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CHAPTER 12 Deep Clastic Seas

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12.1 INTRODUCTION

12.1.1 Historical outline

The systematic study of deep-sea sediments (Fig. 12.1) began with the voyage of HMS *Challenger* (1872–76) which established the general morphology of the ocean basins and the types of sediments they contained (Sect. 11.1). Following this pioneering expedition, the cornerstone of deep-sea sedimentology was, for a long time, the paper on 'Deep-Sea Deposits' by Murray and Renard (1891). The paradigm, put forward by these authors, was that only pelagic clays and biogenic oozes were found in the deep sea and that all coarser-grained clastics were restricted to shallow water or continental environments.

Such belief held sway amongst many geologists for almost a century while several different lines of research were converging to undermine its dominance. In particular, as more and more bottom samples and echosoundings were taken on early oceanographic expeditions in the first half of the 20th century, it was realized that sediments do not become uniformly finer-grained seaward across the continental shelves.

Although the existence of density undercurrents in lakes and reservoirs had been known for some time (Forel, 1885) it was Daly (1936) who suggested that density currents, caused by waves stirring up sediments on the continental shelf during periods of lowered sea level, may have excavated submarine canyons as they flowed down-slope. Johnson (1938) coined the term turbidity current for this type of flow. A series of flume experiments on both dilute and high density flows by Kuenen (1937, 1950), combined with Migliorini's observations on graded sand beds in the Italian Apennines paved the way for their classic paper 'Turbidity currents as a cause of graded bedding' (Kuenen and Migliorini, 1950).

This revolution in clastic sedimentology, as the turbidity current paradigm has been called (Kuhn, 1970; Walker, 1973), immediately solved several apparent anomalies of deep-sea sand deposition (Ericson, Ewing and Heegen, 1951; Natland and Kuenen, 1951) and stimulated an intense period of field, laboratory and oceanographic studies. Some of the key advances (Fig. 12.1) include: the better understanding of deep-sea sedimentation in relation to geosynclinal development and global plate tectonics (Sect. 14.2.5); the recognition of a standard sequence of structures in turbidites (Bouma, 1962), and equivalent sequences in associated coarse-grained (Walker, 1975) and fine-grained facies (Piper, 1978; Stow and Shanmugam, 1980); and an improved knowledge of the physics of such

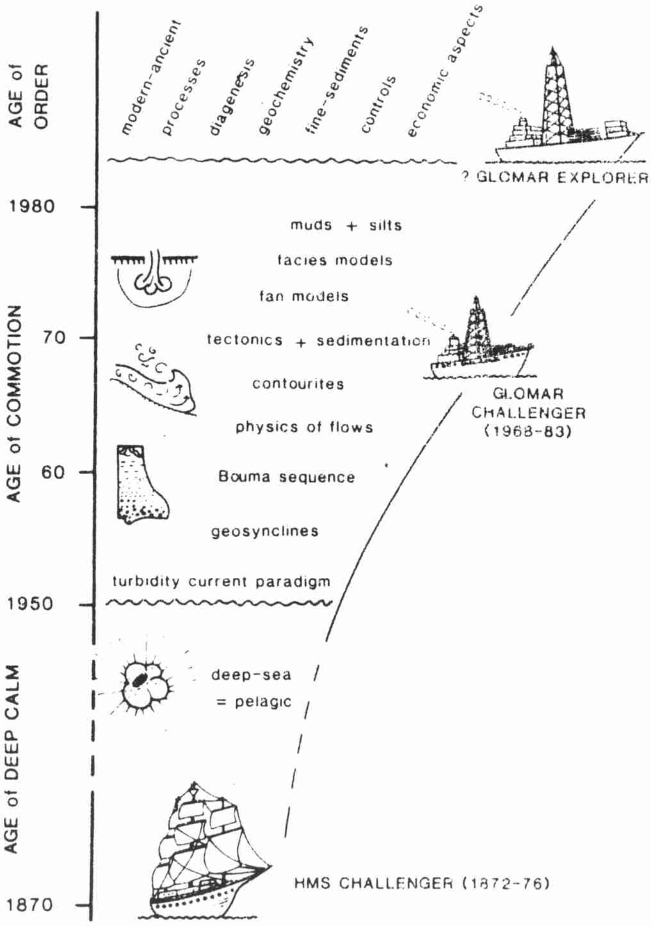


Fig. 12.1. Historical development of important concepts in deep-sea clastic sedimentology (after Stow, 1985).

*Parts of this chapter are based on the first edition, written by N. A. Rupke, who is now active as an historian of science.

flows from experimental and theoretical work (Harms and Fahnstock, 1965; Middleton, 1966, 1967; Komar, 1969, 1970).

Deep-ocean bottom currents were put forward as an important alternative to turbidity currents in the mid 1960s (Heezen, Hollister and Ruddiman, 1966; Hollister and Heezen, 1967) and the characteristics of contourites, those sediments deposited by bottom currents, were established (Stow and Lovell, 1979). After an initial emphasis on sedimentation in basin plains, models for submarine fans were developed in the early 1970s from both present-day oceans (Normark, 1970) and ancient sequences (Mutti and Ricci Lucchi, 1972).

12.1.2 **Geological controls**

Three primary controls on deep-sea sedimentation can be identified: sedimentary supply, tectonics and sea-level fluctuation as well as a number of secondary controls (Howell and von Huene, 1980; Stow, Howell and Nelson, 1984).

(1) *Sedimentary supply* includes the type of sediment (grain size and composition), the volume and rate at which such materials are made available for deposition, and the number and position of input points. Clastic slope-apron systems often differ markedly from carbonate slope-apron systems; mud-dominated fans tend to be elongate whereas sandy fans are radial; and a linear sediment supply from multiple input points along one margin results in a slope-parallel facies distribution (Sect. 12.4).

(2) *The tectonic setting* controls sedimentation by affecting the regional stress regime, rates of uplift and denudation, drainage patterns, widths of coastal plain and shelf, slope gradients, gross sediment budgets, the morphology of receiving basins, and local sea-level changes. The style and frequency of seismic activity and faulting both in the original and transitional source areas are also of primary significance. Tectonic activity varies temporally and spatially within the main tectonic settings, but is most pronounced in convergent, transform and young passive margins. If deposition is slower than subsidence, sedimentary patterns may be controlled by tectonism; if deposition is faster than subsidence, sedimentation may control gradients and the migration of channels, distributaries and terminal lobes.

(3) *Sea-level fluctuation* mainly influences deep-sea sedimentation by affecting sedimentary source areas and, thus, sediment supply. During periods of low sea-level, sediment sources such as rivers and littoral drift cells may have direct access to basin slopes. During periods of high sea-level, access is less direct, commonly via a broad continental shelf. The oceanic circulation pattern and carbonate compensation depth (Sect. 11.4.6) are also affected by sea-level changes. These changes may be global (eustatic) or regional in nature (Vail, Mitchum *et al.*, 1979).

12.2 **PROCESSES**

12.2.1 **Erosion-transport-deposition**

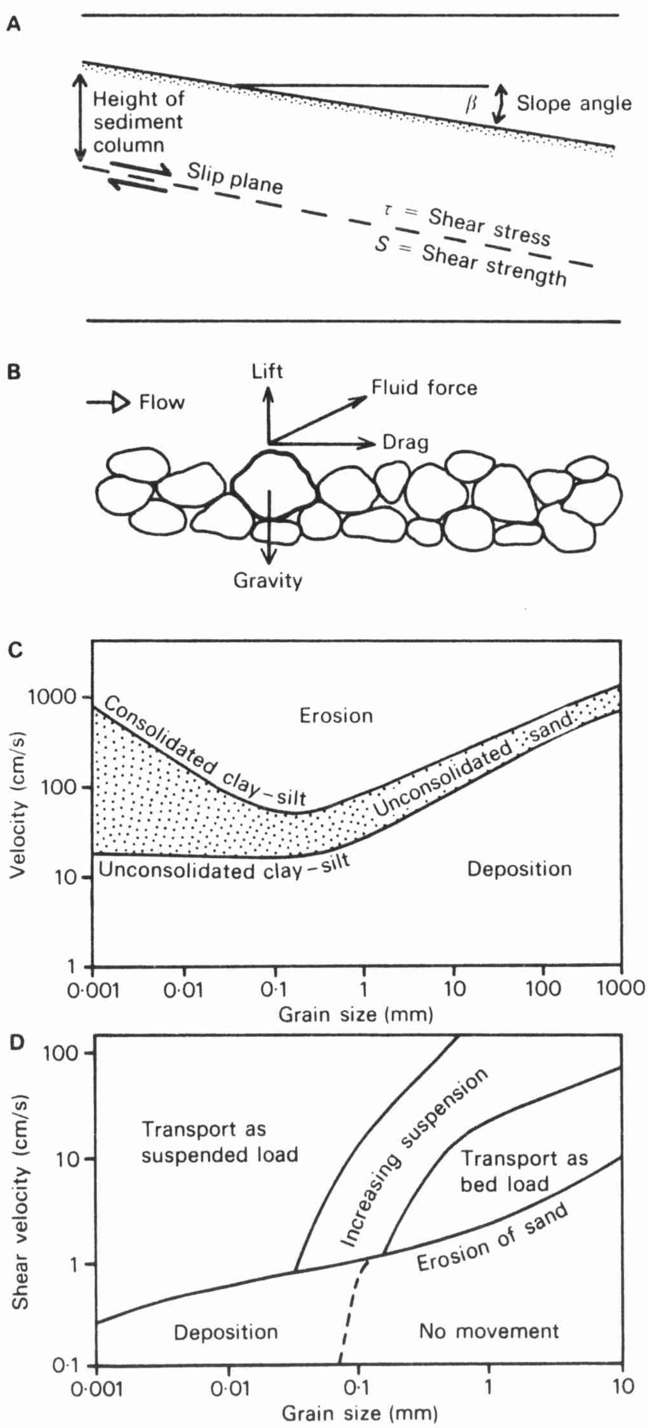
For clastic sedimentary particles to accumulate in the deep sea they must be *eroded* from land or from the sea-floor, be *transported* and then *deposited*. Biogenic material may be similarly eroded, but most is synthesized directly in the oceans, either at the surface or on carbonate banks. Authigenic minerals grow *in situ* at or near the sediment-water interface, and they too may be subsequently reworked.

The physical and chemical weathering and erosion of materials from the land is clearly influenced by the geological controls (Sect. 12.1). Transport to the sea by rivers, glaciers or wind results in most of this material being deposited in paralic or shallow shelf environments, although a small amount of aeolian, ice-rafted and river-plume sediment is transported directly to the open ocean. The upper slope-apron may also become an important transitional sink or source especially during periods of lowered sea level. These shelf and slope sediments then undergo submarine erosion and redeposition in order to reach the deep sea. A third phase of erosion and redeposition may then occur within the deep sea under the influence of bottom currents.

The initiation of sediment movement in the marine realm can be related either to (1) the mechanics of sediment failure on a slope or to (2) the critical shear velocity required to erode and move sedimentary materials on a plane bed. In the first case, sediment deposited on a slope will only begin to move downslope when the shear stress exerted by the force of gravity exceeds the shear strength of the sediment (Watkins and Kraft, 1978; Karlsrud and Edgers, 1982) along a slip plane within the sediment column (Fig. 12.2A). The shear strength is a function of the cohesion between the grains plus the intergranular friction. Sediment failure therefore results either from an increase in shear stress, due to a steepening of the slope or thickening of the sediment pile, or from a decrease in shear strength due to the sudden shock of earthquakes, storms, etc. causing fluidisation or thixotropy in the sediment. The weight of rapidly deposited sediments may exert a similar strain effect.

In the second case, sediment lying on a plane bed will begin to move as the fluid shear stress is increased and the critical threshold for grain movement is reached. Each sediment grain will experience a drag force due to the fluid shear velocity at the bed and a lift force due to the Bernoulli effect. When these combined fluid forces exceed the normal weight force due to gravity the grain will begin to move (Fig. 12.2B). Storms, internal waves, normal bottom currents and turbidity currents can all initiate sediment movement in this way.

Various experimental investigators have attempted to determine the threshold of motion for different grain types and sizes. The much used Hjulström (1939) diagram relating erosion of a particular grain-size to current velocity (Fig. 12.2C) is poorly-established and incorrect for grain-sizes finer than sands. Shields (1936) related the grain Reynolds Number to a dimensionless shear stress, but also had few data for the finer grain sizes, whereas Miller, McCave and Komar (1977) present much better data in this size range. A recent synthesis (McCave, 1984) plots



grain size against shear velocity showing transport-deposition fields for fine sediment and transport-erosion fields for coarser materials (Fig. 12.2D). McCave argues that erosion of fine cohesive sediment is not simply a function of grain size and velocity and cannot therefore be plotted on the same diagram. Bioturbidity can be important for initiating sediment movement because it influences the stability of sediment on some slopes and the erodibility of a plane bed. It can also put fine materials directly into suspension.

12.2.2 **Process continuum**

Three main processes are capable of eroding, transporting and depositing both terrigenous and biogenic material in the deep sea (Fig. 12.3): *resedimentation processes*, *normal bottom currents* and *surface currents with pelagic settling*. Several attempts have been made to classify these processes, so that a plethora of terminology and confusing (partial) synonyms exists (see review by Nardin, Hein *et al.*, 1979). The classification in Table 12.1 is based on the mechanical behaviour of the flow, the transport mechanism and sediment support system (Dott, 1963; Middleton and Hampton, 1976; Moore, 1977; Lowe, 1979; Nardin, Hein *et al.*, 1979).

The fifteen conceptually distinct processes listed are in fact part of a continuum of mechanical behaviour, ranging from elastic through plastic to viscous fluid and viscous settling (Fig. 12.3). The transition from slides to sediment gravity flows involves a change in the physical state of the sediment mass towards greater internal disaggregation by breakdown of the metastable grain packing and incorporation of more fluid. The transition from debris flow to liquefied or fluidized flows and turbidity currents involves further remoulding and dilution of the flow. During any single event of transport and deposition (Fig. 12.4) these various processes may operate at the same time or in temporal sequence, as demonstrated experimentally (Middleton, 1967) and from field evidence (Hein, 1982).

The extreme end member of sediment gravity flows, a very low-concentration, low-velocity turbidity current will be deflected by the Coriolis force from its downslope path to a direction along the slope. At this point it may grade imperceptibly into a normal bottom current, also known as a contour current, which is driven by the deep thermohaline circulation in the oceans (Stow and Lovell, 1979) rather than by the gravita-

Fig. 12.2. (A) Stability of a plane infinite slope: basic instability when $\tau = s$; creep occurs when $\tau > s$; slumping may develop when $\tau \gg s$ (after Watkins and Kraft, 1978). (B) Forces acting on a grain at rest on a non-cohesive granular bed when subjected to fluid flow above it (after Middleton and Southard, 1978). (C) Hjulström's diagram, showing critical velocity for movement of quartz grains on a plane bed at a water depth of *one metre* as modified by Sundborg (1956). (D) Proposed diagram for the transport and deposition of fine-grained suspended sediment, showing also the erosion and transport fields for coarser materials (after McCave, 1984).

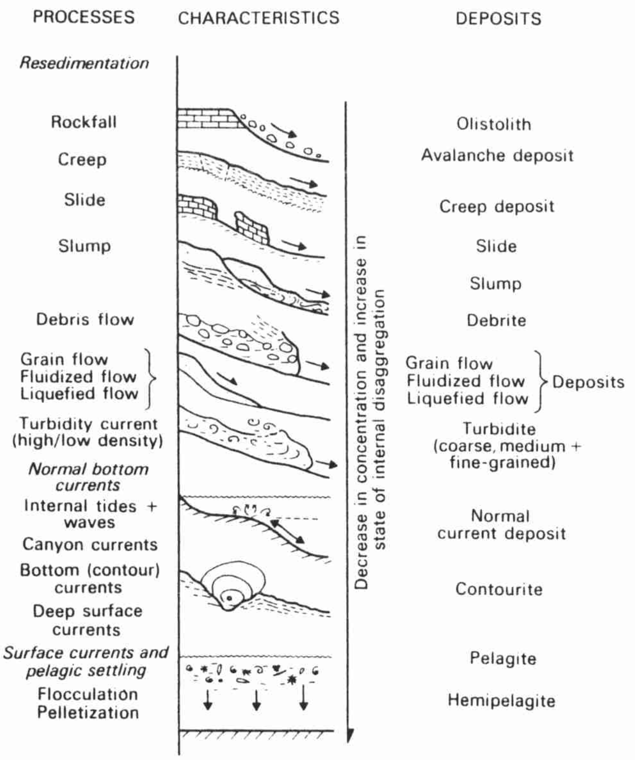


Fig. 12.3. Process continuum of the main transport and depositional processes and deposits in the deep sea.

tional effect of its sediment load. Other bottom currents (Fig. 12.4) are also caused by normal oceanic circulation, and all behave as viscous fluids. When there is no horizontal advection and dilution is extreme, simple vertical settling of particles occurs.

12.2.3 **Resedimentation processes**

Resedimentation processes (synonymous with mass gravity transport) are all those processes that move sediment downslope over the sea floor from shallower to deeper water and are driven by gravitational forces (Fig. 12.3) (Table 12.1).

FALLS, CREEP, SLIDING, SLUMPING

Rock falls are sudden, rapid, freefall events that are common in mountainous areas on land or along sea-cliffs but are relatively rare at sea because the slopes are mostly too gentle. They occur only on steep slopes of faulted or carbonate margins or in the heads of deeply incised submarine canyons, and are initiated by undercutting and erosion, and by earthquake shocks. Displaced clasts (olistoliths) may be very large (> 10 m) and bounce or roll downslope for several tens or hundreds of metres before coming to rest (Abbate, Bortolotti and Passerini 1970; Johns, 1978).

Sediment creep is a process of slow strain due to constant load induced stress that may extend over periods ranging from hours to thousands of years (Watkins and Kraft, 1978). It has not often been described from the deep sea, mainly because of the large scale on which it occurs and small amount of deformation involved, but is probably a widespread phenomenon on even very gentle slopes, depending on the physical properties of the sediment and rate of deposition. A 50 m thick, stratified and compressionally-folded, surface unit on the Canadian Beaufort Sea slope has been interpreted as being slowly displaced downslope by sediment creep along an internal *decollement zone* (Fig. 12.5) (Hill, Moran and Blasco, 1983). At high ratios of shear stress to shear strength, creep deformation may accelerate rapidly to creep rupture and may thus act as a precursor to slide or slump failure.

Sliding and slumping involve downslope displacement of a semi-consolidated sediment mass along a basal shear plane while retaining some internal (bedding) coherence. Sliding emphasizes the lateral displacement along simple or slightly rotational shear planes with little internal disturbance, whereas slumping emphasizes the internal disturbance and folded shear planes. These processes are very widespread on slopes of all gradients greater than about 0.5° and range in volume from less than 1 m^3 to over 100 km^3 and can be several hundreds of metres thick (Morgenstern, 1967; Sakov and Nienwenhius, 1982; Sect. 14.5; Fig. 14.19). They are commonly triggered by earthquake shocks, but depend also on such interrelated factors as sediment shear strength, lithology, rate of deposition, slope angle, and current systems.

A large slump on a gentle slope typically has the morphology shown in Fig. 12.6 (Lewis, 1971). The *head* is characterized by tensional structures such as faults, slump scars and bed deficiency. Above the head area retrogressive slumping may have occurred, involving successive sediment failure and the upslope progradation of unstable slump scar surfaces. The main *body* of the slump mass can be relatively undisturbed, whereas the *toe* area displays compressional structures such as thrusting and overriding of beds.

DEBRIS FLOWS, GRAIN FLOWS, FLUIDIZED/LIQUEFIED FLOWS

Debris flows are highly concentrated, highly viscous, sediment dispersions that possess a yield strength and display plastic flow behaviour (Johnson, 1970; Hampton, 1972). They are slurry-like or glacier-like, slow laminar flows that advance down slopes in excess of only 0.5°, either continuously or intermittently. Commonly, the front of the flow forms a steep scarp up to 30 m or more in height, but on steeper slopes the flow is thinner, more rapid and has a lower elevation to the mud nose (Fig. 12.7). As debris flows advance downslope they load the underlying deposits and may induce secondary failure on the sea bed. They

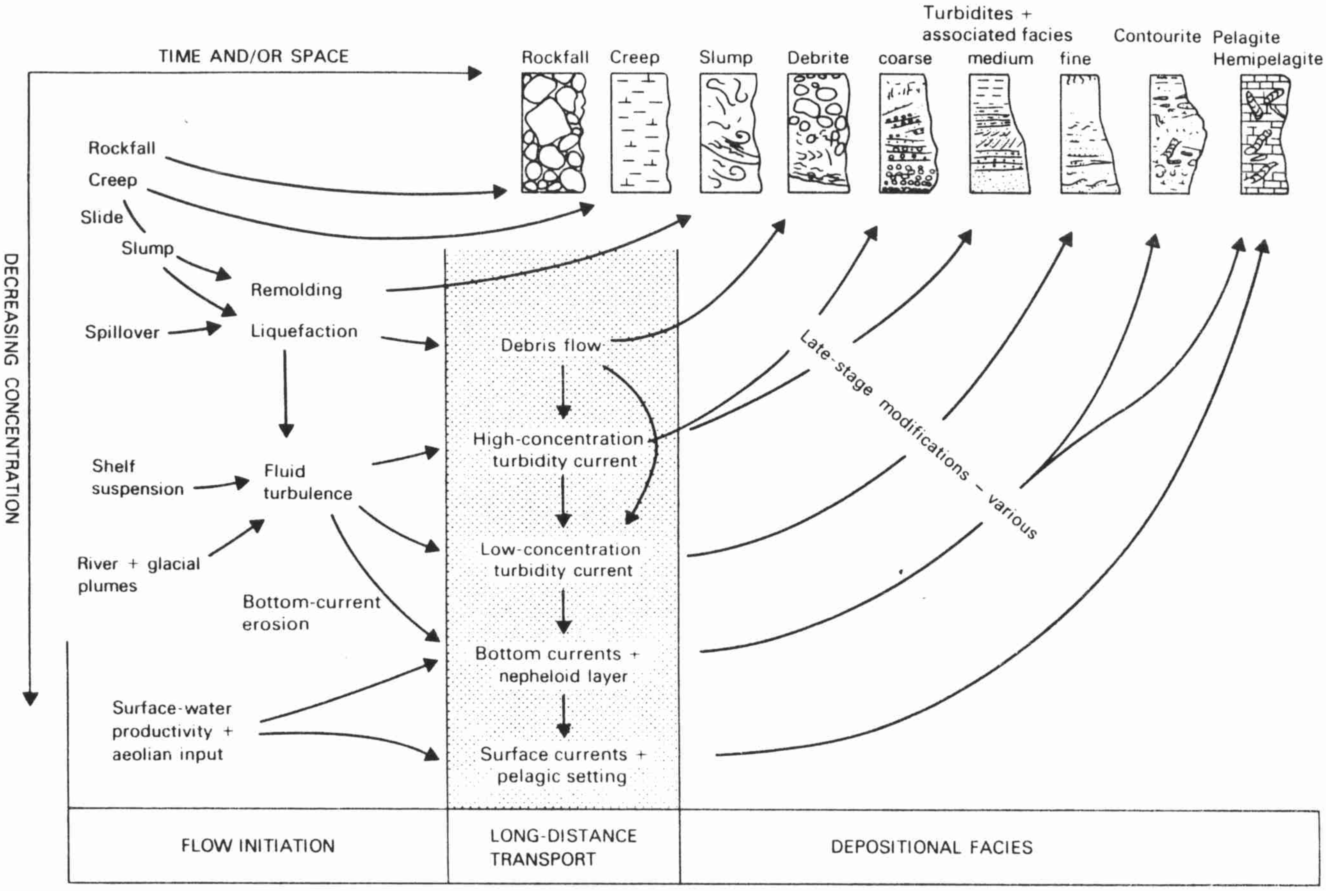


Fig. 12.4. Probable interrelationship of process of initiation, long-distance transport and deposition of sediment in the deep sea. Framework is one of time and/or space, and concentration of flows. Idealized facies models that result from deposition by the different processes are also shown. Post-depositional modification can involve current reworking, liquefaction and bioturbation (from Pickering, Stow and Watson, in press; modified after Walker, 1978).

can also give rise to slumping where either the nose of the flow or the slope of the sea bed becomes oversteepened. They are probably initiated by seismic shock, slumping or sediment creep, but also appear to develop as a result of rapid sedimentation or gas generation creating locally high pore pressures. When the downslope pull of gravity no longer exceeds the shear strength of the debris mass, or when the excess pore pressure is dissipated, the flow comes to a sudden halt or '*freezes*'.

Grain flows are quasi-visco-elastic flows characterized by grain-to-grain collisions that result in a dispersive pressure support mechanism (Bagnold, 1954). They require slopes in excess of about 18° and so are probably a very localized process in the deep sea, perhaps occurring as small-scale sand avalanches in the heads of submarine canyons (Shepard and Dill, 1966). From an analysis of the mechanics of grain flow, Lowe (1976) demonstrated a near-parabolic velocity profile with a thin surficial plug of non-shearing grains moving passively above an active shear plane. Lowe further concluded that sandy grain flows cannot be thicker than a few centimetres and so cannot be solely responsible for the deposition of thick massive sand beds.

Liquefied and fluidized flows are related processes which involve the collapse of a metastable fabric and partial or full grain support by upward-moving pore fluids. The grains become suspended and the sediment strength is reduced to zero. Loosely packed silt and sand are especially susceptible to

Table 12.1 Definitions of depositional processes in the deep sea (modified from Nardin, Hein *et al.*, 1979) and estimates of their chief physical characteristics

Depositional process	Transport and sediment support mechanisms	Slope	Dimensions	Concentration	Velocity (cm/s)	Duration	Transport Distance (km)	Average sedimentation rate
RESEDIMENTATION† Rock fall	**Elastic*** Freefall and rolling of blocks and clasts, no internal deformation of clasts	Very steep	Clasts can be > 10 m	Solid	Freefall	?min to h	< 0.5	High
Sediment creep	Slow strain and downslope movement along decollement zone due to load-induced stress, little internal deformation	Gentle	20–80 m thick	'Solid'	V. slow (imperceptible)	Semi-continuous	? < 0.5	As for background
Slide (glide)	Shear failure along discrete shear planes with little internal deformation	> About 1	Max. 300 km^3, 500 m thick (+ complete range)	Almost 'solid'	?	?h	0.001–?100	High
Slump	Shear failure accompanied by rotation along discrete shear surfaces	> About 1	As above	Almost 'solid'	?	?h	0.001–?100	High
Debris flow (mudflow)	**Plastic*** Shear distributed throughout sediment mass, slow plastic flow, clast buoyancy and matrix strength support mechanisms	> About 1	Up to few 10's of m thick	Dense slurry	?1–20	?h	?Max 350	Moderate to high
Grain flow	**Viscous Fluid (flow)*** Quasi visco-plastic flows of cohesionless grains, dispersive pressure support mechanism, localised, small-scale events	> 18	Up to few cm thick	Few data available	Few data available	?min to h	? < 0.1	Do not usually operate as separate processes
Fluidized flow	High-viscosity, short-lived flow of cohesionless grains, supported by upward-moving pore waters	> 3	< 10 cm thick	Few data available	Few data available	?min to h	? < 0.1	Do not usually operate as separate processes
Liquefied flow	Cohesionless sediment supported by upward-escape of pore waters as flow collapses and freezes, very short-lived	> About 0.5	Basal few 10 cms of flow	Few data available	Few data available	?min to h	? < 0.05	Do not usually operate as separate processes

Turbidity current (high density)	Low-viscosity flow of mixed grains supported by fluid-turbulence (autosuspension)	> About 0.5	Length and width up to 10 s of km, thickness up to 100 s of m	50–250 g l	Max. 250	?h to about 1 day	Up to about 1000	< 5 cm to > 5 m per 1000 year
Turbidity current (low density)	Very low-viscosity flow of mixed grains supported by fluid turbulence (autosuspension)	Almost no slope	As above	0.025–3 g l	Average 10–50	?h to few days	Up to several 1000 s	< 5 cm to to > 5 m per 1000 year
NORMAL BOTTOM‡ CURRENTS								
Internal tides and waves	Medium to large-scale oscillations at density discontinuities within upper few hundred metres of water column, can suspend sediment by fluid turbulence	No slope	Up to few 10 s of m amplitude	?	5–300	Semi-continuous currents often with marked periodicities	?	V. low
	Viscous*							
Normal canyon currents	Essentially 'clear-water' flows, up and down slope canyons and channels, tidal or higher periodicity, minor sediment suspension by fluid turbulence	Up and down slope < few	Up to few 10 s of m thick	? < 0.3 mg l	0–30	Semi-continuous currents often with marked periodicities	?Up to several 100 s	Low
Bottom (contour) currents	Deep, slow, essentially 'clear-water' flows driven by thermohaline circulation, can be associated with bottom nepheloid suspensions (fluid turbulence)	No slope or gentle slopes	Width up to few 10 s of km, thickness up to 100 s of m	0.025–0.25 mg l	Max. 200 Mean 10	Semi-continuous currents often with marked periodicities	Up to several 1000 s	< 10 cm per 1000 years
Deep surface currents	Deep, flow, essentially 'clear-water' flows that are deep parts of surface wind-driven ocean currents	No slope or gentle slope	As above	? As above		Semi-continuous currents often with marked periodicities	? As above	
PELAGIC SETTLING	**Viscous fluid***							
Pelagic settling	Vertical settling of individual grains, flocs and pellets through water column (viscous fluid)	Ubiquitous	Settling through 100 s to 1000 s of m water column	Extremely low	0.002–0.005 settling rate (or more if flocs)	Semi-continuous	No horizontal transport	Mean < 1 cm per 1000 years

* Mechanical Behaviour.
† Resedimentation (= mass gravity transport).
‡ Normal bottom currents (= semi-permanent bottom currents).

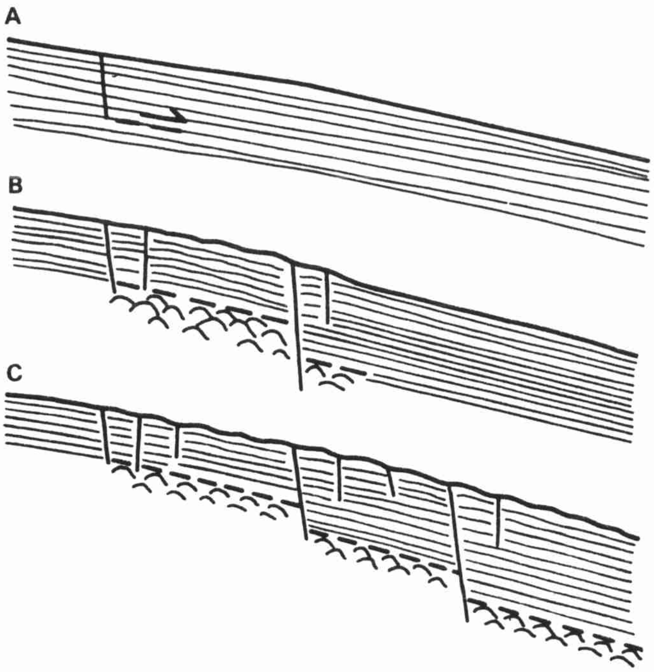


Fig. 12.5. Model for sediment creep on a gentle submarine slope (after Hill, pers. comm., 1983). The three stages (A to C) show the propagation of an internal decollement zone, its vertical displacement along zones of tension, and the development of 'sediment waves' in a horizontally stratified sediment column.

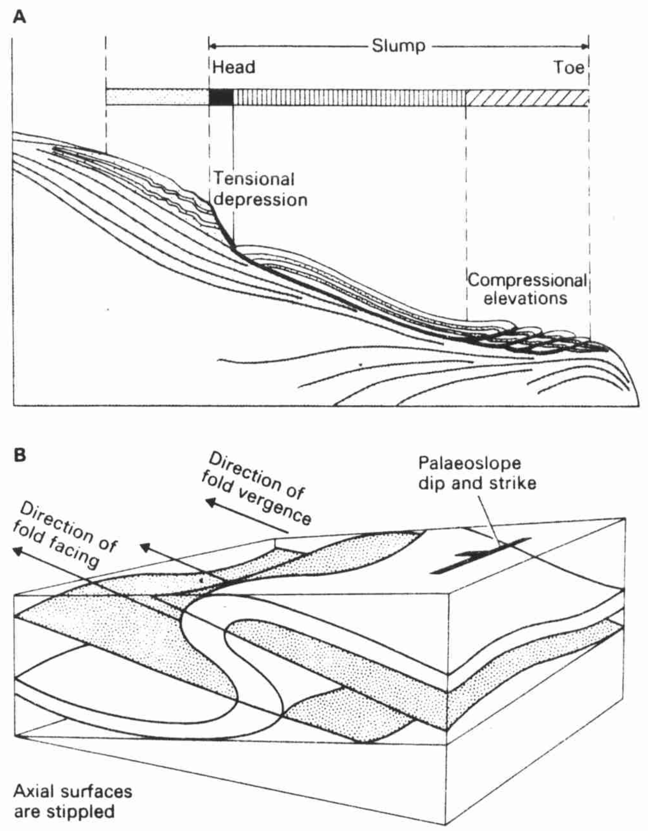


Fig. 12.6. (A) Diagrammatic cross-section of a large submarine slump on a gentle slope (from Lewis, 1971). (B) Relationship between the geometry of slump folds and the direction of slumping (from Woodcock, 1976a).

fluidization, whereas gravel is usually too porous and in muds the cohesive forces resist fluidization (Lowe, 1975, 1979; Middleton and Hampton, 1976). Fluidized sand behaves like a fluid of high viscosity and can flow rapidly down slopes in excess of 2–3°. The excess pore fluid pressures dissipate quickly, from minutes to a few hours depending on flow thickness and grain size. Deposition occurs through a short period of liquefied flow in which the grains settle rapidly and the flow freezes bottom to top. These flows rarely occur alone as a separate process in the deep sea, but commonly take place during the final stages of deposition from a high-density turbidity current (Fig. 12.4).

TURBIDITY CURRENTS: HIGH AND LOW DENSITY

Turbidity currents are perhaps the best known of the resedimentation processes from theory and experiment but they have remained elusive in nature. Nevertheless, from the very common occurrence of their characteristic deposits, *turbidites*, they are assumed to occur widely throughout the deep sea. Both high-density (50–250 g/l) and low-density (0.025–2.5 g/l) currents have been identified, within a presumed continuum of low concentrations (e.g. Middleton and Hampton, 1976; Stow and Bowen, 1980).

High-density turbidity currents are probably initiated in one of four main ways (Sect. 12.2.2; Fig. 12.4): (1) from the transformation of slumps or debris flows by mixing with seawater; (2) from sand-spillover, grain flows or rip-currents feeding sediments into the heads of submarine canyons; (3) by storm stirring of unconsolidated bottom sediments and the build-up of a concentrated shelf nepheloid layer; and (4) directly from suspended sediments delivered to the sea by rivers in flood or by glacial meltwaters.

In all turbidity currents the sediment support mechanism which keeps the sediment particles in suspension is provided primarily by the upward component of fluid turbulence, which is mainly sustained by friction at the boundary between the flow and both the floor and the ambient fluid. It has been argued that turbidity flow can be sustained in the form of *auto-suspension* (Bagnold, 1962; review by Middleton, 1970). Auto-suspension is

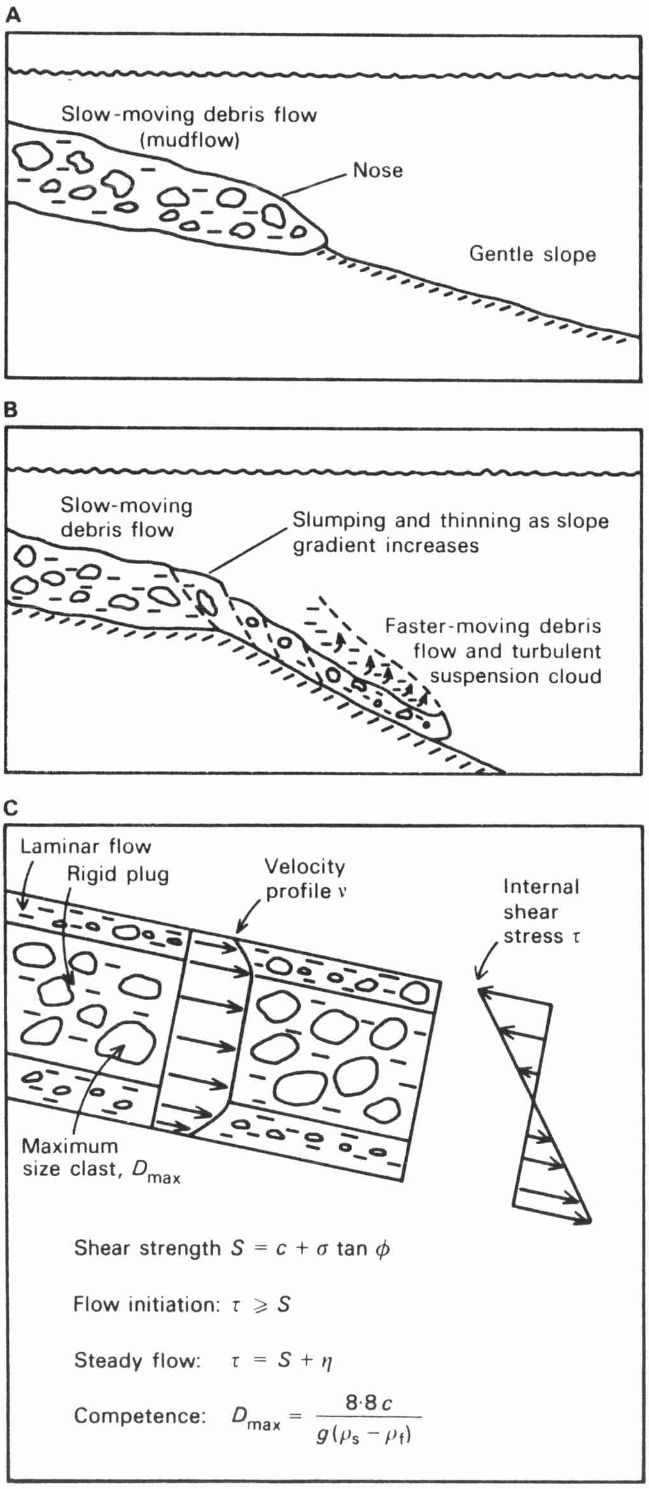


Fig. 12.7. (A) Slow moving debris (= mudflow) moving down submarine slope. (B) Slumping, thinning and velocity increase of debris flow as seafloor slope increases (after Watkins and Kraft, 1978). (C) Hydraulics of submarine debris flows (after Middleton and Hampton, 1976).

a state of dynamic equilibrium in which (1) the excess density of the suspended sediment propels the flow, (2) the flow generates friction and fluid turbulence, and (3) the turbulence keeps the sediment particles in suspension, and so on, i.e. a complete feed-back loop. All that is needed to keep the loop intact is that the loss of energy by friction be compensated for by a gain in gravitational energy as the flow travels downslope. In this theoretical model it is possible for a turbidity current to travel over long distances without appreciable erosion or deposition as long as the slope remains constant.

Experiments have shown that turbidity currents develop a characteristic longitudinal anatomy of head, neck, body and tail (Fig. 12.8) (Middleton, 1966; Middleton and Hampton, 1976). The *head* of a turbidity current has a characteristic shape and flow pattern. In plan view, the head appears lobate with local divergences of flow direction (Allen, 1971). Inside the head a forward and upward sweeping, circulatory flow pattern exists. The coarsest grains tend to become concentrated in the head. The *body* is the part behind the head where the flow is almost uniform in thickness. Deposition may take place from the body while the head still erodes. The *tail* is the part where the flow thins rapidly and becomes very dilute. Mixing between the flow and the ambient fluid produces a dilute entrained layer. On slopes greater than 1.24° the head is thicker than the body, whereas on lesser slopes the body is thicker than the head, (Komar, 1972). This is important for the type of sediment overflow in channelized environments. Mixing of the flow with water, loss of sediment by deposition and by flow separation in the *neck* will slacken and eventually stop the turbidity current. In an average turbidity current most coarse sediment will be deposited in a time-span of hours, though complete settling of the fine-grained tail may take a week (Kuenen, 1968).

Turbidity currents in the oceans are several orders of magnitude larger than those produced in laboratory flumes, so that the extent to which experimental results can be applied to turbidity currents in nature is somewhat problematic. The closest we have come to high density currents in nature is noting the occurrence of sequential breakages of submarine cables. The classic example is the Grand Banks earthquake of 1929 that triggered an enormous slump and an ensuing turbidity current that travelled downslope for hundreds of kilometres on to the Sohm Abyssal Plain (Heezen and Ewing, 1952; Piper and Normark, 1982). The maximum velocity attained by this current was some 70 km/h (25 m/s) (Menard, 1964). Other well documented examples have occurred off the coast of Algeria, from the canyon systems off the mouths of the Congo and Magdalena rivers and in the western New Britain Trench (see

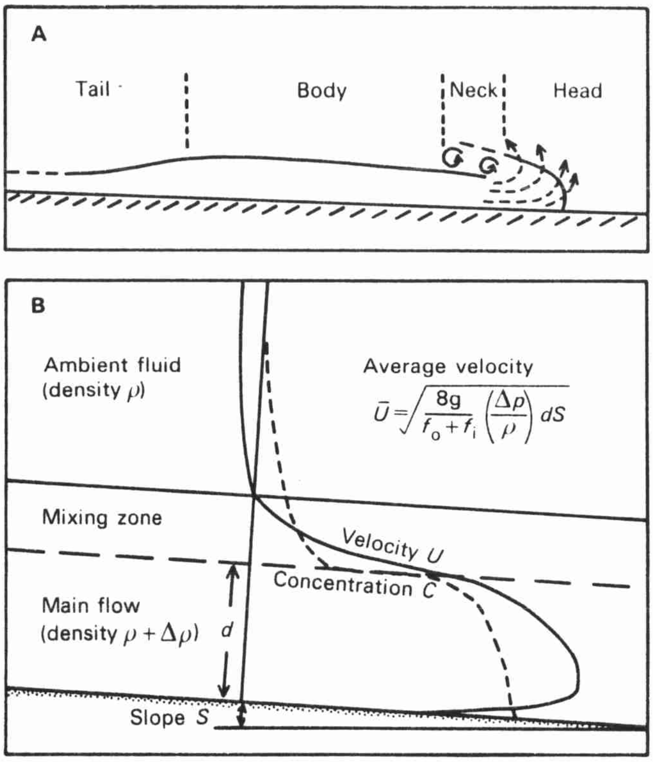


Fig. 12.8. Hydraulics of turbidity currents. (A) Schematic division of a turbidity current into head, neck, body and tail, with the flow pattern shown in and around the head region. (B) Steady, uniform flow of a turbidity current down a slope S. The average velocity of flow U is related to the thickness of the flow d, the density difference, and the frictional resistance of the bottom f_0 and upper interface f_i. (After Middleton and Hampton, 1976.)

summary by Heezen and Hollister, 1971). Estimated velocities were again of the order of tens of kilometres per hour.

Some idea of the width and thickness of turbidity currents and the distances they travel can be deduced from the resulting depositional topography. The natural levees of the submarine channels are believed to be produced from the overflow of channelized turbidity currents. Such currents must therefore be up to several kilometres wide and several hundreds of metres thick (Komar, 1969; Nelson and Kulm, 1973; Stow and Bowen, 1980). The length of deep-sea channels and of the flat expanses of abyssal plains both indicate that turbidity currents can travel as far as 4000–5000 km (Curray and Moore, 1971; Chough and Hesse, 1976; Piper, Normark and Stow, 1984).

The frequency with which turbidity currents are generated and turbidites are emplaced in any particular locality in the deep sea depends on such factors as the nature of the area from where turbidity currents originate, proximity of the area of deposition to the source, seismicity of the source area, and sea level. River-generated turbidity currents produced during periods of high river discharge may occur as often as once every two years (Heezen and Hollister, 1971). These turbidity currents are generally low-density flows. Sands accumulated in the heads of submarine canyons may be flushed out and turbidity currents develop at the same high frequency (Reimnitz, 1971). In proximal parts of active deep-sea fans turbidites may be emplaced once every 10 years (Gorsline and Emery, 1959; Nelson, 1976). However, a more distal slope or basin plain environment receives a turbidite once every 1000–3000 years, though this frequency may vary a great deal (Rupke and Stanley, 1974; Kelts and Arthur, 1981; Stow, 1984). Rise of sea level lowers the frequency of turbidity currents, especially those which are shelf- and slope-generated (Nelson, 1976). Furthermore, the occurrence of carbonate or other biogenic turbidites appears to be an order of magnitude less frequent than clastic turbidities, i.e. once every 20,000–30,000 years on average (Kelts and Arthur, 1981; Stow, 1984).

Low-density turbidity currents carry largely clay- and silt-sized particles in low concentrations and at low velocities. They are probably much more common in the deep sea than high-density currents (Piper, 1978; Stow and Bowen, 1980; Kelts and Arthur, 1981) and occur in several different forms, generated by several different processes. (1) Storm waves on the shelf or at the shelf break can stir up sediment to produce a storm-generated turbidity flow (Shepard, McCloughlin et al., 1977). (2) More continuous transport of fines across the shelf may produce thin (< 1 m) turbid-layer flows that feed downslope and along canyon axes (Moore, 1969). (3) Down-canyon flow of thick nepheloid layers (Drake and Gorsline, 1973; Drake, 1974). (4) The direct discharge into the sea of mud-charged rivers in flood or melting glaciers may also produce, directly or indirectly, low-density turbidity currents (Stow and Bowen, 1980) or lutite flows and a suspension-cascade system (McCave, 1972). (5) Creep, slumps, debris flows and high-density turbidity currents may all develop, in whole or in part, into low-density flows. All these slightly different flow types are in fact members of a spectrum of low-density turbidity currents. They occur intermittently and are of relatively short duration (of the order of days), as distinct from the *very* low density nepheloid flows associated with normal bottom currents that are semi-permanent (Sect. 12.2.4).

Various attempts have been made to estimate the physical features of low-density turbidity currents (Shepard, McCloughlin *et al.*, 1977; Shepard, Marshall *et al.*, 1979; Stow and Bowen, 1980; Bowen, Normark and Piper, 1984) (Table 12.2). They vary in thickness from a few metres to channel-full flows over 800 m thick and have velocities in the range 10 to 50 cm/s.

12.2.4 **Normal bottom currents**

This group of processes includes all those deep currents that actively erode, transport and deposit sediment on the sea floor but are not driven by sediment suspensions, and may therefore flow alongslope and upslope as well as downslope (Fig. 12.3). A selection of data showing the typical characteristics of these currents is given in Table 12.1.

INTERNAL WAVES AND TIDES

Surface waves and tides are some of the most important physical processes affecting sediments and biota in shallow water (Chap. 9). As the sea is clearly a heterogeneous body, undulation swells or internal waves can also form between subsurface water layers of varying density in the upper few hundreds of metres, most notably at the thermocline (Lafond, 1962). Such internal waves are very widespread and vary considerably in amplitude and periodicity. They may exceed surface waves in amplitude, although their speed of progression is usually slow (5–300 cm/s). Similar large-scale oscillations at density discontinuities have been shown to have a tidal period and are known as internal tides (Rattay, 1960).

The breaking and turbulent eddies caused by internal waves and the velocities attained by both internal waves and tides probably cause significant sediment stirring and erosion at the shelf break, on the tops of seamounts or in relatively shallow slope and shelf basins (Shepard, 1973). They are also thought to contribute to up-and-down canyon currents (Shepard, Marshall *et al.*, 1979).

CANYON CURRENTS

Currents rarely cease flowing up and down the axes of canyons and other submarine valleys (Fig. 12.9), even at depths of over 4000 m. (Shepard, Marshall *et al.*, 1979). Current velocities are very variable but are commonly up to 30 cm/s. A tidal periodicity seems most usual in the deeper parts, but a higher frequency of flow reversal normally occurs in the head region. Other flow periods and directions have also been recorded, probably related to internal waves, surface currents, storm surges or cold-water cascading currents.

Whereas some of these flows are clearly low-density turbidity currents (12.2.3) attaining speeds of 50–100 cm/s and capable of transporting large volumes of fine sediment downslope, many are non-turbid normal bottom currents. The frequency and velocity of these currents suggest that they will have considerable effect on the movement of sea-floor sediments and on the moulding of canyon and channel morphologies.

BOTTOM (CONTOUR) CURRENTS

Deep ocean bottom currents (Fig. 12.10) are formed by the cooling and sinking of surface water at high latitudes (Gill, 1973; Killworth, 1973) and the deep, slow thermohaline circulation of these polar water masses throughout the world's oceans (Neumann, 1968). Highly saline but warm water also flows out of the

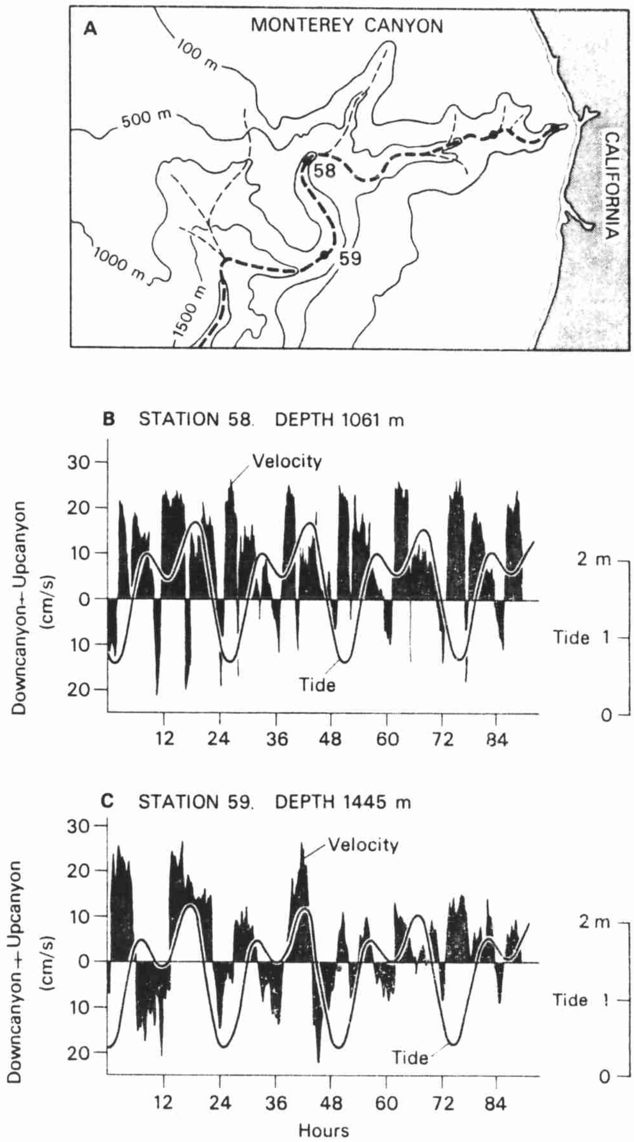


Fig. 12.9. Normal bottom currents within Monterey Canyon, off western California (after Shepard, Marshall *et al.*, 1979). (A) Bathymetric chart showing station locations. (B) Current record for station 58 at 1061 m showing rough tidal relation of upcanyon major flows but of only a few downcanyon flows. (C) Current record for station 59 at 1445 m showing possible diurnal tidal influence on both up- and downcanyon flows.

Mediterranean Sea as an intermediate level contour current. Current intensity is increased by flow restriction through narrow passages and flow concentration on the western margins of basins by the Coriolis force which deflects moving water to

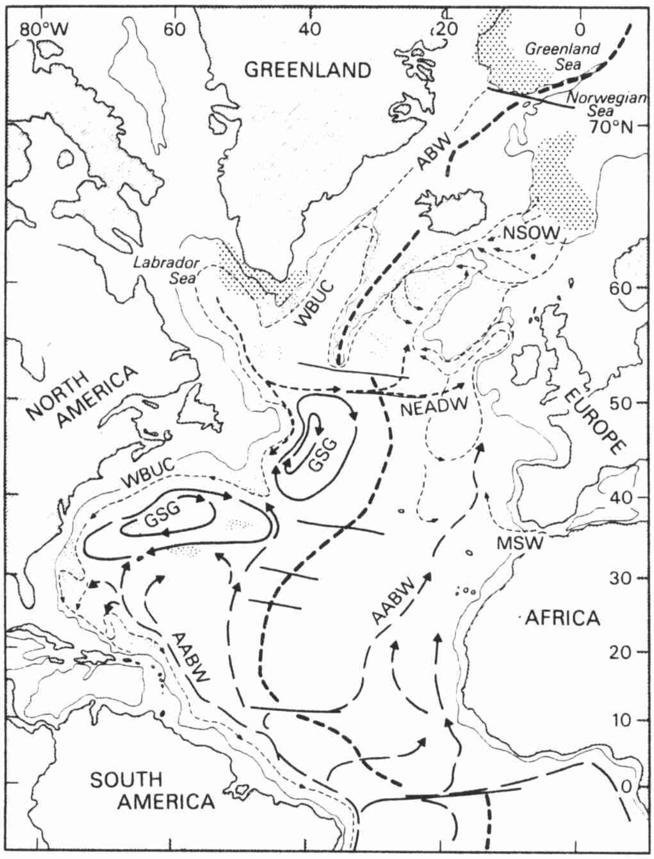


Fig. 12.10. North Atlantic present-day deep-water circulation (arrows), sediment drifts (close stipple), and areas of bottom-water formation (wide stipple). AABW, Antarctic Bottom Water; ABW, Arctic Bottom Water; MSW, Mediterranean Sea Water; NSOW, Norwegian Sea Overflow Water; WBUC, Western Boundary Undercurrent; GSG, Gulf Stream Gyre; NEADW, North-east Atlantic Deep Water. Contour at 2000 m. (After Schnitker, 1980 and Stow, 1982.)

the right in the northern hemisphere (Fig. 12.10) and to the left in the southern hemisphere, thus forming contour currents. The global system is summarized by Stow and Lovell (1979).

Whereas much of the deep-sea floor is swept by very slow currents (< 2 cm/s), the western boundary currents commonly attain velocities of 10–20 cm/s and these may be greater than 100 cm/s where the flow is particularly restricted.

Although these bottom currents are more or less continuous and sufficiently competent in parts of the ocean to erode, transport and deposit sediment, they are clearly highly variable in both velocity and direction (Luyten, 1977; Richardson, Wimbush and Meyer, 1981). Large-scale eddies peel off and move at right angles to the main flow, and the average velocity decreases from the core to the margins of the current. Both seasonal (Shor, Lonsdale *et al.*, 1980) and tidal (McCave, Lonsdale *et al.*, 1980) periodicities have been recorded, and current reversals are common. The currents vary from a few kilometres to tens of kilometres in width and can flow at different levels within the water column depending on the relative densities of adjacent water masses. In addition, surface currents driven directly by the winds may impinge on the sea floor at very great depths (several kilometres), such as the deep Gulf Stream gyres of the North Atlantic (Fig. 12.10) or the deep Kuroshio Current off Japan.

Well-developed *nepheloid layers* (Fig. 12.11), or turbid bottom waters with marked concentrations of suspended matter, are commonly associated with the higher velocity bottom currents in many parts of the ocean basins (Eittreim, Thorndike and Sullivan, 1976; Biscaye and Eittreim, 1977). These currents appear to maintain fine (average size 12 μm) particles in suspension by turbulent eddy diffusion for a residence period of about 1 year (Eittreim and Ewing, 1972). Concentrations of deep-sea nepheloid layers are extremely low (0.01–0.3 mg/l, McCave and Swift, 1976) and their thicknesses vary from less than 100 m to over 1000 m.

The geological effects of bottom currents include the eroding of channels, moats and furrows, the resuspension and transport of fine-grained sediment, the sculpting of current bedforms such as ripples, waves and lineation, and the construction of large elongate or domed sediment drifts (Fig. 12.10) made up of *contourites* (Sect. 12.3) (Hollister and Heezen, 1972; Barrusseau and Vanney, 1978; Stow and Lovell, 1979). Bottom currents also deposit contourites along the continental slope and rise, where they are interstratified with turbiditic, hemipelagic and other sediment facies, and can substantially rework and winnow previously deposited sediments (Carter and Schafer, 1983; Shor, Kent and Flood, 1984). Where particularly strong, bottom currents commonly cause a depositional hiatus in the sediment record, in some cases associated with a coarse-grained lag deposit.

12.2.5 Surface currents and pelagic settling

Slow pelagic settling through the water column (Chap. 11) can be considered one extreme end-member of the process continuum (Fig. 12.3). It is less important for clastic than for biogenic sediments as the materials involved are largely the tests of calcareous and siliceous planktonic organisms and their associated organic matter that have been biosynthesized in the surface layers of the oceans. These form the *pelagic* deposits of the deep sea.

However, in many areas of the deep sea, particularly on slopes and in basins close to land, terrigenous elements (clays, quartz, feldspar, volcanic dust and other minerals) with a high proportion of silt-sized grains can form a significant part of the settling material and hence of the resulting *hemipelagic* deposit. Such materials are transported by surface currents, winds and

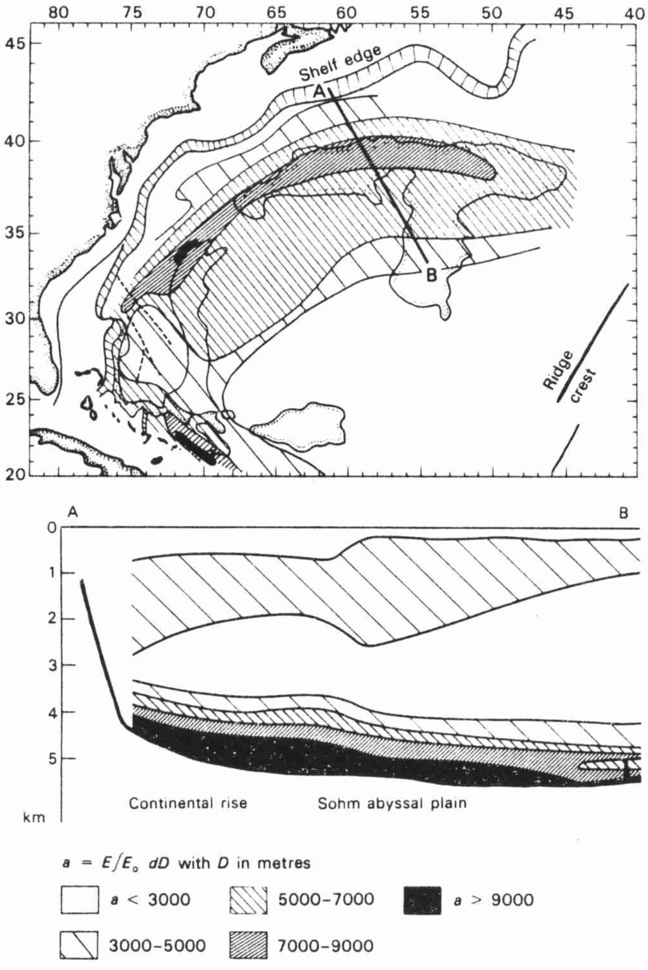


Fig. 12.11. Map and cross-section showing horizontal and vertical distribution of suspended matter concentrations in the nepheloid layer, northwestern Atlantic Ocean. The intensity of shading indicates the relative concentrations from a maximum of about 0.1 p.p.m. (0.2 mg/l) near the sea floor to a minimum of about 0.01 p.p.m. (0.02 mg/l) in the mid-column clear water zone (from Eittreim and Ewing, 1972).

floating ice and mix with pelagic biogenic components during settling.

Vertical settling of the finest particles is extremely slow (10^{-4}–10^{-6} m/s), although much of the material settles more quickly (10^{-2}–10^{-3} m/s)) as flocs and faecal pellets. As it settles and before burial on the sea floor, the material is subject to dissolution of calcareous and siliceous tests, oxidation of organic matter and lateral transport by bottom and turbidity currents.

12.3 FACIES: MODERN AND ANCIENT

12.3.1 Facies characteristics

Deep-sea facies are currently defined on the basis of the following principle features: grain size and other textural attributes, sand/mud ratio, bed thickness and geometry, internal organization of beds, dynamic and biogenic sedimentary structures, fabric, composition and biota. Ideally, each facies so defined should be a unique type that forms under certain conditions of sedimentation, reflecting a particular process.

However, with more than ten distinct depositional processes (Sect. 12.2), with also a large range of environments (Sect. 12.4) and of sediments ranging from huge boulders to the finest clays, there is clearly a very large number of possible facies in the deep sea. The existence of a process-continuum implies there must be a related *facies-continuum* (Figs 12.3 and 12.4), so that the facies identified are members of a more gradual spectrum of deposits.

Whereas the early descriptions of deep-sea facies were based mainly on ancient *flysch* sequences and emphasised their homogeneity (i.e. the absence of abrupt vertical and lateral facies changes), the past 15 years have seen an enormous increase in the number of samples and cores collected from the modern oceans, as well as an appreciation of the heterogeneity of deep-sea rocks on land. More than 50 facies have been identified for deep-sea clastic sediments alone (Fig. 12.12) and several different facies classifications have been proposed.

12.3.2 Facies classification

The classification of deep-sea facies evolved to a relatively sophisticated level about 10 years ago, with particular emphasis on sandstones (Mutti and Ricci Lucchi, 1972, 1975; Walker and Mutti, 1973). Since then several other facies have been recognized for both the coarser-grained (Carter, 1975; Walker, 1975; Watson, 1981) and finer-grained sediments (Piper, 1978; Stow, Bishop and Mills, 1982), and attempts have been made to combine some of these into composite facies models (e.g. Walker, 1978; Stow and Piper, 1984). In the classification scheme (Fig. 12.12) the terms and divisions used are entirely descriptive, although they are designed to aid interpretation of the processes outlined in the facies models of the following sections.

The first order classification and description into *classes* is often adequate for the purpose of regional mapping or reconnaissance work (Figs 12.12 and 12.13). For the second order classification the facies classes A to E can be subdivided into *disorganized* and *organized* facies *groups* (A1, A2 etc.). The disorganized groups essentially lack clear stratification or grading and include thick structureless gravels, sands and muds; irregular, thin-bedded gravel lag or coarse sand layers; and bioturbated, massive or irregularly-layered, silty muds. The organized facies groups show some degree of stratification or

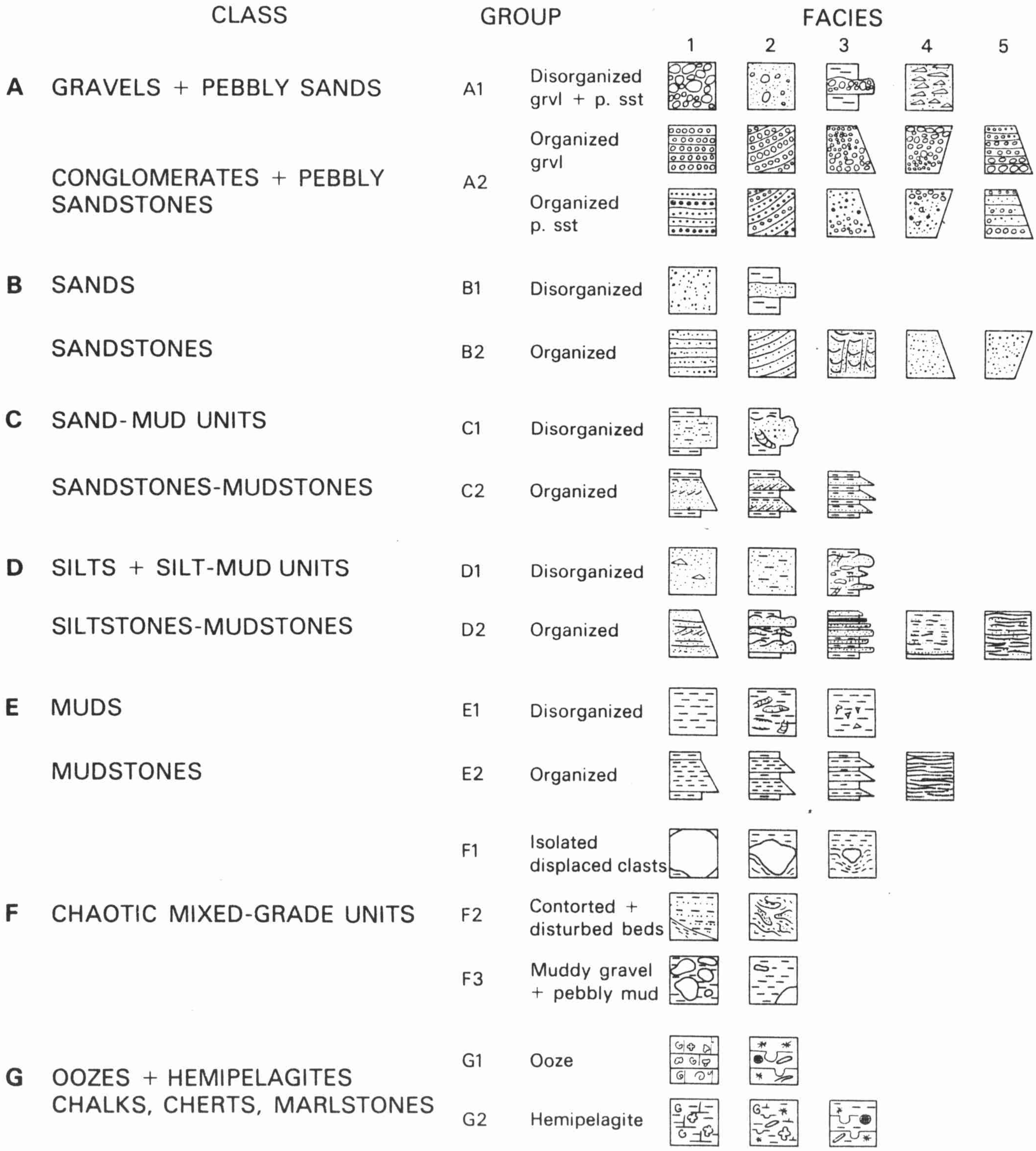
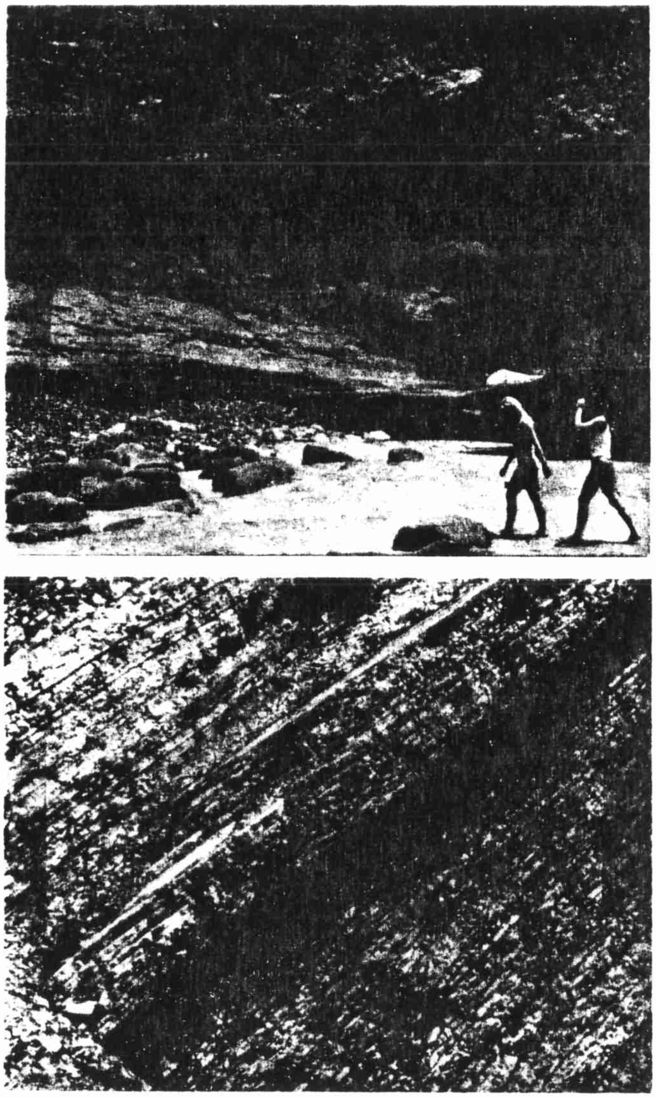


Fig. 12.12. The main classes and groups of sediment facies recognized in the deep sea (from Stow, 1985; modified after Mutti and Ricci Lucchi, 1972). The facies classes are distinguished on the basis of grain size (Facies Classes A–E), internal organisation (Facies Class F) and composition (Facies Class G). Facies groups are distinguished mainly on the basis of internal organization of structures and textures. Individual facies (sub-groups 1–5) are based on internal structures, bed thickness and composition.

(A)



(B)

Fig. 12.13. Selected photographs of typical deep-sea sediment facies (scale approximately the same for both sections): (A) channel fill and associated turbidite facies, Upper Cretaceous Cabrillo Formation, La Jolla, California; (B) silicified black shales and interbedded pelagic limestones, Mid Cretaceous Scisti a Fucoidi Formation, Pietralata, Italy.

marked grading and include regularly laminated, cross-laminated, rippled and graded layers of variable bed thickness and grain size.

Facies class F is mainly disorganized and can be subdivided into three groups: exotic clasts, ranging from giant rock-fall boulders to small glacial dropstones (F1); contorted and disturbed slumps and slide masses (F2); and pebbly muds or muddy gravels (F3). Facies class G comprises both the pelagic biogenic sediments, the calcareous, siliceous and muddy oozes, and the silty biogenic sediments or hemipelagites.

12.3.3 **Facies models (general)**

Most of the separate facies shown in Fig. 12.12 can be interpreted in terms of depositional process by reference to one of the facies models for resedimented, normal current deposited and pelagic sediments (Figs 12.14, 12.16 and 12.18). These facies models show the *standard sequences* of structures and sedimentary characteristics of sediments deposited by single events or particular processes. They rarely occur as complete sequences in the geological record; more commonly a portion of either the upper or lower parts is missing (top-absent and base-absent sequences respectively). Facies D2.2 (lenticular and rippled silt laminae in mud) for example, may be interpreted as a series of repeated top-absent, fine-grained turbidites. Actual examples of some of these facies from both the recent and ancient record are shown in Figs 12.15, 12.17 and 12.19.

12.3.4 **Resedimented facies models (clastics)**

SLUMPS

Slumps and *slides* (facies group F2, Fig. 12.15) can involve any lithology and be very thick (> 100 m) or very thin (< 10 cm). The internal beds are mainly coherent in slides which have moved largely as an undeformed block along a basal shear zone. The toe and the head regions may show compressional and tensional deformation structures respectively (Fig. 12.16). In slumps there is a more pervasive disruption of beds and a variety of deformational structures has been recognized, including several types of folds, thrusts, balls, hook-shaped overfolds, rotational slumps, scars, etc. (Dzulynski and Walton, 1965; Allen, 1982 v 2). No standard vertical sequence of these structures has been identified.

The axes of slump folds in many slump occurrences are preferentially oriented (Rupke, 1976; Woodcock, 1976a). The direction of slumping is generally assumed to be perpendicular to the mean of the azimuths of the slump fold axes and is determined from the sense of overturning (or facing direction) of the folds. However, in a slumping sediment mass, friction along the margin of the mass or internal obstructions may cause slump folds to rotate so that their long axes are turned parallel to the downslope movement. In such instances the axes of slump folds may seem randomly oriented, but the direction of the slumping can be determined using the method of the separation angle developed by Hansen (1971).

An important distinction has to be made between sedimentary slumping and subsequent tectonic deformation. Although a

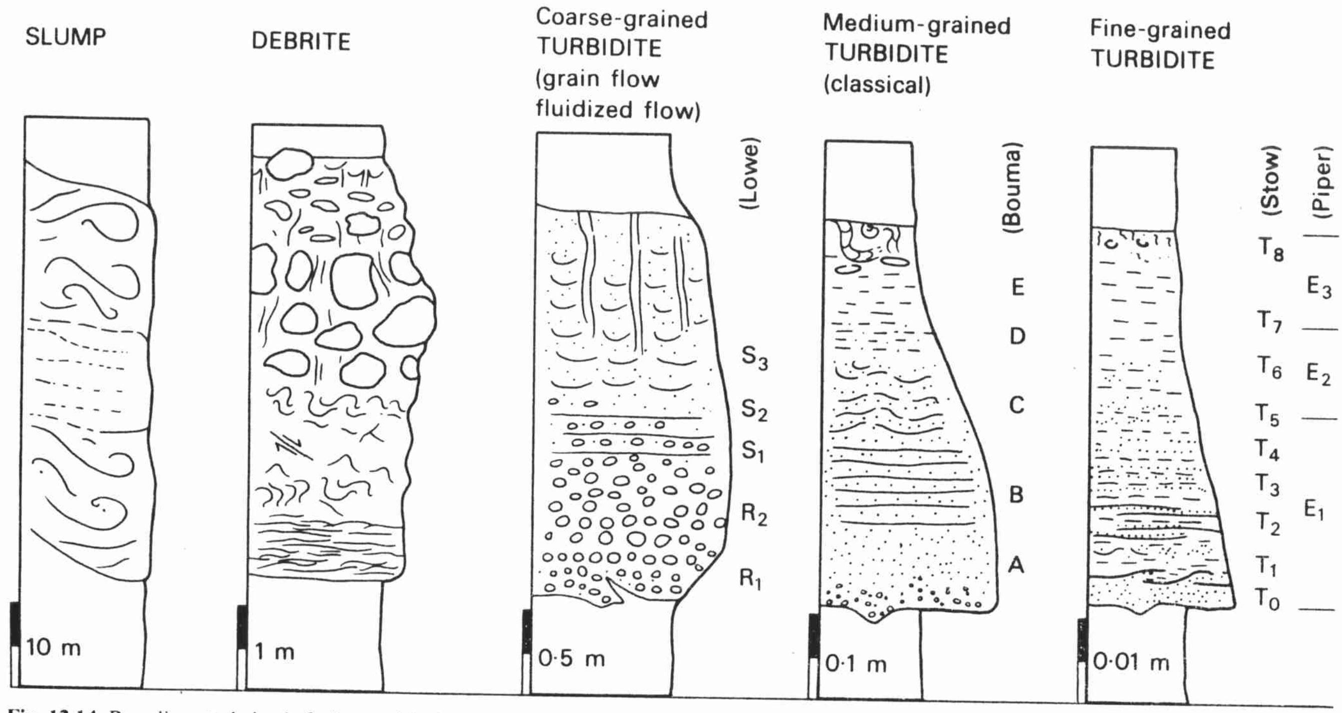


Fig. 12.14. Resedimented clastic facies models for slumps, debrites and turbidites, showing the idealised structural sequences. The scale bars give an indication only of typical unit thickness, which may vary widely in practice. Grain-size increases to the right for each column (from Stow, 1984).

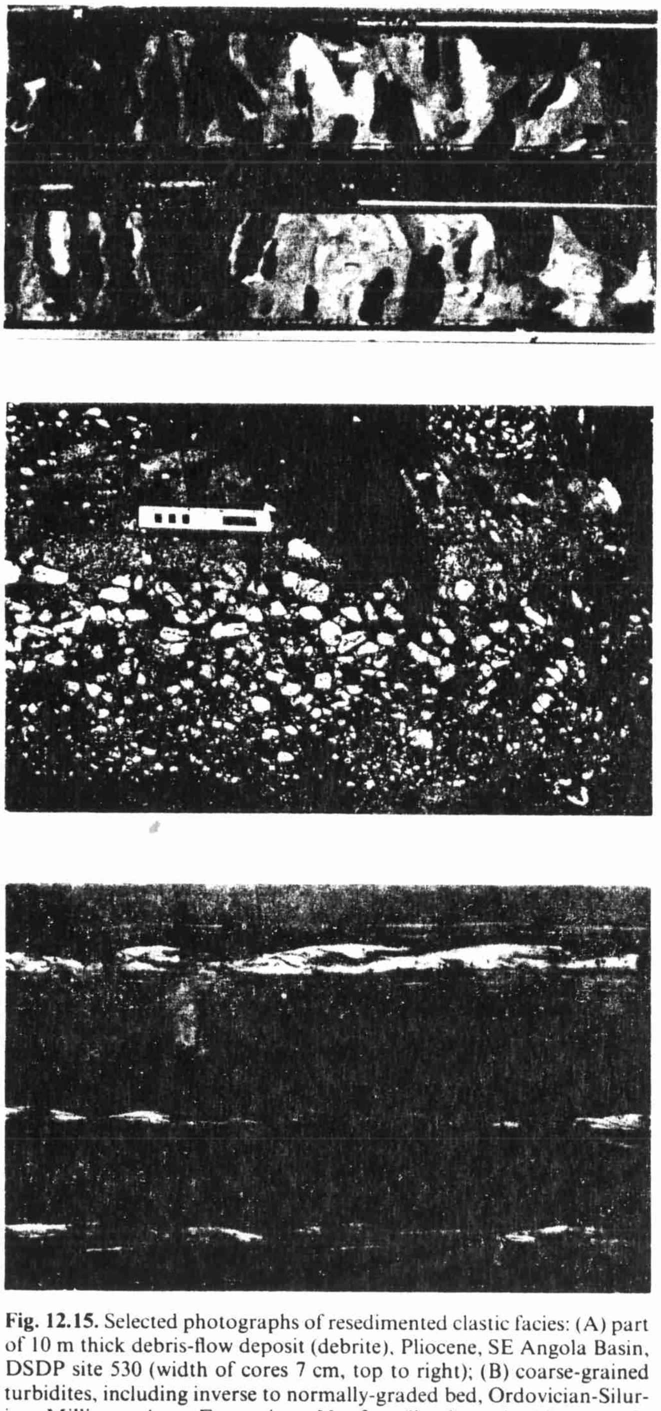
spectrum of deformational structures from purely sedimentary to purely tectonic has been recognized (Maas, 1974), it is nevertheless possible to list the features that characterize sedimentary slumps (Kuenen, 1953; Helwig, 1970; Woodcock, 1976b; Naylor, 1980; Allen, 1982) among which are the following: (1) deformed beds occur as a zone between undisturbed beds; (2) the upper contact of the zone of deformed beds is welded, i.e. a depositional fit occurs between the irregularities of its upper surface and the base of the overlying bed; (3) fold anticlines may be eroded at the upper surface; (4) the preferred orientation of fold axes, if present, may be unrelated to the tectonic strike; and (5) within a single slump the structural style may be irregular and a wide range of deformational structures may occur.

DEBRITES

Debrites (facies group F3, Fig. 12.14), also called *debris flow deposits* and olistostromes, consist of mixed lithologies and range from muds containing only a few sand- to boulder-sized clasts to a bouldery mass containing little mud. The thickness of individual beds can vary widely up to several tens of metres. They may be quite structureless and disorganized or minimally organized with a scoured base, negative basal grading, slight irregular positive grading through the rest of the bed, some horizontal alignment of elongate clasts, and a top that either grades into a muddy turbidite or is sharp with large protruding clasts (Middleton and Hampton, 1976).

The facies model shown in Fig. 12.14 is based on modern examples of slump and debrite deposits from the California borderland basins (Thornton, 1984). A regular vertical sequence can be recognized: a basal sheared sediment zone of lensoid lamination, a middle deformed zone with high-angle faults, minor slump folds and possible convolute bedding, and an upper matrix-supported clast-rich zone (? debrite proper) which may show dewatering pipe and dish structures especially near the top. In the Plambini limestone-shale sequence of the northern Apennines, debrites derived from slumps along a block-faulted margin show folds and boudins occurring as clasts within the debrites (Naylor, 1980, 1981).

Debrites are well known from both the modern deep-sea (Embley, 1976; Moore, Curray and Emmel, 1976) and from



(A)

(B)

(C)

Fig. 12.15. Selected photographs of resedimented clastic facies: (A) part of 10 m thick debris-flow deposit (debrite), Pliocene, SE Angola Basin, DSDP site 530 (width of cores 7 cm, top to right); (B) coarse-grained turbidites, including inverse to normally-graded bed, Ordovician-Silurian Milliners Arm Formation, Newfoundland (scale 15 cm); (C) fine-grained turbidites, with 'fading-ripple' siltstones, Cambro-Ordovician Halifax Formation, Nova Scotia (width of photo 40 cm).

ancient rocks (Abbate, Bortolotti and Passerini, 1970; Cook and Taylor, 1977). In some cases they can be seen to have travelled several hundreds of kilometres over gentle slopes ($\sim$ 1–2°) and cover areas of many thousands of square kilometres. Bed thicknesses range from a few tens of centimetres to a few tens of metres and there appears to be a close relationship between bed thickness and maximum clast size.

TURBIDITES

Three different turbidite models can be recognized, each with its own distinctive standard sequence of structures through a single bed (Fig. 12.14).

The *coarse-grained turbidite* model (after Lowe, 1982) represents many of the facies in facies classes A and B. The main process of long-distance transport is a high-density turbidity current but many of the structures in the $R_{12}S_{123}$ sequence are a result of grain flow, fluidized or liquefied flow mechanisms during the final stages of deposition (Fig. 12.4). The lower part of the sequence can comprise gravel, pebbly sand or sand, overlying a sharp, scoured base. Characteristic structures include, a negatively-graded lower division (R_1), overlain by massive (R_2), stratified (S_1), graded-stratified (S_2) and finally by dish and pipe structured (S_3) divisions. The top is commonly sharp and flat (Walker, 1978; Lowe, 1979, 1982). Some of the facies in our classes A and B (e.g. A2.2, B2.2) may be the result of 'normal-current' traction processes rather than turbidity currents.

The *medium-grained turbidite* model is the classical Bouma (1962) sequence and represents most of our facies class C and parts of B and D (there being some overlap between the three turbidite models). The five structural divisions overlying a sharp, erosive or loaded base, are: massive to graded sand (Ta), parallel-laminated sand (Tb), cross-laminated and convolute sand (Tc), parallel-laminated fine sand and silt (Td), and massive to bioturbated mud (Td).

The *fine-grained turbidite* model, represents much of facies classes D and E. A graded silt-laminated mud division (E_1) passes upward into a graded mud (E_2) and a nongraded mud (E_3) (Piper, 1978). The graded laminated unit (E_1) can be further subdivided into a thick, often lenticular basal silt laminae with fading ripples at the top (T_0), a relatively thick mud layer with convolute silt laminae (T_1), low amplitude ripples (T_2), parallel distinct (T_3), parallel indistinct (T_4) and wispy silt laminae (T_5). These are overlain by graded mud (T_6), nongraded mud (T_7) and a thin microbioturbated zone (T_8). (Rupke and Stanley, 1974; Nelson, Normark *et al.*, 1978; Stow and Shanmugam, 1980; Kelts and Arthur, 1981).

These idealized turbidite sequences can be interpreted hydrodynamically as resulting from a single resedimentation event that deposited progressively finer-grades of sediment and gave rise to different sedimentary structures as the flow velocity and carrying power decreased (Harms and Fahnestock, 1965;

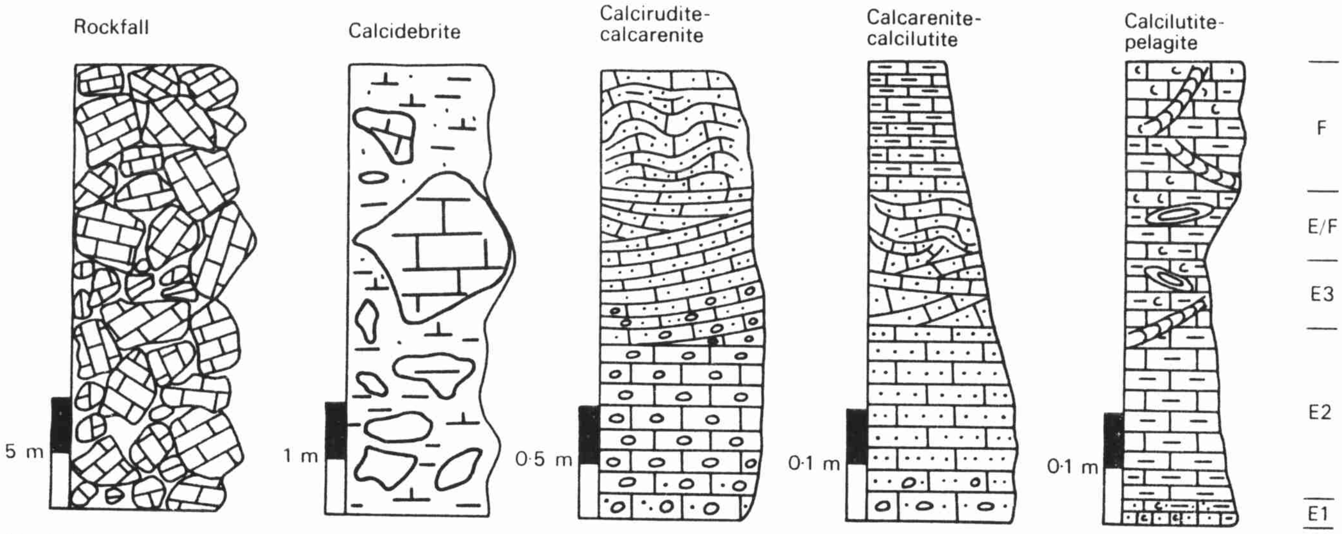


Fig. 12.16. Resedimented carbonate facies models for rock falls, debrites and turbidites. Scale bars give an indication only of typical unit thickness, which may vary widely in practice. Grain size increases to the right for each column.

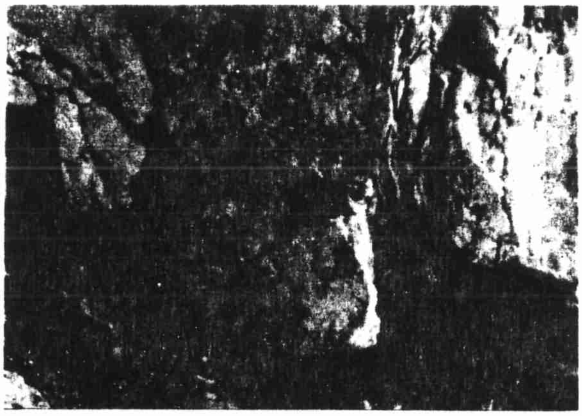
Walker, 1965, 1975; Stow and Shanmugam, 1980; Lowe, 1982). A *complete sequence* is very rarely deposited and *partial sequences* are the rule (top-absent, base-absent, mid-absent etc.). These partial sequences give rise to the many possible facies shown in Fig. 12.12. For example, deposition of top-absent classical turbidites (Bouma divisions A, AB, ABC, or Ta, Tab, Tabc turbidites) produces massive sands (facies B1.1) parallel-laminated sands (facies B2.1) or thick-bedded turbidites (facies C2.1), whereas base-absent fine-grained turbidites (Piper divisions E_{23}, Stow divisions T_{678}) give massive and graded mud turbidites (facies E1.1, E2.2 and E2.3).

There are several other characteristics, in addition to the dynamic sedimentary structures, that are important for recognition and interpetation of turbidites. Positive grading is very common in the coarse and medium-grained turbidites as well as in the silt-laminated and pure mud turbidites. Negative grading is also common at the base of the many beds. Various attempts have been made to characterize turbidites in terms of the shape of the grain-size distribution curves (e.g. P... 1977, Kranck, 1984), statistical parameters such as [...]ting (i.e. high matrix content) [...]itive skewness, an[...] [...]cal cross-plots (e.g. 'C-M diagr[...] [...]assega, 1964). The [...]ral attributes, in particular, a[...] readily applied [...]cient lithified sediments, partly [...]ause accurate analys[...] [...]e difficult and partly because a mud matrix in turbidite sandstones (greywackes) may be caused by diagenesis rather than primary deposition.

Fabric studies have shown that elongate particles (sand grains, plant fragments, graptolites, etc.) are often aligned parallel to current flow ([...]n, 1967). From the base to top of a [tu]rbidite the alignmen[t] [...] increasingly diverge from the [or]ientation of the s[...]ks (Scott, 1967; Parkash and [M]iddleton, 1970), wh[...] be due to a meandering flow [pa]ttern in turbidity cur[...] [g]rain imbrication dipping upcurrent also occurs. Curre[...] [ali]gnment of silt grains was used by Stow (1979a,b) to distinguish between turbidites deposited by downslope currents and contourites by alongslope currents on the Nova Scotian continental rise. It appears that mud fabrics differ between turbiditic and hemipelagic muds (O'Brien, Nakazawa and Tokuhashi, 1980): turbidites have larger, more randomly arranged clay particle clusters, whereas hemipelagites have more single clay particles aligned parallel to bedding.

Biogenic structures of different types are present in some turbidites. Mostly, they are restricted to or more abundant towards the tops of individual beds and within the intervening pelagic intervals (dwelling and resting traces). They also occur on bedding planes between turbidite beds, commonly as sole structures (crawling, grazing and feeding traces). There is an important bathymetric control on burrow (or trace fossil) assemblages (Seilacher, 1967) (Fig. 12.30), although several other factors can be equally significant in determining the burrowing activity in deep-sea sediments, including biotic density and diversity, ecologic stress, grain-size, depositional environment, sediment composition and turbidity-current frequency (Crimes, Goldring *et al.*, 1981; Werner and Wetzel, 1982).

(A)



(B)

Fig. 12.17. Selected photographs of resedimented carbonate facies: (A) megaflute at base of calcirudite turbidite, Cretaceous-Tertiary Scaglia Rossa Formation, Italy. (B) calcarenite turbidite, Cretaceous-Tertiary Scaglia Rossa Formation, Italy.

Muddy sediments have the most abundant and diverse trace fossils and the highest degree of bioturbation. Basinal muds can be intensely bioturbated with many grazing and associated traces of the *Nereites* assemblage (Fig. 12.30). The slow sedimentation rates permit even small numbers of organisms to completely rework the bottom muds, except where thicker turbidites are introduced with a high frequency or where anoxic conditions prevail. Slope and turbidite-dominated muds commonly have *Zoophycos* and *Nereites* trace fossil assemblages (Fig. 12.30), and the interbedded turbidite sands may show abundant escape traces as well as crawling and grazing traces on the soles of beds. By contrast, more shallow-water associations include the shelf and shoreface, stable, low-sedimentation rate, highly-bioturbated facies with the *Cruziana* assemblage (Fig. 12.30) and intertidal to non-marine, highly-stressed environments with *Skolithos* and *Scoyenia* assemblages (see also Sect. 9.9.1).

The composition of turbidites is dependent primarily on the sediment source and may be extremely variable. Compositions can be used as a criterion for distinguishing between turbidites and interbedded non-turbidite sediments (e.g. Hesse, 1975; Stow, 1979a), for recognizing grading in turbidites where grain-size variation is minimal (e.g. Kelts and Arthur, 1981; Stow, 1984), or for inferring the tectonic setting and depositional environment (e.g. Dickinson and Suczek, 1979).

12.3.5 **Resedimented facies models (biogenics)**

Resedimented carbonates occur off many modern carbonate platform and reef margins and on the flanks of seamounts and mid-ocean ridges (e.g. Mullins and Neuman, 1979; Faugeres, Gayet *et al.*, 1982). They are equally well-known from the ancient record (e.g. Cook and Enos, 1977; McIlreath and James, 1978, Fig. 11.33). Resedimented siliceous biogenic sediments have been described from modern ocean basins, particularly near areas of upwelling and high surface productivity such as offshore SW Africa (Stow 1984) and the Gulf of California (Curray, Moore *et al.*, 1980); while ancient cherts have in certain cases been interpreted as largely turbiditic (Nisbet and Price, 1974; Folk and McBride, 1978).

Resedimented biogenic facies are in many respects similar to the equivalent clastic sediments. However, some of the main differences are emphasized below (Figs. 12.16, 12.17).

(1) *Rockfall* deposits are more abundant in carbonates than in clastics, probably as a result of the steep slopes associated with reef and carbonate platform margins (Cook, McDaniel *et al.*, 1972; Conaghan, Mountjoy *et al.*, 1976; Johns, 1978). They are typically poorly sorted, with a mixture of angular to subangular small clasts and large blocks, and are chaotic in appearance. There is little fine-grained matrix so the deposit is wholly clast-supported. The fabric is quite random and both grading and stratification are absent. Large isolated boulders can also fall as individual clasts into finer-grained sediment, thereby distorting the original fabric. *Carbonate slumps* and *debrites* are also widespread and do not markedly differ from the clastic equivalents (Hubert, Suchecki and Callahan, 1977; Shanmugam and Benedict, 1978).

(2) The structures and facies of the *calcirudite-calcarenite turbidite* sequence are equivalent to those of coarse-grained clastic turbidites. However, the dune cross-bedded division is more common (Hubert, Suchecki and Callahan, 1977).

(3) In the Bouma C division convolute bedding is more common in *calcarentite-calcilutite turbidites* and ripple cross-bedding in clastic sand-mud turbidites (Hesse, 1975; Enos, 1977). In relatively pure carbonate systems there is commonly a sharp, but often irregular and disturbed break between the calcarenite and calcilutite parts of the turbidite, and the calcilutite appears quite structureless (Stow, Wezel *et al.*, 1984).

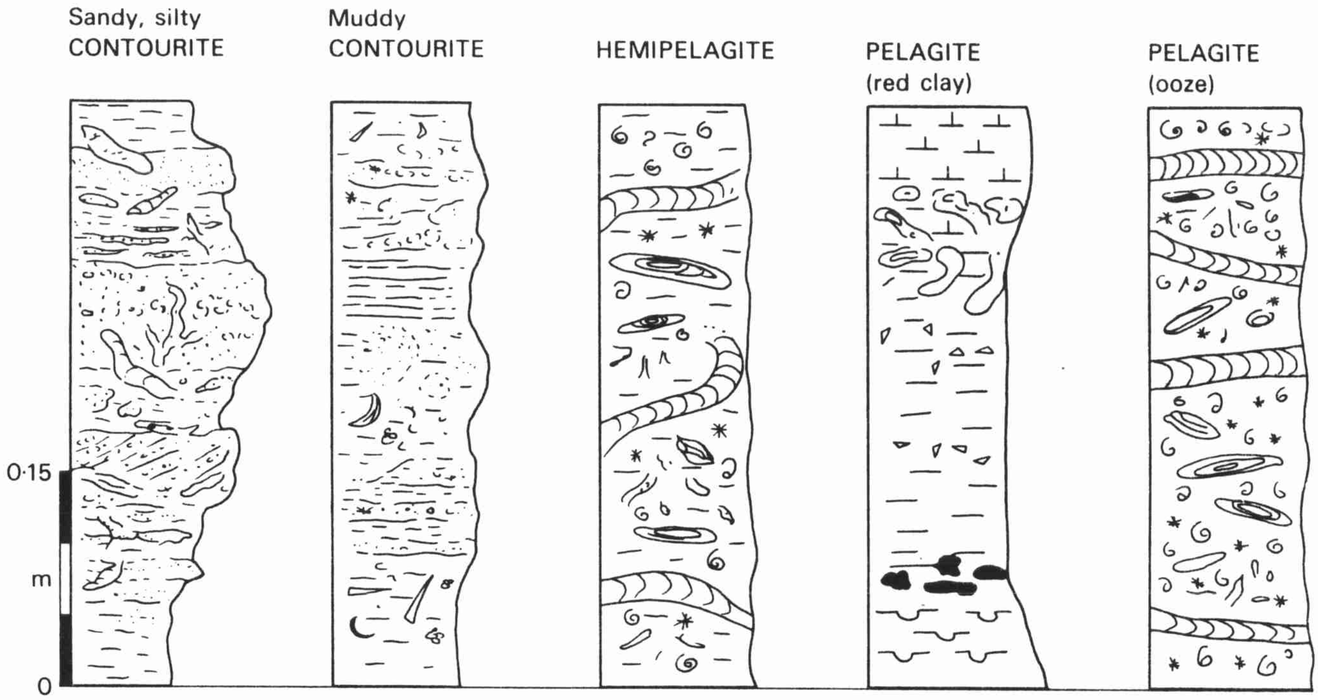


Fig. 12.18. Normal sedimentation facies models for contourites, hemipelagites and pelagites. Grain size increases to the right for each column.

In many cases it is therefore difficult to determine whether the interbedded calcilutites are turbiditic or pelagic. In less pure carbonate systems, where there is a significant admixture of terrigenous clays, a more normal clastic-like Bouma sequence is developed.

(4) The *calcilutite turbidite-pelagic* model emphasizes the subtle differences between the turbiditic and pelagic division in fine-grained resedimented carbonates. If there is a small amount of silt or sand grade material in the flow this will deposit first as a thin (laminated) basal division (E_1) and be overlain by a featureless calcilutite that may show very slight positive grading (E_2). Most of the calcilutite is extremely fine-grained, massive and ungraded (E_3), with burrowing and bioturbation becoming more evident upwards. There is a gradual transition (E/F) showing negative size-grading into the overlying more thoroughly bioturbated, coarser-grained pelagite (model and divisions after Stow, Wezel *et al.*, 1984; examples, Kelts and Arthur, 1981; Schlager and Chermark, 1979).

The lack of distinction in pure carbonate systems between turbiditic and pelagic calcilutites probably results from the tendency of fine-grained carbonate material being transported by turbidity currents to disperse into the water column and settle out with the background pelagic biogenics. Terrigenous clayey material, on the other hand, flocculates readily and deposits more rapidly directly from the transporting turbidity current.

12.3.6 **Bottom current facies models**

There are two distinct types of facies that have been affected by deep-sea bottom currents: *reworked channel deposits* and *contourites*. In addition, bottom currents are probably responsible, in part, for the distribution of fine-grained hemipelagite facies.

REWORKED CHANNEL DEPOSITS

Within canyons and channels distinctly fluvial characteristics and facies may be developed by more or less constant bottom currents, (e.g. McGreggor, Stubblefield *et al.*, 1982; Damuth, Kowsmann *et al.*, 1983; Bouma, Stelting and Coleman, 1984). Large-scale dune cross-bedded sandstones (facies A2.2, B2.2) and thin gravel lag deposits (facies A1.3) probably represent normal-current (or turbidity-current reworked) facies (Fig. 12.12) of submarine canyons and channels. The reworked sediments occur as thin to thick isolated beds within a channel sequence and not as one part of a larger resedimented bed. They show characteristics of traction transport and reworking and an

(A)

(B)

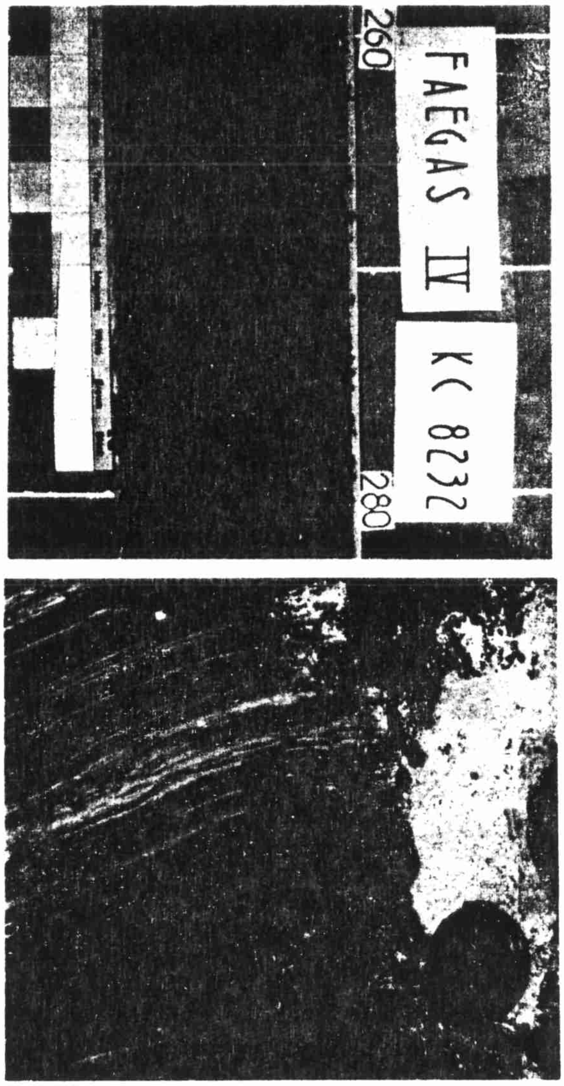


Fig. 12.19. Selected photographs of contourite and pelagite facies: (A) core section through late Quaternary muddy-silty contourite facies, Faro Drift, Gulf of Cadiz (scale in cm); (B) finely laminated organic-carbon-rich siliceous pelagite, Miocene Monterey Formation, California.

absence of features related to instantaneous deposition such as fluid-escape structures. They are well documented from ancient successions (e.g. Winn and Dott, 1977; Scott and Tillman, 1981; Hein and Walker, 1982).

CONTOURITES

Two main contourite facies result from *deposition* by bottom currents: *muddy contourites* and *sandy contourites* (Fig. 12.18). Facies models for these types have been developed from Tertiary to Recent contourite drifts in the deep sea (Stow and Lovell, 1979; Stow, 1982), but it has proved particularly difficult to recognize ancient contourites on land with any degree of certainty (e.g. Anketell and Lovell, 1979; Lovell and Stow, 1981).

Muddy contourites (Facies E1.2 and D1.3 on Fig. 12.12) are fine-grained, poorly-sorted clay- and silt-sized sediment with up to 15% sand fraction. They are mainly homogeneous or structureless and thoroughly bioturbated, more rarely having irregular layering, lamination and lensing. They range from finer-grained homogeneous muds to siltier mottled silts and muds. Their composition varies with the primary source material, but is most commonly mixed biogenic and terrigenous. They mostly closely resemble hemipelagites.

Sandy contourites (Facies C1.2 on Fig. 12.12) occur either as thin irregular layers (< 1–5 cm) or thicker beds (5–25 cm), that are either structureless and thoroughly bioturbated or have some primary horizontal and cross-lamination preserved. They can show both negative and positive grading, or both, and have sharp or gradational bed contacts. Grain size is commonly fine sand, more rarely medium sand, with poor to moderate sorting. In many cases the mean grain size is in the coarse silt grade and the facies may be more accurately termed 'silty to fine sandy' contourites. The composition is variable, commonly mixed terrigenous and biogenic. The facies may sometimes be confused with fine-grained turbidites.

Muddy and sandy contourites commonly occur together in characteristic vertical 'sequences', in some ways analogous to the standard turbidite sequences (Faugeres, Stow and Gonthier, 1984). A complete sequence (Fig. 12.19b) shows negative grading from a fine homogeneous mud, through a mottled silt and mud, to a fine sandy contourite facies and then positive grading back to a muddy contourite. The grain-size changes and concomitant changes in sedimentary structures and composition are probably related to long-term fluctuation in the mean current velocity, of the order of 2000–10,000 years for a 50 cm sequence.

The effects of *winnowing* and *reworking* by bottom currents can result in contourite facies with rather different characteristics. Thin, irregular, poorly-sorted, structureless, mixed-composition, iron-manganese coated, coarse-sand and *gravel-lag contourites* (Facies B1.2 and A1.3 on Fig. 12.12) are formed by the winnowing and removal of all fines from a coarse-grained sediment by powerful bottom currents. The reworking more or less *in situ* of sandy turbidites can result in a bottom-current modified turbidite sand. These are believed to be common on continental slopes and rises. In the central parts of ocean basins, bottom currents are known to construct large sediment drifts out of almost pure biogenic material (Stow and Holbrook, 1984; Kidd, Ruddiman *et al.*, 1984). These *biogenic contourites* are often very similar to pelagites.

12.3.7 Hemipelagite and pelagite facies models

Summary facies models for hemipelagites and pelagites are also shown in Fig. 12.18 (examples, Fig. 12.19) (see also Chap. 11).

Hemipelagites (facies group G2) are compositionally very similar to muddy contourites, being composed of mixed biogenic and terrigenous material; they also appear homogeneous, massive and thoroughly bioturbated. However, they do not show any evidence of current-control during deposition, probably have a somewhat different ichnofacies and show no vertical 'sequence' of facies or textures (Hesse, 1975; Cook and Enos, 1977; Hill, 1981).

The two contrasting *pelagite* models are for oozes (facies group G1) comprising more than 70% biogenic material, and for red clays (facies E1.3) that commonly have less than 10% biogenic material (see Chap. 11; Hoffert, 1980; Thiede, Strand and Agdestein, 1981).

12.4 MODERN DEEP-SEA ENVIRONMENTS

12.4.1 Environmental models and their components

Within the marine realm we can identify three fundamentally different environments of clastic deposition. These are *slope-aprons*, *submarine fans* and *basin plains*. Slope-aprons accumulate over gently inclined surfaces, submarine fans are large constructional mounds developed at the bases of slopes, and basin plains are the deepest, flattest parts of the deep sea.

The sedimentary features that characterize each environment are best studied by bottom sampling, coring, drilling and underwater photography. The tectonic features and geometric distribution of sedimentary or seismic facies are most easily identified on seismic reflection profiles (Brown and Fisher, 1977; Vail, Mitchum *et al.*, 1979). In general, high frequency seismics give greater penetration but low resolution, whereas lower frequencies give better resolution of only the surface few tens to hundreds of metres (Fig. 12.20).

The *morphological elements* within the different environments include canyons, channels and slump scars, lobes, mounds, drifts and irregular masses, wedges and levees, and interchannel, open slope and open basin regions (Fig. 12.21). They are commonly a few hundred metres to several kilometres in width, a few metres to a few hundred metres in elevation, and may be approximately equidimensional or markedly elongate (up to several thousand kilometres). Such features are readily observed with normal bathymetric sounding techniques and have also been characterized to some extent in terms of their echocharacter (e.g. Damuth, 1975; 1980) and seismic facies (Sangree and Widmier, 1977; Vail, Mitchum *et al.*, 1977; Nardin, Hein *et al.*, 1979) (Figs 2.6 and 12.21).

There are many still smaller-scale erosional, depositional and irregular morphological features in the deep sea that we have only recently begun to resolve using deep-tow instrument packages (Spiess, Lowenstein *et al.*, 1976; Normark, Hess and Spiess, 1978). These are on the scale of features that are more easily recognised in ancient outcrops, and Normark, Piper and Hess (1979) have drawn attention to the differences in scale between the features observed at sea and those we observe on land.

12.4.2 Slope-aprons

Slope-aprons make up the region between the shelf and the basin floor, surrounding both small shelf basins and the large ocean basins. They include both the continental slope and continental rise. Slope-aprons also occur on the flanks of oceanic ridges, isolated seamounts and plateaus. The marginal ocean slopes are particularly important as major depocentres, and also as the sites of erosion and of the initiation of resedimentation processes towards deeper base-of-slope, fan and basin environments.

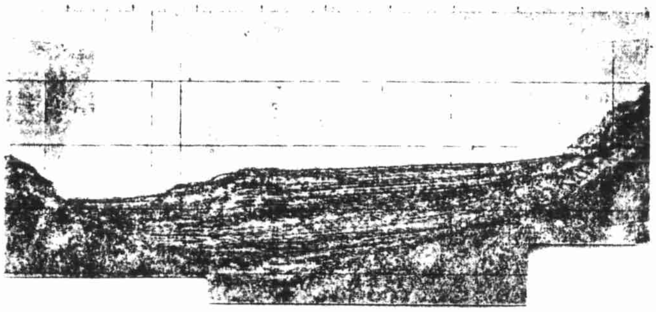
Slope-aprons vary in width from less than 1 km to over 200 km and commonly have gentle gradients from 2° to 7°, rarely exceeding 10°. They may be erosional or depositional, smooth or rugged, and comprise a complete range of clastic and biogenic facies. The main morphological elements include, a relatively abrupt shelf-break, fault-scarp and reef-talus wedges, slump and slide scars, irregular slump and debris flow masses, small straight or slightly sinuous channels and gullies, more complex dendritic canyons, isolated lobes, mounds and drifts, and broad areas of smooth or current-moulded surface.

At least ten different types of slope-apron can be distinguished on the basis of their primary morpho-tectonic settings (Emery, 1977; Bouma, Moore and Coleman, 1978; McIlreath and James, 1978; Doyle and Pilkey, 1979). However, on the basis of the main sedimentary features and morphological elements we recognize three composite slope-apron models (Fig. 12.22).

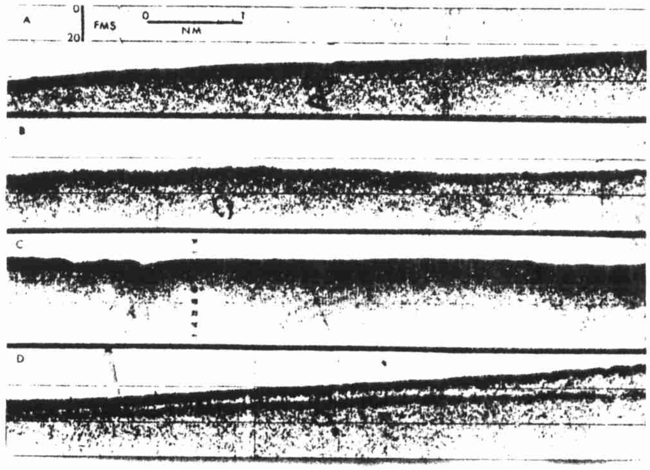
NORMAL (CLASTIC) SLOPE APRONS

These have a relatively smooth convex-concave profile built upwards and outward by slope progradation (Fig. 12.22A). In a low-energy setting they have a *sigmoid-progradational* cross-section pattern, whereas in a high-energy setting they have an *oblique-progradational* cross-section (Vail, Mitchum *et al.*, 1979; Fig. 2.6). In the former case the slope surface tends to be smooth or current-moulded, sometimes with the development of distinct elongate contourite drifts near the base of the slope, whereas in the latter case the slope may be gullied, or irregular and slump-scarred with sediment lobes, debris-flow masses and slump blocks at the foot of the slope. Larger canyons and channels cutting right across the slope may occur at intervals.

Where the sediment supply is particularly high and subsidence keeps pace with accumulation the shelf builds out as a



(A)



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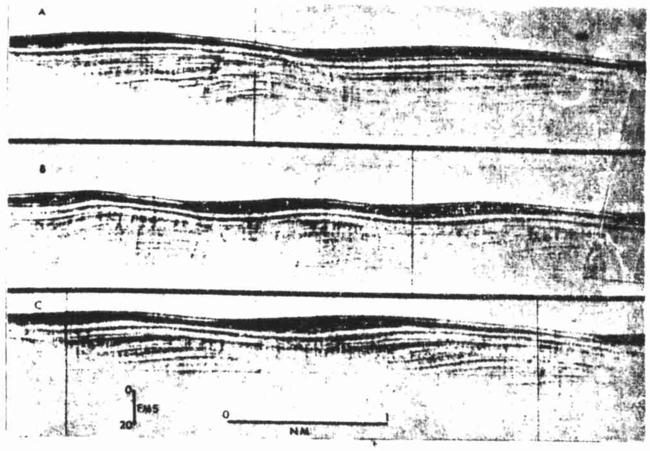
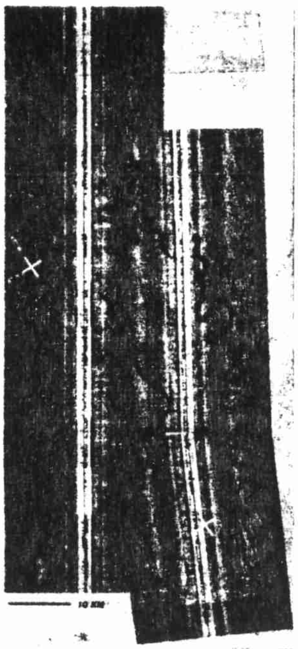


(C)



(D)

Fig. 12.20. Different types of acoustic profiles from the deep-sea. (A) Seismic reflection profile showing sediment drape over the Walvis Ridge, southeast Atlantic Ocean and mainly resedimented basin fill in the adjacent Angola Basin (from Stow, 1984). (B) 3.5 kHz echogram with indistinct prolonged fuzzy characteristic due to high proportion of coarse-grained sediment near the surface (from Damuth, 1975). (C) 3.5 kHz echogram with broad low-amplitude sediment waves and semi-parallel subbottom reflectors, indicating current control of fine-grained sediment deposition (from Damuth, 1975). (D) *GLORIA* sidescan sonograph showing high and low-sinuosity meandering channels on the Amazon deep-sea fan at about 3000 m water depth (from Damuth, Kolla *et al.*, 1983).

wedge that thickens towards the shelf edge. *Erosion* on the face of the slope, possibly as a response to sediment instability and rotation during downwarping, results in a somewhat steeper profile and a more irregular surface than for the progradational slopes. Older sediments and reflectors outcrop on the slope face.

The distribution of facies is highly irregular and dominated by fine-grained sediments (silts, muds, oozes and hemipelagites). Commonly, there is a *mudline* (Stanley and Wear, 1978) dividing the shallow, higher-energy, sandy shelf facies from the muddier slope sediments. A certain amount of *sand spillover* occurs along the shelf-break and sands are funnelled down canyons or gullies to form isolated depositional lobes. Slump and debrite facies are common on some slope-aprons.

ECHO TYPE	SKETCH	MORPHOLOGICAL PROVINCE + INTERPRETATION
DISTINCT single		Continental shelves, coarse surface layer
multiple		Slopes, fans, basin plains, fine turbidites, contourites and pelagites
PROLONGED strong		Channels and canyons, coarse facies, small-scale surface irregularities
medium + sub-bottoms		Minor channels, lobes etc. mixed coarse and fine facies
weak		Slopes and channels, debris flow or slide mass
HYPERBOLIC large ± sub-bottoms		Slopes, irregular surface of slumps, slides and scars
medium		Slopes, irregular surface, slumps etc. or bedforms
small		Lower slopes, fans, channels, small current bedforms
WAVES standing ± regular		Lower slopes, channel levees, basin plains, large current bedforms
migrating ± regular		As above More regular flow, migration commonly up current
SCARPS erosional		Slopes, channels, slide scar face or channel margin
fault		Slopes, channels, sedimentary or tectonic instability
flexure + piercement		Slopes, usually result of diapirism

Fig. 12.21. Schematic representation of typical echocharacteristics for different morphological environments in the deep sea (after Damuth, 1975; Jacobi, 1982).

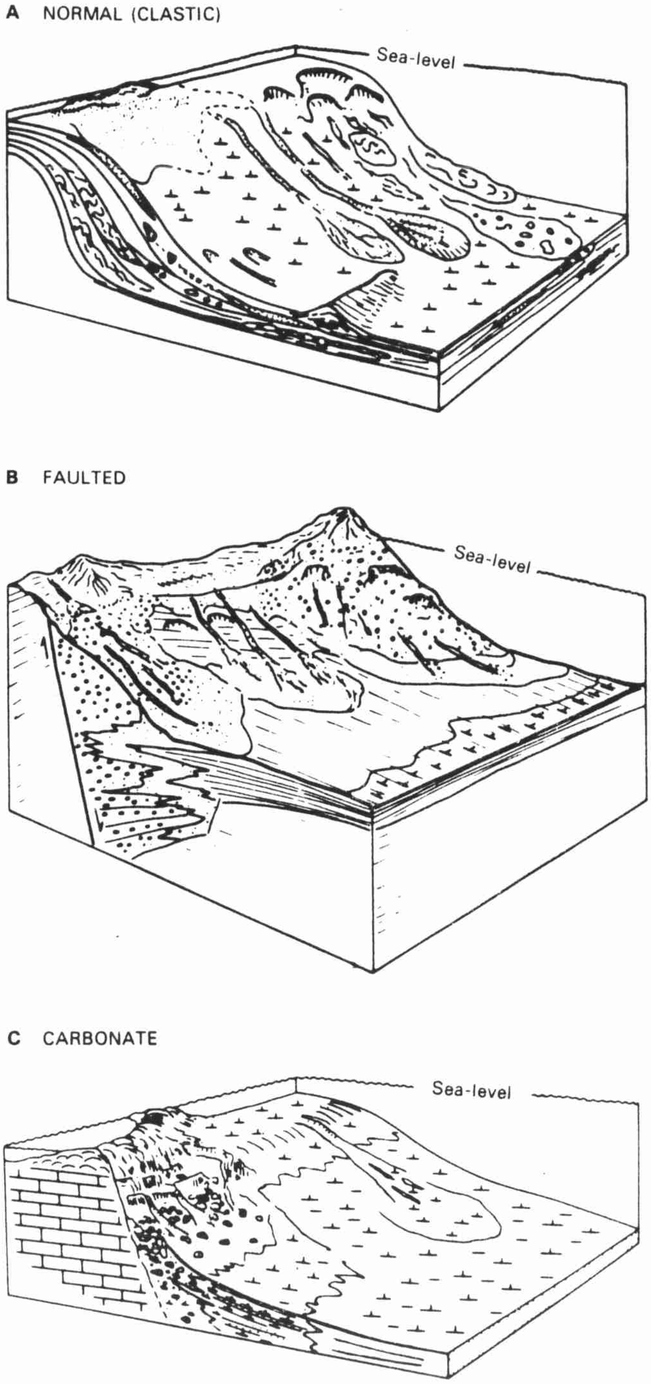


Fig. 12.22. Sedimentary environment models for submarine slope-aprons: (A) normal (clastic); (B) faulted, and (C) carbonate. Scales are variable, slope widths vary between 1 and 500 km, slope gradients are commonly from 1°–7° (from Stow, 1985).

Many slopes from around the world fit into this model for normal clastic slope-aprons, including those from passive and active margins, marginal seas and shelf basins.

The Nova Scotian slope off eastern Canada (Fig. 12.23) is an example that shows both constructional and destructional elements (progradational-erosional; King and Young, 1977), with an overall accumulation of 10–12 km of sediments since early Mesozoic time.

During the Plio-Quaternary glacial-interglacial cycles there were marked differences in sea-level and in sedimentary response. When sea-level was low (Fig. 12.23A) much of the continental shelf was exposed to subaerial processes. Ice covered the eastern provinces of Canada and, possibly, the present day shelf, while a major ice stream from the Laurentide ice cap fed through the Laurentian Channel to form a floating ice margin (Alam and Piper, 1978). Parts of the downslope received large amounts of sediment; other parts underwent significant slumping and erosion.

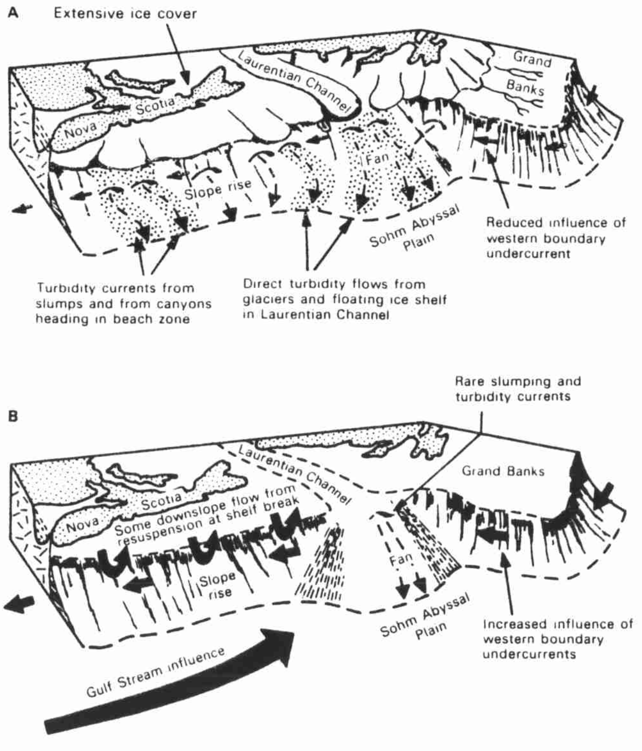


Fig. 12.23. The Nova Scotia normal slope-apron margin off eastern Canada, showing late Quaternary sedimentation history (after Stow, 1981). (A) Glacial: lowered sea-level, sediment supply direct to upper slope, turbidity current resedimentation dominant, slumping common, reduced bottom current activity. (B) Postglacial: rapid rise in sea-level, retreat of ice, broad submerged shelves, sediment resuspension at shelf break, hemipelagic sedimentation dominant, bottom currents active.

At the present day (Fig. 12.23B), the slope to the west is relatively smooth and gentle, whereas further east it is steeper and dissected by numerous channels and gullies. Several canyons indent the shelf break, but most of the channels head in waters greater than 400 m in depth and die out on the rise (Stanley, Swift *et al.*, 1972; Piper, 1975). The channels may be the result either of greater sediment supply leading to more slumping or to being kept open by rip-currents which supplied littoral sands to the canyon heads during low stands of sea-level. Slumps and slump masses of all sizes occur throughout, but are most common on the gullied slope and on the slope above the Laurentian Fan.

There is a patchy and irregular distribution of mainly fine-grained sediment facies over the Scotian slope. Sands and gravel are restricted to spillover sands on the upper slope off Sable Island, to the axes of canyons and channels crossing the slope and to the isolated lobes on the lower slope. A strong bottom current, the Western Boundary Undercurrent, is active at depths of 2–4 km, so that muddy contourites and turbidites are intimately interbedded in the sediments of the lower slope (Stow, 1979a; Shor, Kent and Flood, 1984).

Where thick slope-apron sequences are underlain by evaporites or low-strength mobile muds, the formation and intrusion of salt and shale diapirs can completely modify and control slope development. The slope profile becomes very irregular with highs and lows, disconnected channel segments, small isolated basins and slump and slide masses, and is constantly changing. Sediment facies distribution is equally irregular.

The best-known examples of normal slopes modified by diapiric activity are in the Gulf of Mexico (Bouma, Moore and Coleman, 1978; Bouma, 1981), off Angola-Gabon (Driver and Pardo, 1974) and off Nigeria (Whiteman, 1982; Fig. 14.11). The upper continental shelf in the northern Gulf of Mexico is a region of hummocky topography, with local high angle slopes, underlain by a multitude of salt and shale diapiric structures, some of which reach to within tens of metres of the sea floor. Growth faults, tensional faults, slumps and slides occur just below the shelf breaks and are closely associated with diapirism. This diapirism blocks canyons and causes interdomal lows and collapse depressions to form.

FAULTED SLOPE-APRONS

Where slopes are actively fault-controlled they commonly develop relatively steep portions alternating with flatter perched basins forming a complex stepped profile (Fig. 12.22B). Slump scars, slump masses and short-lived shallow channels are widespread. There is commonly an abrupt change of gradient at the foot of the slope to a flat basin floor, with little development of lower slope or rise. A thick *fault-scarp wedge* of sediment accumulates in a narrow trough at the foot of the slope. Sediment facies vary laterally as a result of the non-uniform, periodic nature of fault activity and the presence of faults perpendicular to the margin. These latter may serve as sites of long-term canyons and channels funnelling sediment out to submarine fans or basin plains beyond the foot of the slope.

Provided there is an adequate source of materials of all size grades, faulted slope-aprons may develop a slope-parallel distribution of coarse to fine-grained sediment. Near-fault, proximal facies can include rockfalls, debrites, gravel-rich turbidites and associated deposits. These die out rapidly away from the fault zone and interdigitate with sandy, muddy and biogenic facies. Lateral variability of facies commonly makes this simple slope-parallel distribution much more complex in reality. Intermittent synsedimentary tectonic activity has a marked effect on the vertical arrangement of facies. Fining-upward sequences from gravels to sands to muds occur repeatedly, following each new phase of tectonic movement.

Examples of fault-controlled slope-aprons are most common on strike-slip margins (Figs 14.51 and 14.52) and on early rifted margins such as those of the Red Sea rift system. They also occur around marginal basins such as the Caribbean (Case, 1974) and Tyrrhenian Sea (Wezel, Savelli *et al.*, 1981).

On the western margin of the Tyrrhenian Sea (Fig. 12.24) the relatively steep slope was controlled in the early stages by rifting and later by basin subsidence. It is erosional and step-like, with the steeper parts exposing older basement rocks or having a thin veneer of muddy sediments much affected by slumps, slides, sediment creep and debris flows. The coarse-grained facies occur in channel axes cutting across the slope, and in lobes and small fans that partly coalesce along the base of the slope to form a slope-parallel fringe. Distally these pass into sandy, silty and then muddy and pelagic sediments. The margin is not uniform along its length because the width of the shelf clearly affects sediment supply to the slope apron. Basinwards the slope system is interrupted to a greater or lesser extent by upfaulted basement ridges that also have an important effect on facies distribution, and may locally provide a secondary source of material for resedimentation.

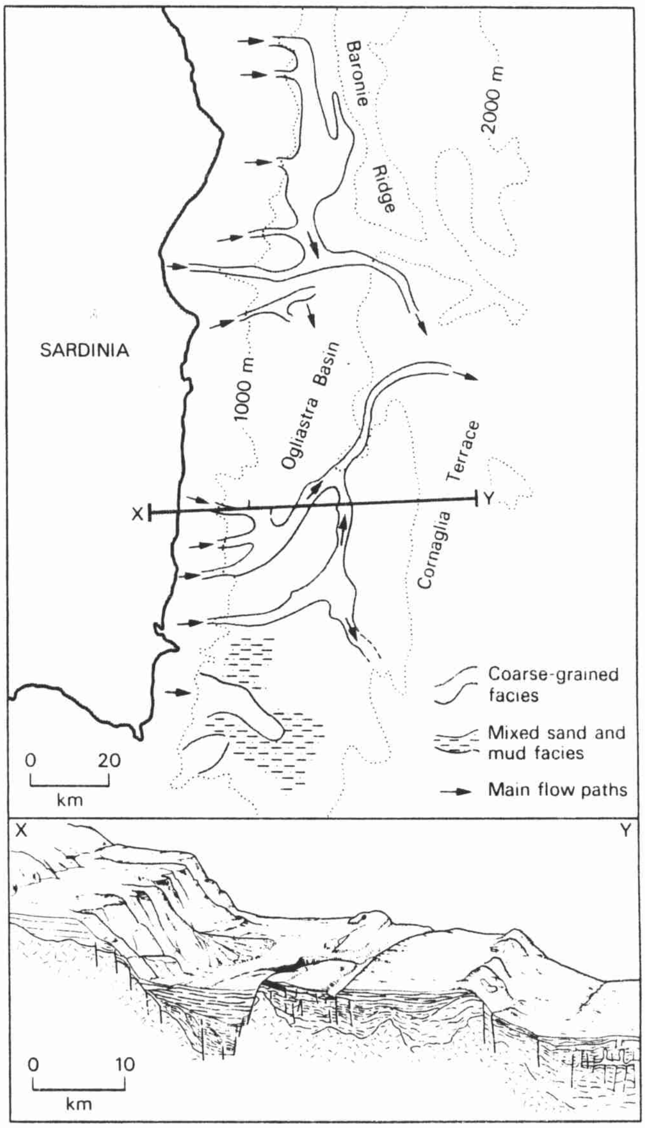


Fig. 12.24. The Sardinia faulted slope-apron and marginal basin in the Tyrrhenian Sea west of Italy (from Wezel, Savelli *et al.*, 1981).

CARBONATE SLOPE-APRONS

Three types of carbonate margin, can be identified: (1) the abrupt, reef-edge or carbonate-shoal *by-pass margin*; (2) the more gentle reef or carbonate-shoal *depositional margin*; and (3) the very gentle, highly-dissected *ridge-flank slopes* of oceanic ridges and other mid-basin highs (see Sect. 11.3). The composite carbonate slope-apron model (Fig. 12.22C) combines characteristics of by-pass and depositional margins.

By-pass margins can have very steep and often stepped portions of submarine cliff, where sediment is thin or absent, fringed by a peri-platform calcirudite talus wedge that grades rapidly downslope into calcarenites, calcilutites and pelagic/hemipelagic limestones. Locally, channels and canyons dissect the margin and funnel coarser sediments into deeper water, thus breaking up the slope-parallel facies distribution.

The depositional margin is more akin to the normal slope types for clastic sediments, with a gentle convex-concave profile and a more irregular distribution of resedimented-carbonate, bottom-current and pelagic facies. There is a general downslope fining trend, but this is commonly interrupted by slumping, debris flows and channelling of coarse material to isolated slope lobes.

The best known carbonate slope-aprons are around the

A RADIAL

B ELONGATE

C FAN-DELTA

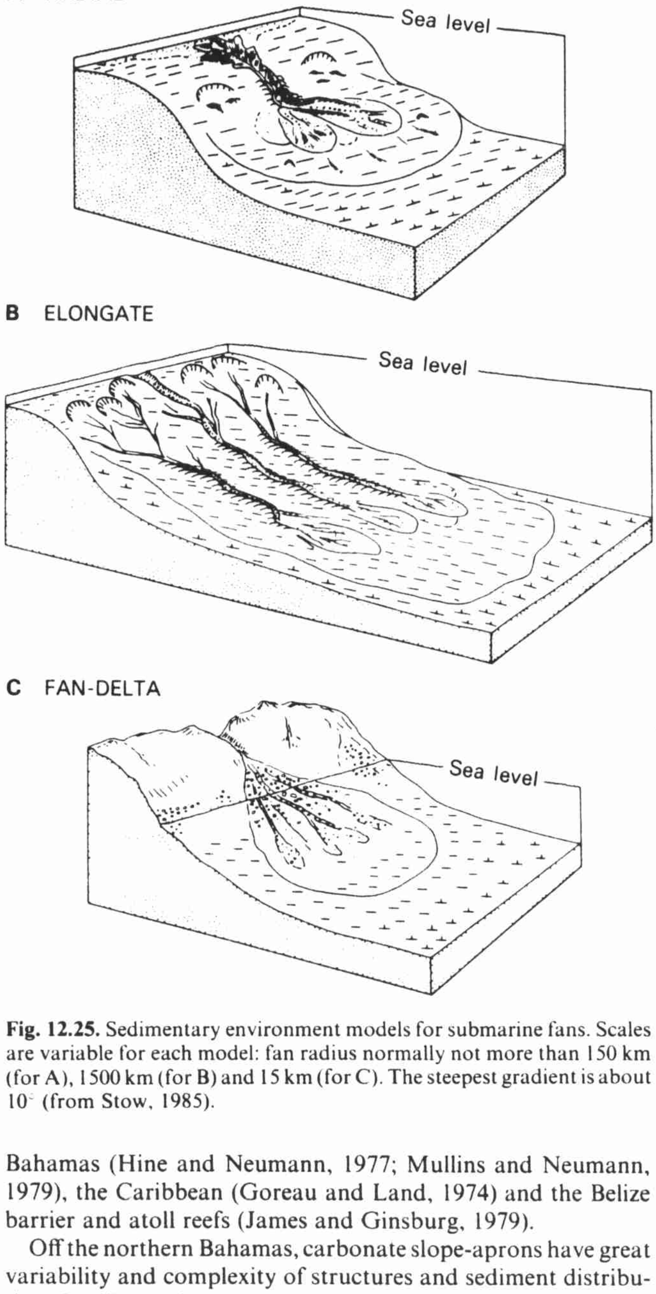


Fig. 12.25. Sedimentary environment models for submarine fans. Scales are variable for each model: fan radius normally not more than 150 km (for A), 1500 km (for B) and 15 km (for C). The steepest gradient is about 10° (from Stow, 1985).

Bahamas (Hine and Neumann, 1977; Mullins and Neumann, 1979), the Caribbean (Goreau and Land, 1974) and the Belize barrier and atoll reefs (James and Ginsburg, 1979).

Off the northern Bahamas, carbonate slope-aprons have great variability and complexity of structures and sediment distribution (Mullins and Neumann, 1979). The by-pass margin model can be subdivided into seven different types that face oceanward or towards a smaller seaway, that are windward or leeward, and that are extended or eroded. Important controls on the types and distribution of the carbonate slope facies are: the nature of offbank supply and sedimentation processes, basement faulting, oceanic circulation, submarine cementation and biological buildups.

The highly dissected ridge-flank slopes of oceanic ridges, linear island chains, oceanic plateaus and isolated seamounts have a distinctive irregular to concave profile with ocean-crust basement highs, perched basins and a thin but irregular sediment cover. They are cut transversely by fracture zone valleys or separated by stretches of flat ocean floor. On the flank of the Mid-Atlantic Ridge, resedimented carbonate debrites and calciturbidites occur adjacent to the Gibbs Fracture Zone (Faugères, Gayet *et al.*, 1982).

12.4.3 **Submarine fans**

Submarine fans are distinctive constructional features at the foot of slopes. Unlike slope-aprons which extend parallel to the margin, fans are isolated bodies that develop seaward of a major sediment source (river, delta, glacier, etc.) or main supply route (canyon, gully, trough, etc.)

They are very variable in size, from a radius of little more than 1 km to a length of more 2000 km, and have gradients similar to those of slopes, decreasing from the upper (2–5°) to lower fan (< 1°) region. The main morphological elements include one or more feeder channels, slump and slide scars and blocks, debris flow masses, broad channel levees, lobes built up at the end of channels and distributaries, and relatively smooth or current-moulded interchannel and inter-lobe areas. Upper, middle and lower or inner, middle and outer subenvironments have been described, although these divisions may not always be distinct and may differ between the small and very large fans.

A number of different fan models have been developed over the past 15 years (Normark, 1970, 1978, 1980; Nelson and Nilsen, 1974; Mutti and Ricci Lucchi, 1972, 1975; Walker, 1978, 1980; Nilsen, 1980; Stow 1981; Howell and Normark, 1982; Bouma, Normark and Barnes, in press) from studies of both modern and ancient systems. However, there appear to be two principal end-member types developed in deeper water, *radial* and *elongate* fans (Fig. 12.25A,B), with all possible gradations between the two (Stow, Howell and Nelson, 1984) and a third shallow-water type, or *fan-delta* (Fig. 12.25C). The 'deeper water' fans may occur in relatively shallow basins but are developed at the base of a slope, whereas fan-deltas are the marine continuation of alluvial fans and hence extend downwards from sea-level.

Submarine fans can rarely conform to the ideal shape because they are commonly constrained by basement relief and local topography. These effects are often amplified during fan growth by synsedimentary tectonic movements and differential sediment compaction. Various confinement modifications of the ideal fan models are therefore common. From work on the Upper Miocene Stevens sandstone of California, Scott and Tillman (1981) have developed an *on-lap* model for turbidites that lap onto a contemporaneously rising anticlinal surface, and a *confinement* model for turbidites that are confined to bathymetric lows between adjacent anticlines (Sect. 12.6.2). Where complete confinement of a fan results in the rapid fill of the entire confining basin, the system is more appropriately represented by a basin plain model (Sect. 12.4.4).

The distribution of facies on fans is also affected by the external influence of the Coriolis force. In the northern hemisphere this acts to deflect turbidity currents to the right, so that channelized flows will tend to construct higher levees on their right-hand bank and this may eventually force the whole channel to migrate to the left (Menard, 1964). The development and positioning of fan lobes will depend on the position and activity of the feeder channels as well as on local relief caused by earlier-formed lobes (Normark, Piper and Hess, 1979). Channel-margin slumps or large debris flow deposits can both serve to block channels and cut off sediment supply to lobes down-fan (Normark, Piper and Stow, 1983). Mutti and Sonnini (1981) have suggested that even the low positive relief presented by a single thick lensoid turbidite bed will slightly deflect the subsequent turbidity current thereby producing small-scale thinning-upward *compensation cycles*.

RADIAL FANS

Radial fans have a true fan-like shape developed concentrically about a single feeder canyon or channel and a concave-convex-concave longitudinal profile (Fig. 12.25A). Their radii range from a few kilometres to a few hundred kilometres at most and sediment thicknesses do not generally exceed 1 km. They are more or less synonymous with the sandy, low-efficiency, canyon-fed, small, restricted-basin and morphologically well-developed fans of other authors.

The upper fan is characterized by a concave-up profile with rugged topography and the presence of a main fan valley which may be straight or sinuous. It has levees which may be from a few tens of metres or over 200 m above the valley floor. The valley floor itself is depositional and may be elevated above the adjacent fan surface by many tens of metres. The valley width ranges from approximately 0.1–10 km. The middle fan environment is characterized by a convex-up profile with hummocky topography where the main fan valley splits up into many distributaries, called fan channels. The channels may meander or braid, be active or abandoned. Their axial depth may be several tens of metres and their width up to about 1 km. At the termination of the channels, in the lower part of the middle fan, depositional lobes occur. The lower fan environment has a concave-up profile and smooth topography with numerous small channels without levees. The boundary between the middle and lower fan environments is gradual and indistinct. In general, fan valleys and channels can be (1) depositional, (2) erosional or (3) mixed depositional-erosional (Normark, 1970).

There is both an elongate and concentric distribution of coarse to fine-grained facies over most radial fans. Slumps, slides and debrites, are confined to areas of the lower slope, upper fan and channel margin. Turbidites and associated facies are dispersed in two main ways across the fan surface. Much of the coarse-grained sediment is transported radially through the channels and is deposited along their length as thick elongate sand bodies or as sandy lobes that spread out at their terminations. Fine-grained sediment is transported either down the channels and then laterally by overflow onto the levees and into interchannel areas, or as thick unconfined low-density flows. Sand-mud ratios are therefore high inside channels and in the proximal parts of lobes and low in interchannel areas and more distal fan environments.

In the upper fan, channel sands generally are thick coarse-grained turbidites characterized by poorly developed Bouma sequences (T_{ae}) and grain flow/liquefied flow features (Lowe sequence). In the middle and lower fan the sands of channels and lobes are thick medium-grained turbidites with well-developed Bouma sequences (T_{abcde} or T_{bde}). Both distally downfan and laterally across the levees and interchannel areas, the turbidites become progressively finer-grained with base absent Bouma sequences (T_{cde} or T_{de}) and typical silt-mud turbidite features (Piper/Stow sequences).

Examples of modern radial fans include many of the smaller ones from the west coast of North America such as La Jolla, Navy, Redondo, Coronado, San Lucas and Nitinat (Normark, 1970; Normark and Piper, 1972; Normark, Piper and Hess, 1979). These are all single canyon or channel-fed, medium input, sand-dominated fans, in which a sandy *suprafan-lobe* system is developed on the middle fan at the termination of a leveed upper fan valley (Normark, 1970, 1980).

The La Jolla fan has long been considered a prime example of this type. Littoral sands moved by longshore drift into the head of Scripps canyon are funnelled downslope by a variety of resedimenting processes and are deposited as a series of lobes at the channel termination. The sand-mud ratio decreases down-fan as well as laterally away from the channel axis. However, extensive seismic reflection profiling has shown that fan development is due as much to the tectonics as to purely sedimentary controls, and that the La Jolla fan is a complex of smaller interwoven radial components (Graham and Bachman, 1983). Normal up-and-down canyon currents (Shepard, Marshall *et al.*, 1979) have further influenced the fan morphology and facies types and distribution.

Perhaps the most intensely surveyed of any modern fan is the Navy fan (Fig. 12.26), which has formed in the deeper water of South San Clemente Basin off southern California and is fed by an overflow channel leading from an upslope basin. It is on this fan that Normark and his colleagues have best documented the nature of lobe switching and growth, the mesotopography (channel segments, distributaries, depressions, hummocks, possible bed-forms etc.), and the fan-wide correlation of individual turbidites (Normark, 1978; Normark, Piper and Hess, 1979; Bowen, Normark and Piper, 1984).

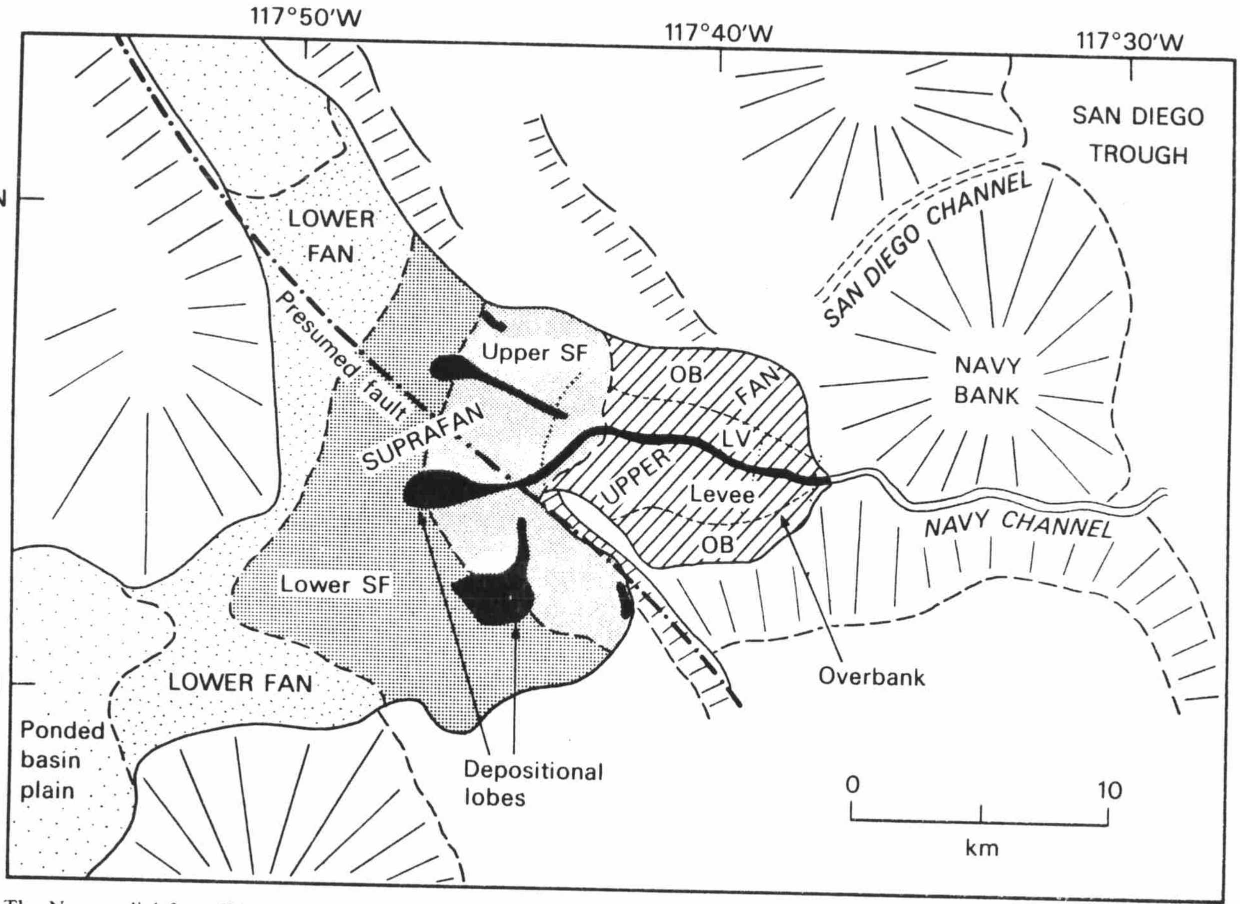


Fig. 12.26. The Navy radial fan offshore California (after Normark, Piper and Hess, 1979).

ELONGATE FANS

Elongate fans extend longitudinally, most commonly in a direction perpendicular to the supply margin; they often have two or more main feeder channels and a concave (irregular to smooth) longitudinal profile (Fig. 12.25B). They range in length from very small (< 5 km) to very large (> 1000 km) and may attain thicknesses in excess of 10 km in the upper regions. They have been variously called muddy, sand-deficient, high-efficiency, delta-fed, large and open-basin fans.

As in radial fans, upper, middle and lower divisions can be recognized, although their characteristics are different (Stow, 1981). The *upper fan* has a broad head that grades imperceptibly into the continental slope and not a distinct apex. It is irregular, slump-scarred, crossed by one or more main channels or troughs and by minor channels and tributaries between slump blocks. Erosion with limited deposition is the dominant process. Several main channels lead across a much extended *middle fan*, piling up large amounts of sediment on the levees and in interchannel areas. The channels may be built up above the fan surface or deeply incised; they are generally sinuous with large-scale meanders and parts that are braided. Many channels show a system of small-scale tight meanders (Fig. 12.20D) (e.g. Damuth, Kolla *et al.*, 1983; Bouma, Stelling and Coleman, 1984). The origin of these fluvial-like features is not yet understood. The channels serve to funnel sediments downslope, producing an elongate fan shape, then die out on the lower middle fan where they construct large *terminal lobes*. Provided that the receiving basin is sufficiently large, a smooth *lower fan* can be built out a long way to merge imperceptibly with the basin abyssal plain.

The pattern of sediment distribution is in many ways similar to that of the radial fan although more elongate than concentric, and with more mud than sand. There are perhaps more slumps, slides and debrites over the upper fan, but a very similar down-channel concentration of coarse-grained sediments and proximal to distal transitions of grain size and structures. The terminal lobes may be sandy or silty. Fine-grained turbidites show both a down-fan and away-from channel evolution of textural, structural and compositional features (Stow, 1981). Hemipelagites and, in some cases, contourites are interbedded with the resedimented facies in areas of lower energy.

Examples of elongate fans in open ocean basins include the giant (3000 km long) Bengal fan (Curray and Moore, 1974), the Indus (Kolla and Coumes, 1984), Mississippi (Bouma, Stelling and Coleman *et al.*, 1982), Amazon (Damuth and Kumar, 1975), Zaire (Weering and Iperen, 1984) and Laurentian (Stow, 1981) fans. In the Mediterranean Sea the Rhône fan (Normark, Barnes and Coumes, 1984) and the Nile fan (Maldonado and Stanley, 1976) both tend towards the large elongate type; whereas the very much smaller (10 km long) Crati fan (Colella, 1981; Ricci Lucchi, Colella, 1983) may be considered as a cross between a highly gullied slope and an incipient elongate fan. Normark (1980) considers the 1 km long Reserve fan (Normark and Dickson, 1976) in Lake Superior as more akin to the elongate than radial type.

The Laurentian fan (Uchupi and Austin, 1979; Piper, Normark and Stow, 1984) extends over 600 km southeastwards from the base of the slope off the Laurentian Channel on the east Canadian margin to merge with the Sohm Abyssal Plain at a depth of 5.2 km (Fig. 12.23). It has been the major depocentre off Nova Scotia since the early Tertiary and has accumulated several kilometres of sediments in that period. Its present morphology and sediment characteristics have been strongly influenced by onshore glacial history and several of the main channels have been incised to over 500 m depth. It is relatively inactive at the present day.

FAN DELTAS

Fan deltas (also called short-headed delta-front fans) are the subaqueous part of alluvial fans that prograde from highland directly into a standing body of water (lake or sea). They are mostly relatively small (< 10 km radius) and thin (< 100–200 m of sediment), pear-shaped in outline and with an ephemeral system of shallow braided channels radiating downslope from the fan head. Channels may be up to some 200 m in width and 30 m in depth, and commonly originate some way down the delta slope. A lateral division of channel, levee and interchannel occurs where the conduit is sufficiently long-lived.

These fans are generally coarser-grained than either of the other types, with gravels and sands dominant in the upper reaches and in the channels. The levees, interchannel areas and lower reaches of the fan receive more muddy sediments, both fine-grained turbidites and hemipelagites. Two main types have been recognised by Westcott and Ethridge (1980). The first type, based on the Yallahs fan delta off southeast Jamaica (Fig. 14.53), is characteristic of truncated subaerial fans that prograde directly onto steep continental or island slopes. Proximal, gravelly braided-stream deposits grade seaward into gravels and sands at the coastline and to muddy gravels and muds on the slope. The second type, based on fans along the southeast coast of Alaska (Boothroyd, 1975; Galloway, 1976; Boothroyd and Nummedal, 1978), is characteristic of more completely developed subaerial fans that prograde onto continental or island shelves. They grade from proximal braided stream deposits through well-laminated nearshore sand to distal burrowed shoreface muds. These two types represent end members of a spectrum of fan deltas that have been described from all types of coasts and margins around the world.

12.4.4 Basin plains

Basin plains are flat and relatively deep. They vary widely in their areal extent from tiny slope basins to the major oceanic abyssal plains (1.5 million km^2), and from quite shallow depths to the deep floors of submarine trenches up to 10 km deep. They generally have a very gentle relief that results from the smoothing and burying of pre-existing topographic irregularities by turbidite fill, and merge gradually or more abruptly with the surrounding slopes and isolated seamounts or other basement highs that remain. They may be elongate, equidimensional or irregular in shape. Sediment thicknesses below most basin plains are a few hundred metres, though in some fault-bounded basins subsidence accompanying sedimentation can lead to thicknesses of several kilometres.

Basin plains function as the ultimate trap for sediments eroded from the continents and from submarine highs, the most extensive basin plains being located seaward of the major drainage basins of the world. A single basin plain may be fed by several sources, including submarine canyons, channels and fans and the surrounding basin slopes. Their main morphological elements include the extreme distal portions of submarine fans, channels and lobes, very large areas of smooth or current-modified sea-floor, as well as isolated intra-basinal channels, ridges and drifts, structurally-controlled graben and morphologically-restricted passages.

Several different basin classifications have been suggested using the criteria of composition (terrgenous versus carbonate), of basin restriction (open versus enclosed), of fill geometry (progradational, mounded, onlap and drape fills), of depth (above and below CCD) and of sediment supply (undersupplied versus oversupplied) (Fig. 12.27).

There are a number of interacting variables that control the sediment supply, facies types and distributions within the basin plains. The most important of these are basin geometry, tectonics and source area (Pilkey, Locker and Cleary, 1980). For example a large basin plain in an area of little tectonic activity has a low sediment supply compared to its size while a small basin plain in a tectonically active area has a high ratio of sediment supply to basin size (Fig. 12.27).

Abyssal plains are extensively developed in the Atlantic and

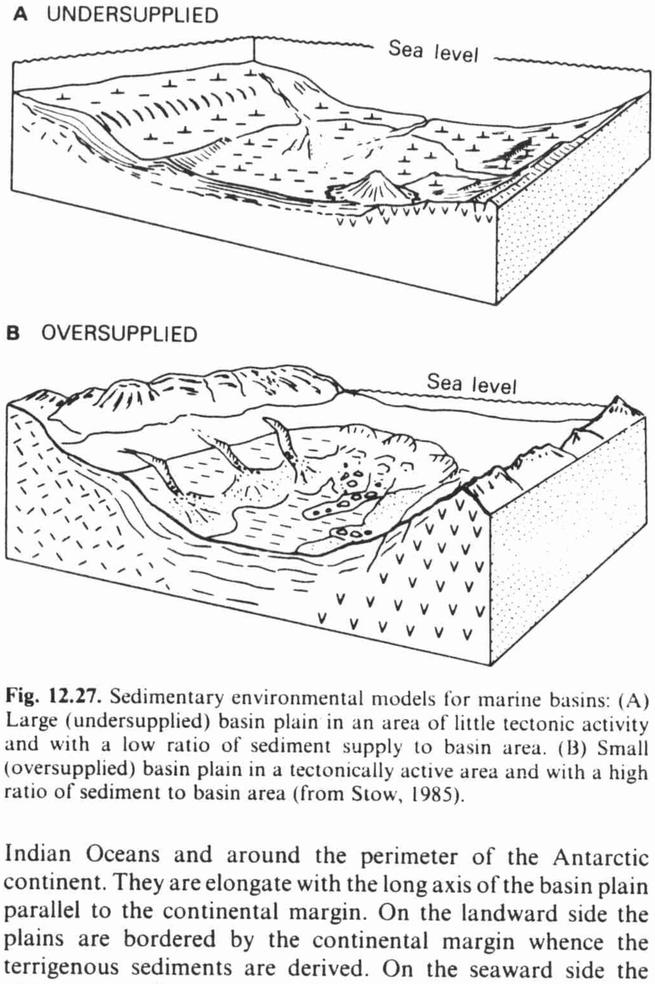


Fig. 12.27. Sedimentary environmental models for marine basins: (A) Large (undersupplied) basin plain in an area of little tectonic activity and with a low ratio of sediment supply to basin area. (B) Small (oversupplied) basin plain in a tectonically active area and with a high ratio of sediment to basin area (from Stow, 1985).

Indian Oceans and around the perimeter of the Antarctic continent. They are elongate with the long axis of the basin plain parallel to the continental margin. On the landward side the plains are bordered by the continental margin whence the terrigenous sediments are derived. On the seaward side the plains encroach upon abyssal hill and mid-ocean ridge provinces. Average rates of sedimentation are of the order of a few centimetres per thousand years. The flat surface of the plains may be interrupted by protruding abyssal hills or seamounts.

The Hatteras Abyssal Plain (Fig. 12.28A) is a large and elongate *primary plain*, having a length of approximately 1000 km and an average width of approximately 200 km and average depth of 5500 m (Pilkey, Locker and Cleary, 1980). The maximum thickness of the sedimentary fill is some 350 m. The main source of terrigenous sediment is at the northern end. The shallow gradient (1:83) and main sediment dispersal are from north to south, and thickness of the sedimentary fill, individual bed thickness and grain size decrease in down-current and across-current directions. The continuity of individual turbidite sand layers is high, covering some 60% of the basin floor, but the

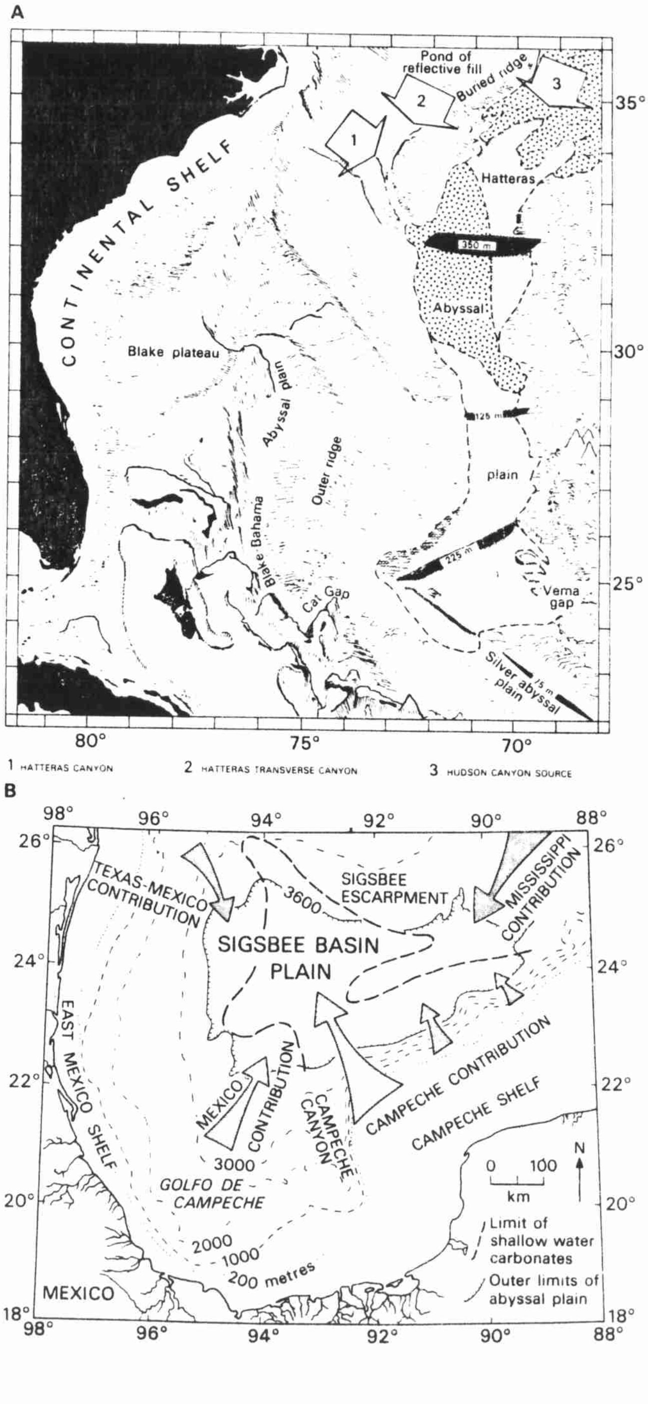


Fig. 12.28. Present-day open ocean abyssal plains: (A) the Hatteras Abyssal Plain in the western North Atlantic; turbidite supply mainly from the north, sands and silts in coarse stipple, silts and muds in light stipple (after Horn, Ewing and Ewing, 1972); (B) the Sigsbee Abyssal Plain in the Gulf of Mexico; centripetal distribution of both terrigenous turbidites from the Mississippi, Texan and Mexican shelves and bioclastic turbidites from the Campeche shelf (after Davies, 1968).

frequency of sand layers is relatively low. There is a distinct longitudinal transition from proximal to distal facies. The turbidity currents that supply the plain must reach an enormous size, but their relative infrequency and the large basin size result in a low ratio of sediment supply to basin size and a sheet-like depositional geometry.

The Nares Abyssal Plain is a *secondary basin plain* which is fed through the Vema Gap at the downcurrent end of Hatteras Abyssal Plain (Heezen and Laughton, 1963; Horn, Ewing and Ewing, 1972). The gradient slopes away from the abyssal gap, through which terrigenous sediment is supplied by turbidity current overflow after the turbidity currents have traversed the primary Hatteras plain for distances up to 1000 km or more. As a result, the turbidite fill of secondary plains consists mainly of graded silts and muds, and interbedded pelagites.

The abundance of large basin plains in the Atlantic, and their scarcity along the perimeter of the Pacific is due to two main features: (1) the Atlantic continental margins are predominantly passive margins, into which many large drainage basins of the world empty (Inman and Nordstrom, 1971); and (2) the Pacific continental margins are predominantly active margins away from which the main continental drainage patterns flow and where volcanic arcs, back-arc basins and trenches act as barriers and traps to terrigenous sediments.

Marginal seas also have basin plains such as the Sigsbee Abyssal Plain in the Gulf of Mexico (Davies, 1968) (Fig. 12.28B), the Balearic Abyssal Plain in the Western Mediterranean Sea (Horn, Ewing and Ewing, 1972; Rupke and Stanley, 1974; Rupke, 1975) and the Black Sea basin plain (Degens and Ross, 1974). As a result of the enclosed nature of the basin, the basin plains are fed by turbidity currents from widely varied sources (Fig. 12.28B). The deepest part of the basin is generally the central part where gradients are very low or where the floor may be level. The dispersal pattern of turbidites on the basin plain is approximately centripetal and ponding of turbidity currents is common. Turbidites from both terrigenous and bioclastic sources may interdigitate, as in the Sigsbee Abyssal Plain, and are commonly composed of fine-grained silt, mud or calcilutite. The thickness and continuity of individual turbidite sands in most marginal basins tend to be low because of the relatively small size of turbidity currents. However, the frequency of turbidites is high as there are several active source areas. Holocene sedimentation rates range from 10 to 20 cm/1000 years, but during the last glacial phase these figures were several times higher (Rupke, 1975).

The floors of some *deep-sea trenches* may consist wholly or in part of basin plains that are elongated parallel to the continental margin, up to hundreds of kilometres in length and, at most, only a few tens of kilometres wide. Transverse lines may break the trench floor into compartments some of which are sediment starved, others well supplied (Fig. 14.29).

Proximity and high relief of source areas and the presence of volcanic activity provide compositionally immature clastics to the trench basin plain (Underwood and Bachman, 1982). Their dispersal by turbidity currents is predominantly longitudinal. Distinct marginal facies may develop alongside the basin plain. In the Middle America Trench terrigenous silty clay predominates on the landward flank, whereas on the seaward flank biogenic sediment accumulates (Ross, 1971). On the landward side of the Aleutian basin plain, parallel to its longitudinal axis, there is a 2.5–6 km wide deep-sea channel with a marked levee on its seaward side (Hamilton, 1967). Sand is present on the channel floor, whereas silt and muds are deposited on the levee and on the trench floor. However, at present the plain has no connection with a terrigenous source area and the turbidite succession is overlain by some 100 m of pelagites. Tectonic activity and steep (up to 10°) inner trench walls have led to the emplacement of slope sediment on the basin plain by slumping. The slumping direction is generally perpendicular to the predominant longitudinal dispersal of turbidites (Piper, von Heune and Duncan, 1973).

In *strike-slip* settings, basin plains occur along the Californian margin, where they are also known as *borderland basins* (Figs 14.50 and 14.51) (Gorsline, Karl *et al.*, 1984), and along the New Zealand margin (Spörli, 1980). Sediment supply to the basin plains may be from several sides, although one active fault margin is usually dominant. The facies types depend on the surrounding source areas. There may be access to coarse sands and gravel, major slumping and debris flow of finer-grained sediments, and a significant input of hemipelagic material from continental sources. Typical proximal to distal changes in grain size and bed thickness occur in a direction away from the margins and towards the basin centre, although ponding of flows may result in increased sediment thickness in the central parts. Turbidite sand layers tend to be discontinuous because turbidity currents are relatively small, whereas the frequency of layers is high due to frequent tectonic triggering of flows (Pilkey, Locker and Cleary, 1980).

Small *slope basins* (Fig. 14.32) along tectonically active or diapirically affected margins can either be sediment starved or fill up rapidly with terrigenous or volcaniclastic sediments and spill over downslope or into lower slope basins. A series of interconnected slope basins have been described from the Hellenic Arc margin (Got, Monaco *et al.*, 1981; Got, 1984) (Fig. 12.29) and from the vertically-faulted margins of the Tyrrhenian Sea (Wezel, Savelli *et al.*, 1981). These basins are well supplied with a mixed sediment load derived partly from slumping on the basin slopes and partly from the interbasinal connecting channels that serve to funnel material through the system to the deepest basin.

12.5 ANCIENT DEEP-SEA SYSTEMS: RECOGNITION

Although we now have well-constrained facies models, relating deep-sea sediments to depositional processes, and relatively

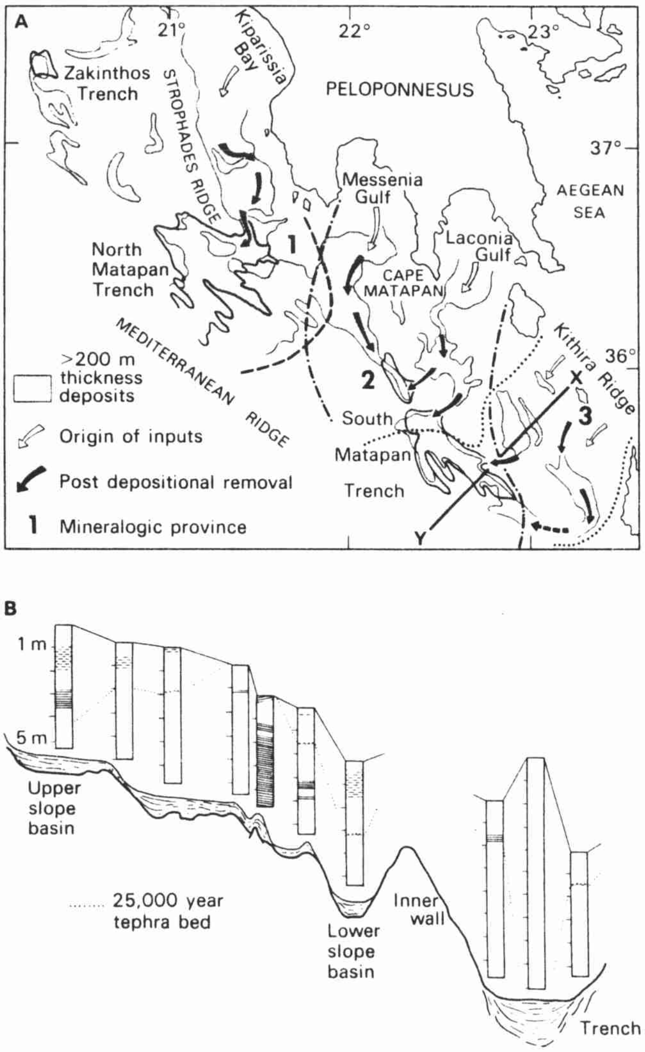


Fig. 12.29. Interconnected slope basins on the southwestern margin of the Peloponnesus, Greece (from Got, 1984): (A) general map showing three distinct systems of interconnected basins based on analyses of heavy minerals and clay minerals; (B) detail of one system from the Kithira Ridge to Matapan Trench (XY) showing correlation of cores along a typical transect.

well-documented environmental models showing the range of variability that exists in modern systems, it is not always easy to recognize and interpret ancient deep-sea sediments. In this section some of the methods and problems of interpreting ancient rocks are illustrated and a summary presented of the main features by which we can recognize ancient slope-apron, submarine fan and basin plain systems.

12.5.1 Scale, preservation and bathymetry

The use of modern analogues in interpreting ancient deep-sea systems has always been limited, to some extent, by the operational limitations of investigations carried out at sea. The problem is largely one of *scale* (Normark, Piper and Hess, 1979). The Bengal fan or the Sohm abyssal plain are each equivalent to or larger than the entire Alpine-Carpathian fold belt of Eurasia. Even a smaller fan such as Navy fan or the slope of a borderland basin off California are an order of magnitude larger than most single continuous outcrops.

Using conventional surface ship surveys, such as echosounding, the limits of definition are morphological features around 20 m deep and 1 km wide. The increased use of deeply-towed instrument packages has permitted resolution of features one or two orders of magnitude smaller than this, and hence is more compatible with good-sized outcrops of ancient rocks. Submarine photography and surface sampling with various coring devices produces data of a similar scale to many outcrop studies. However, the accuracy limits of navigation and of positioning equipment on the sea floor make it difficult to precisely relate these data to actual morphologies.

Studies of ancient rocks on land are also hampered by problems of *preservation-potential* (Sect. 2.2.3). (1) The recognition of individual facies is the starting point for most studies, but both compaction and diagenetic changes may be considerable. (2) Some medium-scale morphological features, such as channels and the subtle bed thickness variations over a sediment lobe, can be detected in outcrop, but the original large-scale topography of slopes, fans and basin plains is rarely preserved. (3) Large-scale features such as these may be uplifted and preserved but usually require careful tectonic unravelling to reconstruct the original palaeoenvironment.

In subsurface studies, the methods and the problems are rather different (Sect. 2.3). (1) The first approach is via remote sensing techniques of which seismic profiling is the most useful sedimentologically. Although very good for recognizing the large- and medium-scale morphological features, as well as synsedimentary and subsequent tectonic movements, seismic facies are not readily interpreted in terms of facies types and sequences. (2) The combination of wireline logs and borehole samples provides a very powerful tool for facies and sequence analysis, but individual wells are of narrow diameter and mostly widely spaced.

Another difficulty with ancient successions, both exposed and subsurface, is one of bathymetric interpretation. There are five main methods by which we may obtain some idea of *palaeobathymetry*. (1) Turbidites and other resedimented facies require a sufficient length of subaqueous slope for the development of the depositional process, and must have been deposited below average wave base to allow for preservation. This generally implies a depth greater than about 50 m, although both wave-modified and storm-modified turbidites have been recognised. (2) Some benthonic foraminiferal assemblages, especially from the Tertiary, can be related to present day assemblages whose depth is known (e.g. Douglas and Heitman, 1979; Gradstein and Berggren, 1982). Resedimented facies commonly display a mixture of reworked shallow water foraminifera or other biogenics, pelagic biogenics and deep-water forms. (3) Trace fossil communities have been proposed that are related to water depth (Seilacher, 1967, 1978) (Fig. 12.30): a shallow shelf assemblage (*Cruziana*), a slope assemblage (*Zoophycos*) and bathyal assemblage (*Nereites*); however, other environmental factors, such as sediment supply, nutrient availability, grain-size and redox conditions, may completely modify this depth zonation (Wetzel, 1984). (4) Actual geometries of rock successions may be preserved, so that a shelf-slope-basin system for example, may allow a minimum palaeodepth to be measured (e.g. Galloway and Brown, 1973). (5) In certain cases, the tectonic history of a region is known or fairly well constrained so that palaeoreconstruction is, to some extent, possible. For ancient rock successions in wells beneath present-day oceans, the back-tracking method based on cooling and subsidence of oceanic crust (Sclater, Anderson and Bell, 1971) can be utilized.

12.5.2 **Horizontal facies distribution**

Lack of exposure and tectonic disturbances make it very difficult to determine the horizontal arrangement of facies in ancient rock successions. Facies models are therefore based largely on other data and many fan models, in particular, have been developed from vertical sequences (e.g. Walker, 1978). Occasionally, however, facies can be traced laterally for up to 2–3 km and broad palaeogeographies can be reconstructed by correlating isolated sections. One method of constructing horizontal facies patterns is to document the systematic lithological changes from ***proximal*** (close-to-source) to *distal* (far-from-source) turbidite facies as observed in a downcurrent direction (Walker, 1967) (Table 12.2; Fig. 12.31): sandstone/shale ratio, sandstone thickness, grain size and erosive features (amalgamated sandstones, channels) all decrease; scour marks

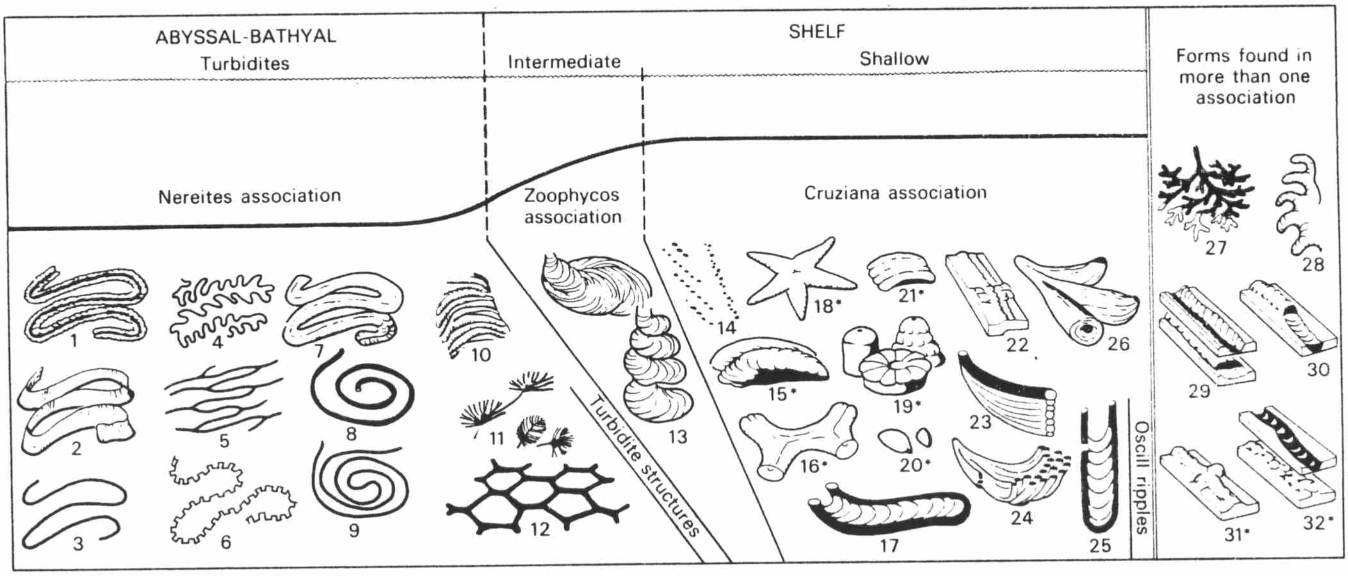


Fig. 12.30. Common associations of trace fossils and their environmental significance, particularly with regard to water depth (after Seilacher, 1967). 1, *Nereites* S. Str.; 2, *Dictyodora*; 3, *Helminthoida*; 4, *Cosmorhaphe*; 5, *Urohelminthoida*; 6, *Paleomeandron*; 7, *Scolicia* (Meandering); 8, *Spirophycus*, *Spirodesmos*; 9, *Spirorhaphe*; 10, *Lophoctenium*; 11, *Oldhamia*; 12, *Palaeodictyon*; 13, *Zoophycos*; 14, Arthropod tracks; 15, *Cruziana*; 16, *Thalassinoides*, *Ophiomorpha*; 17, *Rhizocorallium*; 18, *Asteracites*; 19, *Bergaueria*, *Conostichus*, *Solicycl*; 20, *Lockieia* (= *Pelecypodichnus*); 21, *Curvolithus*; 22, *Gyrochorte*; 23, *Teichichnus*; 24, *Phycodes*; 25, *Diplocraterion*; 26, *Asterosoma*, *Rosselia*; 27, *Chondrites*; 28, *Phycosiphon*; 29, *Scolicia*; 30, *Taenidium*; 31, *Fucusopsis*; 32, *Nerites* (*Scalarituba*).

Table 12.2 Characteristics of proximal, medial and distal turbidites

	Proximal (coarse-grained)	Medial (medium-grained)	Distal (fine-grained)
Bed thickness	Thick	Medium and thin	Thin beds and laminae
Bed shape	Irregular; lensing, channels and washouts common	Parallel-sided; regularly bedded	Parallel-sided beds and laminae, also discontinuous laminae
Sand/Mud ratio	SS/MD high, amalgamation of sandstones, thin mudstone partings and layers	SS/MD medium, rare amalgamation, well-developed mudstone layers	SS/MD low, mudstone dominant
Grading	Beds often ungraded or poorly-graded, some negative grading	Grading commonly well-developed	Grading often subtle and on very small scale
Facies models	Bouma AE sequences and Lowe sequences common	Classical Bouma sequences common ($_{abcde}$, $_{bcde}$, etc)	Stow/Piper sequences common, Bouma $_{(c)de}$ and $_{e}$ divisions only
Stratification	Large-scale parallel and cross-stratification common	Lamination, ripples and convolute lamination common	Interlaminated siltstone and mudstone common, micro-cross-lamination etc.
Top and bottom structures	Base sharp, commonly scoured; top often sharp	Base sharp, minor scours; top usually graded	Base sharp, more rarely gradational, micro-scours; top sharp or gradational
Bioturbation	Mostly absent	Can be well developed in mudstone layers	Can be well-developed; micro-bioturbation also common
Deformation structures	Slump and dewatering structures common	Minor slump and dewatering structures	Siltstone loading and balling in mudstone layers can occur
Grain size	Gravel and coarse-sand size dominant	Medium-fine sand size and interbedded mud-grade	Very fine sand and silt-size with mud-grade dominant
Sorting	Often poor	Moderate	Moderate to well-sorted
Composition	Immature and mixed components	Moderate maturity, compositional grading common	Mature, compositionally well-sorted
Associated facies	Slumps and debrites	Fine-grained turbidites, some hemipelagites	Medium-grained turbidites, contourites, hemipelagites and pelagites

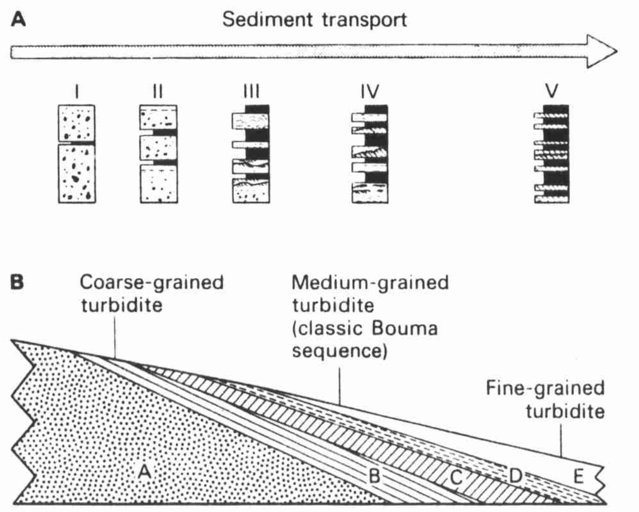


Fig. 12.31. Proximal-distal characteristics in turbidites: (A) Schematic illustration of downcurrent decrease of bed thickness, grain size and sand-mud ratio in a turbidite sequence (from Einsele, 1963); (B) idealized lateral variations in a single turbidite bed (after Corbett, 1972).

such as flutes become fewer and tool marks such as grooves become more in number; beds become more regular, parallel-sided, better-graded, laminated and cross-laminated. Walker (1967) also proposed a proximality (ABC) index based on the percentage of beds in a turbidite sequence of which the base begins with Bouma division $_a$, $_b$ or $_c$ (Fig. 12.31B).

These proximal to distal changes (Walker, 1967) can be extended further downcurrent to include very fine-grained, thin-bedded turbidites (Mutti, 1977; Piper, 1978; Stow and Piper, 1984). (Table 12.2). Walker's distal facies are in fact intermediate (or *medial*) between his proximal turbidites and truly fine-grained turbidites.

Fine-grained turbidites also have their own characteristic proximal-to-distal evolution of grain size, sorting, structures and thickness on the open slope, that parallels the changes observed in coarser-grained sediments confined to channels. Exactly comparable changes are observed in silt and mud turbidite laminae over hundreds of kilometres downslope, tens of kilometres across levees, and a few centimetres upwards through a graded laminated turbidite unit. (e.g. Nelson and Nilsen, 1974; Mutti, 1977; Nelson, Normark *et al.*, 1978; Stow, 1981).

It is important to remember, however, that (1) turbidites of both proximal and distal aspect may be laterally juxtaposed in certain deep-sea settings. For example, sand or gravel-filled channels may cut through mud-dominated slope-aprons and, therefore, because of the radial shape of many fans, a 'distal' slope facies may pass basinwards into a 'proximal' fan facies. (2) The grain-size and bed thickness of the deposit is partly controlled by the type and amount of material in the source area. Therefore fine-grained 'distal' turbidites may be deposited close to source if no coarser sediment is available. Unless true proximal-distal relationships can be demonstrated from palaeocurrent or other evidence, it is preferable to use the terms coarse-grained (thick-bedded) and fine-grained (thin-bedded).

12.5.3 **Palaeocurrents and palaeoslopes**

Critical to our understanding of the geometry and disperal pattern of an ancient rock series is an analysis of *palaeoslope inclinations* and *palaeocurrent directions*. The orientation of slump folds are particularly useful for interpreting the palaeoslope, whereas palaeocurrent direction may be inferred from measurements of grain orientation, clast and fossil alignment, clast imbricaton, sole marks (grooves and flutes), current lineation and ripple marks. These primary indications can then be supplemented with data on grain size, bed-thickness, compositional and other trends, bearing in mind problems of facies juxtaposition.

Many such measurements from ancient turbidite sequences have been carried out (e.g. Potter and Pettijohn, 1963). Very commonly, a dominant longitudinal dispersal pattern has been documented in an elongate basin, with several minor marginal sources bringing sediments downslope at right angles to the basin trend (Enos, 1969; MacDonald and Tanner, 1983). Rather less commonly, analysis reveals a radial dispersal pattern that might be related to deposition on a radial fan (e.g. Kruit, Brouwer *et al.*, 1975; Rupke, 1977; Link and Nilsen, 1980). The interpretation of the actual palaeoslope direction from slump-fold axes is a particularly important addition to simple palaeocurrent measurements (e.g. Woodcock, 1976a, b, 1979).

Palaeocurrent patterns are often complex (Lovell and Stow, 1981). Parkash and Middleton (1970) have shown meandering trend lines for turbidity current flow in the Ordovician Cloridorme Formation in Quebec. The 90° disparity found in current directions within a single rock series may be the result of alongslope reworking of turbidites by bottom (contour) currents (e.g. Stow and Lovell, 1979).

12.5.4 **Vertical facies sequences**

Within ancient deep-sea successions, there are many kinds of *vertical sequence* of different orders of magnitude (Ricci Lucchi, 1975; Rupke, 1977; Shanmugam, 1980; Stow, Bishop and Mills, 1982). At one extreme is the basin-fill sequence, from several hundreds of metres to several kilometres thick, and at the other extreme is the individual graded bed, from less than one centimetre to over a metre thick. The former is primarily tectonic in origin, the latter sedimentary. In between these two extremes, sequences of a few decimetres to a few tens of metres exist that result from an interplay of sedimentary, topographic, tectonic and sea-level effects. These sequences have been recognized on land, in wells and from DSDP sites in the deep sea (Figs 12.32 and 12.33). The sequences are based on grain-size, bed thickness and facies changes and appear to be characteristic of specific morphological elements. There has as yet been relatively little drilling of modern examples to confirm or refine these (but see Bouma, Coleman *et al.*, 1983).

Several types of canyon or channel-fill are recognized: a blocky, massive coarse-grained fill of a canyon or proximal channel; more regular fining (thinning)-upward sequences of mid-slope or mid-fan channels; packets of sands deposited in

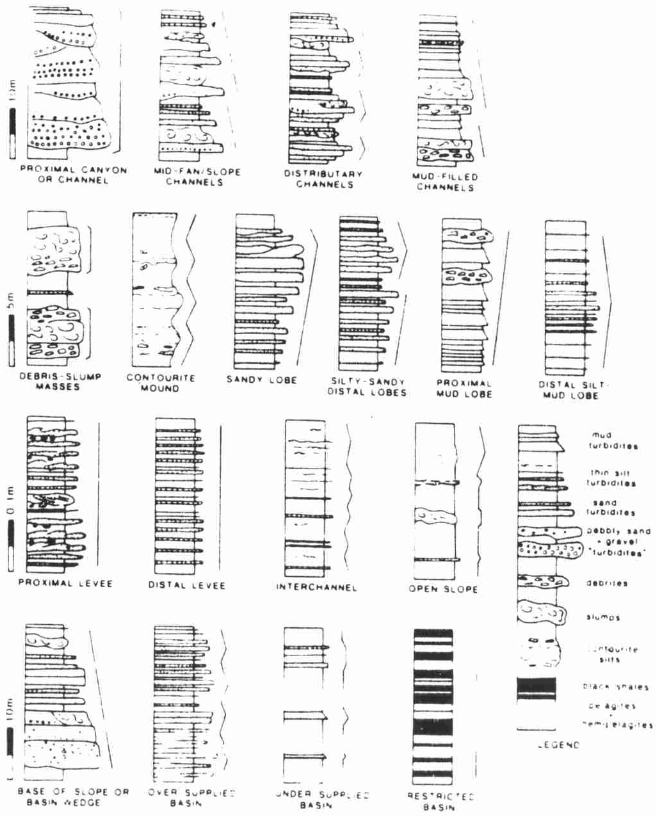
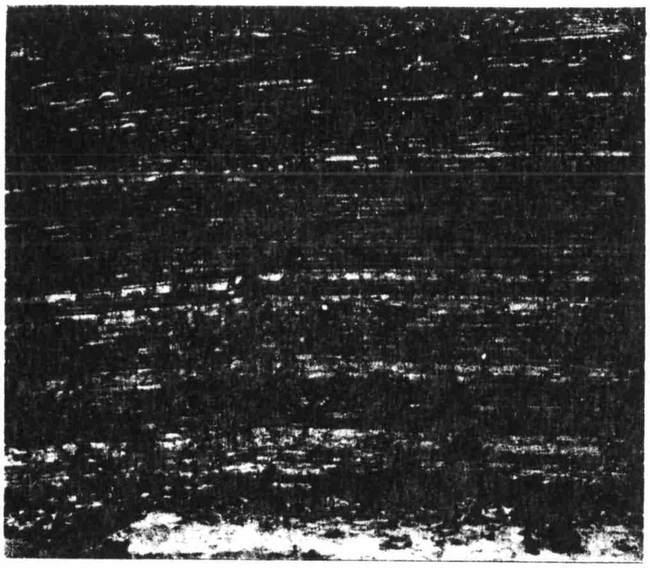


Fig. 12.32. Vertical sequences of turbidites and associated sediments for various morphological elements in the deep-sea. Fining (thinning)-upward, coarsening (thickening)-upward, blocky, symmetrical and irregular sequence types are indicated by the lines to the right of lithological columns (from Stow, 1985).



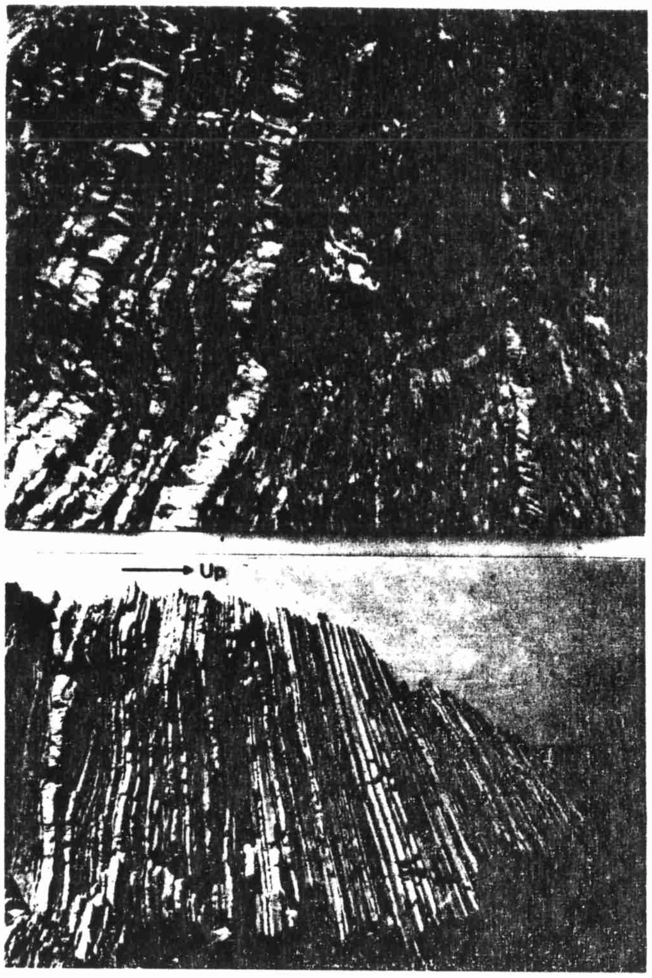


Fig. 12.33. (A) Fining/thinning-upward sequence over 12 m, Cretaceous-Tertiary calciturbidites, Scaglia Rossa Formation, Italy. (B) Coarsening/thickening-upward sequence over 15 m, Cambro-Ordovician clastic turbidites, Meguma Formation, Canada. Sequence youngs to left. (C) Regularly-bedded turbidites (dark) and pelagites (light) with no distinct sequences, Palaeocene flysch, NW Spain, near Zumaya.

distributary channels; and a blocky to fining-upward mud-dominated channel-fill. Regular coarsening (thickening)-upward sequences appear typical of more proximal sandy (suprafan) lobes and probably also of proximal muddy lobes, whereas more symmetrical sequences characterize distal (terminal) sandy and muddy lobes. Other mounds on the sea floor include contourite drifts with an irregular variation of more or less sandy and silty hemipelagic-like muds, and slump or debris flow masses with a chaotic assemblage of slumps and debrites. Either irregular sequences or fining-upward/coarsening-upward sequences can be characteristic of levee, interchannel, smooth slope and several basinal environments. The main differences between these settings are in the relative proportions of the dominant facies types: sandy, silty or muddy turbidites, hemipelagites and pelagites, or black shales. Tectonically-controlled fining-upward sequences are common on faulted slope-aprons or in fault-controlled basins.

The above sequences are generalized ones associated with specific elements in the deep sea. Since they can all show variations of scale, sedimentary materials and regularity it is not always easy to make a definitive interpretation as to their origin. Many studies have related some of these vertical sequences directly to parts of an idealized fan model (e.g. Mutti and Ricci Lucchi, 1972; Walker and Mutti, 1973; Walker, 1978). However, it should be remembered that isolated channel, lobe or other sequences, for example, may derive from any of the three main environments: slope-apron, fan or basin plain. An environmental interpretation does, therefore, require a knowledge of the position of the sequence in a broader context.

12.5.5 Environmental facies association

Facies associations have been proposed for each of the three major environments (Mutti and Ricci Lucchi, 1972) (Fig. 12.34). Although highly schematized and simplified they summarize some of the major attributes, in each case showing the association of different facies types and the superposition of different vertical sequences (see also Table 12.3).

The *upper slope-apron facies association* is mudstone and marlstone dominated, of mainly hemipelagic origin but with some resedimented facies. Slump scars and evidence of erosion or higher energy are common. The *lower slope-apron facies association* is also mudstone and marlstone dominated, perhaps with rather more fine-grained turbidites, cut through by fining-upwards or irregular channel-fill sequences and interbedded with isolated debrites, slumps and slide masses. The lower part of the slope may also contain contourite drifts and turbidite lobes.

Three *fan facies associations* have been defined. The upper fan is characterized by a thick-bedded, coarse-grained, lenticular sandstone-conglomerate facies and a laminated or bioturbated mudstone-marlstone facies representing, respectively, channel and interchannel deposits. Thin-bedded, fine-grained turbidites represent the levee facies. The middle fan is characterized by thinning-upward sequences (distributary channels) overlying thickening-upward sequences (prograding lobes). Medium and coarse-grained discontinuous turbidite facies are dominant with secondary fine-grained turbidites and crevasse-splay sandstones in interchannel and distal lobe areas, and minor hemipelagites throughout. In the upper and middle fan palaeocurrent direc-

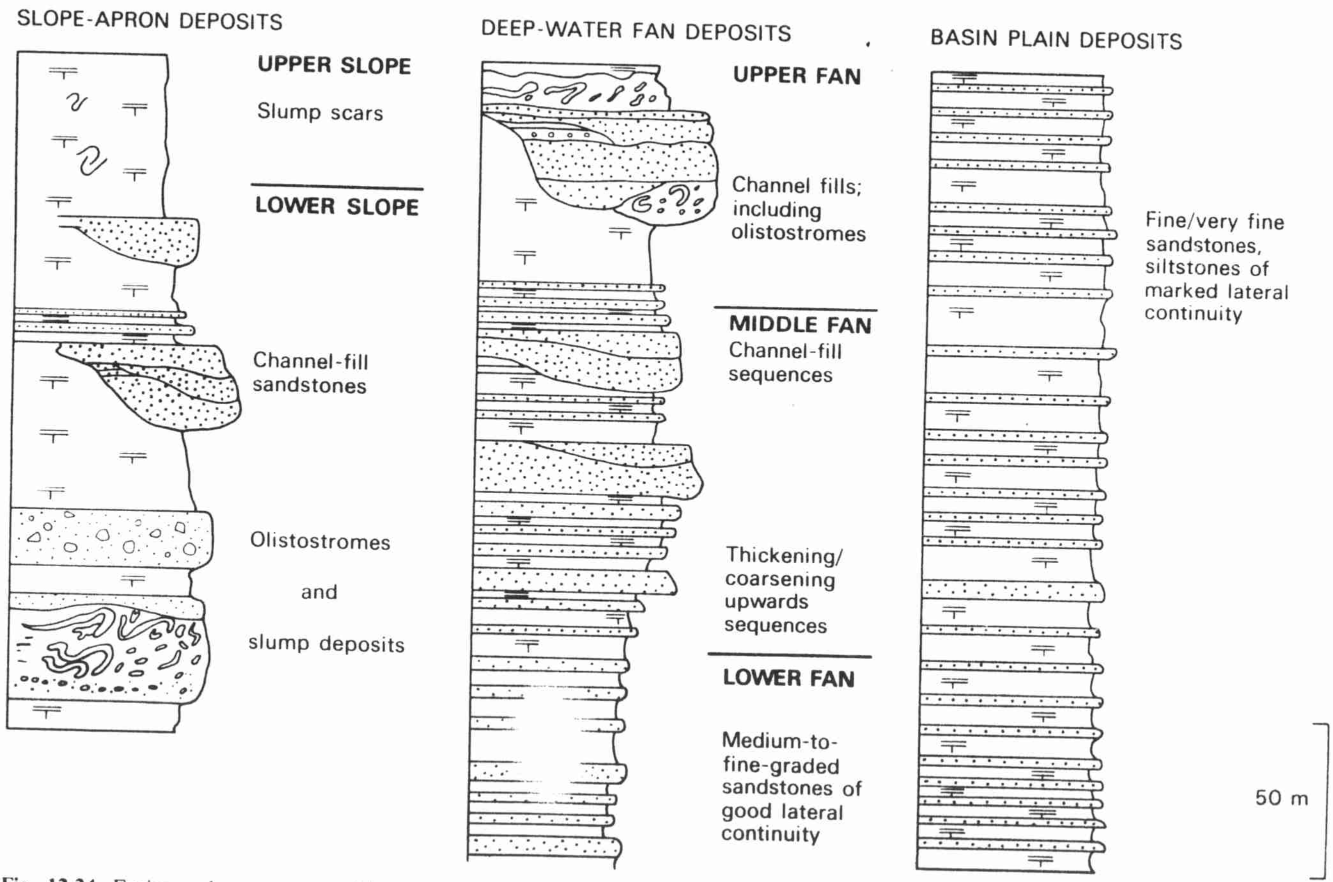


Fig. 12.34. Facies and sequences making up the facies associations characteristic of slope-apron fan and basin environments (after Mutti and Ricci Lucchi, 1972).

Table 12.3 Main characteristics of modern and ancient slope-aprons, submarine fans and basin plains

	Slope-aprons			Submarine fans			
	Normal (clastic)	Faulted	Carbonate	Radial	Elongate	Fan-deltas	Basin plains
Occurrence	Between shelf and basin floor, margins surrounding ocean basins and shelf, slope or marginal sea basins Widespread	Tectonically active regions	Low latitudes and ocean-ridge flanks	Extend from upper slope to basin floor, widespread discrete sediment accumulations Single canyon-fed fans, low to medium sediment supply	Commonly delta-fed fans, large sediment supply	Alluvial fan-fed fans, off high-relief areas	The flattest and deepest parts of a sedimentary system; the ultimate trap for terrigenous material
Shape	Narrow, rectilinear, elongate parallel to margin			Fan-shaped about feeder canyon	Elongate usually perpendicular to slope	Pear-shaped about alluvial-fan apex	Very variable shapes; but often elongate parallel to continental margin
Dimensions	Width variable 5–300 km	Width variable 5–50 km	Width variable 5–300 km	Radius variable typically 5–250 km	Radius (length) can range from 5–2500 km	Radius usually small 5–50 km	Very variable areas <100 km^2 to >1.5 mill km^2
Gradient	From 10° (upper) to 1° (lower)	Can be stepped with steeper parts >45°	Steep off reefs v. gentle off oceanic ridges	From 5–10° (upper) to ~1° (lower)	From 5–10° (upper) to <1° (lower)	From 5–10° (upper) to 1–2° (lower)	Very gentle to horizontal
Chief morphological elements	Shelf break, smooth open slope, gullies, channels, canyons, slump scars, slump & debris flow masses, base-of-slope sediment drifts and isolated lobes, transverse fracture zones across ocean-ridge flanks			Upper, middle and lower divisions based on gradient and morphological character; canyon or trough, channels (tributary & distributary), slump and debris flow scars and masses, channel levees, overbank and smooth open fan, depositional lobes			Extensive flat or undulating sediment-draped basin floor; isolated intrabasinal channels, ridges and drifts; merge with distal fans, slope-aprons, oceanic ridges and seamounts
Processes and dispersal pattern	Linear distribution of sediment input points; resedimentation processes (all types) mainly downslope; bottom currents (all types) up-and-down channels and along-slope; pelagic settling widespread			Usually point source or broad apex of sediment input; sedimentation processes (all types) dominant, downslope in channels with radial overbank dispersal; up-and-down canyon currents common; other bottom currents also active; wave and tidal effects on fan-deltas; pelagic settling widespread between resedimentation events			Several distinct input points usually present; one may be dominant, longitudinal, lateral and centripetal dispersal patterns; different basins dominated by resedimentation, pelagic or hemipelagic processes
Facies associations	Fine-grained turbidites and hemipelagites dominant; slides, slumps and debrites common; minor coarse-grained turbidite channel and lobe facies; interbedded contourites and isolated contourite drifts	Coarse, medium and fine-grained turbidites dominant; slides, slumps and debrites common; hemipelagites and pelagites common to minor; contourites minor to absent	Coarse, medium and fine-grained calciturbidites, pelagites and hemipelagites all important; slides, slumps and debrites common	Resedimented facies dominant on all fans; minor interbedded pelagites and hemipelagites; rare contourites Sand-rich	Mud-rich	Gravel-rich	Variable admixture of resedimented, pelagic and hemipelagic facies; rare contourites; ferromanganese nodules and red clay facies may be distinctive; resedimented megabeds and reflected turbidites also occur; fine-grained facies usually dominant
Horizontal facies distribution	Fairly irregular	May have a slope-parallel distribution of coarse to fine facies	Fairly irregular	Marked horizontal facies segregation; coarse-grained facies in channels and lobes; slumps and debrites on upper-middle fan and channel margins; fine-grained facies widespread		Channel-lobe system and facies segregation not well-developed; proximal to distal fining	Large basin plains show regular proximal to distal fining and thinning related to input points; small basin plains may be less regular
Vertical facies sequences	Coarsening-upward and fining-upward sequences may occur related to slope progradation, sea-level fluctuation or tectonic activity; similar sedimentary-related sequences can occur in channels and lobes			*Medium-scale* sequences (av. 20–80 m thick) throughout; channels: coarsening and blocky sequences lobes: coarsening-upward, fining-upward, symmetrical and blocky sequences can all occur; levees: coarsening-upward, fining-upward and irregular sequences; other subenvironments: more irregular sequences common; *Small scale* (av. 2–8 m thick) compensation cycles common in mid and lower fan regions			Vertical sequences mostly irregular, blocky or symmetrical

tions in the fine-grained and coarse-grained turbidites are often markedly divergent. The lower fan comprises medium and fine-grained laterally continuous turbidites and interbedded hemipelagites, often with more uniform palaeocurrent directions.

The *basin plain facies association* is the most monotonous, with an irregular vertical arrangement of fine-grained, thin-bedded turbidites. Hemipelagic and pelagic mudstones and marlstones may be more important. In reality, subtle lobe sequences, overall progradational sequences and isolated packets of thicker bedded turbidites may occur.

12.6 ANCIENT DEEP-SEA SYSTEMS: EXAMPLES AND CONTROLS

There are many well-documented examples of ancient deep-sea systems based either on detailed mapping of extensive onshore outcrop or on subsurface mapping during hydrocarbon exploration. Several of these examples are outlined in this section. They are related, as far as possible, to the primary controls that have influenced the facies types and distribution, and the morphology and geometry (Sect. 12.1.2), and are interpreted in terms of the slope-apron, submarine fan and basin systems recognized for present day sediments (Sect. 12.4).

12.6.1 Sediment supply and related controls

A large point source of clastic sediment, such as a major fluvio-deltaic system, commonly leads to the development of a submarine fan in deeper water, providing the shelf width is small and there are no other barriers to downslope resedimentation.

In the Carboniferous of northern England the Pennine Delta fed across a narrow shelf onto the adjacent slope. The prodelta slope-apron sequence (Grindslow Shales) is cut through by channels filled with coarse-grained turbidites which supplied sediment to small radial fans at the base-of-slope (Shale Grit) (Fig. 6.38) (Walker, 1966; McCabe, 1978). Part of a larger, probable elongate fan has been described from the Precambrian Kongsfjord Formation in northern Norway (Pickering, 1982a, b). This is overlain by an upper slope/prodelta sequence and then a fluvio-deltaic system that is believed to have provided the main sediment supply to the fan.

The Forties Palaeocene hydrocarbon-bearing submarine fan of the North Sea (Carman and Young, 1981, Rochow, 1981) was also probably fed by a large delta that prograded out into the Moray Firth Basin (Figs 14.13 and 14.17). It shows an overall shape and facies distribution similar in many respects to the elongate fan model (Sect. 12.4.3). Its development was partly controlled by early Tertiary tectonic activity leading to basin subsidence and synsedimentary faulting. Further north, in the Viking Graben, the Frigg, East Frigg and Odin gas/oil fields are developed in a Palaeocene-Eocene submarine fan complex (Fig. 12.35A) (Héritier, Lossel and Wathne, 1981). Seismic facies mapping shows a single sandy feeder channel leading from the Beryl Embayment in the southwest. This channel may have tapped a source of shallow water shelf sand or may have been supplied, in part, from a fluvial-deltaic system. Synsedimentary tectonics probably played some part in controlling sediment supply. A period of tectonic rejuvenation and channel incisement may have led to the emplacement of additional sandy lobes (Odin, N.E. Frigg and East Frigg) beyond the main radial fan (Frigg). Both Coriolis effects during fan growth and post-depositional compaction effects help explain the position and present geometry of the former channel levees (Fig. 12.35B).

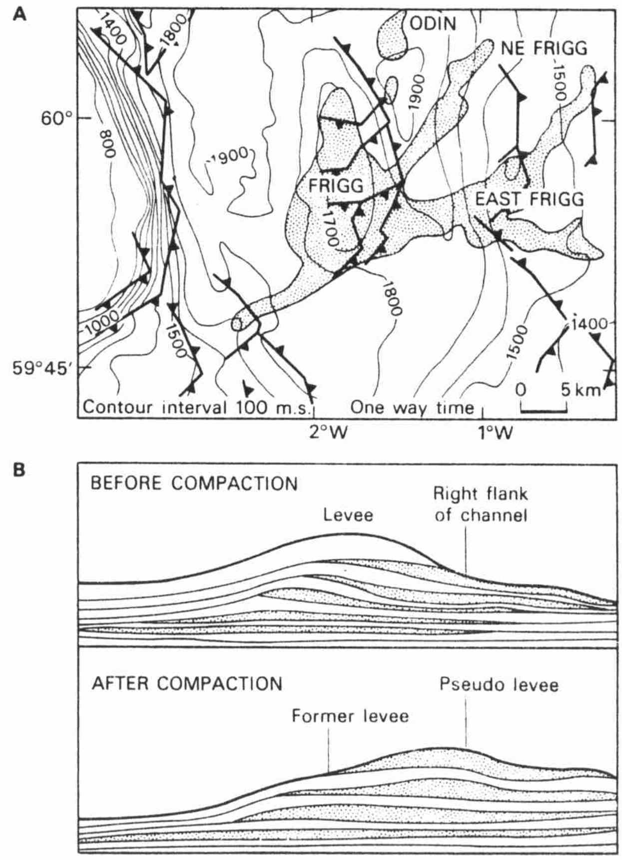


Fig. 12.35. Frigg Fan complex North Sea (after Héritier, Lossel and Wathne, 1981) (A) Outline of Frigg, East Frigg and Odin oil and gas fields, showing relationship of reservoir to deep (approx. base Cretaceous) structure. Interpreted as Palaeocene-Eocene radial submarine fan complex with feeder channel from the southwest. (B) Compaction effect on deep-sea fan channel and levee as seen on the Frigg fan complex, looking upcurrent to SW. Notice (1) channel sands migrate to left side of channel as right levee is built higher than left levee (2) After compaction former levee shales lie lower than channel sands.

Relatively shallow-water fan-deltas supplied by alluvial fans that feed directly onto a basin slope are known for several ancient sequences (review in Westcott and Ethridge, 1983). One of the best-documented examples given by these authors occurs in the Wagwater Trough of east central Jamaica. Here, some 7 km of conglomerates, sandstones and shales of the Wagwater Group have accumulated on humid-region fan deltas that have prograded into the basin from adjacent highlands (Fig. 12.36). There was coastal reworking of coarse-grained braided-fluvial deposits into sand and gravel beaches. The slope sediments on the steep submarine slope adjacent to the coast were frequently remobilized downslope as slumps, debris flows and turbidity currents. At the foot of a second slope into the basin proper the fan delta sediments are resedimented into true submarine fans (?radial type). This dual fan system has a close analogue in the modern Hope-Liguanea system off southeast Jamaica (Sect. 12.4.3, Fig. 14.53).

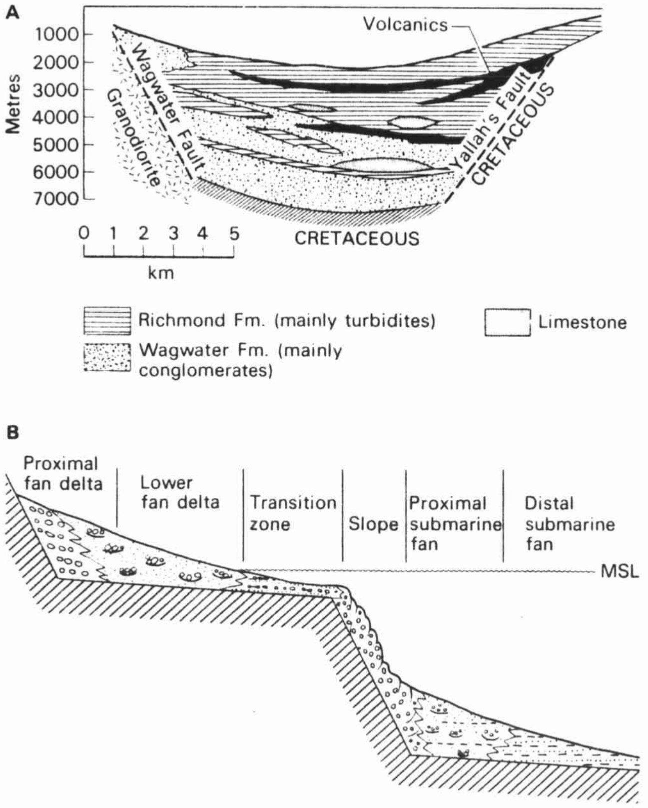


Fig. 12.36. Ancient fan-delta system—the Eocene Wagwater Group Jamaica (after Westcott and Ethridge, 1983). (A) Diagrammatic cross-section through Wagwater Trough prior to mid-Miocene uplift; (B) depositional model for the Wagwater and Richmond Formations (fan-delta systems) showing generalized facies relationships and interpreted environments.

A linear sediment supply, rather than point-source of material, results in the deposition of a margin-parallel slope-apron system. This often occurs off shallow carbonate shelves or reefs. For example, in the Cambro-Ordovician of central Nevada (Cook and Taylor, 1977; Cook, 1979) (Fig. 11.33) a depositional carbonate slope succession, 150 m thick, is composed of dark fine-grained limestones, mainly hemipelagic deposits, interbedded with coarser-grained calciturbidites and associated slump and debrite facies which make up at least 25% of the succession. The slope material was derived from a coeval thick biostromal and biohermal shelf sequence, now exposed in eastern Nevada.

By-pass carbonate slopes with steep escarpments fringed by spectacular rock-fall talus, debrites and coarse calcirudites are also known from ancient rock successions. The Cambro-Ordovician Cow Head Breccia in western Newfoundland (Hubert, Suchecki and Callahan, 1977; James, 1981) is a 310 m thick slope sequence that consists of limestone mega-breccias and debrites with giant carbonate clasts interbedded with calcarenites, lime mudstone, marlstones, shales and radiolarian-sponge spicule cherts. Some of the thin-bedded calcarenites have been interpreted as contourites deposited by southeastward flowing bottom currents, though the evidence for these is equivocal.

In many ancient deep-water successions only the feeder canyon or channel can be identified with any certainty (review by Whitaker, 1974). The Cambro-Ordovician Cap Enragé Formation in Quebec is interpreted as a submarine channel

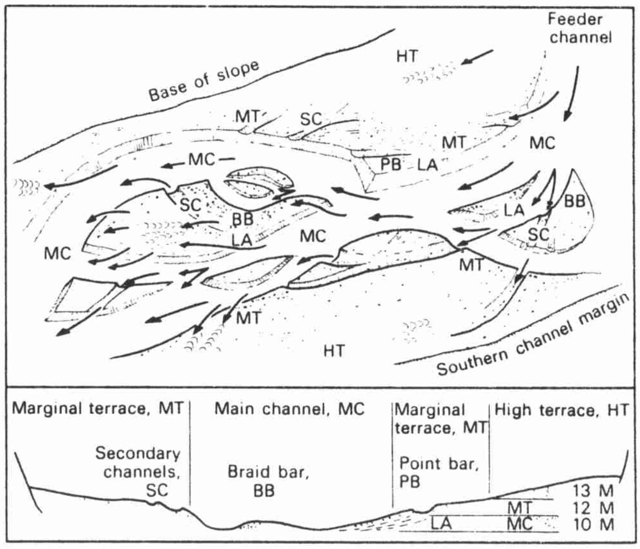


Fig. 12.37. Schematic reconstruction of the Cambro-Ordovician Cap Enrage channel, Quebec. Letters explained in the interpretative cross-section (below); LA, lateral accretion. Typical fining- and thinning-upward sequence due to lateral channel migration is shown in cross-section with average thicknesses of different facies. (After Hein and Walker, 1982.)

complex over 50 km long, at least 10 km wide and 300 m deep, filled with conglomerates, pebbly sandstones and massive sandstones. Internal sedimentary controls such as channel switching and abandonment have led to the deposition of these coarse-grained resedimented facies in submarine braided channels, braid bars, point bars, secondary channels and on marginal terraces (Fig. 12.37).

The Lago Sofia conglomerates and sandstones in the Upper Cretaceous of southern Chile (Winn and Dott, 1979) were deposited in north–south oriented channels up to 125 km long, 6–10 km wide and 350 m deep (Fig. 12.38), that formed axially in a narrow retro-arc basin. (Sect. 14.7.4; Figs. 14.38, 14.39c). Most of the channel facies have features developed by traction, probably occurring at the base of powerful turbidity currents. These facies are enclosed by finer-grained overbank turbidites and hemipelagites.

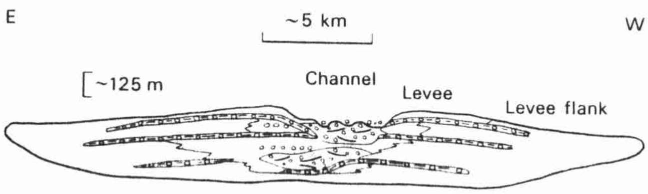


Fig. 12.38. Schematic east-west cross-section through a Lago Sofia lens in the Cerro Toro Formation showing inferred channel (conglomerates and sandstones), levee (sandstones) and levee flank (thin-bedded turbidites and debrites) facies. (After Winn and Dott, 1979.) Notice hypothetical rightward shift of channel.

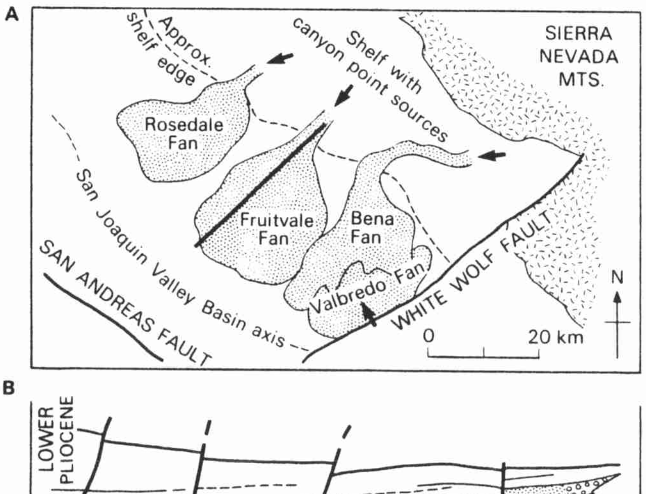


Fig. 12.39. Ancient fan-basin system—upper Miocene Stevens Sand (after Macpherson, 1978): (A) sketch map of fan distribution in Southern Great Valley, California; (B) schematic NE to SW radial cross section over area of Fruitvale Fan.

12.6.2 Tectonic controls

The regional tectonic setting exerts a primary control on the type of deep-water system developed. More specifically, there are many examples in which synsedimentary tectonic activity

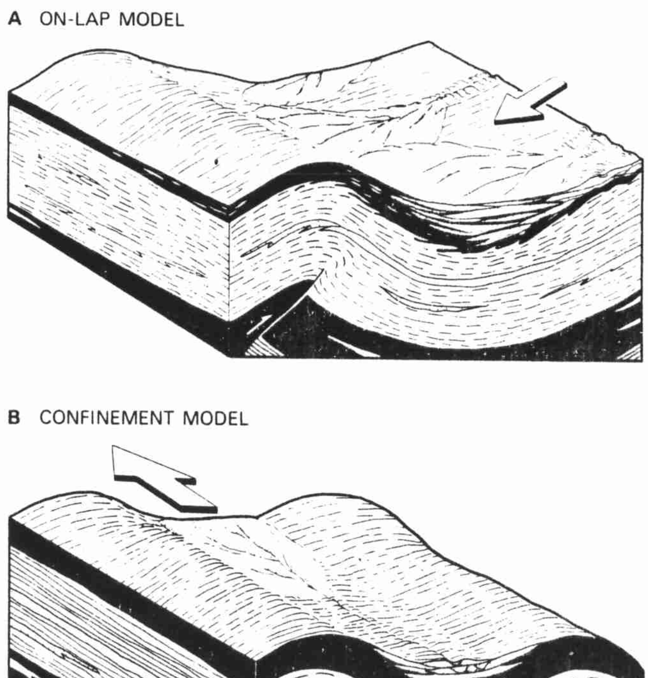


Fig. 12.40. Fan models developed for Miocene Stevens Sandstone Formation, San Joaquin Basin, California showing effects of synsedimentary tectonic control on deposition (from Scott and Tillman, 1981). (A) Onlap model—showing a series of sand bodies that lap onto and stack vertically against a contemporaneously rising anticline. (B) Confinement model—showing a series of sand bodies that accumulate in a synclinal low between adjacent anticlines. Vertical stacking and along-axis progradation can occur.

has played an important role in facies distribution and geometry, often in conjunction with other factors such as sediment supply (e.g. the North Sea Tertiary fans discussed above, Sect. 12.6.1).

The oil-producing Upper Miocene Stevens Sandstone in the San Joaquin Basin of California is a complex of submarine fans fed by channels from the eastern margin (Macpherson, 1978; Scott and Tillman, 1981). (Fig. 12.39). Facies include massive amalgamated coarse sandstone, thick sandstones showing traction (trough cross-bedding) structures, classical medium-grained turbidites, and finer-grained thin-bedded turbidites. These are arranged in coarsening-upward, fining-upward and irregular vertical sequences. Several radial fan-like bodies have been identified especially in the east, as well as lobate, irregular and channel-like accumulations of turbidites.

However, Scott and Tillman (1981) suggested that the conventional fan models do not adequately describe the turbidite facies association in the central and western portions of the basin where synsedimentary tectonics affected bottom topography. They therefore proposed *onlap* and *confinement* models (Fig. 12.40). The onlap turbidite body is a vertically thickened fan-like construction stacked against a contemporaneously rising anticline. Internal sequences are characteristic of fan progradation, but externally the sandstones pinch out crestally and may not be fan shaped. The confinement turbidite body has a channel-like morphology confined to bathymetric lows between adjacent growth anticlines. The facies and associations are not necessarily those commonly ascribed to channels.

Large-scale normal faulting along old lines of weakness fragmented the east Greenland shelf into several westerly tilted fault blocks in the late Jurassic (Fig. 12.41) (Surlyk, 1978). Syntectonic sediments of the Wollaston Forland Group were deposited down the steep fault scarps to form a thick slope-apron or wedge parallel to the fault trend. Rockfall breccias closest to the fault pass laterally into thick resedimented conglomerates and sandstones and then pass rapidly into siltstones and mudstones further to the east. Repeated fault activity led to deepening of the basins, thickening of the slope wedge, and arrangement of facies in several hundred metres thick fining-upward vertical sequences. Each of these sequences

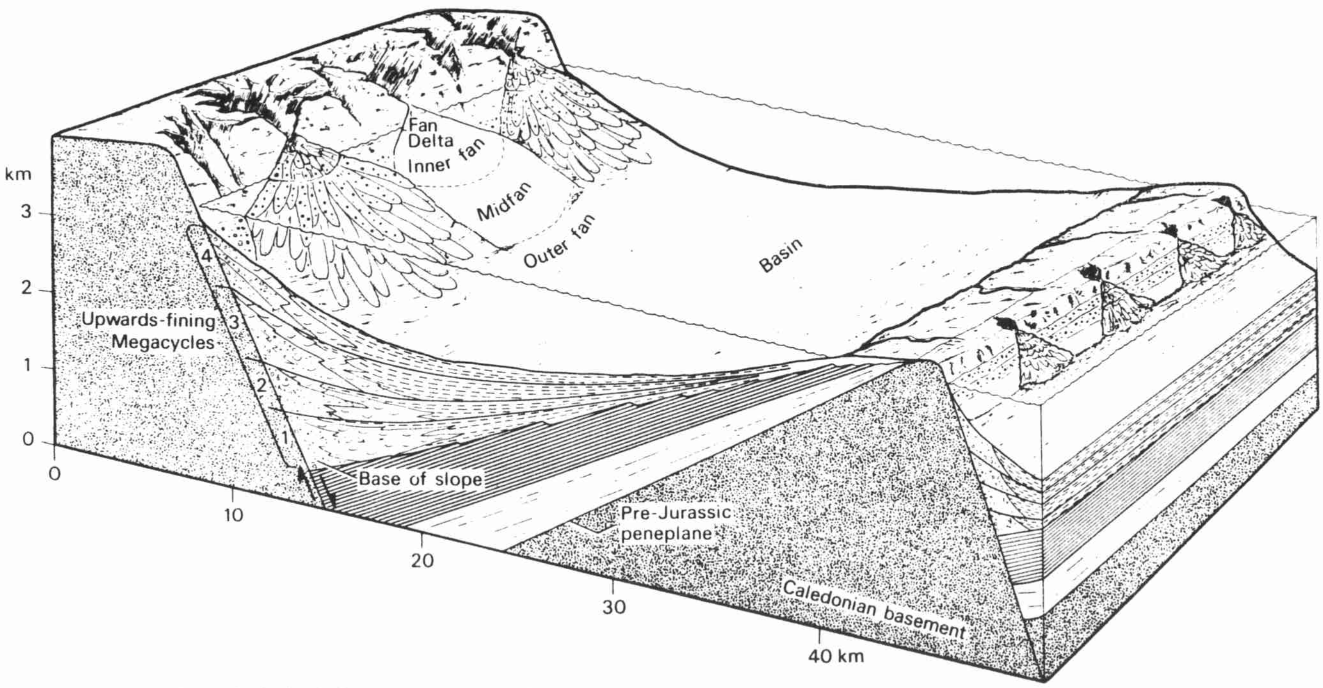


Fig. 12.41. Upper Jurassic faulted slope-apron system from eastern Greenland (after Surlyk, 1978). Block diagram constructed to demonstrate facies distribution and the interpreted palaeoenvironments of the Wollaston Forland Group. The clastic wedge is built of four fining-upwards megacycles. Notice subsidiary fans developed from temporarily emergent tilted fault block.

comprises smaller-scale (metres to tens of metres thick) fining-upward sequences, possibly related to filling and abandonment of temporary channels. The overall 4 km thick slope sequence is overlain by a fine-grained transgressive sequence of turbidites and hemipelagites. This faulted slope-apron system may have been fed by a series of fan-deltas along the coastline.

The Upper Jurassic Brae oilfield in the North Sea is very similar to the Wollaston Forland Group. It comprises resedimented conglomerates and sandstones interbedded with thin-bedded organic-rich mudstone and siltstone turbidites (Stow, Bishop and Mills, 1982; Stow, 1983). The system represents a 600 m thick slope-apron accumulation of sediments deposited in a narrow (< 10 km wide) elongate belt along a penecontemporaneous active fault zone. The main control on development of this system was tectonic but sediment supply and sea-level changes were important secondary controls.

In the Eocene-Oligocene of the Santa Ynez Mountains, California, several basin fill sequences pass from the mudstone through turbidite sandstone into shallow water facies (van de Kamp, Harper *et al.*, 1974). Variations in lateral and vertical arrangements of facies can be related to three environmental settings with differing degrees of tectonic control. Type 1 (Fig. 12.42A) lacks coarse-grained turbidites and has only thin-bedded turbidites which pass directly up into shallow marine sandstones. This relationship is interpreted as progradation of a small delta-front fan into relatively shallow water under stable tectonic conditions. In type 2 (Fig. 12.42B) there are coarse-grained turbidites in the middle of the sequence that probably indicate greater slope instability at the delta front. Type 3 (Fig. 12.42C) has a full suite of thin-bedded turbidites, coarse-sandstone turbidites and conglomerate wedges separated from the shallow water facies by a non-depositional sequence. This succession is thought to be due to slope instability produced by basin margin faulting.

The well-documented basin-plain successions from the Oligo-Miocene foreland basins of the periadriatic region of Italy (Figs. 12.43, 14.66) (Sect. 14.9.2) were considerably affected by tectonic activity during their accumulation (Ricci Lucchi, 1975; Ricci Lucchi and Valmori, 1980). The overall dimensions of these basins were large (400 × 50 km) and they were filled rapidly by relatively coarse-grained and thick-bedded turbidites as well as the associated finer-grained facies. Correlation of individual turbidites and mega-turbidites over wide areas is possible. Palaeocurrent patterns indicate both longitudinal filling and lateral supply. Lobe and channel sequences and facies have been identified and are commonly interpreted as parts of submarine fans that fed laterally into the basin plain system.

The Pliocene Cellino Formation, also from the periadriatic foredeep, has recently been described as a hydrocarbon-bearing submarine fan system (Casnedi, 1983). Although more proximal northern parts of the sequence, with rapid lateral facies changes between channellized sandstones and interchannel mudstones may show fan-like morphology, its elongate geometry (60 × 20

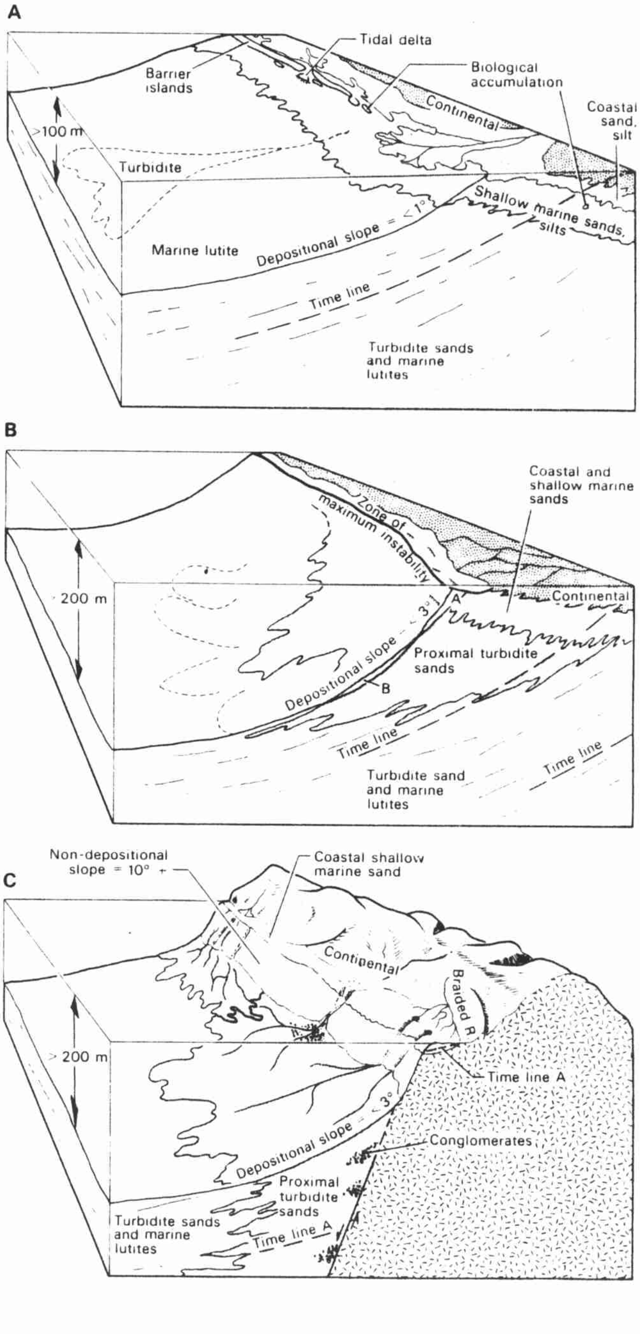


Fig. 12.42. Facies interrelationships in the Eocene-Oligocene of the Santa Ynez Mountains, California (from van de Kamp, Harper *et al.*, 1974), developed under (A) relatively stable basin conditions, (B) unstable slope and delta progradation, and (C) unstable slope and basin margin faulting.

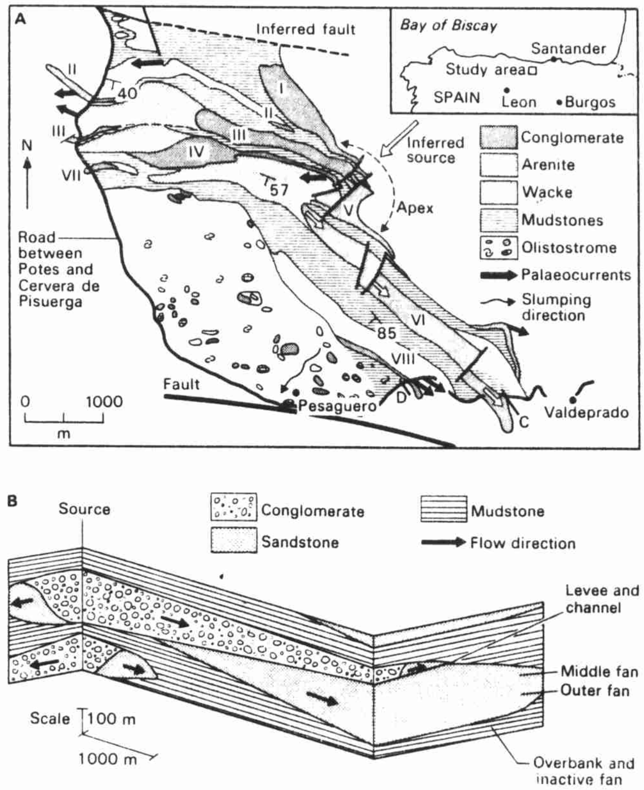


Fig. 12.44. Ancient radial fan system—the Upper Carboniferous Pesaguero fan from the Cantabrian Mountains, Spain (after Rupke, 1977). (A) General map showing facies distribution and palaeocurrent dispersal pattern; (B) Model of the three-dimensional shape and interrelation of the three main facies types of the Pesaguero Fan. A facies triplet (mudstone blanket, sandstone lobe, conglomerate tongue) represents a complete cycle of progradation of a major fan lobe. A new lobe forms by lateral avulsion.

km) in a narrow basin with axial palaeocurrent dispersal and the flat-bedded laterally-continuous nature of many of the sandstone turbidites are more indicative of a basin fill sequence, closely analagous to the older periadriatic basin fill turbidites.

12.6.3 **Sea-level fluctuations**

Although sea-level fluctuation during the Plio-Quaternary can be seen to have had major effects on the growth of, for example, modern oceanic fans, it is far more difficult to isolate sea-level

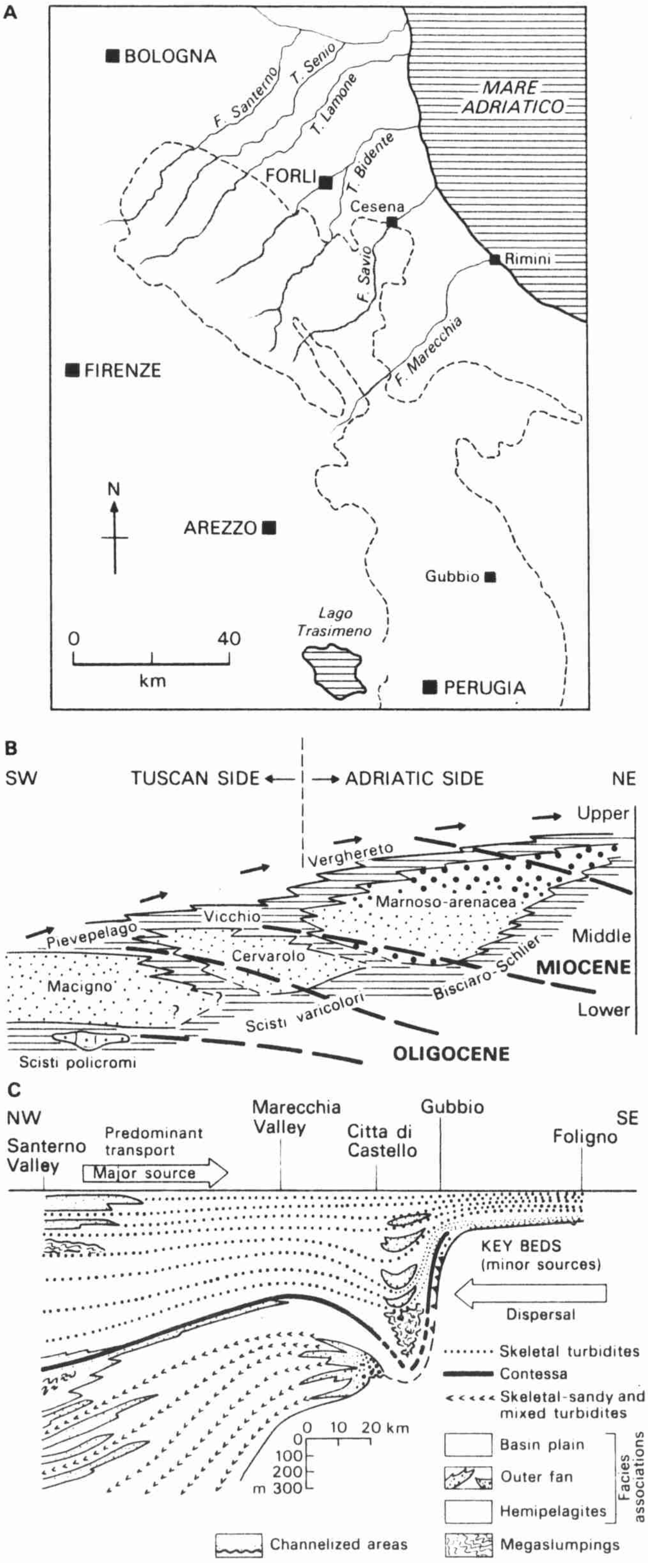


Fig. 12.43. Miocene Marnoso-Arenacea Formation of northern Italy (after Ricci Lucchi and Valmori, 1980). (A) Outline of Marnoso-Arenacea basin plain; (B) presumed stratigraphic relationships between different periadriatic turbidite basins; (C) schematic longitudinal section of Marnoso basin plain.

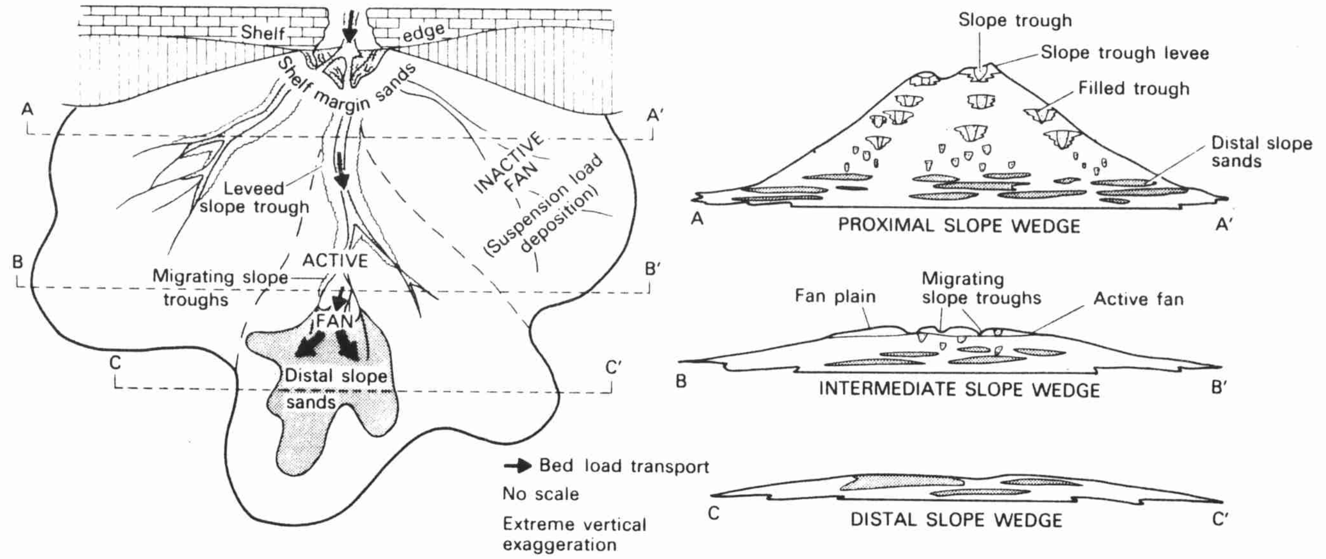


Fig. 12.45. Carboniferous-Permian Sweetwater Slope Group of the Midland Basin Texas (after Galloway and Brown, 1973).

changes as a controlling factor in the development of ancient deep-water systems. In the following two examples, both tectonic and sedimentary controls are known to have been important as well as sea-level fluctuation.

The Pesaguero fan developed in a small fault basin in the Upper Carboniferous of the Cantabrian Mountains, northern Spain (Fig. 12.44) (Rupke, 1977). No vertical passage into shelf or a delta can be documented. The fan is composed of several sequences each a few hundred metres thick and consisting of a mudstone blanket, a sandstone lobe and a conglomerate tongue. Each of these sequences represents progradation of an active part of the fan (coarse-grained) over an inactive part (fine-grained), with a systematic variation of facies types and depositional processes. New facies sequences form by avulsion and progradation. Together, they show a distinct radial palaeocurrent pattern, closely analogous to the radial fan model. A possible effect of sea-level rise is the upward change from lithic-wacke sandstones (low sea-level, immature river detritus) to quartz arenites (intermediate sea-level, mature shelf sands) and subsequently to a mudstone blanket over the entire fan (high sea-level). Progradation of the slope and tectonic activity may be indicated by an overlying olistostrome.

In latest Carboniferous and early Permian strata of Texas a shelf/slope-basin system has been identified in the subsurface over an area of more than 25,000 km^2 on the basis of wireline logs, core and isopach maps (Galloway and Brown, 1973) (Fig. 12.45). A fluvio-deltaic system prograded across the mixed carbonate terrigenous Eastern Shelf and locally extended through the shelf-edge bank onto the upper slope. Preserved relief between the shelf-edge and the Midland Basin floor is between 180 and 300 m. The Sweetwater Slope system is composed of shelf margin sandstones grading basinward into several slope wedges or overlapping fans. Resedimented sandstone facies occur in slope troughs (channels) and in more distal slope lobes. Debrites and slumps are common on the proximal slope, whereas mudstone and siltstone turbiditic and hemipelagic facies are dominant throughout. At times of lower sea-level the fan sediment was derived from the shelf.

FURTHER READING

Doyle, L.T. and Pilkey, O.H. (Eds) (1979) *Geology of Continental Slopes*, pp. 374. *Spec. Publ. Soc. econ. Paleont. Miner.* **27,** Tulsa.

Stanley, D.J. and Kelling, G. (Eds) (1978) *Sedimentation of Submarine Canyons, Fans and Trenches*, pp. 395. Dowden, Hutchinson and Ross Stroudsburg, Pa.

Stow, D.A.V. and Piper, D.J.W. (Eds) (1984) *Fine-grained sediments: deep-water processes and facies*, pp. 659 Spec. Publ. geol. Soc. Lond., 15.