offshore and deposited in deeper water. Basinal processes such as waves, tides and oceanic currents may assist in the dispersal of sediment and also rework sediment deposited by the fluvial currents. Many of the characteristics of deltas stem from the results of this interplay between fluvial and basinal processes. The tectonic settings of major deltas are varied (Audley-Charles, Curray and Evans, 1977), but the most common are passive continental margins (e.g. the Niger delta), marginal basins or foreland basins associated with mountain belts (the Irrawaddy and Ganges-Brahmaputra deltas) and cratonic basins (the Rhine delta).

As deltas are often major depocentres, they produce exceptionally thick sequences which have been recognized throughout the geological column. In addition to their palaeogeographic significance, ancient deltaic successions are also important as sites of oil, gas, and coal reserves in many parts of the world.

## 6.2 DEVELOPMENT OF DELTA STUDIES

Rigorous sedimentological studies of modern deltas commenced with Johnston's (1921, 1922) account of the Fraser River delta and classic work on the Mississippi delta (Trowbridge, 1930; Russell, 1936; Russell and Russell, 1939; Fisk, 1944, 1947). The exemplary nature of early research on the Mississippi delta caused it to be regarded as *the* delta model for a period, and this position was fortified by later work which provided further insight into this delta (Fisk, 1955, 1960; Coleman and Gagliano, 1964; Coleman, Gagliano and Webb, 1964). Other deltas were described at this time, with an emphasis on demonstrating similarities between deltas. However, van Andel and Curray (1960) recognized the need for a critical comparison of modern deltas. Whilst noting basic similarities between deltas, they stressed that 'the striking variation in structure and lithology . . . as exemplified by the Rhône and Mississippi deltas, should not be underestimated. A study of comparative morphology and lithology of modern deltas appears highly desirable'. Subsequent publications on individual deltas concluded with a comparison of the described delta with other examples (e.g. Allen, 1965d; Van Andel, 1967), thus consolidating the trend towards the variability of deltas which has recently dominated studies of modern deltas (Fisher, Brown *et al.*, 1969; Wright and Coleman, 1973; Coleman and Wright, 1975; Galloway, 1975).

Studies of deltaic facies commenced in ancient successions rather than modern deltas with Gilbert's (1885, 1890) descriptions of Pleistocene deltaic facies in Lake Bonneville. Glacial streams transporting coarse sediment produced a series of fan-shaped lacustrine deltas exposed by subsequent lake-level changes and channel dissection. The deltas have a three-fold structure which generated a distinctive vertical sequence of bedding types during progradation (Fig. 6.1). Barrell (1912, 1914) subsequently proposed criteria for the recognition of ancient deltaic deposits based on Gilbert's descriptions and later applied these criteria to the Devonian Catskill Formation. The terms topset, foreset and bottomset were used to describe the structure of the delta, and the bedding, texture, colour and fauna of each component were discussed, thus initiating the facies approach in deltaic deposits. Although Barrell stressed that not all deltas exhibit this Gilbert-type structure, the concept conditioned thinking on modern deltas for several decades, and the presence or absence of large-scale inclined foresets was considered an important criterion in the recognition of ancient deltaic successions. Barrell also referred to a deltaic cycle of sedimentation, but at this time the cycle was strictly Davisian in relating to the physiographic 'age' of the hinterland. High rates

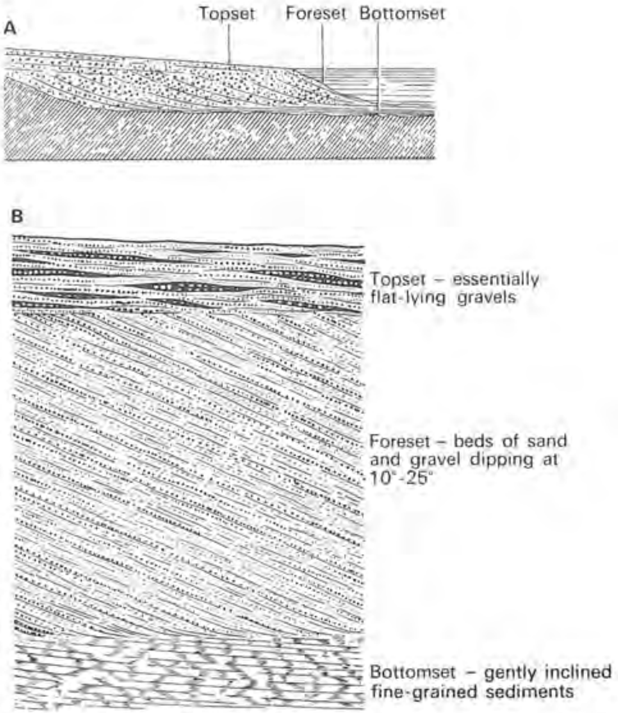



**Fig. 6.1.** (A) Section through a 'Gilbert-type' Pleistocene delta in Lake Bonneville; (B) vertical facies sequence produced by delta progradation (after Gilbert, 1885; Barrell, 1912).

of sediment supply associated with a 'youthful' hinterland resulted in delta progradation, but as the hinterland matured and passed into 'old age' reduced sediment supply resulted in marine planation of the delta.

In the USA during the late 1940's, outcrop and subsurface information began to be interpreted in terms of palaeo-environments, and it was gradually realized that significant amounts of coal, oil and gas were located in ancient deltaic systems. As these studies were concerned with locating and tracing sandstone bodies, they tended to concentrate on lateral facies relationships within well-defined stratigraphic intervals, thus permitting a deltaic interpretation to be offered with some degree of confidence (e.g. Pepper, de Witt and Demarest, 1954). A parallel development in the USA and Europe was the recognition of coarsening-upwards cycles or cyclothems which reflected a passage from marine facies upwards into terrestrial facies and were often attributed to delta progradation. Although this approach focussed attention on the vertical arrangement of facies it was not entirely beneficial to the development of delta studies. In many Carboniferous examples, where the approach was most eagerly applied, debates on the definition and genesis of the cycles often took precedence over analysis of the rocks. Controversy raged over the horizon at which cycles commenced and pleas were issued for more objective definitions of the cycles using statistical techniques. Discussions on the genesis of cycles were often focussed on an idealized cycle rather than actual successions, and a variety of tectonic, climatic or sedimentological controls was invoked to explain these idealized cycles. During the reign of this approach the sedimentary facies and their relationships were often neglected, with a tendency to view the succession as 'rocks that occurred rather than . . . sedimentary processes which happened' (Reading, 1971, p. 1410). In many cases, it is only recently that the facies characteristics of these cycles have been scrutinized.

The economic importance of deltaic facies stimulated extensive borehole programmes sponsored by oil companies in the Mississippi, Rhône and Niger deltas (Fisk, McFarlan *et al.*, 1954; Fisk, 1955, 1961; Oomkens, 1967, 1974; Weber, 1971). These studies demonstrated the wide variety of vertical facies sequences in deltaic successions, with the type of sequence varying not only *between* deltas but also at different locations *within* a delta. Current attitudes towards ancient deltaic successions stem largely from these studies. Facies-sequences are studied rather than idealized cycles, and ancient deltaic successions are discussed in terms of different types of deltas, in harmony with the current emphasis on the variability of modern deltas.

## 6.3 A CONCEPTUAL FRAMEWORK FOR DELTAS

In order to comprehend the variability of modern and ancient deltas, a framework is required which summarizes interactions between variables which control the development of deltas and defines causal relationships between these variables. The framework adopted in this chapter regards delta regime as a general expression of the overall setting and relates the regime of the delta to its morphology and facies pattern (Fig. 6.2).

The variables affecting deltas stem from the characteristics of the hinterland and receiving basin. Since the hinterland supplies sediment, hinterland characteristics are largely reflected in the fluvial regime and the transported sediment load. The most important feature of the receiving basin is the energy regime which contests the introduction of river-borne sediment. The basinal regime depends on several major features of the basin such as shape, size, bathymetry and climatic setting and reflects these features. Interaction between the sediment-laden fluvial waters and basin processes at the river mouth defines the delta regime which dictates the dispersal and eventual deposition of sediment in the delta area, and therefore is the focal point of this framework. An important feature which emerged from comparisons between modern deltas is a relationship betweeen delta regime and morphology. Initially, the classic birdfoot-lobate-arcuate-cuspate spectrum of delta types was related to increasing wave influence over fluvial processes (Bernard, 1965), and

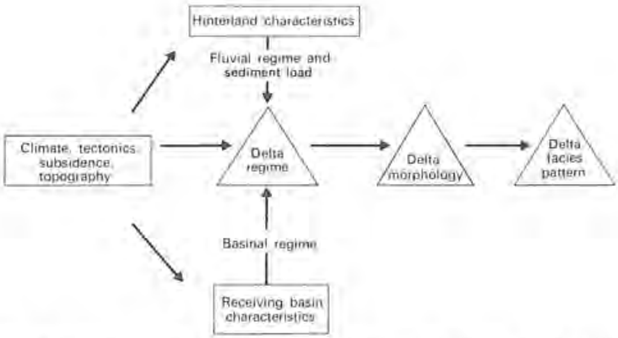



**Fig. 6.2.** Conceptual framework for the comparative study of deltas, applicable to modern deltas and ancient deltaic successions.

more recently this relationship has been extended by using data from a wide range of modern deltas, including those significantly influenced by tidal processes (Fisher, Brown *et al.*, 1969; Wright and Coleman, 1973; Coleman and Wright, 1975). The final link between delta morphology and facies patterns derives from drilling in several modern deltas which reveals that contrasts in overall facies patterns are related to differences in the regime and morphology of the deltas. Causative links therefore exist between delta regime, morphology and facies pattern.

This framework can also be applied to ancient deltaic successions, where studies are based on a partial record of the facies pattern gained from measured sections, subsurface cores or electrical logs, often widely scattered. Recognition of subenvironments and the processes which operated in them permits the nature of the delta to be reconstructed. In thick basinal successions comprising a series of delta complexes, it may be possible to detect temporal variations in delta type, reflecting evolution of the hinterland and/or receiving basin (Belt, 1975; Galloway, 1975; Whitbread and Kelling, 1982). Also, as the nature of the delta is inferred from a partial record of the facies pattern, predictions can be made on the remainder of the facies pattern in unexposed or unexplored areas. This is important in the exploration and development of hydrocarbons located in deltaic sandstone bodies as the postulated delta type provides a predictive model which can be tested by subsequent investigations.

### 6.3.1 Hinterland and receiving basin characteristics

The *hinterland* comprises the drainage basin and fluvial system where variables such as relief, geology, climate and tectonic behaviour interact to determine the fluvial regime and sediment supply which feed the delta (see Chap. 3). With regard to deltas, important features include the following.

(1) The total amount of sediment supplied in relation to the reworking ability of the basinal processes.

(2) The calibre of sediment supply which influences the dispersion and deposition of sediment in the delta. Coarse-grained bedload sediment tends to be deposited in the immediate vicinity of the distributary mouth and either forms distributary mouth bars, or is reworked by wave and tidal processes into beach-barrier systems or tidal current ridge complexes. In contrast, finer grained suspended load sediment is generally transported offshore and dispersed with the aid of basinal processes over a wide area of the basin. Deposition produces an extensive mud-dominated platform in front of the delta which may be over-ridden by delta front sands as progradation continues, resulting in extensive synsedimentary deformation of the succession (Sect. 6.8). The ratio of bedload to suspended load sediment is therefore an important control on deltaic sedimentation, and changes in this ratio can radically alter the characteristics of a delta and its facies pattern.

(3) Fluctuations in discharge can be significant in determining the calibre of sediment supply. For example, rivers with erratic or 'flashy' regimes characterized by brief, episodic high discharge periods are more likely to supply coarse sediment to the delta than more stable regimes which tend to sort sediment prior to its reaching the delta.

(4) The timing of fluctuations in fluvial discharge relative to fluctuations in the basin energy regime also influences deposition in the delta area. If the maxima are in phase, basinal processes continually redistribute the river-borne sediment, but, if the maxima are out of phase, periods of virtually uncontested delta progradation alternate with periods of reworking by basinal processes (Wright and Coleman, 1973).

(5) Tectonic events can also dictate sediment supply from the hinterland. For example, in the Gulf of Mexico, Tertiary tectonic events in the Rocky Mountains caused periodic, large-scale reorganization of the drainage basins supplying streams bound for the Gulf Coast. These changes produced a series of major depositional centres of differing age in the Gulf Coast, each with a distinctive bulk petrography and perhaps a unique assemblage of delta types (Winker, 1982).

Characteristics of the *receiving basin* which influence the development of deltas include water depth and salinity, the shape, size, bathymetry, and energy regime of the basin, and overall basin behaviour in terms of subsidence rates, tectonic activity and sea-level fluctuations.

The relative density of river and basin waters is an important first-order control on the manner in which the sediment-laden river discharge is dispersed in the basin, and this is partly a function of the salinity of the basin waters (Bates, 1953). Where rivers enter freshwater basins, there is either immediate mixing of the water bodies at the river mouth or the river discharge flows beneath the basin waters as a density current. In contrast, where rivers enter a saline basin, the river discharge may extend into the basin as a buoyantly supported plume due to the higher density of seawater (Sect. 6.5.2).

The basinal regime includes the effects of wave and wave-induced processes, tidal processes, and to a lesser extent semi-permanent currents, oceanic currents and wind effects which may temporarily raise or lower sea-level. The type of basin is a prime control on the nature of the basinal regime. For example, at ocean-facing continental margins, the full range of basinal processes affects the deltas (e.g. the Niger delta, Allen, 1965d), whereas in semi-enclosed and enclosed seas, wave energy is limited due to reduced fetch, and tidal influence is minimal (e.g. the Danube, Ebro, Mississippi, Po and Rhône deltas). Deltas located in narrow elongate basins or gulfs connected to an ocean experience considerable tidal effects as tidal currents are amplified and may therefore transport considerable volumes of sediment (e.g. the Ganges-Brahmaputra delta). Smaller scale deltas prograding into lagoons or lakes are commonly dominated by fluvial processes as the influence of basinal processes is limited (Donaldson, Martin and Kanes, 1970; Kanes, 1970). Basin water depth and the presence or absence of a shelf-slope influence the basinal regime, particularly in terms of the extent of wave attenuation and tidal current amplification.

Finally, as deltas are topographically subdued areas at the margins of basins, they are extremely sensitive to subsidence trends, sea-level fluctuations and basin tectonics. Delta sites may be affected by basement-related tectonics, as in the Ganges-Brahmaputra delta which is located in a downwarped basin with numerous active normal faults, and the Tertiary Niger delta which developed in a triple junction rift system (Morgan and McIntire, 1959; Coleman, 1969; Morgan, 1970; Burke, Dessauvagie and Whiteman, 1971). They may also be affected by sediment-induced or 'substrate' tectonics involving overpressured shales which induce deep-seated lateral clay flowage, diapirism and faulting, as in the Mississippi and Niger deltas (Weber, 1971; Coleman, Suhayda *et al.*, 1974; Weber and Daukoru, 1975; see Sect. 6.8).

## 6.4 DELTA MODELS

In view of the variability of modern deltas, a single delta model is no longer adequate. Instead a series of models is required and several schemes have been proposed, based primarily on the physical processes operative within the delta (Fisher, Brown *et al.*, 1969; Coleman and Wright, 1975; Galloway, 1975).

Fisher, Brown *et al.* (1969) distinguished *high-constructive* deltas dominated by fluvial processes from *high-destructive* deltas dominated by basinal processes. Lobate and birdfoot types were recognized in the high-constructive class, and wave-dominated and tide-dominated in the high-destructive class (Fig. 6.3). Each type has a characteristic morphology and facies pattern, described in terms of vertical sequences, areal facies distribution and sand body geometry. Because facies relationships are stressed the classification can be applied to ancient deltaic successions, but one disadvantage is that it concentrates on end-members of what is in reality a continuous spectrum. In addition, use of the term 'high-destructive' is misleading in this context since all deltas are by definition constructive whilst active, and the term therefore confuses a class of deltas with the destructive or abandonment phase of delta history which follows channel switching and delta abandonment (Scruton, 1960; Sect. 6.6).

An alternative scheme involves analysis of statistical information from thirty-four present-day deltas using a wide range of parameters to illustrate the characteristics of the drainage basin, alluvial valley, delta plain and receiving basin (Coleman and Wright, 1975). Interaction of the variables defines a process setting which is unique to any delta, but multivariate analysis of this information produces six discrete delta models each illustrated by a sand distribution pattern. The models are also described in terms of processes and morphology using representative modern deltas, and facies patterns are summarized by single, idealized vertical sequences. This scheme has an extremely broad data base in terms of the number of samples and the number of parameters, and an additional advantage is that initial description of the models is devoid of specific connotations associated with individual deltas. It is therefore an attractive scheme, but a major weakness is that idealized vertical sequences cannot summarize deltaic facies patterns which are ubiquitously characterized by extreme vertical and lateral variations.

The scheme adopted in this chapter is a modified version of a scheme proposed by Galloway (1975) which uses a ternary diagram to define general fields of fluvial-, wave- and tide-dominated deltas (Fig. 6.4). At present the positions of individual deltas are plotted qualitatively but a quantitative positioning may eventually be possible.

Process-based classifications seem to provide the most valid means of summarizing the variability of deltas, but several cautionary comments are necessary.

(1) The regime of the delta front is used to define delta type as the regime of the delta plain is often different. For example, the delta front of the Rhône delta is wave-dominated, whilst the delta plain is largely fluvial-dominated by virtue of being sheltered from wave action by shoreline beach-barriers which enclose the delta plain. Also, in areas of moderate to high tidal range, the upper delta plain is fluvial-dominated, the lower delta plain may be tide-dominated, and the delta front may be influenced by tide and wave processes (e.g. the Niger delta).

(2) Factors other than physical regime are often important in the formation and nature of deltas. For example, in the fluvial-dominated modern Mississippi delta, the fine-grained sediment load and deep-water, shelf-edge position of the delta are important in determining the characteristics of this delta.

(3) As deltas prograde, they may evolve through a series of different types as the regime, sediment load, climate or basin configuration change. Thus, the appearance of a modern delta may not provide a reliable model for interpreting the earlier

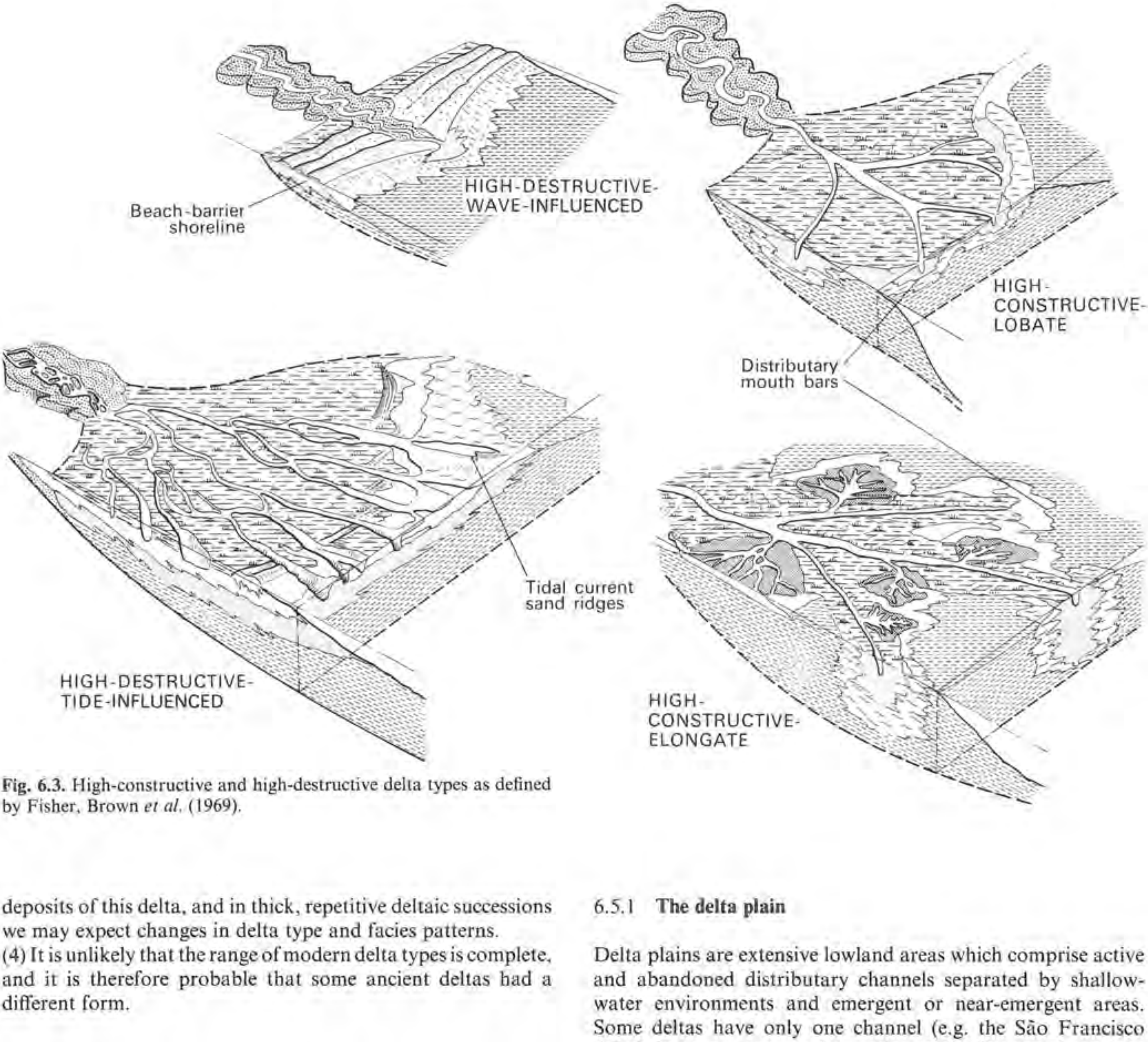



**Fig. 6.3.** High-constructive and high-destructive delta types as defined by Fisher, Brown *et al.* (1969).

deposits of this delta, and in thick, repetitive deltaic successions we may expect changes in delta type and facies patterns.

(4) It is unlikely that the range of modern delta types is complete, and it is therefore probable that some ancient deltas had a different form.

## 6.5 FACIES ASSOCIATIONS IN MODERN DELTAS

Deltas comprise two basic components: the *delta front* which includes the shoreline and seaward-dipping profile which extends offshore and the low-lying *delta plain* behind the delta front. As these components are often characterized by different regimes within, as well as between deltas, they are described separately.

### 6.5.1 The delta plain

Delta plains are extensive lowland areas which comprise active and abandoned distributary channels separated by shallow-water environments and emergent or near-emergent areas. Some deltas have only one channel (e.g. the São Francisco delta), but more commonly a series of distributary channels is spread across the delta plain, often diverging from the overall downslope direction by 60° or more. Between the channels is a varied assemblage of bays, floodplains, lakes, tidal flats, marshes, swamps and salinas which are extremely sensitive to climate. For example, in tropical settings, luxuriant vegetation prevails over large areas of the delta plain as saline mangrove swamps, freshwater swamps or marshes (the Niger, Klang-Langat and Mississippi deltas). In contrast, delta plains in arid and semi-arid areas tend to be devoid of vegetation and are


**Fig. 6.4.** Ternary diagram of delta types, based on the regime of the delta front area (modified after Galloway, 1975).

characterized by calcretes (the Ebro delta) or salinas with gypsum and halite (the Nile delta). Alternatively, arid delta plains are dominated by aeolian dune fields, particularly in sandy, wave-influenced deltas where sand is eroded from active and abandoned beach ridges (the São Francisco delta). Pingos, patterned ground and other cryogenic features occur in the delta plains of high latitude polar deltas, and tundra vegetation accumulates in shallow thaw ponds (the Mackenzie, Colville and Yukon deltas).

Most delta plains are affected by fluvial or tidal processes but only rarely by major waves as wave-influenced deltas are characterized by beach-barrier shorelines which enclose and protect the delta plain. Locally generated wind-waves may, however, operate in delta plain bays with a shallow water cover.

### FLUVIAL-DOMINATED DELTA PLAINS

Fluvial-dominated delta plains either are enclosed by beach ridges at the seaward end (e.g. the Rhône and Ebro deltas), pass downstream into a tide-dominated delta plain (e.g. the Niger, Mahakam and Mekong deltas) or are open at the seaward end and pass directly into the delta front (the Mississippi delta).

*Fluvial distributary channels* are characterized by unidirectional flow with periodic stage fluctuations, and are therefore similar to channels in strictly alluvial systems (Chap. 3). High sinuosity patterns are common, but in certain arid or polar deltas with sporadic discharge and a high proportion of bedload the distributary channels are braided and anastomosing. Distributary channels in the Mississippi delta have a low sinuosity pattern and are not braided, even at low stage. Contrasts with alluvial channels include: (a) the lower reaches of the distributary channels are influenced by basinal processes, even in low energy basins. For example, in the Mississippi delta, flood tides and waves associated with strong onshore winds impound distributary discharge during low and normal river stages. Bedload transport is inhibited and fine-grained sediment may be deposited in the channel (Wright and Coleman, 1973, 1974). This sediment may be eroded during the next river flood, but some may persist to form drapes in the channel sequence. (b) Switching or avulsion of channels is more frequent in distributary channels because shorter and steeper courses are created as the delta progrades into the basin. During and after channel abandonment, basinal processes become more effective in the lower reaches of the former channels. In the Rhône and Ebro deltas, for example, abandoned channel mouths are sealed by wave-deposited beach sands (Kruit, 1955; Maldonado, 1975). (c) Multiple-channel distributary systems rarely, if ever, divide the discharge equally and channels of different magnitude co-exist in the delta plain and also wax and wane in response to avulsion and abandonment.

Facies and sequences of distributary channels resemble those of alluvial channels to a large extent. Cores in the Rhône and Niger deltas reveal erosive-based sequences with a basal lag, followed by a passage from trough cross-bedded sands upwards into ripple-laminated finer sands with silt and clay alternations, and finally into silts and clays pervaded by rootlets (Fig. 6.5). Some are composite or multi-storey sequences which either reflect repeated cut-and-fill within the channel, or minor fluctuations in channel location (Oomkens, 1967, 1974). The overall fining-upwards results either from lateral migration of the channel, or more commonly, from channel abandonment, with the upper fine member representing infilling of the channel by diminishing flow and perhaps later by overbank flooding from an adjacent active channel. In the lower reaches of the channels the introduction of sand by basinal processes during channel abandonment may suppress the fining-upwards trend. Well-sorted, evenly-laminated sands with a marine fauna may terminate the channel sequence. In the Mississippi delta, Coleman (1981) considers that the fine-grained nature of the sediment load causes the distributary channel sequences to be dominated by clays and silts deposited principally during abandonment of the channel. Sequences comprise a thin, basal

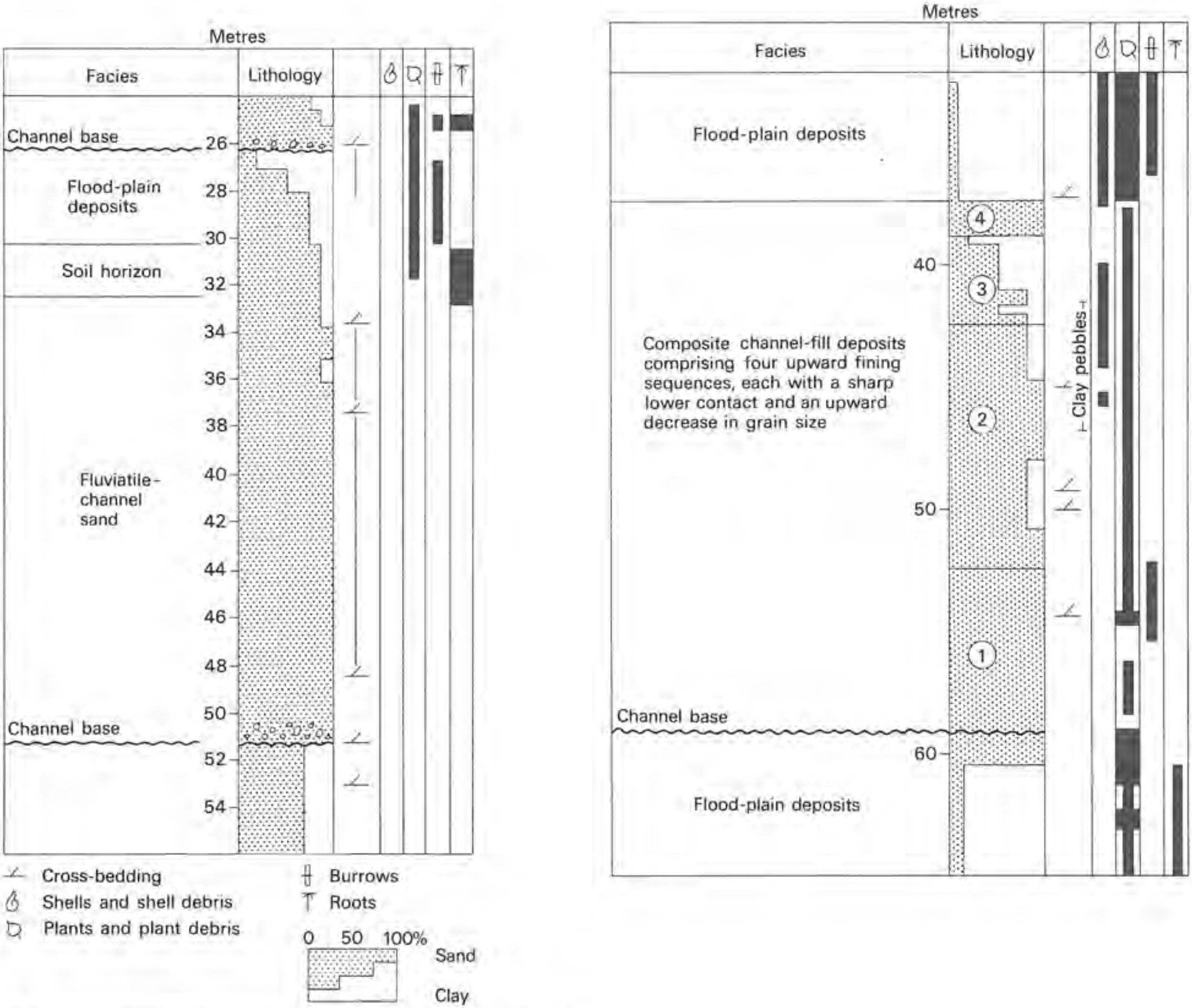



**Fig. 6.5.** Fluvial-distributary channel sequences from the Niger delta (after Oomkens, 1974).

unit of poorly sorted sands and silts which passes upwards into finer grained, bioturbated silts and clays. Plant debris is abundant throughout the sequences, and beds of *in situ* and derived peat are common in the upper parts of sequences.

Large-scale channel bank slumping can be an important feature of distributary channels as they often have fine-grained, cohesive bank materials. Scour during high river stage over-steepens the banks, inducing failure of the wetted sediments along rotational slump planes during low river stage (see Fig. 3.20). Often the entire bank is slumped, and if the basal shear plane extends beneath the base of the channel, slumped sediments may be preserved below the channel facies (Stanley, Krinitzsky and Compton, 1966; Turnbull, Krinitzsky and Weaver, 1966; Laury, 1971).

The switching or avulsive behaviour of fluvial distributary channels causes them to be short-lived channels relative to upstream alluvial equivalents. Sand bodies of the distributaries often have a lower width to depth ratio as a result. For example, in the Rhine and Rhône deltas the ratio decreases from 1000 for the alluvial channels to 50 for distributary channels near the shoreline (Oomkens, 1974). This trend is also apparent in the Mississippi delta, although in this case it is mainly related to a downstream change from freely migrating, high sinuosity channels to rather fixed, low sinuosity channels incised into earlier mud deposits.

*Interdistributary areas* of fluvial-dominated delta plains are generally enclosed, shallow water environments which are quiet or even stagnant, although locally generated wind-waves may induce mild agitation and produce isolated ripple form sets and lenticular laminae (Coleman and Gagliano, 1965). This generally placid regime is frequently interrupted during flood periods as excess discharge is diverted from distributary channels into the bays. Flood-generated processes are the principal means of sediment supply to the interdistributary areas and features which result from these processes include levees, various types of crevasse lobes, and crevasse channels. These features collectively fill large areas of the shallow bays and provide a platform for vegetation growth, gypsum and halite precipitation, or calcrete development, depending on the prevailing climate. The interdistributary areas therefore accumulate a wide range of facies and sequences which reflect infilling by a variety of flood-generated processes (Elliott, 1974b; Fig. 6.29).

(1) *Overbank flooding*: involving sheet-flow of sediment-laden waters over the channel banks. Fine-grained, laminated sediment is deposited over the entire area, although frequently the laminations are destroyed by subsequent bioturbation. Coarser sediment is confined to the channel margins and contributes to the growth of levees. As a result, levee facies comprise repeated alternations of thin, erosive-based sand beds representing sediment-laden flood incursions and silt-mud beds deposited from suspension. The sands may be parallel- or current-ripple-laminated, whilst the finer sediments are frequently affected by rootlets, indicating repeated emergence or near emergence. Levee facies fine away from the channels, and lateral encroachment of the levees associated with channel migration or alluvial ridge build-up may therefore produce a coarsening-upwards sequence characterized by increasing thickness of the sand beds upwards.

(2) *Crevassing*: flood waters flow into the interdistributary area via small crevasse channels cut in the levee crest. There are two distinct mechanisms.

(a) *Crevasse splay*: a sudden incursion of sediment-laden water which deposits sediment over a limited area on the lower flanks of the levees and the bay floor, producing locally wide levee aprons. The sediment may be deposited in many small, anastomosing streams, in which case the deposit comprises numerous small channel lenses. Alternatively the flow may be sheet-like and deposit an erosive-based lobe of sand which may be either a few centimetres thick or up to 1–2 m thick (Kruit, 1955; Arndorfer, 1973). The thicker splay lobes often infill the bay and are overlain by facies reflecting emergence or near emergence.

(b) *Minor mouth bar/crevasse channel couplets*: in the Mississippi delta, couplets comprising semi-permanent crevasse channels and small-scale mouth bars are an important feature of the interdistributary areas (Coleman and Gagliano, 1964; Coleman,

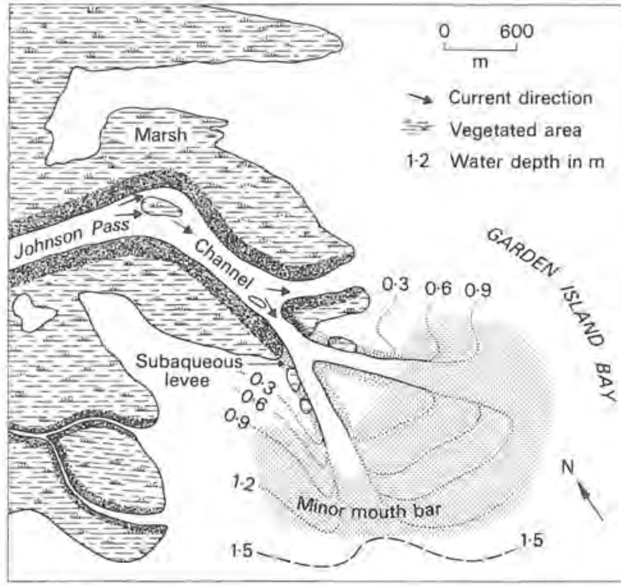



**Fig. 6.6.** A minor mouth bar crevasse channel couplet in an interdistributary bay of the modern Mississippi delta (after Coleman, Gagliano and Webb, 1964).

Gagliano and Webb, 1964; Fig. 6.6). Shallow crevasse channels bounded by subaerial levees flare at the mouth and deposit minor mouth bars which form shoal areas dipping gently into the bay. As a couplet progrades into the bay, proximal facies progressively overlie distal facies, and shallow borehole descriptions can be used to construct a series of vertical sequences. Bioturbated muds and silts deposited on the bay floor pass upwards into interbedded silts and sands with multi-directional trough cross-lamination which is considered to reflect current and wave action on the mouth bar front. The upper silts and sands are frequently eroded by the base of the crevasse channel as progradation continues. Stage variations are particularly important in the crevasse channels as the channels may be temporarily abandoned at low river stage, resulting in complete cut-off of sediment supply until the next river flood. Crevasse channel facies may therefore comprise sands with unidirectional, current-produced structures, together with numerous reactivation surfaces and fine-sediment drapes.

Close spacing of the minor mouth bars results in a laterally continuous front which advances into the bay, producing the sub-deltaic lobes of the Mississippi delta (Coleman and Gagliano, 1964). They develop from a crevasse break in the levee of a major distributary and have a life cycle of initiation, progradation and abandonment which spans 100–150 years. Following an initial period of subaqueous crevassing during which the initial break is enlarged and made semi-permanent, numerous minor mouth bars prograde rapidly across the bay infilling areas

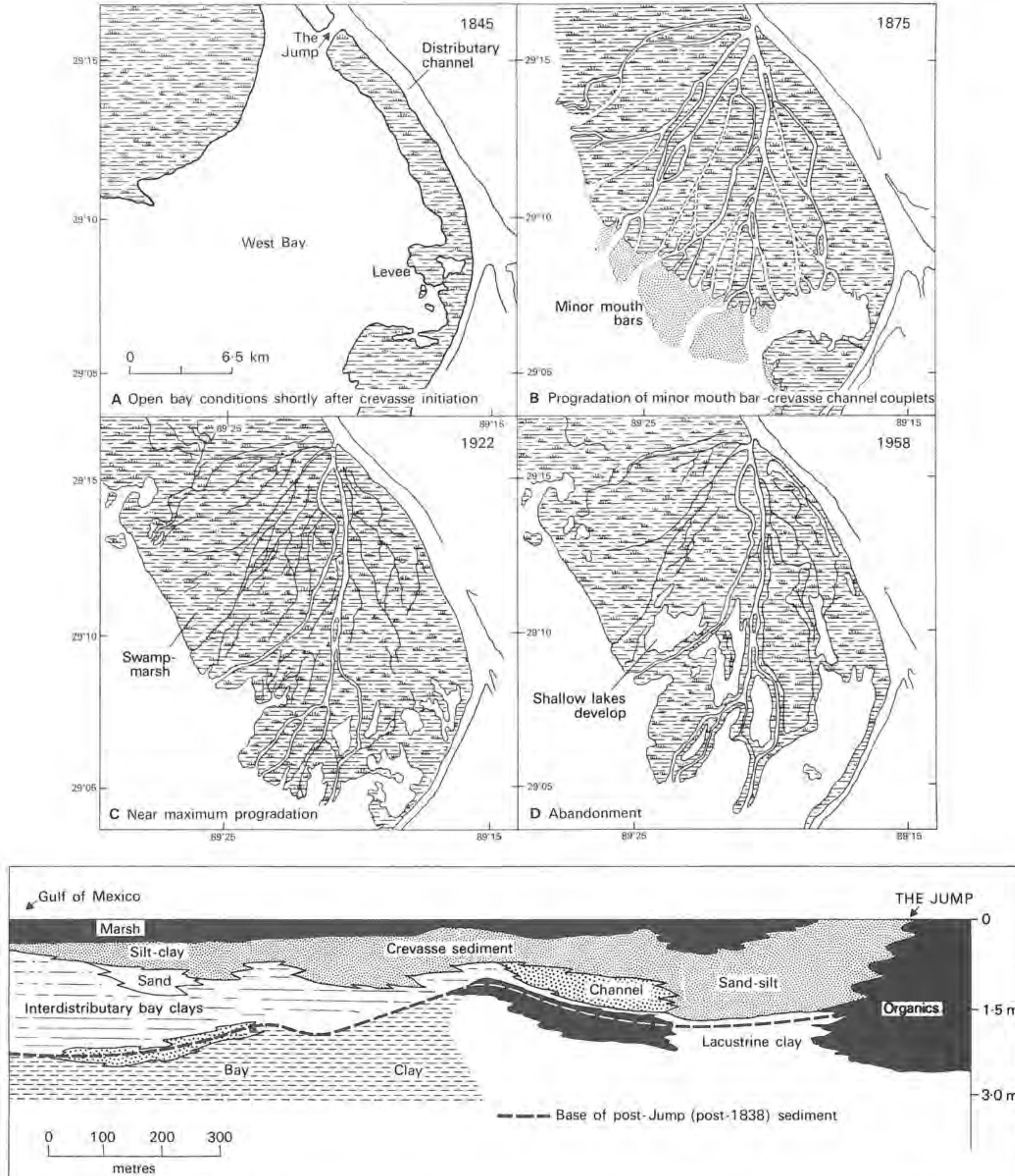



**Fig. 6.7.** Development of the West Bay subdelta, Mississippi delta and the resultant facies pattern (after Gagliano and van Beek, 1970).

of 300–400 $km^2$ (Fig. 6.7). When infilling is accomplished or the crevasse point is healed, the sub-delta is abandoned. Compaction, subsidence and coastal erosion take over, producing large bays on the marsh surface. Oyster reefs form on distal levee ridges, and the front of the sub-delta is reworked by wave action. However, despite modification during abandonment, the sub-deltas have a high preservation potential (Gagliano, Light and Becker, 1971). Following a period of subsidence the infilled bay reverts to an open bay several metres deep, and the infilling process recommences. Cores through bay areas reveal vertically stacked coarsening-upwards sequences 3–15 m thick, each representing infilling of the bay (Coleman, 1981).

The spatial distribution of processes operating in a fluvial-dominated interdistributary area is determined by the distance from active distributary channels. Near-channel facies will be dominated by levee sequences and numerous crevasse splay lobes, whilst distal or central facies may comprise fine-grained bay floor sediments or crevasse channel/minor mouth bar couplets.

In the open bays of the Mississippi delta, waves rework the upper part of crevasse sands into minor sand spits, and sediment may be reworked directly from the distributary mouth as large-scale sand spits extending back into the interdistributary bay (Fisk, McFarlan *et al.*, 1954). These features are likely to produce small and large-scale coarsening-upwards sequences which terminate in wave-dominated sand units comprising well-sorted sands with flat lamination, wave ripples and perhaps low-angle accretion surfaces.

Details of these sequences are discussed later, using ancient examples (Sect. 6.7.1; Fig. 6.29).

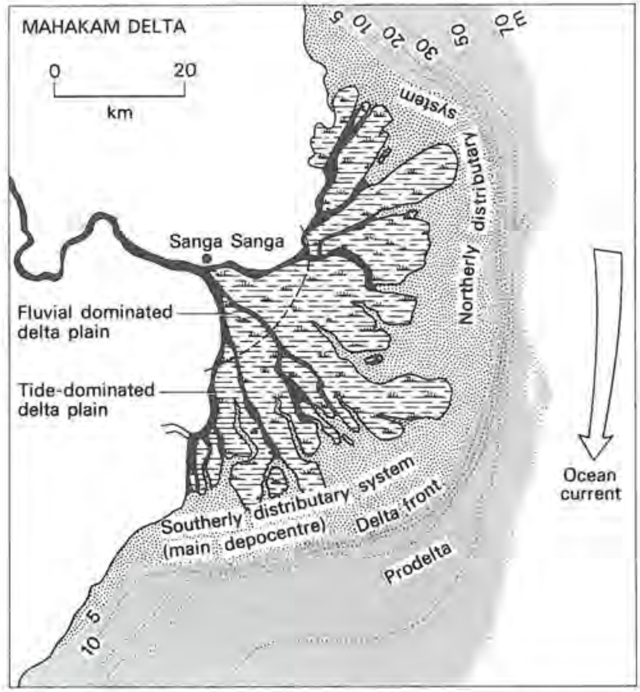



**Fig. 6.8.** Mahakam delta, Indonesia; a fine-grained, tide-dominated delta with an extensive area of tidal flats, estuarine channels, tidal channels and creeks dominating the delta plain (after Allen, Laurier and Thouvenin, 1979).

### TIDE-DOMINATED DELTA PLAINS

In areas of moderate to high tidal range, tidal currents enter the distributary channels during tidal flood stage, spill over the channel banks and inundate the adjacent interdistributary area. The tidal waters are stored temporarily and subsequently released during the ebb stage. Tidal currents therefore predominate in the lower distributary courses and the interdistributary areas assume the characteristics of intertidal flats (Fig. 6.8).

*Tidally influenced distributary channels* have a low sinuosity, flared and sometimes funnel-shaped form with a high width to depth ratio which contrasts with the almost parallel-sided nature of fluvial distributary channels in areas of low tidal range. The properties of the tidal wave determine the rate at which the banks converge upstream, with standing tidal waves inducing an exponential rate of convergence whilst progressive tidal waves induce a linear rate (Wright, Coleman and Thom, 1973). In the Niger delta more than twenty tidal inlets ranging in depth from 9 to 15 m dissect the beach-barrier shoreline (NEDECO, 1961; Allen, 1965d). Dune bedforms predominate in the channels and complex 'inner deltas' comprising a maze of sand bars and mudflats occur at confluences between distributary channels, upstream from their outlet to the sea. The 'inner deltas' probably result from deposition of a large proportion of the sediment load as the channels flare at the confluence and the fluvial currents are impounded by tidal currents. In the Mahakam delta, Indonesia, the fluvial and tidally influenced channels are characterized by side-attached, alternate bars or 'elongate lateral accretion bars' in low sinuosity channels (Allen, Laurier and Thouvenin, 1979). In the lower part of the tidally influenced reaches of these channels, flow-aligned bars occur in the central channel area. These features resemble the linear tidal current ridges of other tidally influenced deltas such as the Mekong, Irrawaddy, Gulf of Papua and Ganges-Brahmaputra deltas (Coleman, 1969; Fisher, Brown *et al.*, 1969; Coleman and Wright, 1975). In some deltas the ridges are several kilometres long, a few hundred metres wide and 10–20 m high, and reflect tidal current transport of sediment supplied by the river system. They resemble the tidal current ridges of shallow shelf seas (Sect. 9.5.3), but details of their morphology, short- and long-term behaviour, and facies characteristics have not been examined.

Sequences from tidally influenced distributary channels in the Niger delta commence with a coarse, intraformational lag with a fragmented marine fauna, overlain by sands which exhibit a

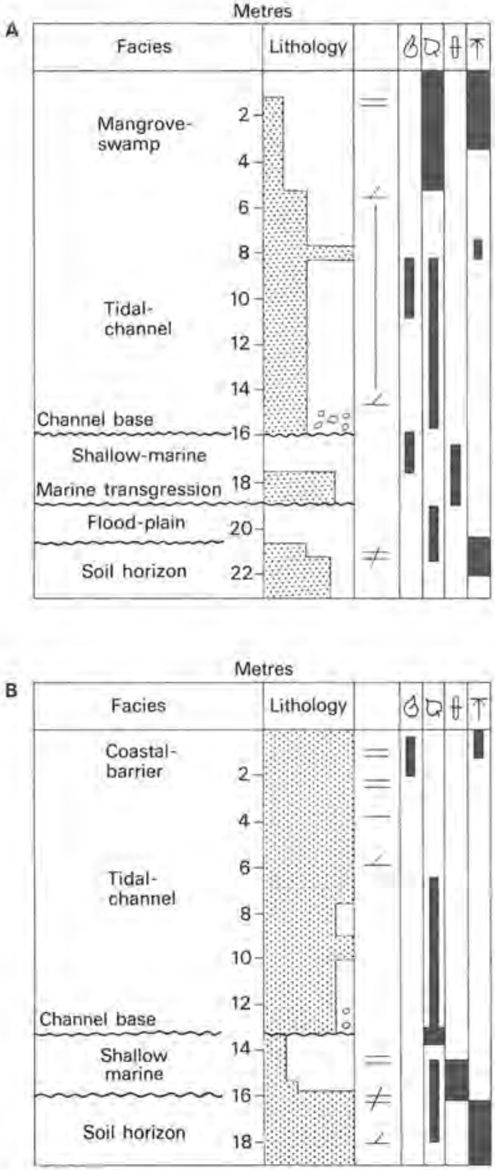



**Fig. 6.9.** Tidal-distributary channel sequences from the Niger delta: (A) an upper delta plain sequence terminating in mangrove swamp facies; (B) a shoreline sequence terminating in coastal barrier facies deposited as the channel migrates laterally alongshore (after Oomkens, 1974); key as for Fig. 6.5.

transition from decimetre-scale trough cross-bedding into centimetre-scale cross-lamination. The sand becomes finer upwards and there is also an increase in the clay content and the number of burrows. Facies in the upper part of the channel sequences vary in accordance with channel position. Upper delta plain distributaries pass upwards into rootlet-disturbed, organic-rich clays of the mangrove swamp, whereas near-shoreline tidal channel sequences terminate in flat-laminated coastal barrier sands (Weber, 1971; Oomkens, 1974; Fig. 6.9). Detailed observations in trenches cut through sub-Recent tidal distributary channels of the Rhine delta reveal the complexity of these sequences, in contrast to the limited observations possible in borehole cores (Oomkens and Terwindt, 1960; de Raaf and Boersma, 1971; Terwindt, 1971b). Trough cross beds with reversed palaeoflow directions generally pass upwards into heterolithic facies comprising linsen and flaser bedding, but this trend is frequently complicated by smaller scale fluctuations. Characteristic features include bimodality in flow direction and the frequency of small-scale vertical facies variations, both reflecting the fact that tidal currents fluctuate in direction and strength on a small time scale.

Tidal distributary channels are less prone to switching and abandonment, but can migrate laterally. Sand body shape and dimensions are therefore a function of the size and form of the channels and the degree of lateral migration. In the Mahakam delta, the channels are characterized by elongate, flow-aligned sand bodies 4–5 km long and 0.5–1.5 km wide. Sand thickness varies along the bodies, with pods up to 10 m thick reflecting the sites of alternate bars which migrated laterally to some extent (Allen, Laurier and Thouvenin, 1979). In contrast, the larger and more freely migrating tidal channels of the former Rhine delta produced sand bodies 20 km wide and 50 km long (Oomkens, 1974).

*Interdistributary areas of tide-dominated delta plains* include lagoons, minor tidal creeks and intertidal-supratidal flats which are sensitive to the climate. In the Niger delta, interdistributary areas are dominated by mangrove swamps (vegetated intertidal flats) dissected by tidal distributary channels and a complex

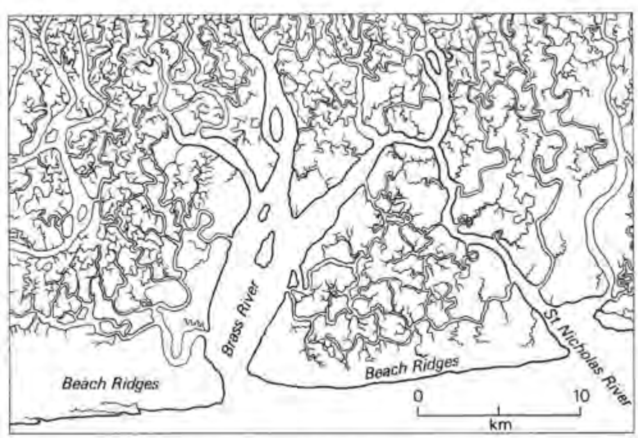



**Fig. 6.10.** Tide-dominated lower delta plain of the Niger delta comprising extensive mangrove swamps dissected by tidal-distributary channels and a maze of minor tidal creeks (after Allen, 1965d).

pattern of meandering tidal creeks (Allen, 1965d; Fig. 6.10). Sands are deposited by laterally migrating point bars in the tidal creeks and mangrove swamps develop on the surface left by the point bars. The entire delta plain probably comprises a sheet-like complex of small-scale, erosive-based sequences which pass upwards from point bar sands-silts into the mangrove swamp facies, with localized clay plugs representing infilled channels. In the Mahakam delta, the fluvial-dominated delta plain is restricted to 10–20 km downstream of the point where the distributary channels branch off the alluvial trunk stream (Fig. 6.8). The high proportion of fine-grained sediments in this delta and the equatorial climate cause the tide-dominated delta plain facies to comprise organic rich muds with abundant plant debris derived largely from the nipah palm and mangroves of the tidal flat (Allen, Laurier and Thouvenin, 1979). The delta plain of the Colorado River delta at the head of the Gulf of California is also tide-dominated, but the arid climate causes the interdistributary areas to be desiccated mud- and sand-flats with localized salt pans, particularly near the supratidal limit (Meckel, 1975).

Tide-dominated delta plains therefore comprise tidally influenced (or dominated) distributary channel sequences and tidal flat facies which also reflect the prevailing climate. Despite the deltaic setting it is possible that there will be no evidence for fluvial processes in this association, except perhaps for a relatively abundant sediment supply in excess of that normally associated with non-deltaic tidally dominated areas. However, in a prograding succession this association will overlie a tide-dominated or wave-dominated shoreline sequence, and will itself be overlain by a fluvial-dominated upper delta plain association.

### 6.5.2 The delta front

This is the area in which sediment-laden fluvial currents enter the basin and are dispersed whilst interacting with basinal processes (Fig. 6.11). The radical changes in hydraulic conditions which occur at the distributary mouth cause the flow to expand and decelerate, thus decreasing flow competence and causing the sediment load to be deposited. Basinal processes either assist in the dispersion and eventual deposition of sediment, or rework and redistribute sediment deposited directly as a result of flow dispersion.

Modern and ancient deltas cannot be understood without consideration of river mouth processes and sediment load. Of prime importance is the precise manner in which fluvial outflow and basin waters mix at the distributary mouth. In an early example of the application of hydrodynamic principles to geological problems, Bates (1953) contrasted situations in which the river waters were equally dense, more dense and less dense than the basin waters (homopycnal, hyperpycnal and hypopycnal flow; Fig. 6.12). If the water bodies are of equal density, immediate three-dimensional mixing occurs at the river mouth

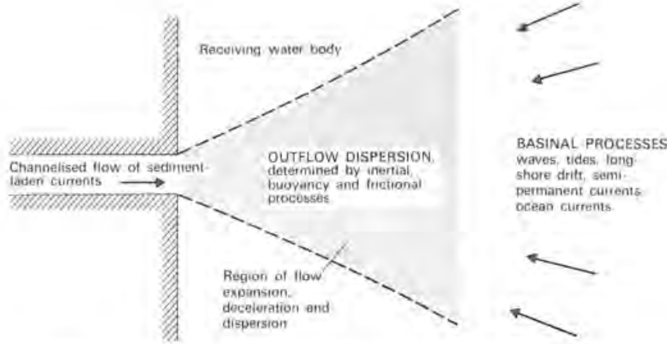



**Fig. 6.11.** Processes involved in the interaction between sediment-laden river waters and basin waters at the delta front (based on Wright and Coleman, 1974).

causing appreciable sediment deposition at this point. High density outflow tends to flow beneath the basin waters as density currents, causing sediment to by-pass the shoreline, thus restricting the development of a delta. If the outflow is less dense than the basin waters it enters the basin as a buoyantly supported surface jet or plume. This latter situation, hypopycnal flow, has been observed off the Mississippi and Po deltas (Scruton, 1956; Nelson, 1970) and is considered to operate wherever river water enters marine basins as seawater is slightly denser than freshwater. Thus hypopycnal flow is the main method of interaction in marine deltas but the possibility of other mechanisms operating, at least briefly during flood periods, should not be neglected.

Central to the idea of hypopycnal flow is the buoyancy of the outflow, but other important factors neglected until recently include inertial processes related to outflow velocity, and frictional processes which result from the outflow interacting with the sediment surface at the distributary mouth (Wright, 1977). Differing combinations of inertial, frictional and buoyancy processes at the river mouth produce a series of outflow dispersion models which are typified by distinctive river mouth configurations in areas where rivers debouch into basins with limited wave or tidal energy (Fig. 6.13).

*Inertia-dominated* river mouths form where high velocity, bedload rivers enter a freshwater basin. Sediment is dispersed as a turbulent jet under homopycnal conditions and produces an elongate, steep-fronted Gilbert-type mouth bar (Fig. 6.1). In this pure form, inertial processes are therefore of limited significance in major river deltas in marine basins, but they can be important at river mouths during brief high discharge periods of river floods. *Friction-dominated* river mouths occur where rivers enter basins with shallow inshore waters. Frictional interference between the flow and the sediment surface increases the spreading and deceleration of the jet and produces a triangular 'middle ground bar' in the mouth of the river which

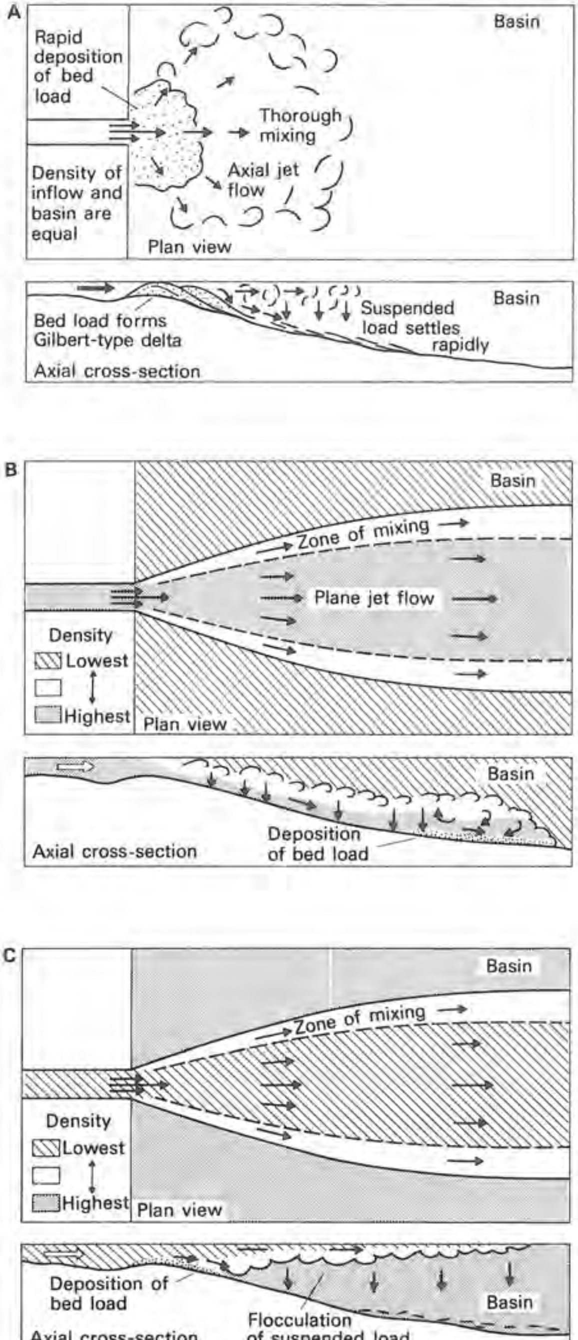



**Fig. 6.12.** Different modes of interaction between sediment-laden river waters and basin waters, determined by the relative density of the water bodies: (A) homopycnal flow; (B) hyperpycnal flow; (C) hypopycnal flow (after Fisher, 1969; originally from Bates, 1953).

causes the channel to bifurcate (Fig. 6.13). As progradation continues, new bars form at the mouths of the bifurcated channels and the delta spreads into the basin in this fashion. *Buoyancy-dominated* river mouths form where river waters extend into the basin as a buoyantly supported plume and are therefore restricted to marine basins. In this mechanism, a salt-water wedge intrudes into the lower part of the channel. This is favoured by the presence of relatively deep water channels at the mouth and moderately deep water fronting the river mouth thus reducing the role of frictional mixing of the water masses. Turbulent exchange across the boundaries of the plume causes expansion, mixing and deceleration of the plume and thus results in deposition of the sediment load. Mixing is particularly intense near the river mouth as internal waves are commonly generated between the plume and the underlying salt wedge. An appreciable amount of the sediment load, particularly the coarser sand fraction, is deposited at the river mouth, whilst the finer grained sediment is transported further into the basin and deposited from suspension as the plume disperses. A dominance of buoyancy processes at the river mouth produces an elongate mouth bar which projects a considerable distance into the basin with a gently dipping (0.5–1°) slope (Fig. 6.13).

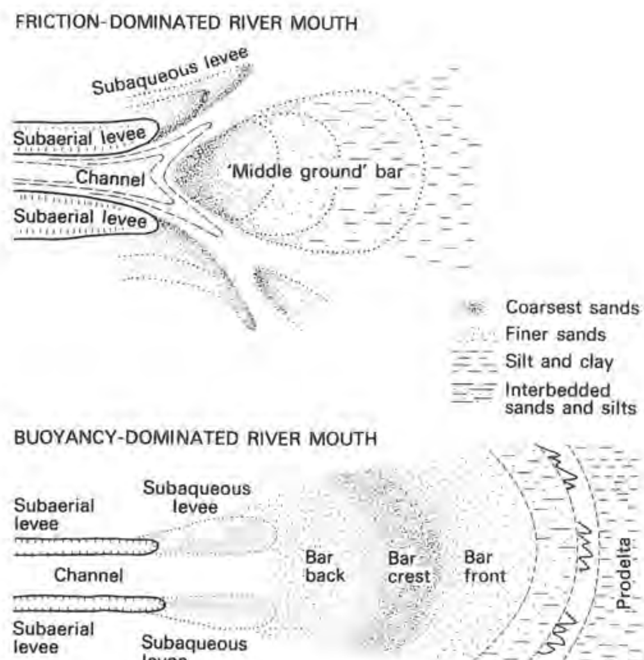



**Fig. 6.13.** Friction-dominated and buoyancy-dominated river mouth bars which respectively develop in shallow-water and deep-water areas of fluvial-dominated deltas, for example in the east and south of the modern Mississippi delta (modified after Wright, 1977).

These mechanisms are end-members of a spectrum, and deposition at river mouths often involves a blend of inertial, frictional and buoyancy processes. Discharge fluctuations of the rivers are particularly important in this respect. It is common for river mouths to be dominated by buoyancy processes when discharge is low and be influenced more by frictional and inertial processes during high discharge periods (e.g. South Pass, Mississippi delta; Wright and Coleman, 1974).

Moderate *wave action* does not unduly interfere with operation of the primary river-dominated outflow dispersion mechanism but may rework sediment deposited at the distributary mouth. Intense and persistent wave energy directly affects outflow dispersion, and sediment is distributed according to the wave-induced circulation pattern (Wright, 1977). High wave energy impounds the discharge and increases mixing of the water masses, causing sand-grade sediment to be concentrated at the shoreline whilst finer grained sediment is dispersed offshore. Localized mouth bars ornamented by landward-oriented swash bars may form, but commonly most of the sand is transported alongshore by longshore currents to produce a continuous fringe of beach sands to the upper delta front (Fig. 6.14). In some cases the course of the distributary channel may be impeded by the growth of a beach-spit, causing the channel to flow parallel to the shoreline in the direction of longshore drift for several kilometres. Major rip current systems (Sect. 7.2.1) also form on the flanks of deltas and can transport considerable volumes of upper delta front sediment offshore into slightly deeper waters (Wright, Thom and Higgins, 1980).

The effectiveness of wave processes in redistributing sediment supplied by rivers has been examined by Wright and Coleman (1973). A calculation of wave power at the shoreline was combined with river discharge data to provide a 'discharge effectiveness index' which describes the relative effectiveness of river discharge against wave reworking ability, and can be compared between deltas. Derivation of this index for the Mississippi, Danube, Ebro, Niger, São Francisco and Senegal deltas demonstrates a close correlation between delta front sedimentation and discharge effectiveness index.

The manner in which *tidal processes* operate in the delta front area has not been studied rigorously. Sediment transport in the lower part of distributary channels and distributary mouth areas may be influenced or dominated by the tidal currents. Once again, the most important effect of tidal current processes is to increase mixing between the water masses and therefore promote sediment deposition in the river mouth area. Where tidal currents are significant they often confine direct fluvial discharge to the upper part of the delta plain, and sedimentation in the lower delta plain and delta front is then largely a response to the tidal current regime, except perhaps during major river

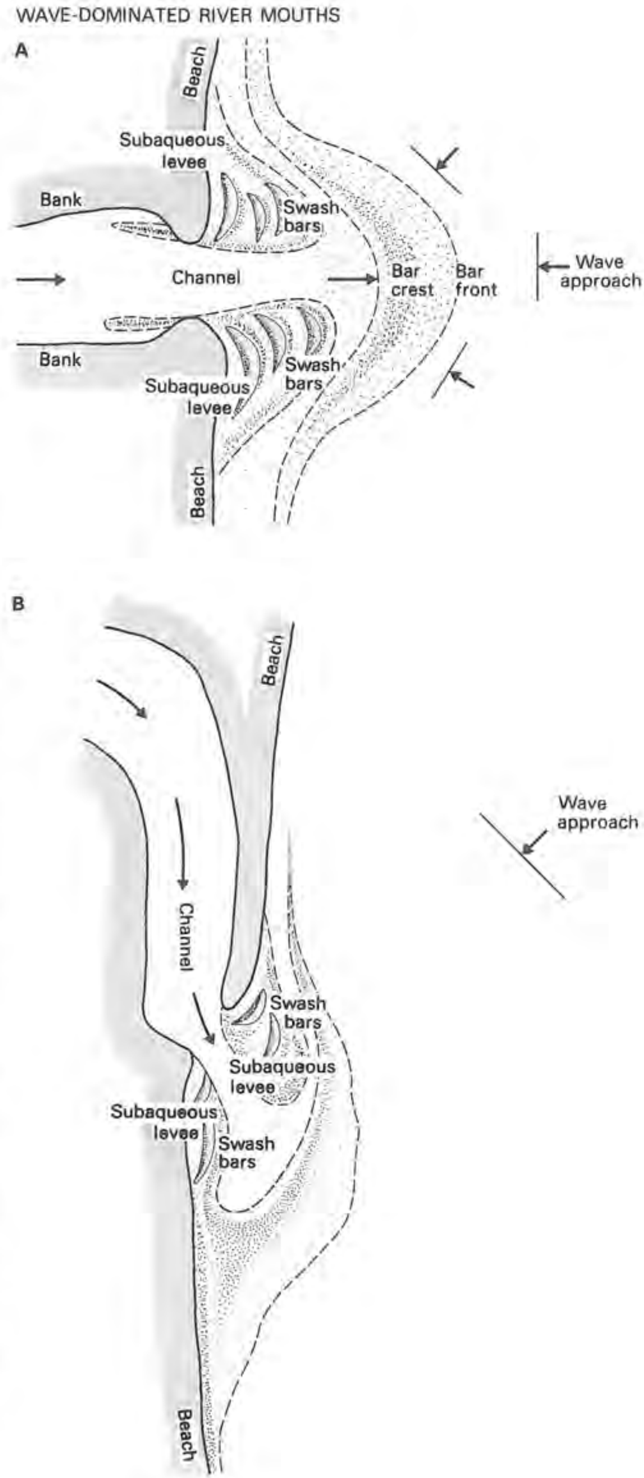



**Fig. 6.14.** Wave-dominated river mouth settings for: (A) direct onshore wave approach; (B) oblique wave-approach and associated dominant longshore drift direction (modified after Wright, 1977).

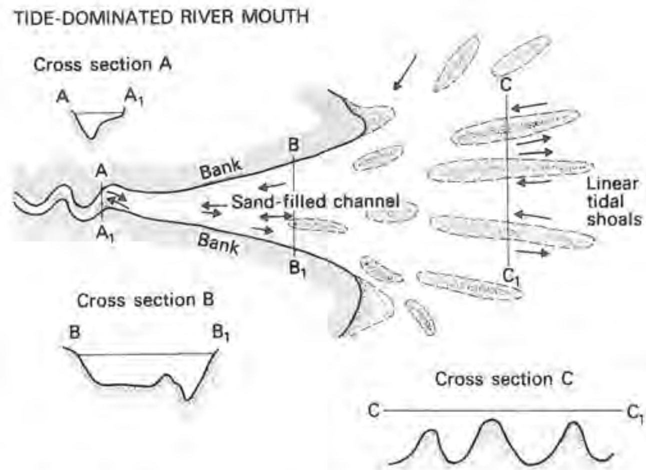



**Fig. 6.15.** Tide-dominated river mouth illustrating the funnel-shape of the lower distributary channel, the predominance of linear tidal current ridges or shoals in the channel and a zone of intense meandering at the head of the funnel-shaped channel (modified after Wright, 1977).

floods. In these areas the lower part of the distributary channel assumes a funnel shape, often with a zone of intense meandering above the head of the funnel. Bi-directional sediment transport along preferred ebb- and flood-dominated pathways prevails in the channels and the river mouth, and these regions are dominated by fields of linear tidal current ridges (Fig. 6.15; Wright, 1977).

In the delta front, coarse sediment tends to be deposited at the distributary mouth, whilst finer sediment is transported further into the basin and deposited in deeper water offshore environments. Sediment deposition therefore constructs a seaward-dipping profile which slopes gently into the basin, generally at an angle of less than 2°, and fines progressively into the basin. The delta front progrades offshore in response to continued sediment supply so that former offshore areas are eventually overlain by the shoreline, producing a relatively large-scale coarsening-upwards sequence which reflects infilling of the receiving basin. However, a point worth emphasizing is that delta front progradation is rarely uniform and the facies patterns may not therefore be as orderly as portrayed in the limited descriptions from modern deltas. In addition to the vagaries of sediment-laden discharge entering a standing water body with its own regime, sediment supply varies from point to point around the delta front according to the position of channel mouths and the proportion of the alluvial discharge in a particular distributary channel. Furthermore, sediment supply is constantly changing as individual distributary channels wax or wane, and supply may abruptly increase or decrease at a point as delta front progradation continues.

FLUVIAL-DOMINATED DELTA FRONTS

The Mississippi delta is the only major river delta in which delta front sedimentation is dominated by fluvial processes with minimal interference from basinal processes. The present delta was established 600–800 years ago and has since prograded rapidly across a platform of previously deposited clays to occupy a position at the edge of the continental shelf in comparatively deep waters. An extremely fine-grained sediment load comprising 70% clay, 28% silt, and 2% fine sand is supplied to the delta via a series of radiating distributary channels (Fig. 6.16). Sediment deposition at the mouth of a distributary constructs a discrete mouth bar which projects into the Gulf of Mexico, deflecting the bathymetric contours seawards. Two types of mouth bar are recognized (Fig. 6.13): (a) on the eastern side of the delta, distributaries empty into relatively shallow waters and are characterized by bifurcating channels and middle ground bars which result from a predominance of frictional processes at the mouths and (b) on the southern side of the delta, distributaries enter deeper water which favours the operation of buoyancy processes during periods of low to intermediate river discharge. Buoyancy-dominated mouth bars are the most extensively studied in the Mississippi delta. They comprise a *bar back* area which includes minor channels, subaqueous levees and bars superimposed on a gently ascending platform, a narrow *bar crest* located a short distance offshore from the

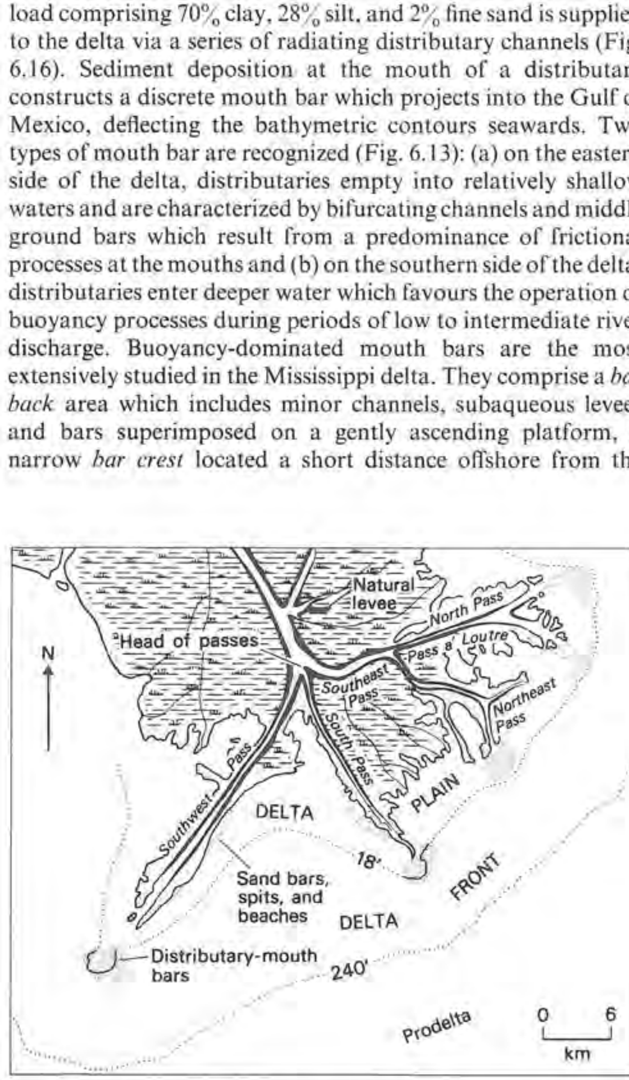



**Fig. 6.16.** Mississippi delta; a fine-grained, fluvial-dominated delta in which the delta front is composed of widely separated river mouth bars which are deep-water, buoyancy-dominated types in the south (Southwest Pass and South Pass) and shallow-water, friction-dominated types in the east (Northeast Pass and North Pass–Pass a' Loutre); see Fig. 6.13 for details of river mouth bars (modified after Gould, 1970).

distributary mouth and a *bar front* which slopes offshore to the prodelta (the term bar front as used here incorporates the distal bar of certain workers). This morphology results from a dominance of buoyancy processes at the river mouth, but during high river discharge the salt wedge is forced out of the channel and frictional and inertial processes therefore predominate at the river mouth with sediment-laden traction currents driving across the bar back and crest (Wright and Coleman, 1974). The manner in which the mouth bars prograde was revealed at South Pass during the extreme flood of 1973. The bar-crest *aggraded* rapidly during the flood with up to 3 m of sediment being deposited. As the flood diminished this sediment was reworked by river currents and transferred to the bar-front causing the 10 m depth contour to advance 90–120 m. Mouth bar *progradation* was therefore most marked immediately after the flood peak and bar front facies therefore have the highest preservation potential (Coleman, Suhayda *et al.*, 1974).

Progradation of these mouth bars produces large-scale (60–150 m) coarsening-upwards sequences which record a transition from prodelta clays upwards into sands of the upper bar front and bar crest (Fisk, McFarlan *et al.*, 1954; Fisk, 1955, 1961; Coleman and Wright, 1975; Coleman, 1981; Fig. 6.17). The prodelta clays vary in thickness from 20 to 100 m and are banded due to slight differences in grain size or colour. Bioturbation is generally slight due to the rapidity of deposition, but more intensely bioturbated horizons are produced when sedimentation rates decrease temporarily. The fabric of these clays comprises a framework of randomly oriented domains of clay particles with large voids and hence high porosity, except for thin crusts of more tightly packed clays which reflect shear zones related to synsedimentary deformation (Bennett, Bryant and Keller, 1977; Bohlke and Bennett, 1980). The prodelta clays are deposited from suspension and are devoid of current-produced laminae, but higher in the sequence, parallel and lenticular silt laminae and eventually thin, cross-laminated sands become intercalated with the clays, reflecting a combination of waves, sediment-laden current incursions from the distributaries and deposition of sediment from suspension. The bar crest is characterized by relatively well-sorted sands with cross-lamination, climbing ripple lamination and some flat lamination, deposited principally during river flood periods. As the introduction of sediment is virtually uncontested by basinal processes and the distributary channels are fixed by the cohesive muds into which they erode, progradation of distributary channel-mouth systems has produced a series of radiating 'bar finger sands' which directly underlie the present distributaries and provide the birdfoot framework of the delta. These sand bodies are bi-convex, elongate bodies up to 30 km long, 5–8 km wide with an average thickness of 70 m (Fisk, 1961; Fig. 6.18).

This classic view of the facies pattern of the Mississippi delta has limitations. The facies characteristics of these mouth bars cannot be summarized in a single, coarsening-upwards sequence as there are bound to be large- and small-scale variations in

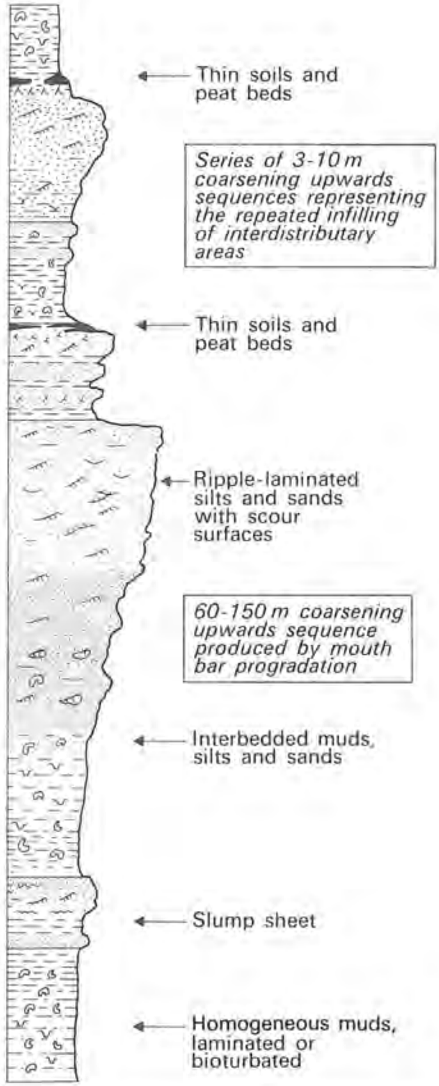



**Fig. 6.17.** Composite, idealized sequence produced by mouth bar progradation in the Mississippi delta (after Coleman and Wright, 1975).

facies as one moves from a near channel, axial area to a lateral area distant from the channel. In the axial area, the upper parts of the coarsening-upwards sequences may be eroded slightly by minor channels of the bar back, or more substantially by deeper, upstream channels as progradation continues. Where preserved, the upper mouth bar sand will be thick in the axial area and thin towards the lateral area, and numerous changes in facies details will accompany this trend as processes vary so markedly across the mouth bar. In addition, a great diversity of deformational

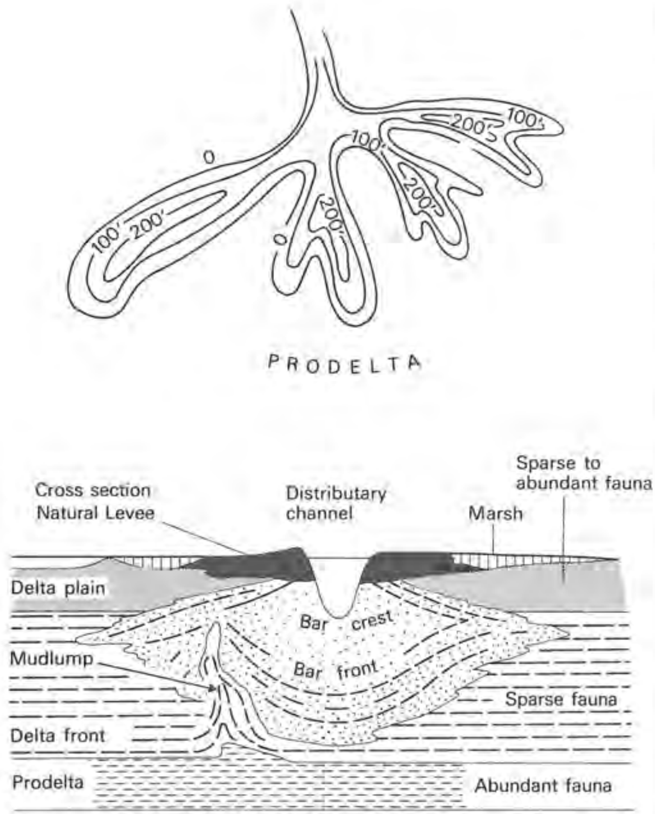



**Fig. 6.18.** Bar finger sands of the Mississippi delta as described by Fisk (1961); the influence of the diapiric mudlumps on sand body shape is now regarded as more significant than originally depicted and the bar fingers are thought to be composed of thick pods of sand between diapirs and thin connecting sand intervals above diapirs (see 6.8.2 and Fig. 6.45).

processes operate as the delta front progrades. These processes influence sedimentation directly by creating preferred areas of sand or mud deposition, or by translating previously deposited sediments downslope as debris flows, slumps or slides (Sect. 6.8). Considering the variety of these processes and the frequency with which they operate, mouth bar progradation is unlikely to produce a regular, orderly, coarsening-upwards sequence. This point also applies to the bar-finger sand body pattern described by Fisk (1961) as some of the deformation processes operate on a scale which can influence the overall facies pattern of the delta. For example, the bar fingers were originally thought to have a relatively uniform thickness of 70 m, but recently it has been demonstrated that this primary shape has been modified by mud diapirism into a series of thick pods of sand separated by thin strands (Sect. 6.8; compare Fig. 6.18 with 6.45).

A different type of fluvial-dominated delta front was formerly displayed by the small-scale Colorado River delta in East Matagorda bay, Texas (Kanes, 1970). The earlier lobe of this delta (pre-1930) was characterized by closely-spaced distributary channels and a continuous delta front composed of coalesced mouth bar sands. This important alternative to the classical birdfoot, fluvial-dominated form is also discernible in the pre-modern 'shoal water' lobes of the Mississippi delta (Fisk, 1955; Frazier, 1967; Sect. 6.7).

### FLUVIAL-WAVE INTERACTION DELTA FRONTS

In general, this type of delta front is characterized by a smooth, cuspate or arcuate, beach shoreline. Localized protuberances in the vicinity of the distributary mouth are composed of subdued mouth bars flanked by beach ridge complexes and reflect the fact that wave processes are capable of partially redistributing the river-borne sediment. Present-day examples occur in the Danube, Ebro, Nile and Rhône deltas, all of which are located in enclosed seas with moderate wave action but minimal tidal processes.

The most thoroughly described example is the Rhône delta (Kruit, 1955; van Straaten, 1959, 1960; Fig. 6.19). The delta front comprises laterally extensive beach ridges fronted by a relatively steep offshore slope (up to 2°) which descends to 50 m depth. Progradation is by beach ridge accretion and mouth bar progradation, and is most pronounced in the vicinity of the main distributary – the Grand Rhône. Elsewhere, the beach zone of the delta front is thin and in some places is retreating landwards (van Straaten, 1960), for example west of the Grand Rhône mouth where an earlier lobe of the delta is being reworked by wave action after abandonment.

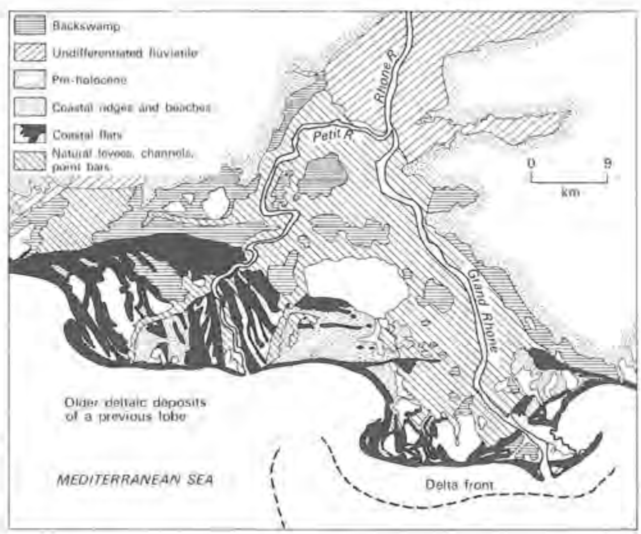



**Fig. 6.19.** Rhône delta; a sandy, wave-influenced delta with a continuous fringe of coastal barrier sands (after Van Andel and Curray, 1960).

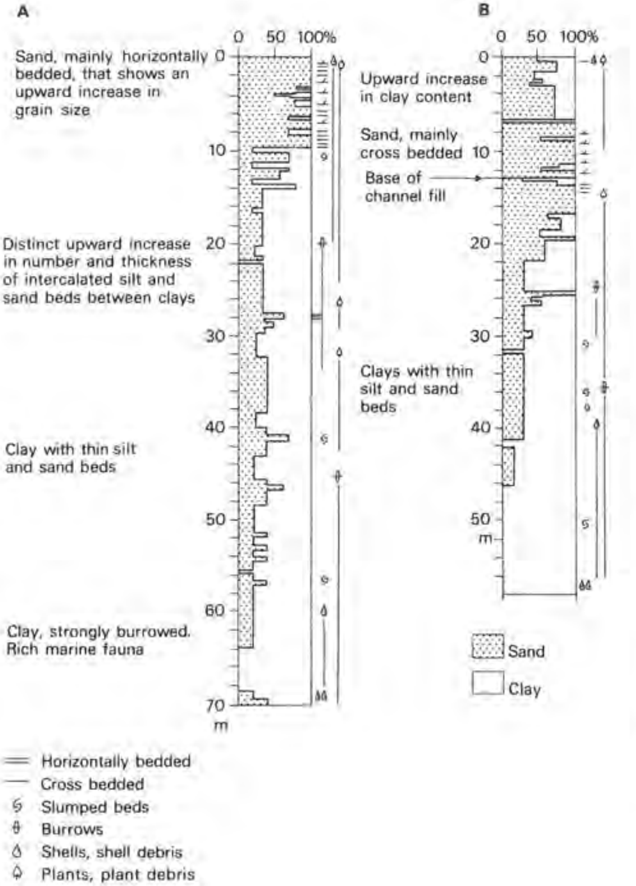



**Fig. 6.20.** Delta front coarsening-upwards sequences of the Rhône delta: (A) coarsens upwards gradationally into a coastal barrier sand, whereas (B) is truncated at 13 m by an erosive-based fluvial-distributary channel sequence (after Oomkens, 1967).

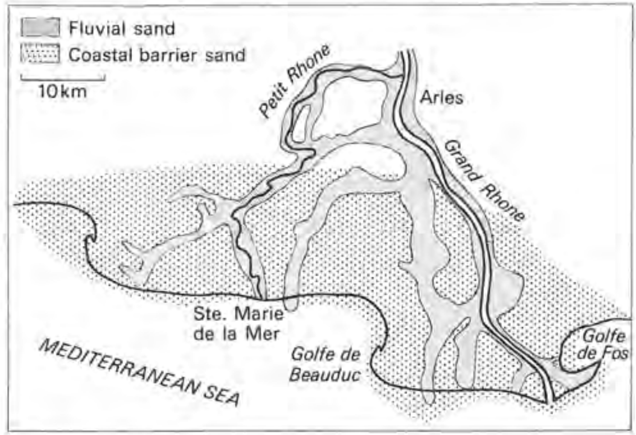



**Fig. 6.21.** Sand body pattern of the Rhône delta, illustrating a laterally extensive, slightly lobate coastal barrier sand cut locally by fluvial-distributary channel sands (after Oomkens, 1967).

Delta front coarsening-upwards sequences have been described from the Rhône and Ebro deltas (Lajaaij and Kopstein, 1964; Oomkens, 1967, 1970; Maldonado, 1975). Bioturbated offshore clays pass upwards into finely laminated silts which gradually acquire discrete beds of silt and sand in the intermediate part of the sequence. Ripple-lamination is common at this level, but the sand member at the top of the sequence consists of well-sorted, horizontally-bedded sand deposited by nearshore wave processes (Fig. 6.20). Oomkens (1967) distinguished *fluviomarine* sequences produced by direct fluvial input of sediment near the active distributary mouth, and *holomarine* sequences in which sediment was supplied by longshore drift from the distributary mouth, but these sequences appear identical lithologically and could only be distinguished by their microfaunal content which was more diverse and abundant in holomarine sequences. The absence of distinctive fluvial-influenced mouth bar facies in the upper part of the fluviomarine sequences may suggest that this facies is reworked after distributary channel abandonment and has a low preservation potential.

The sand distribution pattern in the Rhône delta consists of a laterally extensive beach-barrier sand, cut locally by distributary channel sands (Oomkens, 1967; Fig. 6.21). The maximum dimension of the sheet sand parallels the shoreline trend and subdued lobes are superimposed on this general trend in the vicinity of the present and formerly active distributaries. This pattern may resemble that produced by laterally coalescing distributary mouth bars in lobate, fluvial-dominated deltas, but points of difference include the lower number of distributary channel sands and the wave-dominated nature of the sheet sand.

### WAVE-DOMINATED DELTA FRONTS

In this type, wave processes are capable of redistributing most of the sediment supplied to the delta front. It is therefore characterized by a regular beach shoreline with only a slight deflection at the distributary mouth and a relatively steep delta front slope. Mouth bars do not form and bathymetric contours parallel the shoreline. Progradation involves the entire delta front, rather than particular points and is generally slow by comparison with other types. Abandoned beach-ridges occur behind the active shoreline and the delta plain is often dominated by aeolian dunes and shallow, elongate lagoons between beach ridges. In plan view the ridges are separated into discrete groups by discontinuities which reflect changes in shoreline configuration induced by changes in the direction of

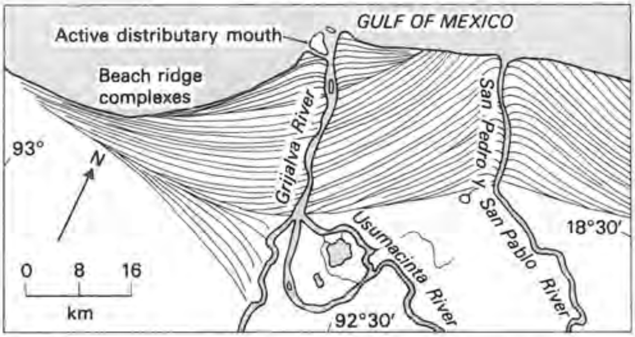



**Fig. 6.22.** Wave-dominated Grijalva delta (after Psuty, 1967).

longshore drift or the position of distributary channels (Psuty, 1967; Curray, Emmel and Crampton, 1969; Fig. 6.22).

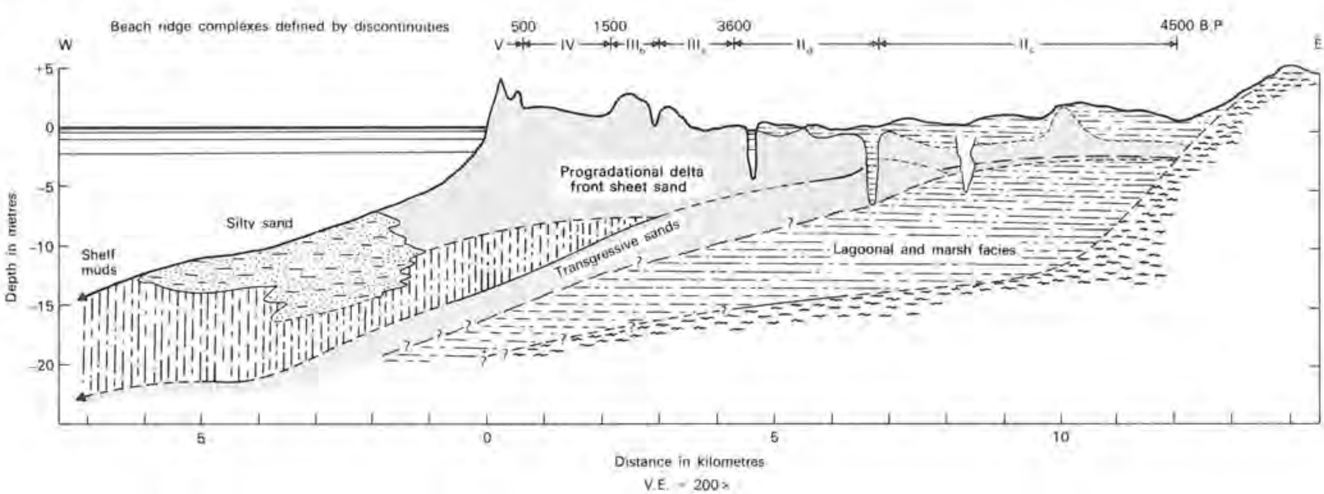
An idealized coarsening-upwards sequence of this type of delta front has been described from the São Francisco delta front (Coleman and Wright, 1975). Bioturbated, fossiliferous muds at the base pass upwards into alternating mud, silt and sand beds with wave-induced scouring, grading and cross-lamination and finally into a well-sorted sand with parallel and low-angle laminations representing a high-energy beach face (see Sect. 7.2). Aeolian sands succeed the delta front sequence, though the preservation potential of these sediments is uncertain. As progradation involves the entire delta front the resultant sand-body is a sheet-like unit which parallels the shoreline. For example, the coastal plain of Costa de Nayarit, Mexico is a broadly arcuate wave-dominated delta which has prograded 10–15 km over a distance of 225 km since the stabilization of sea-level after the Holocene transgression to produce a major, delta front sheet sand (Curray, Emmel and Crampton, 1969; Fig. 6.23).

### FLUVIAL-WAVE-TIDE INTERACTION DELTA FRONTS

Tidal currents frequently operate in conjunction with wave processes at the delta front. Tidal effects are confined to distributary mouth areas whilst waves operate over the remainder of the delta front, and the shoreline is composed of wave-produced beaches or cheniers separated by tide-dominated distributary channels and mouth areas. Offshore, bathymetric contours and facies belts parallel the shoreline, although there may be slight protrusions in the vicinity of distributary mouths. Examples of this type occur in the Burdekin, Irrawaddy, Mekong, Niger and Orinoco deltas, of which the best described is the Niger delta (Allen, 1965d; Fig. 6.24).

More than twenty tide-dominated distributary channels dissect the beach-barrier shoreline of the Niger delta and each distributary mouth has a shallow, sandy bar. These bars vary in shape from linear to arcuate and are deflected by longshore currents around the delta front. They have been consistently described as river mouth bars, but river discharge is minimal at this point (NEDECO, 1961) and it seems more probable that they result from the expansion of tidal currents and more closely resemble ebb-tidal deltas (Sect. 7.3). These features and the beach face descend to an inshore terrace rather than sloping uniformly offshore. This terrace, known as the 'delta front

**Fig. 6.23.** Cross-section through the wave-dominated Costa de Nayarit delta system, Mexico, illustrating an extensive sand body produced by progradation of the delta front following the Holocene transgression (after Curray, Emmel and Crampton, 1969).



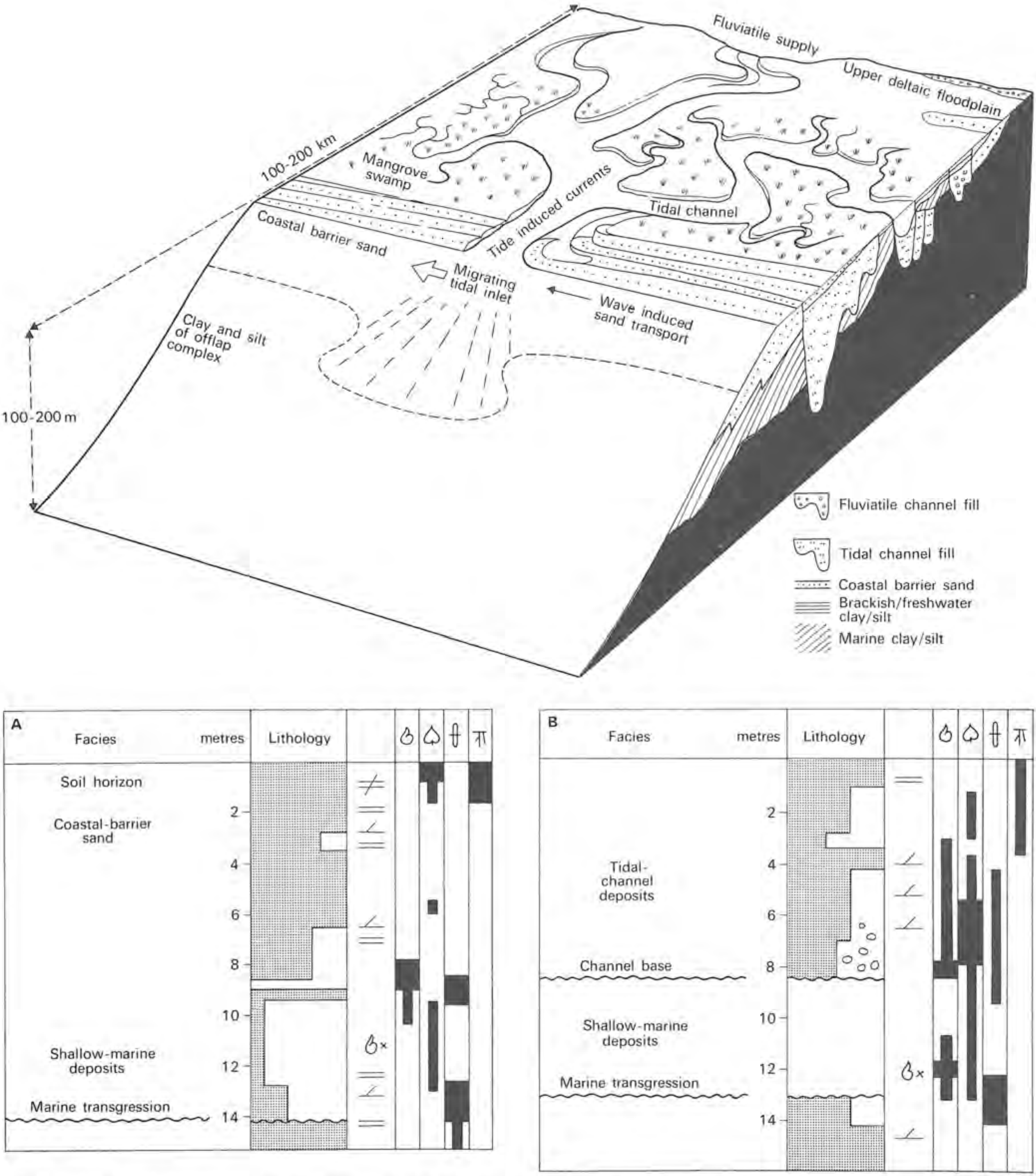



**Fig. 6.24.** Niger delta, a wave- and tide-influenced delta in which the delta front consists of (A) wave-dominated, coastal barrier coarsening-upwards sequences cut by (B) erosive-based tidal channel sequences which sit directly on shallow marine deposits (after Oomkens, 1974). (Key as in Fig. 6.5.)

platform', occurs at 5–10 m water depth, is up to 20 km wide and has a distinct regime produced by the interaction of tidal currents, waves, longshore and semi-permanent currents (Allen, 1965d). Beyond this platform the delta front slopes gently offshore into a low-energy environment mildly affected by waves, tidal currents and the Guinea Current which contours the prodelta slope. Detailed facies descriptions are available for the various subenvironments of the Niger delta front (Allen, 1965d), and vertical facies sequences have also been described from cores (Weber, 1971; Oomkens, 1974). The sequences range in thickness from 10 to 30 m and commence with bioturbated clays with occasional silt and sand lenses which pass upwards into interbedded muds, silts and sands. Towards the top of a sequence there is either a gradational passage into well-sorted, parallel-laminated sands of the beach face, or the sequence is cut by a tidal channel sand (Fig. 6.24). The extent to which the tidal inlets migrate laterally in the direction of longshore drift, as in non-deltaic, wave-tide influenced shorelines, is not known but may be appreciable (Sect. 7.3). Ideally this delta front would be represented by a sheet-like barrier-beach sand body frequently cut by linear tidal channel sand bodies normal to the shorelines and with up-dip and (?) down-dip extensions, but in fact sand body characteristics are extensively controlled by synsedimentary growth faults (Weber, 1971; Sect. 6.8.2).

In the Mekong and Irrawaddy deltas the shoreline consists of discontinuous chenier-like beach ridges rather than substantial beach-barriers. Shoreline progradation in the Mekong delta has produced an area of abandoned beach ridges which extends inland for 56 km. Inland, the ridges become progressively subdued and are eventually overlain by delta plain facies (Kolb and Dornbusch, 1975).

#### TIDE-DOMINATED DELTA FRONTS

In tide-dominated delta fronts the shoreline and distributary mouth areas are often an ill-defined maze of tidal current ridges, channels and islands which may extend a considerable distance offshore before giving way to the delta front slope (e.g. Ganges-Brahmaputra delta; Coleman, 1969). The main features of this type of delta front are the tidal current ridges which radiate from the distributary mouths. In the Ord River delta the ridges are on average 2 km long, 300 m wide and range in height from 10 to 22 m. Channels between the ridges contain shoals and bars covered by flood- and ebb-oriented bedforms (Coleman and Wright, 1975). In an idealized vertical succession from this delta, the tidal current ridge sands at the top of the delta front coarsening-upwards sequence are composed of bi-directional trough cross-beds with occasional clay drapes and numerous minor channels (Fig. 6.25). In terms of sand body characteristics, this type of delta front will probably produce relatively thick, elongate bodies aligned normal to the shoreline trend. In the Mahakam delta, lower tidal range and the mud-dominated nature of the sediment load cause the delta front to comprise an

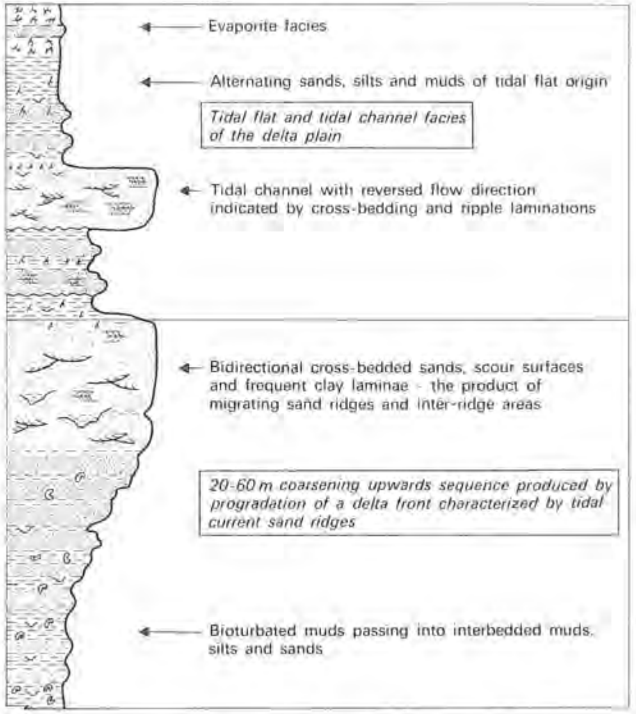



**Fig. 6.25.** Composite, idealized sequence through the tide-dominated Ord delta (after Coleman and Wright, 1975).

extensive, seaward-dipping platform 8–10 km wide of river- and tide-dominated silts-sands with localized wave-reworked concentrations of plant debris (Allen, Laurier and Thouvenin, 1979).

## 6.6 DELTA ABANDONMENT

Deltas often have a two-fold history comprising a constructional phase during which the delta progrades, and a destructional or abandonment phase initiated by a reduction in the amount of sediment supplied to the delta. Although most sedimentation takes place during the constructional phase, consideration of the abandonment phase can greatly assist the interpretation of sub-recent and ancient deltaic successions.

One cause of abandonment is alluvial- or distributary channel switching which results from over-extension of the channel system as the delta progrades into the basin. Shorter, steeper courses are generated, and if a crevasse breach is enlarged during a series of floods it may become a persistent feature of the channel network which gradually accepts an increasing proportion of the discharge of the parent stream until the latter is abandoned (Fisk, 1952a). In the present Mississippi delta the

Atchafalaya River is diverting an increasing amount of discharge from the Mississippi River. From its point of bifurcation the Atchafalaya River flows for only 227 km before reaching the Gulf of Mexico, whereas the Mississippi River flows for 534 km. The gradient advantage of the Atchafalaya River is precipitating the next major abandonment phase of this delta complex. Man-made controls have, to some extent, postponed this abandonment event, but despite this, a new delta is forming rapidly in Atchafalaya Bay (van Heerden and Roberts, 1980; van Heerden, Wells and Roberts, 1981).

The present Mississippi delta was preceded by a series of 'shoal-water' deltas which prograded across the shallow shelf east and west of the modern delta site. Four major pre-modern delta complexes comprising fifteen lobes have been recognized (Frazier, 1967; Fig. 6.26). As these lobes have been successively abandoned during the last 6000 years they are currently undergoing various stages of abandonment and a sequence of events can be demonstrated reflecting progressive changes during abandonment. The deltas were stable and subsided slowly by comparison with the deeper water, shelf-edge, modern Mississippi delta. The shoal-water deltas are also characterized by sheet-like delta front sands which extend over 800 km or more, and therefore subside uniformly rather than differentially (Fisk, 1955; Sect. 6.7.1; Fig. 6.36). Modification of the Lafourche complex is restricted to slight smoothing of the shoreline, accompanied by lateral transport of the reworked sand to form a minor barrier island (Grand Isle). More advanced stages of abandonment are illustrated by the older St. Bernard Complex which is characterized by a narrow, arcuate barrier island (the Chandeleur Islands) produced by wave reworking of the former delta shoreline (Fig. 6.27). This 'delta margin island' confines a shallow bay over the former delta plain in which fossiliferous clays, silts and sands are slowly accumulating. As subsidence continues, the delta margin islands tend to migrate landwards as they are entirely dependent on the underlying abandoned lobe for sediment supply. Finally, in the still older Teche and Maringouin complexes the former delta margin islands are marked by broad, submerged shoal areas several metres below sea-level. Whilst these modifications are taking place in the vicinity of the former delta shoreline, the upstream areas are covered by peat blankets which may extend uninterrupted over several hundred square kilometres, and in

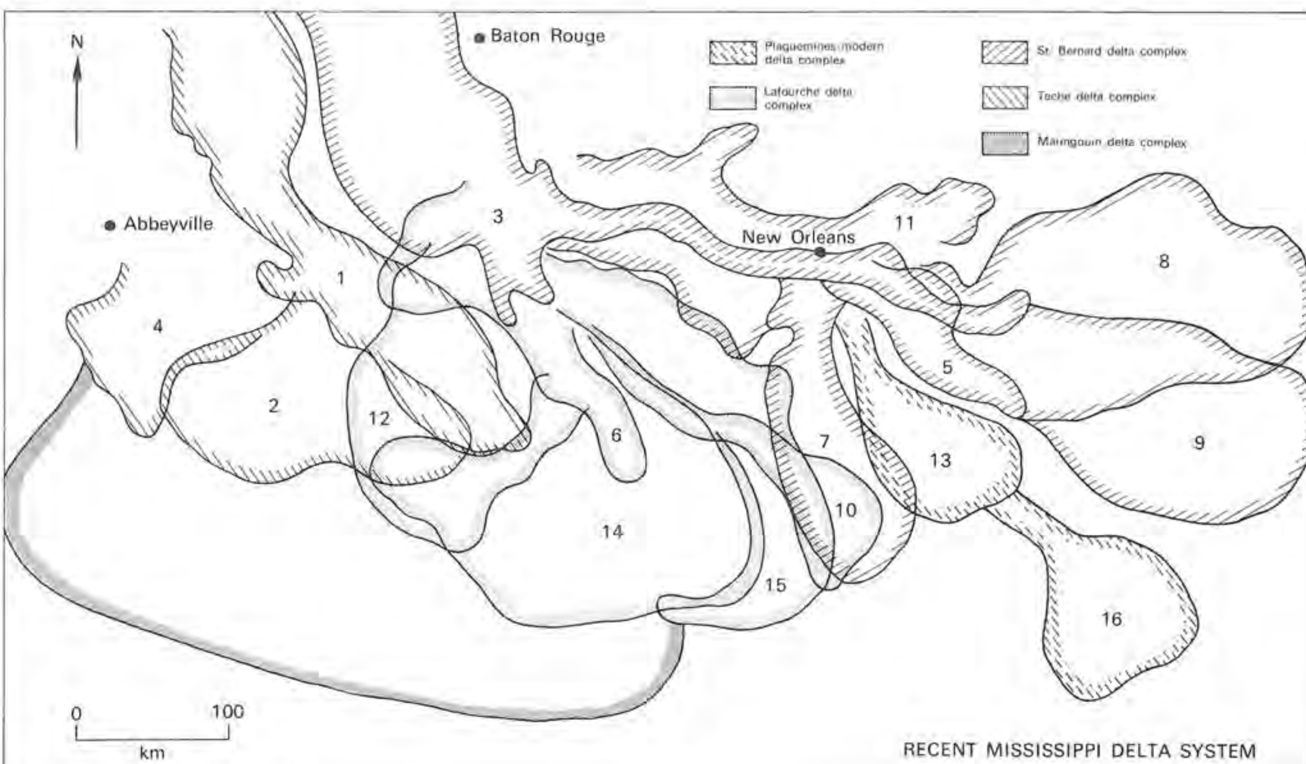



**Fig. 6.26.** Delta complexes and lobes of the pre-modern and modern Mississippi delta (based on Frazier, 1967 and Fisher and McGowen, 1969).

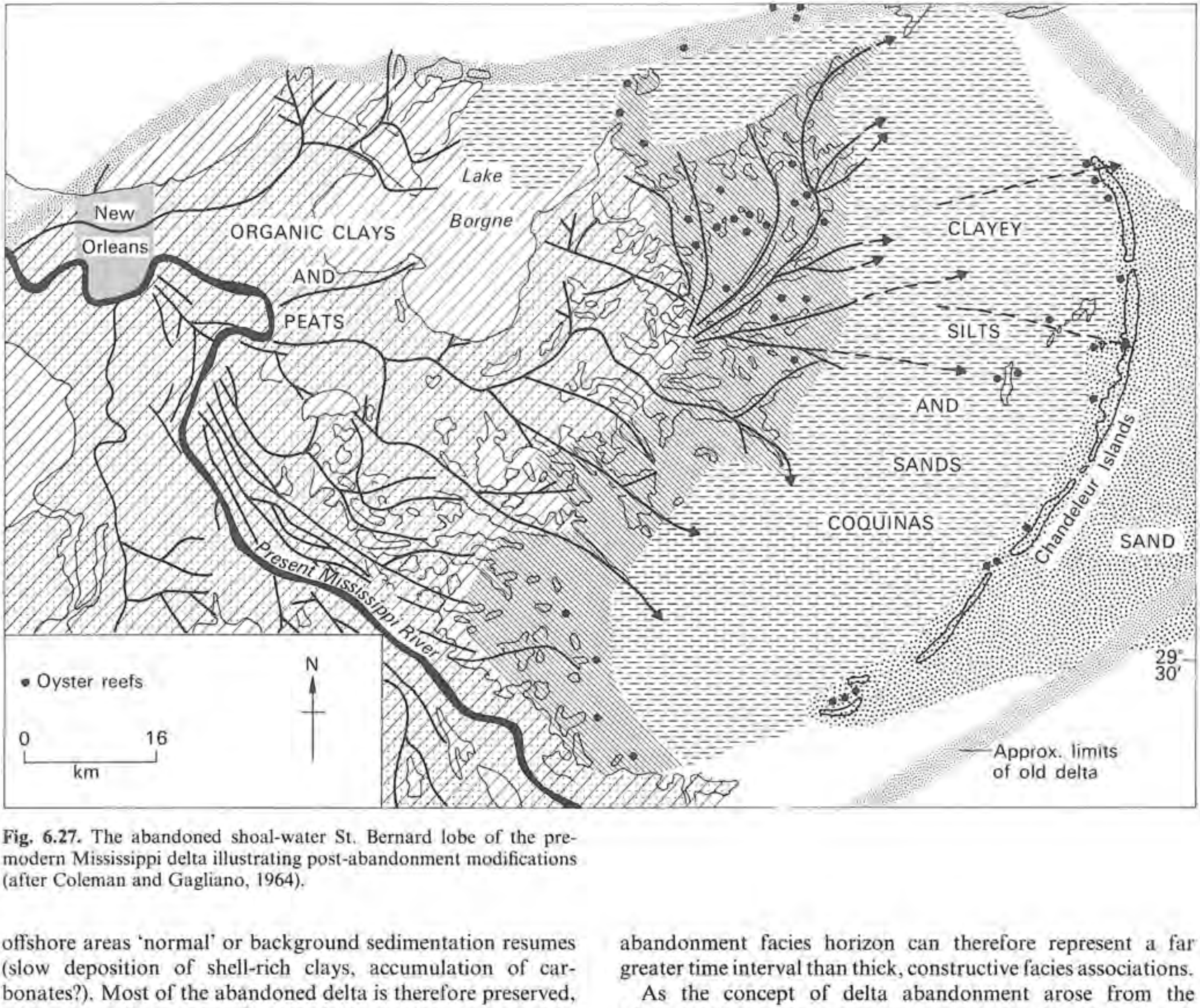



**Fig. 6.27.** The abandoned shoal-water St. Bernard lobe of the premodern Mississippi delta illustrating post-abandonment modifications (after Coleman and Gagliano, 1964).

offshore areas 'normal' or background sedimentation resumes (slow deposition of shell-rich clays, accumulation of carbonates?). Most of the abandoned delta is therefore preserved, with only the former shoreline area being partially reworked. In this example, the abandonment phase produces a thin, laterally persistent unit which varies in facies across the abandoned delta but is generally distinguished by relatively slow rates of sedimentation. An upstream peat blanket passes laterally into fossiliferous clays, silts and sands of the protected bay (restricted fauna?), a thin sheet sand which is the transgressed remnant of a delta margin barrier island, and finally into a thin unit of offshore facies with a diverse and prolific marine fauna.

Radiocarbon dating of the peat blankets reveals that during the abandonment of one lobe another may be initiated, prograde to its full extent and commence its own abandonment phase (Frazier, 1967; Frazier and Osanik, 1969). A thin abandonment facies horizon can therefore represent a far greater time interval than thick, constructive facies associations.

As the concept of delta abandonment arose from the Mississippi delta, its universality must be considered in view of the current emphasis on the variability of deltas. The initiation and abandonment of delta lobes or complexes is related to the frequency of channel avulsion. Rapid progradation in fluvial-dominated deltas produces significant shoreline protuberances and gradient advantages therefore abound. Avulsion occurs frequently and lobes proliferate. However, as wave effectiveness increases, deltas advance more slowly over a broader front. Fewer gradient advantages are created and avulsion is therefore less frequent. In addition, after abandonment a greater proportion of the former delta is likely to be reworked by wave action. For example, in the Rhône delta only three lobes have formed during the same period in which the Mississippi delta lobes were produced. If abandonment occurred in a wave-dominated delta, reworking would probably obliterate the potential lobe and the entire area of the former delta would be transgressed. In tide-influenced deltas avulsion is likely to be confined to the fluvial channel reaches in the upper delta plain or alluvial valley as tidal channels tend to migrate continuously rather than switch direction. As tide-influenced deltas often protrude significantly from the general shoreline trend, avulsion may occur in upstream areas, but many of these deltas are located in narrow basins which restrict the development of discrete lobes.

Channel switching and the development of delta lobes is, therefore, a preferential feature of fluvial-dominated deltas and to a lesser extent of fluvial-wave interaction deltas. However, delta abandonment may also result from a rise in sea-level, from fluctuations in sediment input due to climatic changes in the source area, or from tectonically induced river capture. For example, in the Ganges-Brahmaputra delta, river capture has resulted from basement faulting in combination with erratic major floods. Capture of the Hooghly river resulted in the abandonment of a large deltaic tract now occupied by a dense swamp area (the Sundarbans Jungle).

## 6.7 ANCIENT DELTAIC SUCCESSIONS

Ancient deltaic successions have the following characteristics.

(a) They are thick, predominantly clastic successions which pass from offshore basinal facies upwards into continental, fluvial facies as the delta progrades.

(b) The sediment body of the delta is of restricted lateral extent, forming a depocentre fixed around the mouth of the major river.

(c) Within a depocentre the successions are often repetitive or cyclic due to the repeated progradation and abandonment of the entire delta, or lobes within the delta.

(d) In major, long-lived deltas a series of discrete depocentres occurs in various patterns dictated by long-term fluctuations in sediment supply, subsidence and sea-level.

The recognition of deltas in the geological record requires the identification of three principal facies associations which can be seen to be genetically linked within the same formation: the delta plain facies association, the delta front facies association, and the delta abandonment facies association.

The *delta plain facies association* reflects deposition in distributary channels and interdistributary areas between the channels. The association may be fluvial-dominated or tide-dominated. The former is common to all deltas to some extent, whilst the latter occupies the lower delta plain of tidally influenced deltas.

The *delta front facies association* is generally represented by large-scale coarsening-upwards sequences which record a passage from fine-grained offshore or prodelta facies upwards into a shoreline which is usually sandstone-dominated. These sequences result from progradation of the delta front and may be truncated by fluvial- or tidal-distributary channel sequences as progradation continues. The sequences vary considerably within deltas in relation to the proximity of the distributary channel mouth, and between deltas according to the regime of the former delta front and the nature and extent of synsedimentary deformational processes. Consideration of the processes operating in this sub-environment is usually crucial to a complete understanding of the ancient delta as the interaction between sediment-laden fluvial processes and basinal processes takes place in the delta front.

The *delta abandonment facies association* comprises those facies which accumulate following the abandonment of a delta, or a lobe of the delta. Abandonment facies generally consist of thin but laterally persistent marker beds composed of facies which reflect slow rates of sedimentation. Although volumetrically insignificant, abandonment facies are important for four reasons (Fisher, Brown *et al.*, 1969; Elliott, 1974a): (1) they permit correlation in successions which are otherwise characterized by lateral impersistence of facies, (2) the beds reflect delta (or lobe) abandonment and therefore help in reconstructing the history of sedimentation, (3) the beds only develop in the areas of the abandoned delta (or lobe) and therefore define its areal extent and (4) since fluvial processes are at their weakest, the abandonment facies often provide the best indication of 'background' conditions such as the climate of the depositional area and water conditions (e.g. salinity, temperature) of the receiving basin.

Recognition of these facies associations not only permits a deltaic interpretation to be made, but also enables the type of delta to be debated. Early attempts at distinguishing different types of deltas in the geological record are largely confined to discriminating between lobate and birdfoot Mississippi types. However, as the range of deltas was presumably comparable, if not greater, in the geological past it follows that a more diverse range of delta types should be recognizable in the ancient record.

### 6.7.1 Ancient fluvial-dominated deltas

A wide range of fluvial-dominated deltas has been recognized in the geological record, particularly from the Carboniferous of Europe and the United States, the subsurface Tertiary of the Gulf coast, USA, and to a lesser extent from the Jurassic and Cretaceous of the Western Interior, USA. Following a general introduction to the facies associations of fluvial-dominated deltas several types of fluvial-dominated deltas are discussed.

*Fluvial-dominated delta plain associations* comprise large-scale fluvial-distributary channels, smaller-scale crevasse channels, and facies of the interdistributary bays which often occur as a series of small-scale coarsening-upwards sequences reflecting repeated infilling of the bays (Ferm and Cavaroc, 1968; Elliott, 1974b; Baganz, Horne and Ferm, 1975; Fig. 6.28). These sequences are on average 4–10 m thick and commence with mudstones-siltstones deposited from suspension across the

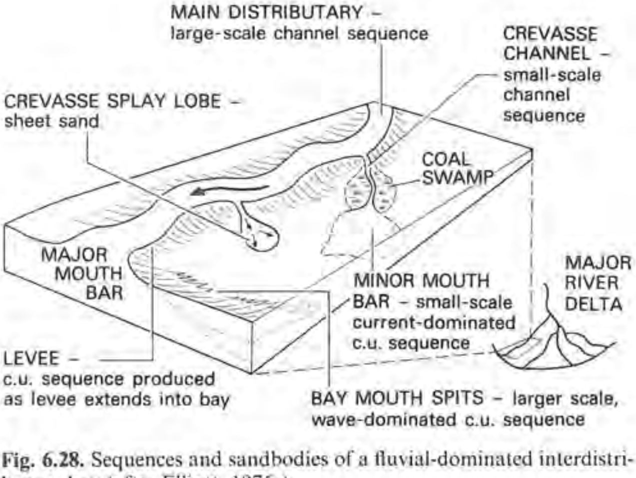



**Fig. 6.28.** Sequences and sandbodies of a fluvial-dominated interdistributary bay (after Elliott, 1976c).

entire interdistributary area during river flood periods. Plant debris is often abundant, along with a brackish or freshwater fauna. The sediments are sometimes finely banded or varved, but are more commonly thoroughly bioturbated. In organic-rich mudstones-siltstones, nodules or thin, impersistent beds of sideritic ironstone are common, having formed by early, pre-compaction diagenesis as described in present-day lake sediments of the Atchafalaya area (Ho and Coleman, 1969). These mudstones-siltstones may constitute the entire bay-fill sequence, but more commonly the sequences terminate in a thin sandstone member. Facies details of the sandstones vary, depending on whether they reflect levee construction by overbank flooding, crevasse splay lobes, minor mouth bar-crevasse channel couplets, or wave-reworked sand spits (Fig. 6.29). Levee sequences are dominated by numerous thin, erosive-based beds of ripple- or flat-laminated sandstones deposited by overbank floods or small crevasse splays (Fig. 6.29B). These beds grade upwards into mudstones-siltstones deposited from suspension as the flood wanes. Individual beds become thinner and finer away from the channel margin, and depositional dips have been observed in some examples (Elliott, 1976c; Horne, Ferm *et al.*, 1978). The coarsening-upwards trend is considered to reflect infilling of the bay by growth or encroachment of the levee into the bay. Major crevasse splay events which deposit 1 m or so of sediment during a single event often accelerate infilling of the bay and thus terminate bay fill sequences (Fig. 6.29C,D). The crevasse splay is a sheet-like or lenticular erosive-based unit characterized by a waning flow sequence of structures and upwards fining of grain size which can easily be mistaken for a turbidite (cf. Sect. 3.9.2; Fig. 3.36). Abrupt shallowing caused by the sudden emplacement of these major crevasse splays can result in ponding and emergence of the splay area, thus completing a bay-fill sequence. Crevasse channel-minor mouth bar systems produce a variety of bay-fill sequences. The mouth bars deposit gradational coarsening-upwards sequences which record increasing influence of traction currents as the mouth bar progrades, but the mid to upper parts of these sequences are often eroded by the crevasse channel supplying the mouth bar (Fig. 6.29F,G,H). Crevasse channel sequences are generally 1–4 m thick and have a channel-fill which exhibits numerous reactivation surfaces, clay drapes and indications of temporary bedform emergence. These features reflect ephemeral flow in the crevasse channels resulting from healing or 'stranding' of the channel during periods of low river stage. Reworking of crevasse-supplied sediment by locally-generated wind-waves produces thin, wave-dominated coarsening-upwards sequences which probably reflect migrating beach-spits (Fig. 6.29E). Each of these sequences terminates in a sandstone body which is thin and impersistent, but they often coalesce into a more extensive sheet sandstone which infills the entire interdistributary area (Fig. 6.30). Palaeosols frequently occur towards the top of the sequences, and coal deposits can accumulate in infilled bays under conducive climatic conditions.

A similar suite of sequences occurs extensively in the Westphalian Coal Measures of northern Europe (Reading, 1971; Scott, 1978). Small-scale coarsening-upwards sequences identical to those produced by minor mouth bars reflect the infilling of shallow lakes by small deltas. These sequences are locally eroded by channels, occasionally with lateral accretion surfaces (Scott, 1978), which represent the fluvial channels which supplied the lake deltas. In general, the setting resembles the Atchafalaya lake-swamp area described by Coleman (1966).

Distributary channel sequences are larger than crevasse channel sequences in the same complex and generally reflect more continuous discharge conditions, though still with stage fluctuations. The sequences are similar to those of fluvial channels. Erosion surfaces at the base of the channels are often lined by intraformational debris such as mudstone-siltstone clasts, derived ironstone nodules and logs. The channel sandstones exhibit a variety of structures such as trough and planar cross-bedding, flat lamination and current ripple lamination which reflect unidirectional traction currents of fluctuating strength. The channel sandstone bodies are either single- or multi-storey, with a tendency within the same delta complex for being multi-storey in the upper delta plain and single-storey in the lower delta plain (see below). Avulsion is a common event on delta plains and leads to the abandonment of channel courses which often produces an overall fining-upwards trend in the channel-fill sequence (Fig. 6.31), with the fine member comprising ripple laminated siltstones, occasional thin crevasse splay sandstones, plant-rich shales and palaeosol-coal units. Aside from these generalizations, distributary channel sandstones display considerable variety. In the Upper Carboniferous of northern England a high sinuosity pattern is inferred in one example from the presence of lateral accretion surfaces (Elliott, 1976b), whereas in a separate example giant cross-bed sets up to

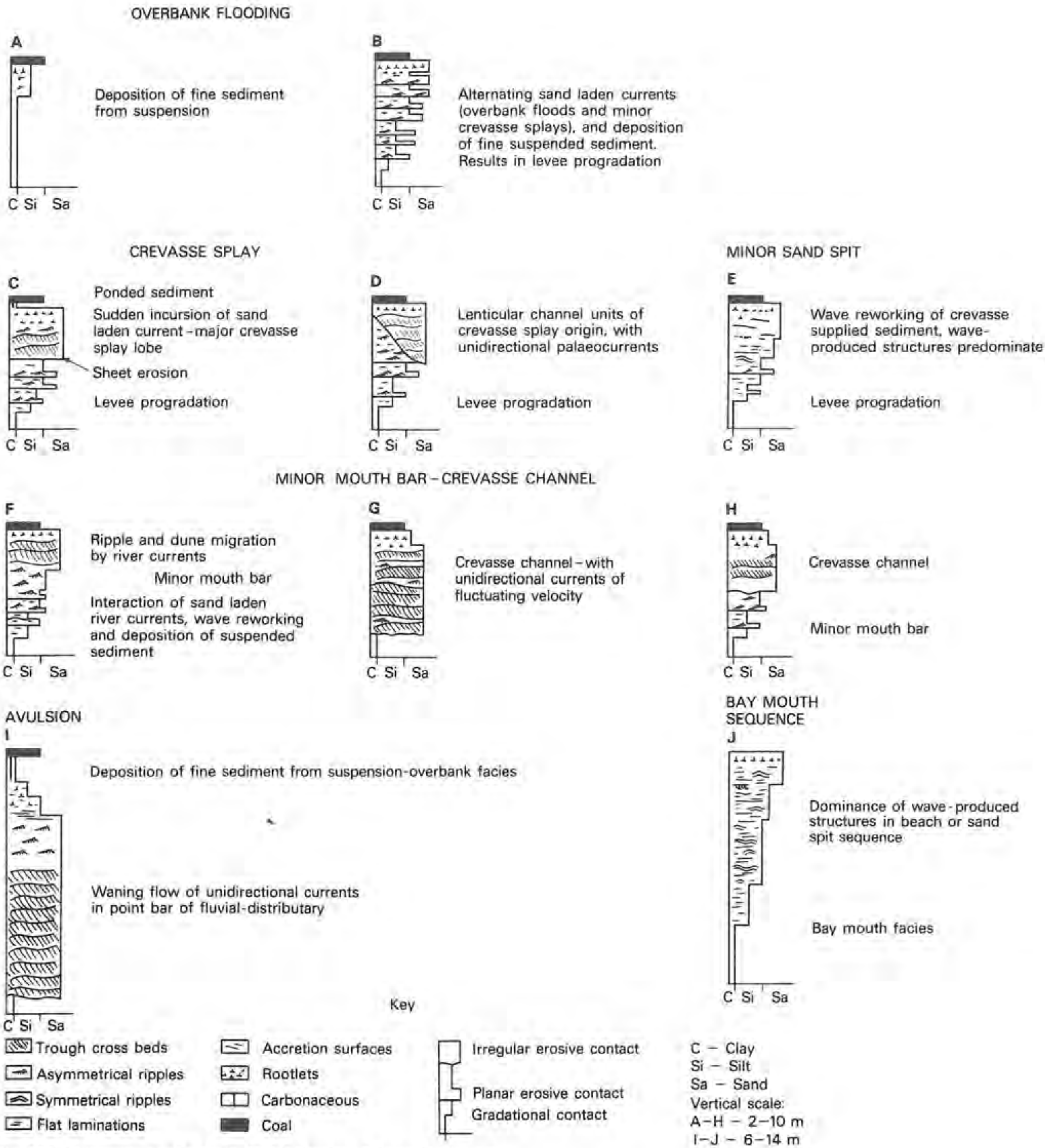



**Fig. 6.29.** Sequences produced in fluvial-dominated interdistributary areas (after Elliott, 1974b).

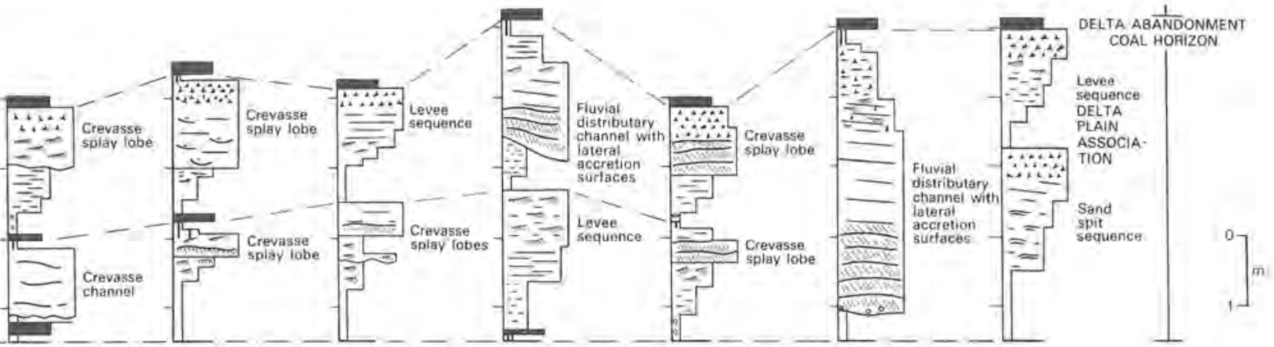



**Fig. 6.30.** Fluvial-dominated delta plain association from the Upper Carboniferous of northern England comprising two bay-fill sequences of laterally coalesced crevasse splay lobes, levees and sand-spits cut locally by crevasse and distributary channels; horizontal distance 10 km (after Elliott, 1975).

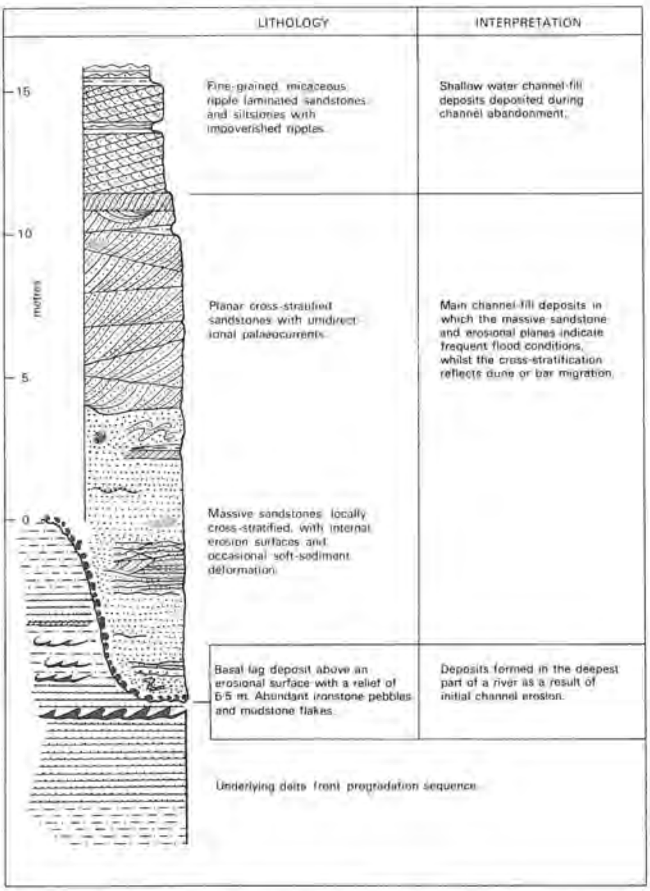



**Fig. 6.31.** Carboniferous fluvial-distributary channel sequence in south-west Wales (after Kelling and George, 1971).

40 m thick are interpreted as side-attached alternate bars in a low sinuosity channel (McCabe, 1977; Sect. 3.9.4; Fig. 3.52). In the Roaches Grit delta in the same area, similar large, bedload-dominated distributary channels with alternate bars shifted laterally in discrete steps to produce a sheet sandstone traceable over 400 km$^2$. This sandstone is composed entirely of channel facies, implying that finer grained delta plain facies were reworked by the shifting channels (Jones, 1980).

It is often possible to distinguish upper and lower delta plain facies associations within the same delta complex (Horne, Ferm *et al.*, 1978). The lower delta plain comprises relatively thick bay-fill sequences dominated by minor mouth bar-crevasse channel couplets deposited in the distal parts of the bay. Distributary channel sandstones in the lower delta plain are thin, single-storey channel units of limited lateral extent, reflecting the small scale of the channels and the frequency with which they avulse (Sect. 6.5). In contrast, the upper delta plain facies association is dominated by thick, multi-storey, laterally extensive distributary channel sandstones separated by interdistributary bay facies composed largely of levee and crevasse splay deposits. In some cases the coals are thicker and more abundant in the upper to mid delta plain (Ferm, 1976; Horne, Ferm *et al.*, 1978), but in other cases they are more abundant in the lower delta plain (Flores, 1979; Ryer, 1981). The latter is particularly pronounced where the delta front is a continuous sandstone body composed either of coalesced mouth bars or wave-built beaches as the coals accumulate immediately behind the delta front.

Fauna is often sparse in interdistributary bay facies and it is therefore difficult to distinguish fresh, brackish and saline bays. Palynological analysis of the shales can, however, assist in this problem (e.g. Hancock and Fisher, 1981).

*Fluvial-dominated delta front sequences* have been widely recognized in the geological record and exhibit considerable

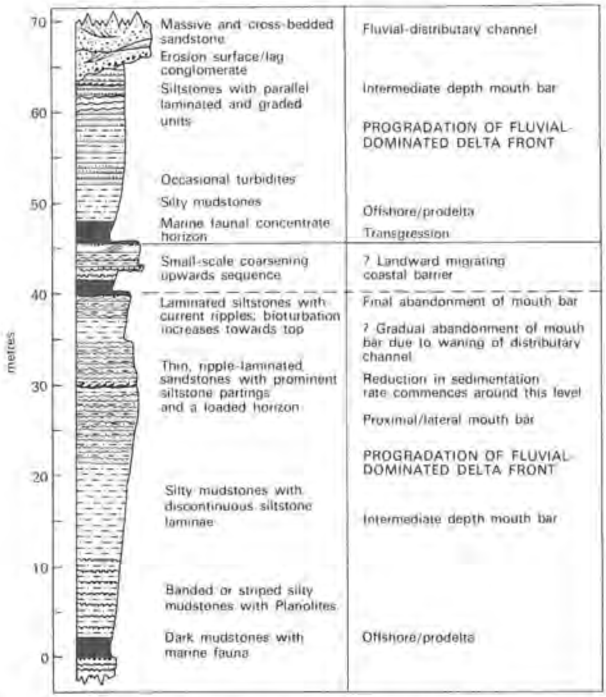



**Fig. 6.32.** Variations in fluvial-dominated delta front sequences: in the first sequence (0–40 m) the upper part is characterized by a reduction in the rate of sedimentation, possibly related to a gradual waning of the distributary channel; in the second sequence (45–70 m) the fluvial-distributary channel cuts into the delta front sequence (modified after Kelling and George, 1971).

variety. Although steep-fronted, Gilbert-type delta fronts have been described on rare occasions where coarse, bedload streams enter freshwater or low salinity basins (e.g. Cotter, 1975) (Sect. 4.6.1), most ancient delta front sequences are mud-silt-sand systems deposited at the margins of marine sedimentary basins. In general, they commence with a thick, uniform interval of mudstones-siltstones deposited from suspension at the base of the delta front and beyond (Fig. 6.32). This facies may appear massive, but more commonly exhibits diffuse banding defined by slight variations in grain size which reflect fluctuations in the supply of suspended sediment. Bioturbation may disrupt this banding and marine faunas occur, but faunal density and diversity are generally low due to the almost continuous fall-out of sediment from suspension. Plant debris occurs in this facies and is presumably an additional consequence of sediment input being direct from the distributaries. The intermediate parts of these sequences comprise mudstone-siltstone background sediment in which coarser siltstone and sandstone beds are repeatedly intercalated. Initially the background sediment is a direct continuation from below, but as the sequence is ascended thin, wave-produced siltstone laminae, small-scale ripple laminations and ripple form sets appear, reflecting the passage of the sequence above wave base. The coarser beds deposited in this mid-delta front setting in most cases have planar erosive bases and exhibit waning flow sequences involving a passage from parallel lamination upwards into asymmetrical ripple laminations. Upper surfaces of these beds are often sharply defined and may exhibit straight-crested symmetrical ripple marks reflecting post-flood, wave reworking of the upper few centimetres. These beds result from waning, sediment-laden traction currents, but the origin of these currents is problematic. Do they originate directly from the distributary in a friction- or inertia-dominated river mouth, or do they represent minor, tractional density currents generated on the upper delta front as river currents emerge from the channel mouth? In either case, is it necessary to argue that salinity values at the river mouth were reduced during flood periods in order to enable the currents to 'escape' from the river mouth? Other beds have gradational bases and indicate increasing flow velocity and sediment transport upwards, perhaps related to a different mechanism of outflow dispersion and more gradual flood rise. Towards the top of the sequence the coarse beds become thicker and are often amalgamated. Individual sandstone beds are laterally continuous, but lenticular units representing minor subaqueous extensions of the distributary channel may occur. Sedimentary structures in these beds reflect high rates of sediment transport and deposition by traction currents which prevail close to the distributary mouth during flood periods.

Variability in these sequences is considerable, with numerous examples departing substantially from the generalized account above. This variability arises from a number of causes.

(1) The sediment load supplied the delta front can be dominated by mud-silt, sand or gravel.

(2) The processes operating in fluvial-dominated delta fronts can vary in the relative importance of inertial, frictional and buoyancy processes (Sect. 6.5.2) and frequently show evidence of sediment gravity flow processes, particularly where synsedimentary slumping and faulting operated on the delta front.

(3) The number and spacing of distributary channels varies. Widely spaced distributary channels produce localized mouth bar sandstone bodies which display marked lateral variations in thickness and facies between near-channel, axial and lateral areas (Elliott, 1976a). Conversely, more numerous and closely spaced channels cause mouth bars to coalesce laterally and produce a continuous delta front sandstone. In an area influenced by two distributary mouths the sequence is likely to be more complicated.

(4) Delta front sedimentation can vary at a site because multiple distributary systems do not divide the discharge of the alluvial system equally, and also because individual distributaries wax and wane with time and may migrate laterally. Delta front coarsening-upwards sequences can, therefore, include abrupt

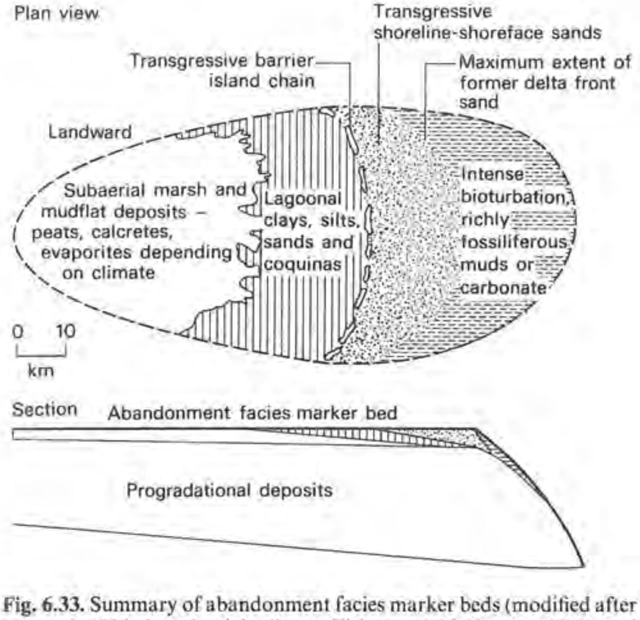



**Fig. 6.33.** Summary of abandonment facies marker beds (modified after Heward, 1981; based originally on Fisher and McGowen, 1967; and Elliott, 1974a).

changes to finer-grained or coarser-grained facies within the overall trend.

(5) Synsedimentary deformational processes can not only influence primary deposition of sediment on the delta front, but can also disrupt and displace previously deposited sediments and thereby modify the original facies pattern (Sect. 6.8).

*The abandonment facies association* in fluvial-dominated deltas usually comprises a thin but distinctive marker bed which can be traced across the abandoned delta or lobe (Fig. 6.33). In the low to mid delta front, abandonment is marked by an intensely bioturbated zone, sometimes accompanied by thin, highly fossiliferous shale or limestone beds reflecting the reduced sedimentation rates. In the upper delta front, the abandonment facies comprises a thin unit of intensely bioturbated, quartz-rich sandstones interpreted as the transgressed remnants of Chandeleur-type barrier islands. These sandstones are often calcite-cemented due to the dissolution of shell fragments, and may also contain glauconite. The units vary in thickness from 0.5 to 3.0 m, or may consist merely of a sharp erosional surface overlain by a thin, coarse-grained transgressive lag, akin to the shoreface erosion surface discussed in Sect. 7.4. In the mid to lower delta plain, a thin unit of mudstones and siltstones with an abundant marine to brackish water fauna is deposited, whilst the upper delta plain becomes an extensive emergent area subject to palaeosol and other surface processes dictated by the prevailing climate. In humid-tropical settings laterally extensive lignite or coal beds blanket the entire upper delta plain and produce correlative beds which transgress all underlying facies variations and may extend over several thousand square kilometres (Fisher and McGowen, 1967; Elliott, 1974a; Tewalt, Bauer & Mathew, 1981; Flores and Tur, 1982). In the Tertiary Lower Wilcox Group of the Gulf Coast, USA the recognition and correlation of abandonment facies marker beds enables major delta complexes and lobes to be mapped (Fisher and McGowen, 1967).

### DEEP-WATER, FLUVIAL-DOMINATED DELTAS

These deltas are characterized by large-scale delta front coarsening-upwards sequences with thick prodelta muds reflecting progradation into relatively deep water. In several cases mouth bar sandstones are laterally discontinuous, implying that they were localized around widely spaced distributary channels as in the modern, birdfoot Mississippi delta. However, not all deep-water deltas have this configuration as the Upper Wilcox Group of the Gulf Coast, USA includes lobate, deep-water deltas, produced by partial wave reworking of sands from the distributary mouths (Edwards, 1981).

The *Bideford Group* deltas in the Westphalian of north Devon have also been interpreted as fluvial-dominated, elongate deltas, but this is inferred solely from vertical sequences (de Raaf, Reading and Walker, 1965; Elliott, 1976a; see also Sect. 2.1.2, Figs 2.1, 2.2). The succession comprises nine coarsening-upwards cycles, each representing the progradation of a delta into a moderately deep basin. Thick progradational facies can be distinguished from thin abandonment facies horizons, and delta front and delta plain facies can be differentiated in the progradational facies (Fig. 6.34). The delta front is represented by large-scale (50–100 m) coarsening-upwards sequences and two distributary channel sandstones (20–26 m thick). The delta plain facies includes small-scale interdistributary bay-fill coarsening-upwards sequences and small-scale crevasse channels. Abandonment facies are represented by thin horizons of bioturbated siltstone, and, in one case, a thick horizon of impure coal. Comparison of the cycles reveals that: (1) the upper sandstone member is frequently absent; (2) abandonment facies only occur where there is a substantial sandstone member at the top of the cycle; and (3) delta plain sequences are preferentially developed above mudstone-siltstone dominated delta front sequences devoid of a significant sandstone member. These contrasts are explicable in terms of differing locations in an elongate, birdfoot delta. The frequent absence of the upper sandstone member suggests that bar finger sands were impersistently developed in individual deltas. Lateral margins of the bar fingers provided shallow platforms on which delta plain facies were deposited, thus explaining the preferential development of delta plain facies. After abandonment the bar fingers were shallow, elevated areas where deposition was slow and abandonment facies accumulated. The adjacent mud-silt dominated

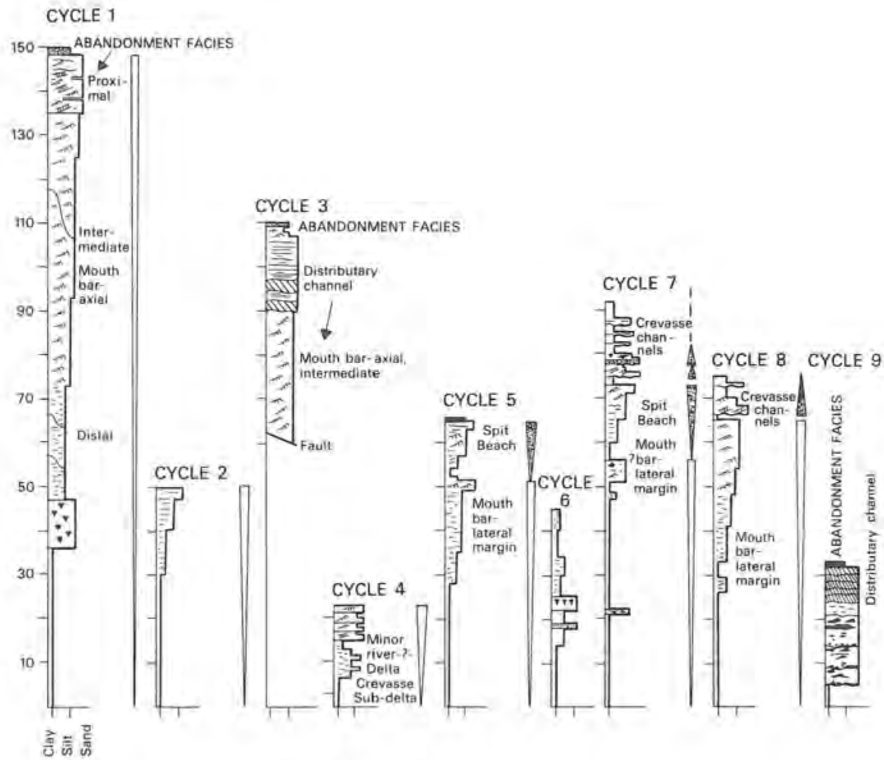
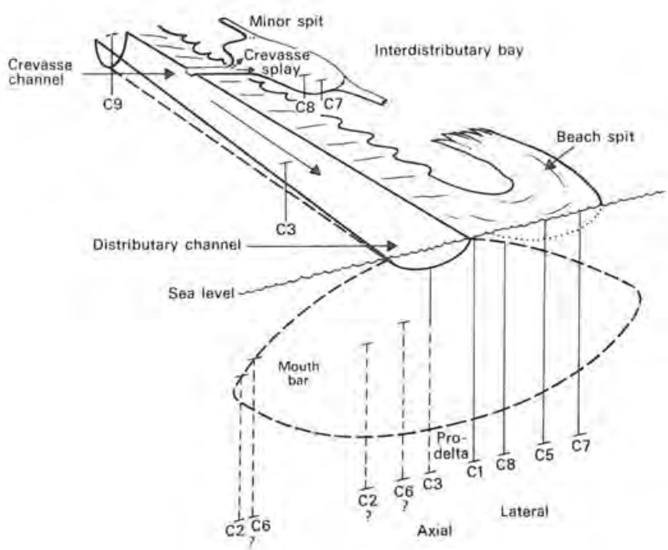



**Fig. 6.34.** Interpretation of Carboniferous cycles in the Bideford Group, north Devon, in terms of differing positions in a bar finger system of a fluvial-dominated delta (after Elliott, 1976a).



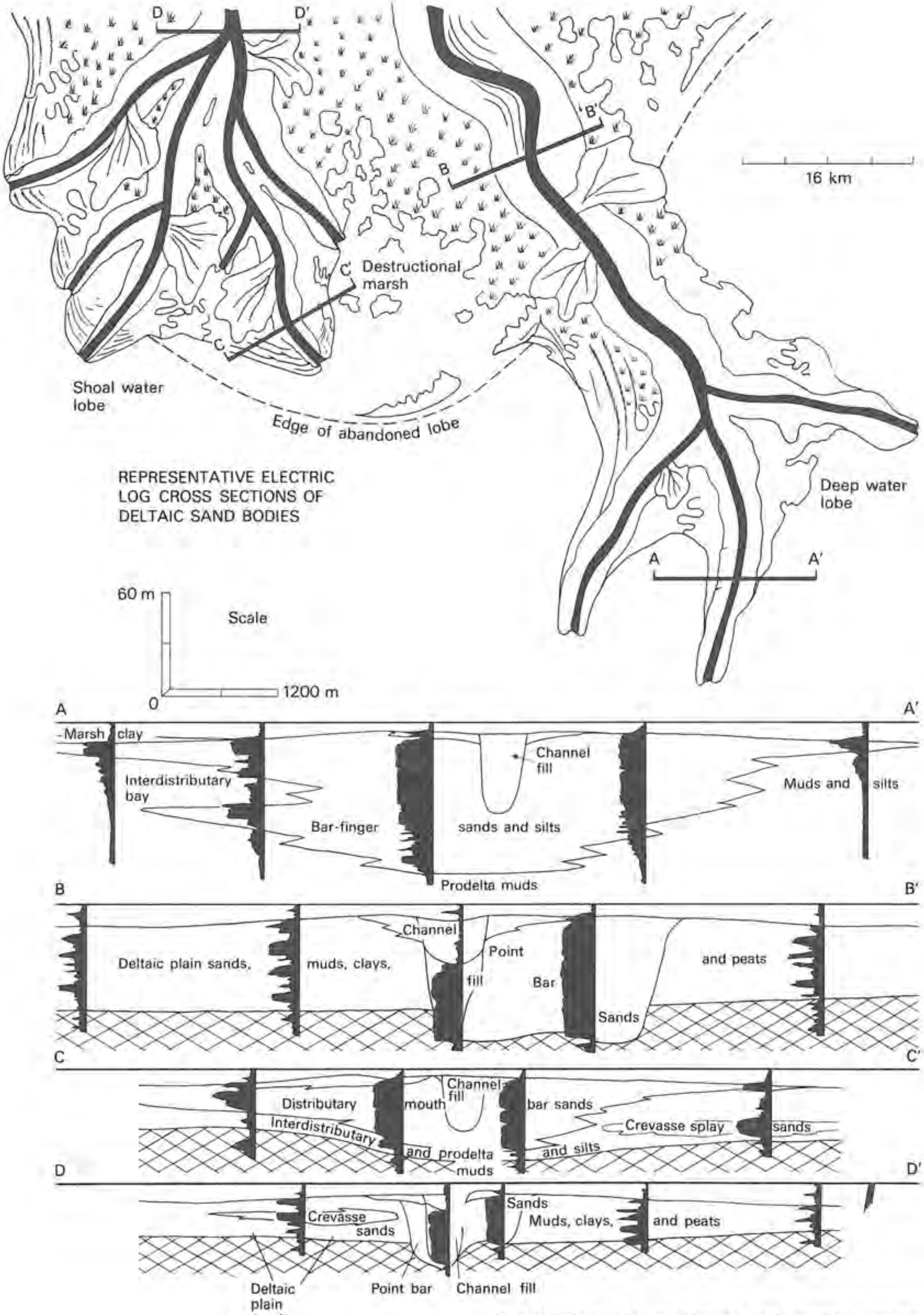



**Fig. 6.35.** Electric log profiles of deltaic sandstone bodies based on the Tertiary fluvial-dominated Holly Springs delta system in the Gulf Coast, USA (after Galloway, 1968).

areas subsided more rapidly and experienced only a brief cessation in deposition.

Numerous birdfoot and slightly lobate deep-water deltas have been recognized in sub-surface studies of the Tertiary of the Gulf Coast, USA using vertical sequence and sandstone isopach data derived from self-potential and resistivity logs (Fisher and McGowen, 1967; Galloway, 1968; Fisher, 1969; Edwards, 1981; Fig. 6.35). In cases where these deltas prograded onto the shelf edge they were substantially influenced by large-scale growth faulting during sedimentation (Edwards, 1981; Winker, 1982; Sect. 6.8). These 'shelf-margin' deltas thicken dramatically in the vicinity of the growth faults and may also be vertically stacked in rapidly subsiding depocentres defined by the growth faults.

### SHALLOW WATER, FLUVIAL-DOMINATED DELTAS

These deltas are characterized by thinner delta front sequences with a higher sandstone to shale ratio and generally have a lobate form, with a continuous delta front sandstone. In some cases, these delta front sandstones are composed of coalesced mouth bar deposits implying the presence of numerous, closely spaced distributary channels which may have resulted from the prevalence of friction-dominated outflow with channels repeatedly bifurcating around middle ground bars in the channel mouth. The 'shoal-water' lobes of the Mississippi delta which prograded across the shallow shelf prior to the modern birdfoot delta appear to be of this type (Frazier, 1967). Numerous distributary channels traverse the delta plain, frequently bifurcating towards the shoreline. Lateral coalescing of the mouth bars associated with these channels is considered to account for the continuity of the sheet sand, although wave processes may have contributed to this as the recently abandoned Lafourche delta exhibits clusters of beach ridges adjacent to distributary mouths. Delta front sheet sands can be traced laterally for more than 800 km$^2$ and in vertical section are represented by 10–30 m coarsening-upwards sequences which are often deeply eroded by fluvial-distributary channel sands (Fisk, 1955; Fig. 6.36).

Analogous delta systems have been described in the Tertiary subsurface of the Gulf Coast, USA (Fisher, 1969; Fisher, Brown *et al.*, 1969). Delta front sequences comprise prograding mouth bar sequences and subordinate wave-dominated, beach face coarsening-upwards sequences which can be distinguished in self-potential and resistivity logs. The delta plain facies association is dominated by multi-storey channel sandstone bodies up to 60 m thick, separated by lignite-bearing, interdistributary muds-silts with crevasse splay and levee facies. Growth faults may occur in these shallow water deltas, but they are not common.

In the *Yoredale Series delta system* in northern England three shallow water delta lobes separated by inter-deltaic clastic embayments can be recognized (Elliott, 1974a, 1975; Fig. 6.37). Lobes are distinguished by tracing abandonment facies marker horizons and one lobe extends over 700 km$^2$ from upper delta plain to delta front. The delta plain of this lobe is dominated by two small-scale (4.5 m) coarsening-upwards sequences which reflect the repeated infilling of shallow interdistributary bays by levees, crevasse splay lobes and minor beach spits. Small-scale crevasse channels and larger-scale distributary channels locally dissect these sequences, and one of the distributary channels is represented by a 1.5 km wide sandstone body with large-scale

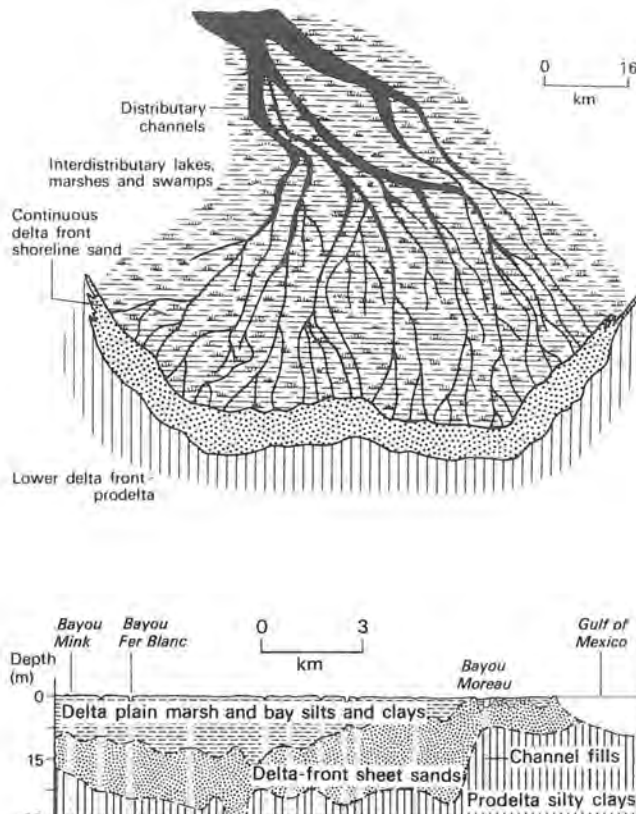



**Fig. 6.36.** Reconstruction of the abandoned 'shoal water' Lafourche lobe of the pre-modern Mississippi delta and cross-section through the lobe (after Fisher, Brown *et al.* 1969 and Gould, 1970).

lateral accretion surfaces produced by the lateral migration of point bars. The delta front is represented by a single 9–17 m coarsening-upwards sequence which records the progradation of a fluvial-dominated, mouth bar shoreline. Downcurrent changes in the progradational phase are paralleled by facies changes in the abandonment phase marker horizon from coal to marine sandstones and limestones (Fig. 6.37).

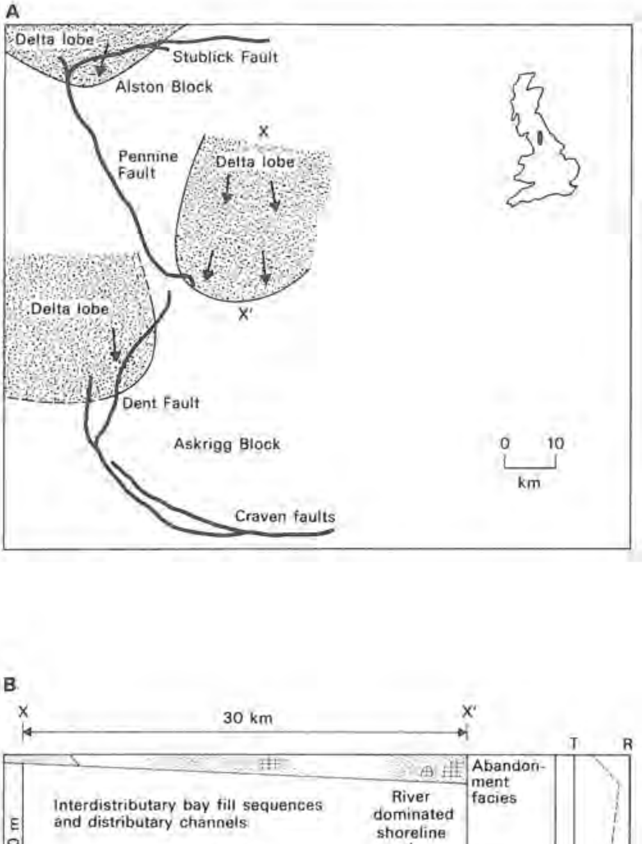



**Fig. 6.37.** Plan view and cross-section of fluvial-dominated delta lobes from a Carboniferous Yoredale cyclothem in northern England (after Elliott, 1975).

### FLUVIAL-DOMINATED DELTAS INFLUENCED BY TURBIDITY CURRENT PROCESSES

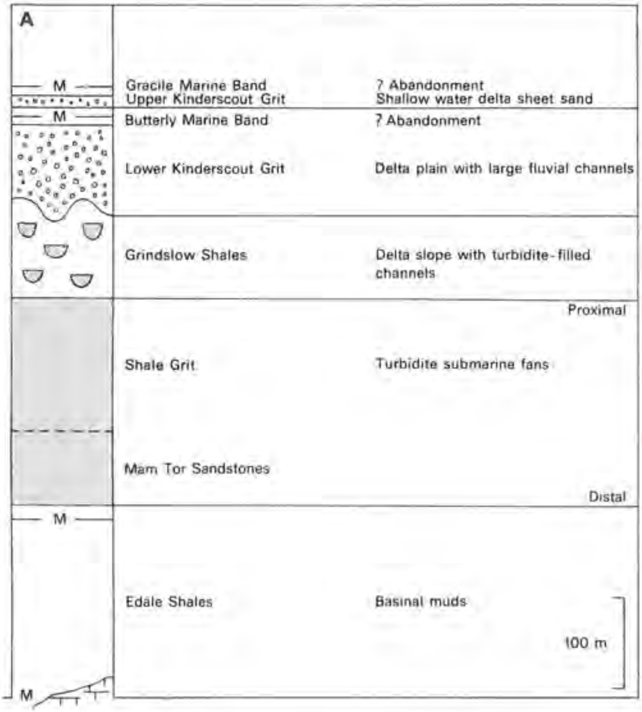
The lower parts of many delta front sequences, particularly those of deep-water deltas, include thin turbidite beds generated either directly from the distributary mouth, or by slump events on the upper delta front (de Raaf, Reading and Walker, 1965; McBride, Weidie and Wolleben, 1975). Turbidites are particularly common where synsedimentary slumping and faulting were common on the delta front. For example, in a delta system in the Cretaceous Recôncavo basin, Brazil, the delta front facies is dominated by a wide range of slump and sediment gravity flow facies induced by growth faulting and oversteepening of the delta front by diapirism (Klein, De Melo and Della Favera, 1972). Sand- and gravel-dominated 'short-headed stream deltas' or fan deltas (Sect. 12.4.3) also commonly exhibit abundant evidence of turbidity current deposition on the delta front (e.g. McBride, 1970a; Flores, 1975).

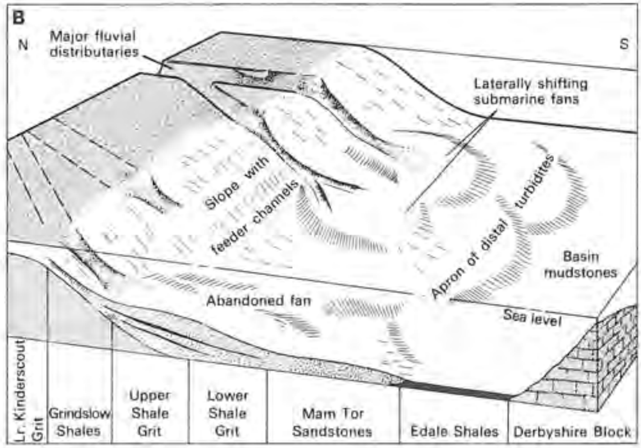
Some ancient deltas have thick accumulations of turbidites interpreted as submarine fans underlying the delta front sequence (Walker, 1966; Galloway and Brown, 1973). The Namurian *Kinderscout Grit delta system* in northern England exhibits a passage from basinal muds into a distal turbidite apron, a submarine fan complex (Sect. 12.5.6), a delta front slope and finally the delta plain, with a total thickness of 700 m (Fig. 6.38; Reading, 1964; Walker, 1966; Collinson, 1969; McCabe, 1977, 1978). The delta front slope is represented by a 100 m coarsening-upwards sequence which is dominated by mudstones and siltstones but includes steep-sided, turbidity current channels at various levels and also displays numerous slump scars and slump units (Fig. 6.38C). The delta plain facies association includes bay mudstones and siltstones, crevasse splay sandstones and small-scale (4 m) minor mouth bar-crevasse channel couplets, but it is dominated by fluvial-distributary channels up to 40 m deep and 0.5–1.0 km wide filled by coarse-grained sandstone with giant cross-bed sets up to 40 m thick interpreted as large-scale alternate bars in a low-sinuosity channel (Sect. 3.9.4; Fig. 3.52). These major Brahmaputra-like rivers poured large volumes of water into a small, confined basin and may therefore have lowered its salinity. As a result, the delta front received relatively fine-grained sediment from suspension, whilst sand largely by-passed the delta front and was discharged directly down the delta front in subaqueous channels. Turbidity currents generated in these channels supplied sediment directly to the submarine fan at the foot of the delta front.

The *Roaches Grit delta system* in the same basin is very similar. It differs in that the delta front is dominated by ripple laminated turbidites rather than muds and silts, and synsedimentary faulting in the upper delta front is considered important in the generation of the turbidity currents (Jones, 1980).

### 6.7.2 **Ancient wave-dominated deltas**

Wave-dominated deltas are well known in the Cretaceous epeiric seaway of North America and the subsurface Cretaceous-Tertiary of the Gulf Coast, USA (Hubert, Butera and Rice, 1972; Hamblin and Walker, 1979; Balsley, 1980; Weise, 1980; Leckie and Walker, 1982). However, they have not been widely recognized and some sequences interpreted as prograd-



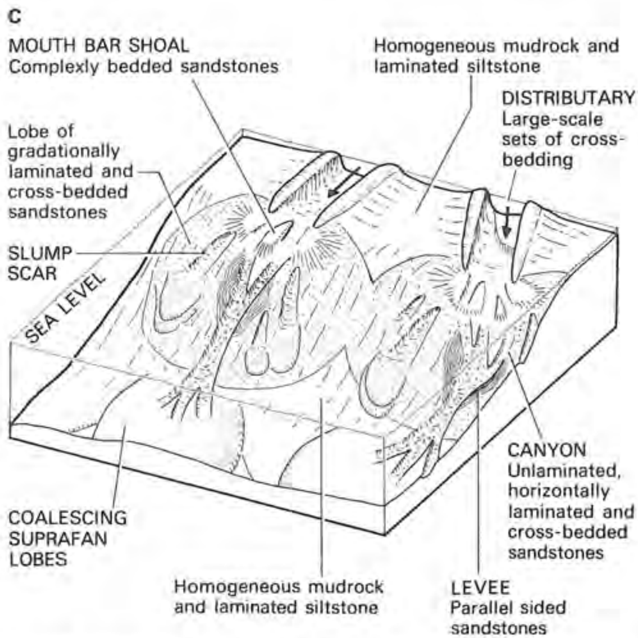




**Fig. 6.38.** Kinderscout submarine fan/delta system, Upper Carboniferous, northern England. A and B, Vertical succession and generalized interpretation (after Reading, 1964; Walker, 1966; Collinson, 1969); C, detailed model of delta slope, mouth bars, distributaries and canyons feeding the underlying submarine fan (after McCabe, 1978).

ing beaches or barrier islands may be in fact part of wave-dominated delta systems (Sect. 7.2).

The delta plain facies of wave-dominated deltas comprise fluvial-distributary channel sandstones and interdistributary bay facies, whereas the delta front is represented by coarsening-upwards sequences which resemble those of prograding beach fronts (Sect. 7.2). The fine member at the base of the delta front sequence may include storm-generated turbidites and the mid- to upper parts of the sequence are dominated by well-sorted sandstones which exhibit hummocky-cross-stratification, wave-produced ripple lamination, flat lamination and cross-bedding. Bioturbation varies in intensity, but can be appreciable and results in extensive disruption of the laminations. Sequences are eroded by fluvial-distributary channel sandstones, at least locally, and upper beach face sandstone facies may be better developed on the immediate flanks of these channels implying clustering of beach ridges around the channels as in modern wave-dominated deltas.

Subsurface studies in the Gulf Coast have recognized wave-dominated deltas using geophysical logs and, more particularly, sand isopach data. In some cases strike-aligned sandbodies of beach face and distributary mouth bar origin can be linked

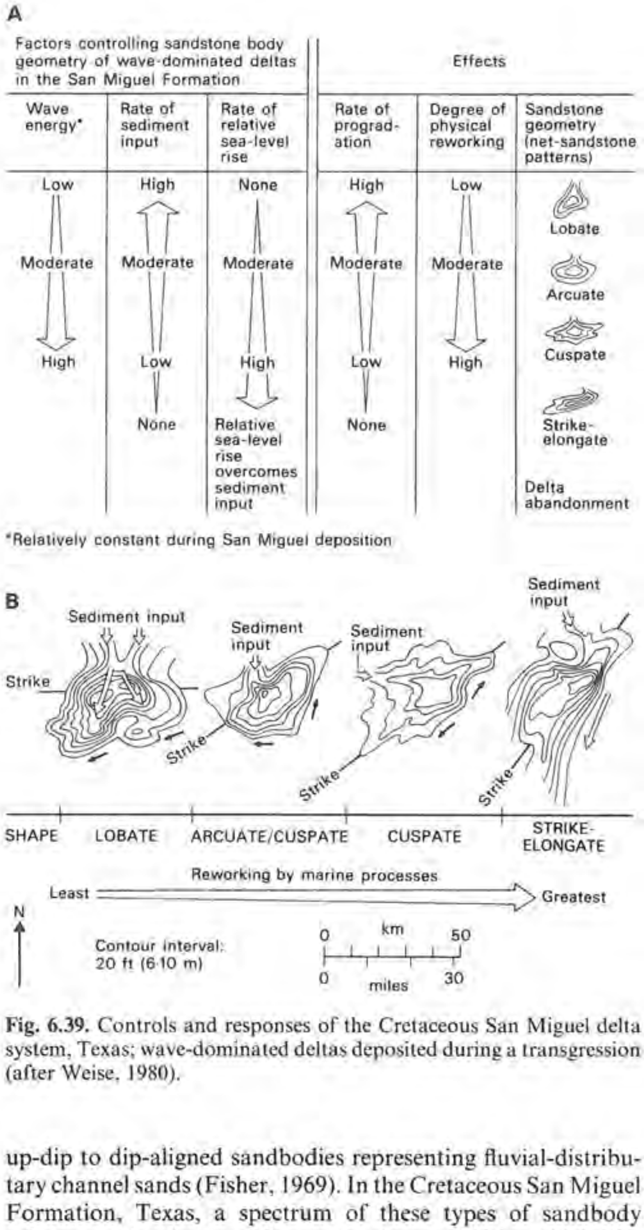



**Fig. 6.39.** Controls and responses of the Cretaceous San Miguel delta system, Texas; wave-dominated deltas deposited during a transgression (after Weise, 1980).

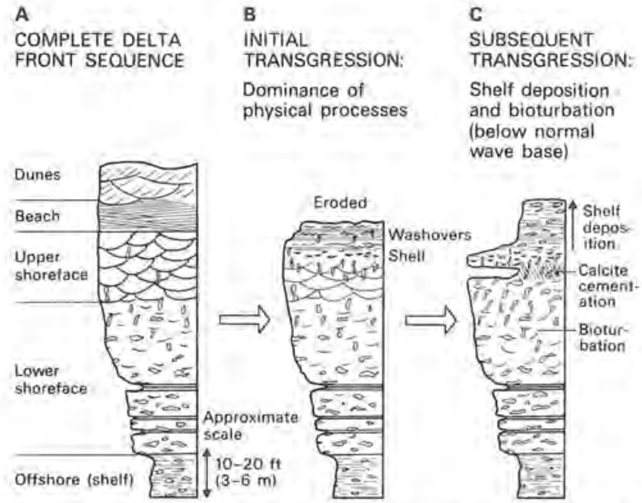



**Fig. 6.40.** Development of delta front sequences in the Cretaceous wave-dominated San Miguel delta system, Texas; (A) complete delta front progradation sequence, (B) modification of the sequence during the ensuing transgression, (C) the final, preserved sequence (after Weise, 1980). The profiles of these graphic logs reflect SP response and not grain size *per se*.

up-dip to dip-aligned sandbodies representing fluvial-distributary channel sands (Fisher, 1969). In the Cretaceous San Miguel Formation, Texas, a spectrum of these types of sandbody reflects varying degrees of wave-dominance (Weise, 1980; Fig. 6.39). This formation accumulated during a transgression and was deposited by a succession of wave-dominated deltas which prograded intermittently during periods of appreciable sediment supply. As a result of the transgression, successive deltas occur progressively landward and project less into the basin. The formation of a delta and its characteristics were a function of: (1) sediment supply which was variable in response to hinterland tectonics; (2) wave energy which was relatively constant in an absolute sense, but varied in effectiveness; and (3) the rate of sea-level rise. During periods of high sediment input and a low rate of sea-level rise the effectiveness of waves to redistribute sediment was limited, and lobate and wave-dominated deltas resulted. In contrast, during periods of low sediment input and high rates of sea-level rise, wave reworking was extensive and the delta was characterized by an elongate, strike-aligned sandstone body. Wave reworking after abandonment of the delta was appreciable with much of the upper shoreface, foreshore and delta front facies being reworked. This resulted in a predominance of incomplete, attenuated delta front sequences (Fig. 6.40), with only local preservation of fluvial distributary channel sandstones.

The Middle Jurassic *Brent Sand delta system* in the North Sea (Figs 14.15, 14.16) is, in part, interpreted as a wave-dominated delta (Eynon, 1981; Parry, Whitley and Simpson, 1981; Johnson and Stewart, 1985). In general, the Brent Sand represents a northerly prograding, wave-dominated sandy coastline which includes a wave-dominated delta developed in a rift-graben (the Viking Graben) and flanked by barrier island-lagoonal shorelines in areas adjacent to the graben (e.g. Budding and Inglin, 1981). A large scale, coarsening-upwards sequence dominated by wave-produced structures represents the delta front and in the graben this sequence is usually cut by composite, multistorey channel sandstones representing the distributary channels.

### 6.7.3 Ancient tide-dominated deltas

Ancient tide-dominated or tide-influenced deltas have been only sparsely recognized in the geological record so far. Recognition depends on the character of the mid to upper part of the delta front facies and the lower delta plain facies as tidal effects are most pronounced in these sub-environments. In the delta front, gradational coarsening-upwards sequences of tidally influenced facies may result from the progradation of ebb-tidal deltas or tidal sand ridges, but erosive-based tidal channel or inlet sequences will also be common. The lower delta plain is likely to comprise tidal flat sequences, small-scale channels produced by tidal creek systems, and larger-scale tidal-distributary channel sequences.

A small-scale wave- and tide-influenced delta, possibly analogous to the Niger delta, has been recognized in the Middle Jurassic Cloughton Formation of Yorkshire, England (Livera and Leeder, 1981). A wave-influenced though intensely bioturbated coarsening-upwards delta front sequence is cut locally by a unit of thinly bedded, bioturbated sandstones and siltstones with pronounced dipping surfaces interpreted as lateral accretion surfaces. This unit is considered to reflect a laterally migrating tidal channel of the lower delta plain which is eroded into upper delta front beach facies. Both the beach and tidal channel facies are then erosively overlain by a major fluvial channel sandstone, also with lateral accretion surfaces, interpreted as the upper delta plain distributary channel.

Tidally influenced channel sequences have been inferred from bimodal current patterns in Cretaceous delta plain associations in the Western Interior, USA (Hubert, Butera and Rice, 1972; van de Graaff, 1972), and from the abundance of clay partings in channel sandstones of the subsurface Tertiary of the Niger delta (Weber, 1971).

## 6.8 SEDIMENT-INDUCED DEFORMATION

The theme of the chapter so far has been that deltaic facies patterns are controlled largely by depositional processes operating in the delta, but several studies have demonstrated that the facies pattern of a delta can be significantly influenced by synsedimentary deformation operating on a wide range of scales. Deformation may be related to basement tectonics, as in the Ganges-Brahmaputra delta, but there is also a discrete class of deformational processes related solely to sedimentary factors. These factors stem largely from the instability of muds deposited rapidly on the lower to mid-delta front and continental slope, and their continuing instability during early burial. In the Mississippi delta these processes have produced a diverse range of deformational features including mud diapirs, rotational slumps, delta front gullies, surface mudflows and deep-seated faults (Figs 6.41 and 6.42). It is estimated that 40% of the sediment supplied to this delta is involved in some kind of mass movement after initial deposition (Coleman, 1981). In view of this, the validity of the orderly vertical sequence and sandbody patterns described from this delta should perhaps be questioned (Sect. 6.5.2; Figs 6.17 and 6.18). Similar synsedimentary deformational features have been described from the Fraser, Magdalena, Mackenzie, Niger and Orinoco deltas, and large-scale, synsedimentary growth faults have been described from numerous sub-surface deltaic successions. Thus, a full understanding of the facies patterns of deltas, particularly those with a high mud content, requires awareness of the full range of synsedimentary deformational processes and features.

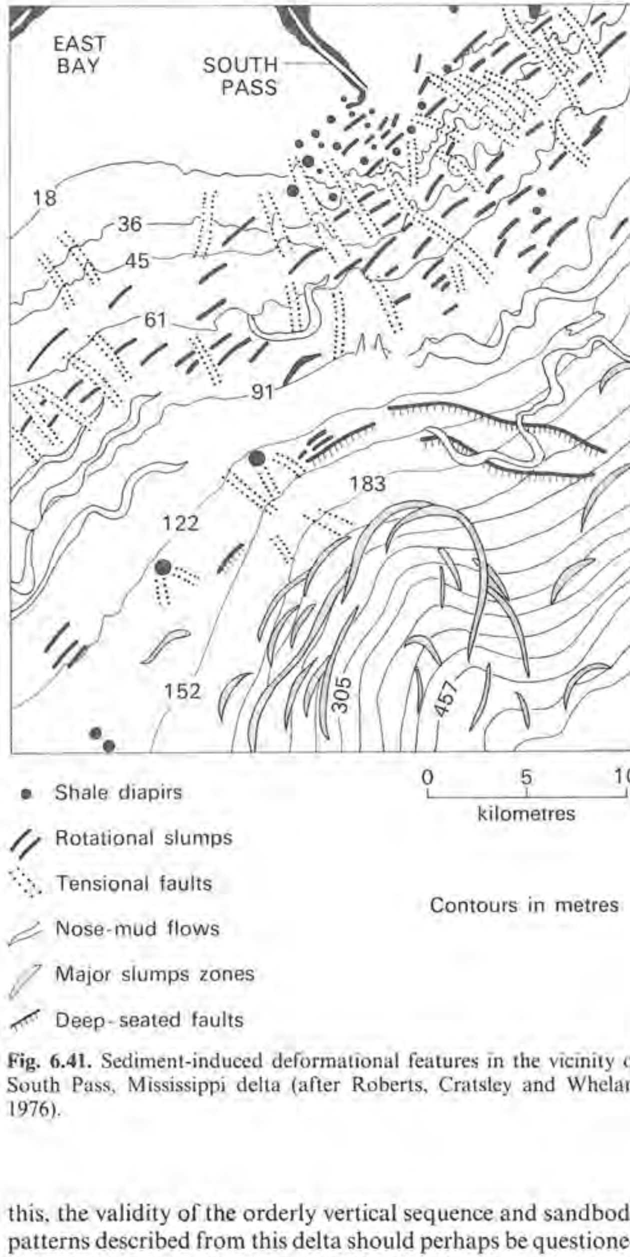



**Fig. 6.41.** Sediment-induced deformational features in the vicinity of South Pass, Mississippi delta (after Roberts, Cratsley and Whelan, 1976).

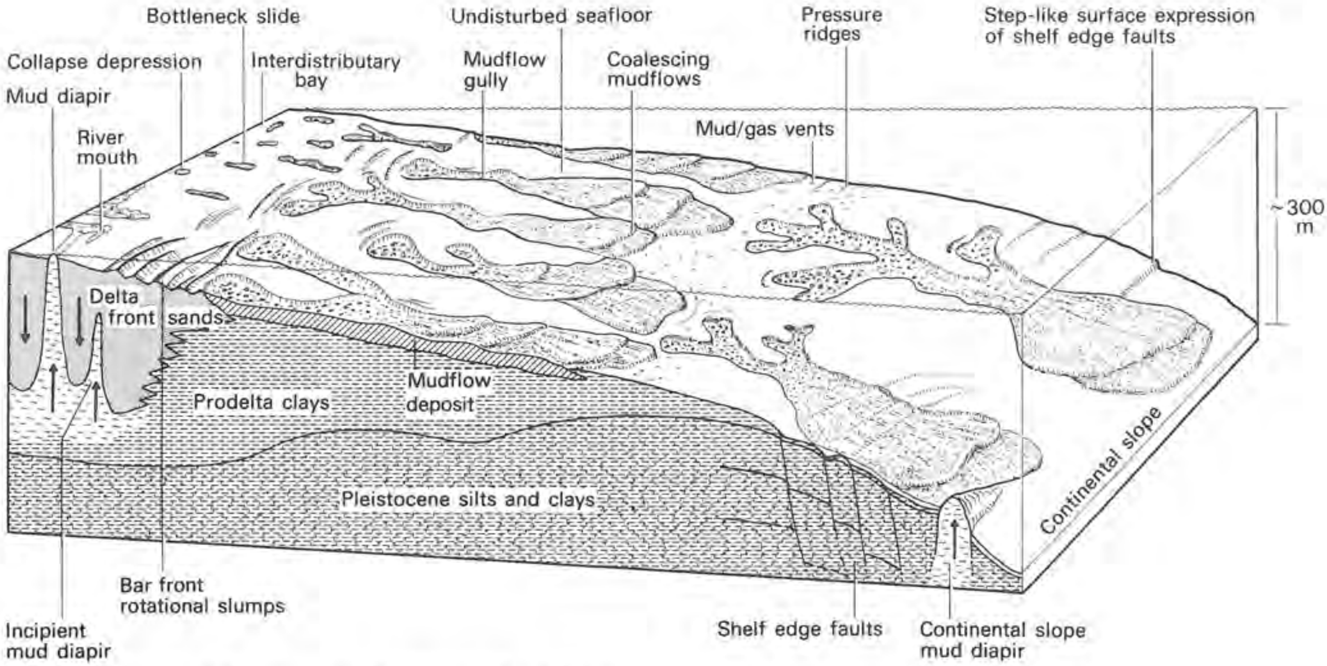



**Fig. 6.42.** Summary of the main types of sediment-induced deformational features arising from surface instability of sediments and deep-seated flowage of overpressured clays in the Mississippi delta (after Coleman, 1981).

### 6.8.1 Deformational processes

Synsedimentary deformation principally affects the delta front which dips gently seawards at an angle of 0.2°–2° on average. Surface instability of these slopes occurs in response to over-steepening and loading of the slope produced by the higher sedimentation rates which characterize the upper part of the slope (Coleman, Suhayda *et al.*, 1974). Mass movement often results from this surface instability, aided by wave pounding during storms which induces short period and often intense fluctuations in bottom pressures which can temporarily exceed sediment shear strength and initiate downslope movement of sediment. The effects of wave pounding are enhanced if high concentrations of methane gas exist in the sediments due to bacterial decomposition of organic matter. The presence of gas reduces the shear strength of the sediment and may also cause the sediments to degassify and undergo a brief, liquefied or fluidized phase when subjected to high bottom pressures. Surface slumps and mudflows may form in areas with minimal slopes as a result of these processes (Whelan, Coleman *et al.*, 1976).

An additional feature of delta front or prodelta muds is that they often possess high pore pressure and low compaction values which cause them to be extremely unstable in the sub-surface as burial and loading continues. In the initial stages of compaction, water is easily expelled but as compaction continues the rate of water expulsion decreases as permeability falls. Eventually the orderly expulsion of water is prohibited and high pore-water pressures develop in the clays. At this point, compaction ceases and the clays are then overpressured and undercompacted. Methane gas may also contribute to the development of overpressured conditions (Hedberg, 1974). In addition to being out of pressure equilibrium with their surroundings, overpressured clays have low viscosity and sediment strength and are therefore unstable and potentially mobile when loaded. Originally it was felt that these conditions were generated only at considerable burial depths, but *in situ* measurements of pore pressure in the upper Mississippi delta have revealed that high pore pressures occur within 15 m of the present sediment surface (Bennett, 1977). Thus, overpressured/undercompacted conditions commence early in the burial history of the muds and develop as burial continues. This leads to pronounced *sub-surface* instability of the muds which can result in a slow, but continuous, deep-seated flowage of clays away from the depocentre and into the basin. In the Mississippi depocentre, for example, deeply buried Pleistocene muds have flowed basinwards in response to loading by recent deposits of the Mississippi delta. This deep-seated flowage of overpressured clays is a prime mechanism of subsidence in the depocentre and also produces a wide range of intermediate to large-scale deformational features such as mud diapirs and growth faults. A similar situation exists in the Niger delta depocentre where the Akata shales are migrating basinwards by deep-seated clay flowage, creating mud ridges in the offshore area and assisting in the formation of growth faults in the depocentre by creating a tensional regime (Evamy, Haremboure *et al.*, 1978; Fig. 14.11).

### 6.8.2 **Deformational features**

#### ROTATIONAL SLUMPS

For a long time it was assumed that the bar front areas of distributary mouth bars in the Mississippi delta possessed smooth seaward-dipping profiles, but fathometer studies off South pass reveal frequent abrupt 'stairstep' changes in slope (Coleman, Suhayda *et al.*, 1974; Coleman, 1981; Fig. 6.43). Seismic studies indicate that these irregularities are a surface expression of fault or slump planes along which large blocks of sediment are translated downslope. The slump planes strike across the mid to upper bar front and initially dip seawards at gentle angles of 1°–4° before flattening into slope-parallel shear planes. Individual slump blocks average 90 m in width, 6 km in length and move downslope for distances in excess of 1.5 km. They are preserved intact and do not exhibit flowage structures, producing seemingly anomalous shallow water sand facies in deeper water mud-silt facies. The blocks may, however, dip landwards by up to 30°. This rotational slumping and downslope translation of slump blocks is an integral part of progradation in the Mississippi delta and may make a substantial contribution to the final, preserved facies pattern.

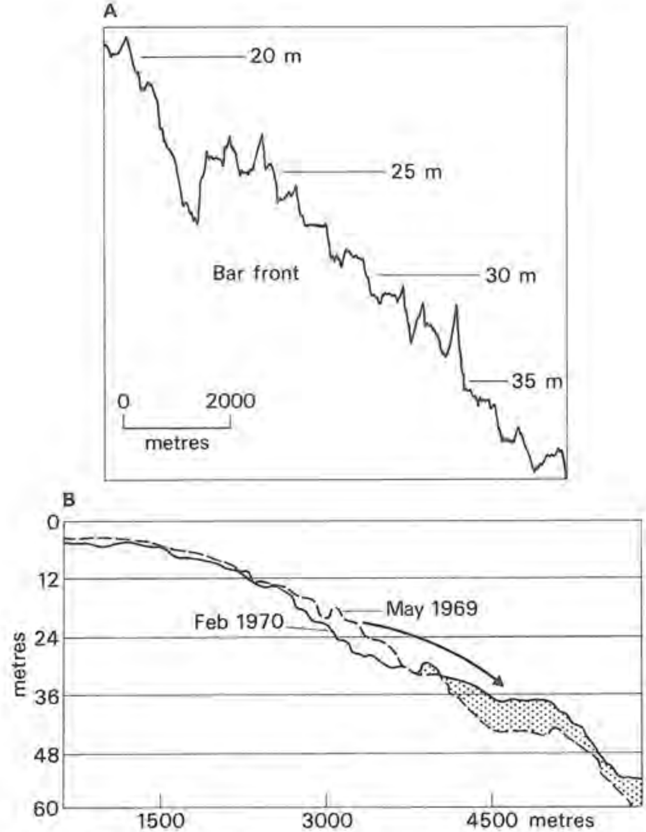



**Fig. 6.43.** Rotational slumps on the delta front of the modern Mississippi delta. (A) Irregular profile of a river mouth bar front reflecting the presence of rotational slump planes; (B) offshore slumping associated with the slump planes, revealed by time-separated fathometer profiles (after Coleman, Suhayda *et al.* 1974).

#### COLLAPSE DEPRESSIONS, DELTA FRONT GULLIES AND MUDFLOWS

Collapse depressions are bowl-shaped depressions 100 m or so in diameter and 1–3 m deep which occur in the distal interdistributary bay area and are formed by localized liquefaction/fluidization of sediment by storm waves (Coleman and Garrison, 1977; Prior and Coleman, 1978; Coleman, 1981; Roberts, 1980; Fig. 6.44). In some cases the depressions are closed, circular features rimmed by small, listric fault scarps and with a chaotic mass of isolated blocks of sediment in the central area. More commonly, however, the depressions are open at their downslope side and pass into 'bottleneck slides' or 'delta front gullies' which were originally described by Shepard (1955). These gullies trend down the delta front as long, slightly sinuous features bounded by sharp, rotationally slumped walls. They extend over several kilometres from shallow water depths (7–10 m) down to 100 m in depth in some cases, and are 3–20 m deep. The origin of these graben-like gullies is not clear, but they act as conduits for mudflows or debris flows which emanate from the collapse depressions at the head of the gullies. At the mouths of the gullies, mudflows emerge onto the prodelta surface and produce a virtually continuous fringe of mudflow lobes around the lower prodelta (Fig. 6.44). Individual lobes are 10–15 m thick and contain erratic blocks of sediment which are commonly 30 m in diameter and can be substantially larger. Once again, downslope transference of sediments by mudflows is a frequently recurring event and is therefore part of the normal sedimentation of the Mississippi delta.

#### DIAPIRS AND SHALE RIDGES

Diapirism has been well documented in the Mississippi delta as 'mudlumps' frequently emerge near the distributary mouths and form temporary islands (Morgan, 1961; Morgan, Coleman and

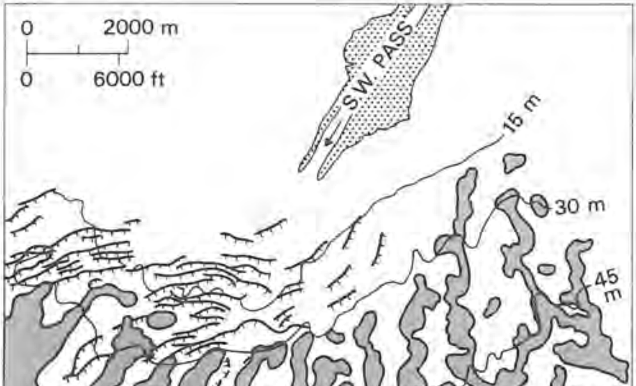



**Fig. 6.44.** Map of part of the delta front of the modern Mississippi delta illustrating the distribution of near-surface synsedimentary deformational features (after Suhayda and Prior, 1978).

Gagliano, 1968). Surface exposures of the mudlumps reveal steeply dipping, stratified delta front sediments with numerous small anticlines, *en echelon* normal faults, reverse faults, radial faults and thrusts. Other features include small mud cones formed by extrusion of methane-rich muds from fault planes, and planation horizons produced by wave erosion of the exposed mudlump. Clays involved in the diapirism exhibit intense brecciation and later fractures (Morgan, Coleman and Gagliano, 1963; Coleman, Suhayda *et al.*, 1974).

The mudlumps are considered to be thin spines superimposed on linear shale folds or ridges, with large-scale, high-angle reverse faults in the mudlump crests producing most of the uplift. Up to 200 m of uplift can be demonstrated in some cases, and rates of 100 m uplift in 20 years have been documented. There is a close relationship between diapiric activity and distributary mouth bar sedimentation with the appearance of mudlumps invariably coinciding with rapid sedimentation during river flood periods, and the site of mudlump activity migrating seawards in concert with mouth bar progradation (Morgan, Coleman and Gagliano, 1968). Isopach maps reveal that these strongly diapiric mudlumps have substantially modified the bar finger sands. Instead of being linear bodies with a uniform thickness of approximately 70 m as originally described (Fisk, 1961), they comprise a series of discrete sand pods up to 100 m in thickness separated by areas of minimal sand thickness (Coleman, Suhayda *et al.*, 1974; Fig. 6.45).

All other examples of diapirs or shale ridges in the Fraser, Magdalena, Niger and Orinoco deltas are submerged and occur considerable distances in front of the delta (Nota, 1958; Mathews and Shepard, 1962; Shepard, Dill and Heezen, 1968; Shepard, 1973a; Weber and Daukoru, 1975). In the Niger delta, for example, shale ridges occur along the continental slope in front of the extensively growth-faulted depocentre. These offshore shale ridges are probably a surface expression of overpressured clays which flowed away from the depocentre; in which case the Mississippi mudlumps may reflect the re-activation of pre-existing offshore shale ridges by direct *in situ* loading as the delta progrades onto the continental slope.

### SHELF EDGE SLUMPS AND FAULTS

Large, arcuate slump scars and faults are common at the shelf-slope edge on to which the prodelta of the Mississippi delta is currently prograding (Coleman and Garrison, 1977; Coleman, 1981; Figs 6.41 and 6.42). These features form scarps on the sea floor up to 30 m high and are preferred sites of deposition on the downthrown side of these faults. Some examples are large-scale slump scars which are being passively infilled by mudflow deposits, but others are faults which show evidence of growth during sedimentation. These faults affect 700–800 m of sediment and exhibit increasing fault throw with depth from 5–10 m near the surface to 70–80 m at depth. Deposition on the downthrown side is by mudflows which often thicken across the fault and, to a lesser extent, by slump sheets initiated by upper delta front rotational slumps. In some ways these faults resemble growth faults (see below), but they are developed in a prodelta setting and do not, as yet, involve *in situ* deposition of upper delta front sands.

### GROWTH FAULTS

These are a discrete class of synsedimentary faults produced by processes which operate within the sediment pile as rapidly deposited muds are buried and develop overpressured/undercompacted conditions. As a result, they form preferentially where prodelta muds are well developed and, more particularly, where deltas prograde over thick, mud-dominated basin slope deposits. In the Gulf of Mexico, *stable shelf deltas* contrast with

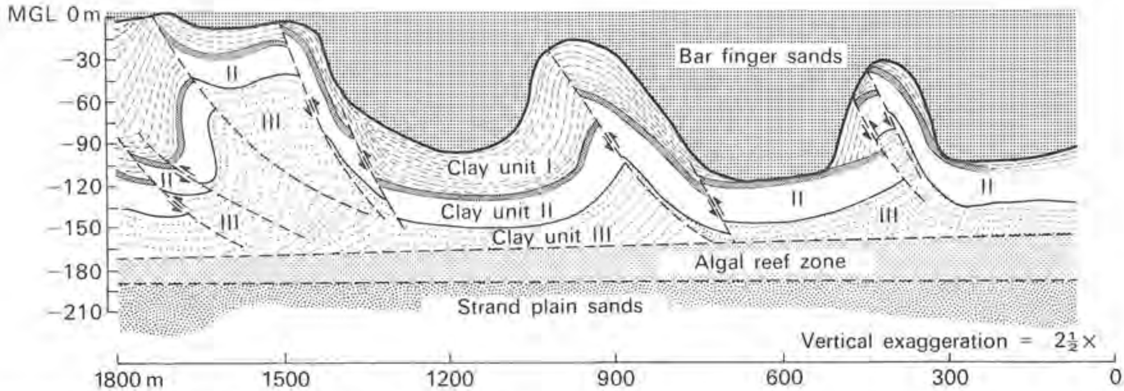



**Fig. 6.45.** Diapiric mudlumps in the modern Mississippi delta with high-angle reverse faults in the diapir crest and exceptional thicknesses of distributary mouth bar facies between the diapirs (after Morgan, Coleman and Gagliano, 1968).

*unstable shelf-edge deltas* in the extent of growth faulting which is far greater in the latter (Edwards, 1981; Winker, 1982; Winker and Edwards, 1983). Deep-seated clay flowage and gravity sliding of the continental slopes promotes extensional thinning and rapid subsidence at the shelf margin which results in the formation of large-scale listric faults. Compression at the toe of the slope initiates diapirism in shale and salt, if present, and as progradation of the slope and shelf-edge deltas continues the diapirs may be reactivated. The shelf-edge deltas are then influenced by simultaneous diapirism and growth faulting, thus increasing the structural complexity (Winker and Edwards, 1983).

Growth faults have been extensively described in sub-surface studies of deltaic successions in the Gulf of Mexico, Niger delta and Mackenzie delta (Ocamb, 1961; Weber, 1971; Evamy, Haremboure *et al.*, 1978). Their initiation, development and decay are intimately related to sedimentation in the deltaic depocentre and they exert considerable control on facies patterns. The faults parallel the shoreline, with active faults situated in the vicinity of the shoreline or shelf edge, incipient faults in front of the shoreline and a zone of decaying faults behind the shoreline. In plan view individual faults often have curved traces concave to the basin and are of limited lateral persistence, though they frequently merge to form extensive fault lines (Fig. 6.46). In cross-section the faults have a listric profile which flattens with depth, passing into bedding plane faults. The scale of the faults varies but the thickness affected by a fault is often 1–7 km and downthrow can be as much as 1 km. The faults are normal, with downthrow consistently into the basin, and an essential characteristic is that the amount of vertical displacement varies from almost zero at the top of the fault to a maximum at some mid-point in depth, and finally decreases as the fault plane flattens. The faults create preferred sites of deposition for delta front sediments, and in particular for upper delta front sands. Thicker successions with a higher sand content occur on the downthrown side, either as a single delta front coarsening-upwards sequence or in large-scale faults as a series of vertically stacked sequences. Often growth faults define preferred depositional centres within the overall depocentre to the extent that it is not possible to correlate units between adjacent fault blocks (Walters, 1959; Weber, 1971; Evamy, Haremboure *et al.*, 1978). Rotation of the downthrown strata along the curved fault planes often produces broad anticlines (rollovers), accompanied by smaller, antithetic faults. As the rollovers occur in thick, sand-dominated successions they often form excellent hydrocarbon traps (Weber, 1971; Busch, 1975). Linear ridges of overpressured shale or 'shale masses' occur at depth on the upthrown side of the faults, and are felt to be important in the formation and maintenance of the faults by some workers (Bruce, 1973; Weber and Daukoru, 1975).

The site of active growth faults migrates as the delta front progrades. New faults are initiated progressively basinward and higher in the stratigraphic succession (Fig. 6.47). The faults remain active whilst in the vicinity of the delta front and grow in response to continued sedimentation. As the delta front migrates farther into the basin faults decay and exhibit progressively smaller amounts of throw until they are eventually over-ridden by un-faulted sediments. The fault pattern and, in some cases, the style of faulting can be related to the rate of deposition or progradation (Rd) versus the rate of subsidence (Rs) (Curtis, 1971; Evamy, Haremboure *et al.*, 1978). In the Niger delta the entire depocentre is influenced by growth faults and it is possible to recognize a hierarchy of faults in terms of scale and influence on sedimentation. Important 'structure-building' growth faults define major structural units in the depocentre which appear to have independent histories of

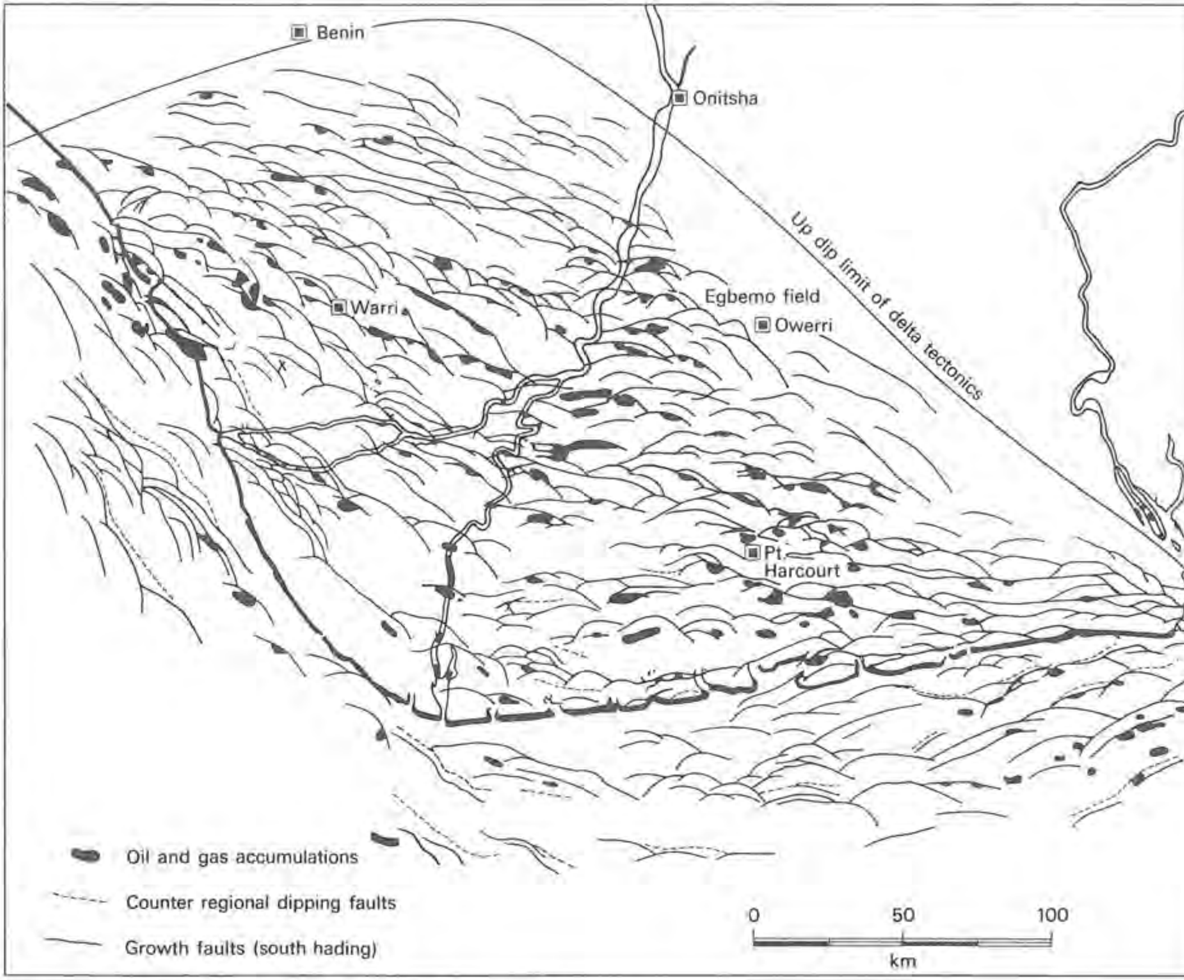
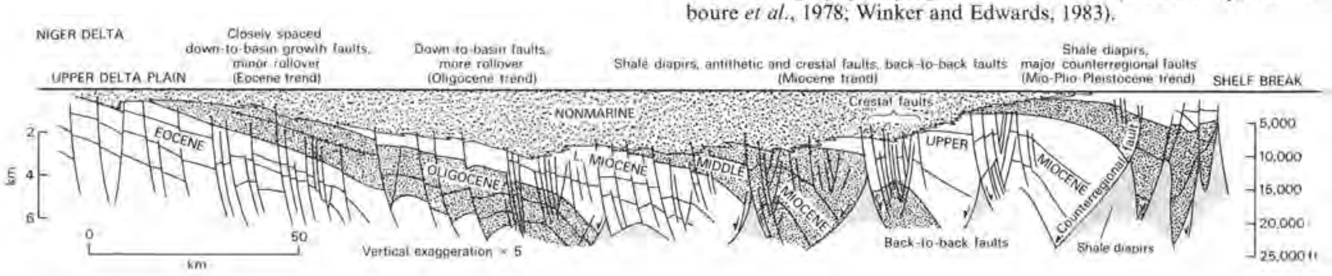



**Fig. 6.46.** Plan view of growth faults in the Niger delta illustrating their lateral impersistence, slightly curved concave-to-basin trace, and their general parallelism with the delta front (after Weber and Daukoru, 1975).

**Fig. 6.47.** Cross-section through the Niger delta depocentre illustrating extensive growth faulting occurring progressively basinward and higher in the stratigraphy as progradation continues (after Evamy, Haremboure *et al.*, 1978; Winker and Edwards, 1983).



sedimentation, faulting and hydrocarbon formation. Fluctuations in Rd:Rs are felt to produce these units (Evamy, Haremboure *et al.*, 1978).

Numerous mechanisms have been proposed to account for growth faults, many of which are now untenable. The majority view at present is that the faults mainly result from 'thin-skinned' extensional gravity sliding (Crans, Mandl and Haremboure, 1980; Mandl and Crans, 1981). The sliding is superficial and takes place along a plane which is approximately parallel to the gently dipping delta front slope due to the presence of overpressured/undercompacted conditions in the shallow subsurface. The precise depth at which sliding commences and the listric shape of the fault plane are both determined by the overpressured conditions. Bruce (1973) favours a deep-seated, 'thick-skinned' deformation mechanism which stresses the role of overpressured shale masses at depth on the upthrown side. In this mechanism, differential compaction in the depocentre produces subdued shale masses in the vicinity of the gross sand-to-mud transition. Growth faults form on the seaward side of a shale mass, defining a subsiding depocentre which is perpetuated as sedimentation continues. Both mechanisms are feasible, even within the same depocentre, and the aim should be to distinguish between faults produced by each mechanism.

### 6.8.3 Sediment-induced deformational features in exposed deltaic successions

Certain of the deformational features described above have been identified in exposed deltaic successions. Rather surprisingly, very few moderate- to large-scale slumps or diapirs have been described in ancient, exposed deltaic successions, but growth faults have been recognized in the Cretaceous of Colorado, the Triassic of Svalbard, and the Carboniferous of northern Europe (Weimer, 1973; Edwards, 1976a; Chisholm, 1977; Rider, 1978; Elliott and Ladipo, 1981). All these examples describe, or infer, listric, synsedimentary faults with a thicker, sand-dominated succession on the downthrown side. In Svalbard, a series of growth faults is exposed approximately normal to fault strike (Edwards, 1976a). The faults dip at 20°–50° in the direction of delta progradation and affect 120–150 m of section composed of four delta front coarsening-upwards sequences. Horizontal spacing between the faults is variable but has an average value of 500 m. The sand-dominated facies on the downthrown side dips into the fault by as much as 20° and exhibits gentle rollover anticlines. Shale diapirism is not very pronounced in the Svalbard examples, but is more evident in Carboniferous examples in western Ireland where slightly diapiric masses of contorted mudstones-siltstones are common at depth on the upthrown side of the faults (Rider, 1978). The Svalbard growth faults are the largest examples described so far and are the only examples which affect more than one delta front coarsening-upwards sequence. In all examples the scale of the faults is very small (10–150 m) by comparison with sub-surface equivalents, but is commensurate with the scale of exposures available. Two questions can therefore be posed: firstly, do smaller scale growth faults exist in the large-scale, sub-surface growth fault complexes and secondly, if so, in surface studies of exposed successions, can small-scale faults be used to infer the presence and nature of larger scale growth faults?

## FURTHER READING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