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Deep-sea clastics: where are we and where are we going?

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SUMMARY: The transition from our belief in a deep calm ocean to a recognition that deep-sea clastics other than pelagic clays exist in the oceans, spanned nearly a century. In the last three decades enormous strides have been made in understanding these sediments and their deposition. There is a continuum of processes that transfer material from shallow to deep water and rework sediments within the deep sea. These include: (1) resedimentation processes, ranging from giant rockfalls and slumps to low-density turbidity currents; (2) normal bottom currents; and (3) pelagic settling through the water column. More than fifty distinct facies have been described from the deep sea and these can be interpreted in terms of depositional process via ten standard facies models for resedimented, normal bottom current and pelagic sediments. Environmental models can be constructed for: (1) normal, faulted, carbonate and ridge-flank slope-aprons; (2) radial, elongate and fan-delta submarine fans; and (3) under- and oversupplied basin-plains. These show the generalized horizontal and vertical distribution of facies and the chief morphological elements in each of the three major deep-sea settings. Sedimentary, tectonic and sea-level changes are the main groups of factors that control deep-sea sedimentation within these separate environments. Part of the interest in deep-sea clastics stems from their demonstrable economic importance for the generation and entrapment of hydrocarbons. Many areas of deep-sea sedimentology remain to be investigated and earlier models to be refined; these advances will depend significantly on improvements in our methodology.

The systematic study of deep-sea sediments (Fig. 1) began with the voyage of the HMS *Challenger* (1872–1876) which established the general morphology of the oceans and the types of sediments they contained. Following this pioneering expedition, the cornerstone of deep-sea sedimentology was, for a long time, the paper on 'Deep-sea deposits' by Murray & Renard (1891). However, 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 several decades while several different lines of research were conspiring to undermine its dominance. In particular, as more and more bottom samples were collected on early European and American oceanographic expeditions in the first half of the Twentieth Century (Böggild 1916; Andrée 1920; Stocks 1933; Shepard 1932, 1948; Bramlette & Bradley 1940; Arrhenius 1950) it was realized that sediments do not become uniformly finer-grained seaward across the continental shelves, and that the ocean floors have irregularities as great as any other part of the globe.

Although the existence of density undercurrents in lakes and reservoirs had been known for some time (Forrel 1885; Grover & Howard 1938), it was Daly (1936) who suggested that density currents, formed by waves stirring up

sediments on the continental shelf during periods of lowered sea level, may have excavated submarine canyons as they flowed downslope. 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 of graded sand beds in the Italian Apennines paved the way for their classic paper 'Turbidity currents as a cause of graded bedding' (Kuenen & Migliorini 1950).

This revolution in clastic sedimentology, as the turbidity current paradigm has been called (Kuhn 1970; Walker 1973), stimulated an intense period of systematic field, laboratory and oceanographic studies that has continued to the present day, and that has shown the deep sea to be anything but calm! Some of the key advances are listed below: there was early confirmation of graded sands in the deep sea (Ericson *et al.* 1951) and of the occurrence and nature of major turbidity currents (Heezen & Ewing 1952); palaeo-turbidity-current directions were measured to document the pattern of basin fill in ancient sequences (Pettijohn 1957); the understanding and interpretation of geosynclinal sediments was much improved (Kay 1951; Drake *et al.* 1959) and, following the plate tectonic revolution in geology as a whole, in the early 1960s, patterns of deep-sea sedimentation were better related to global tectonics (Mitchell & Reading 1969); the classical sequence of structures in turbidites was developed by Bouma (1962); the physics of

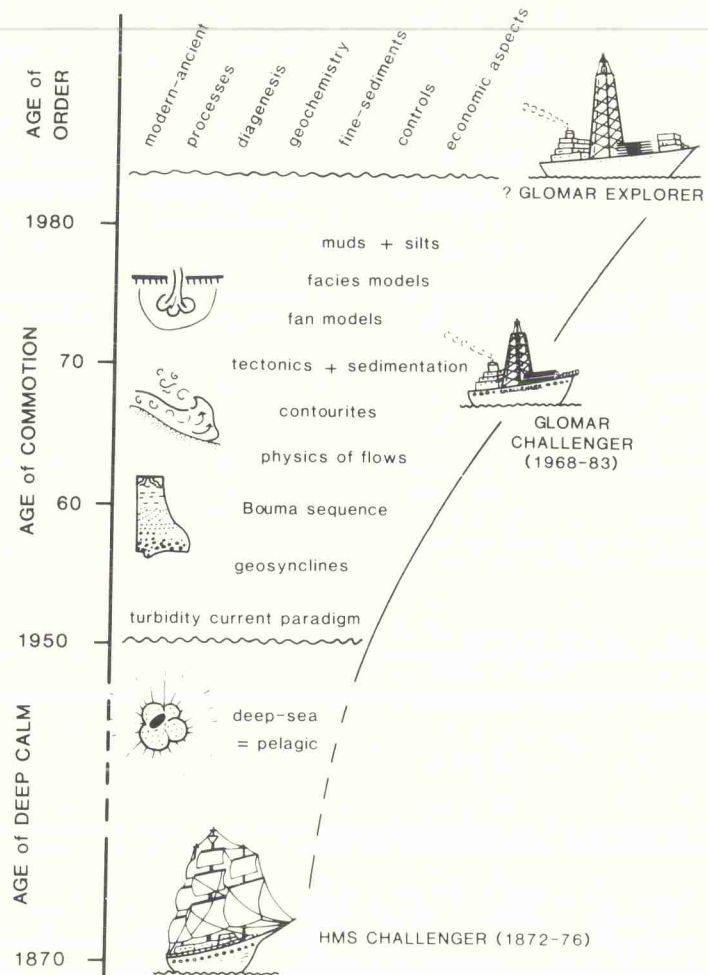


FIG. 1 Historical development of the main concepts in the study of deep-sea clastic sediments.

turbidity current flow has been much better appreciated following the experimental and theoretical work of Harms & Fahnstock (1965), Middleton (1966, 1967) and Komar (1969, 1970); deep-ocean currents were put forward as an important alternative to turbidity currents in the mid-1960s (Heezen *et al.* 1966; Hollister & Heezen 1972), and the characteristics of contourites more firmly established by Stow & Lovell (1979); submarine fans were characterized in detail from the present-day oceans (Normark 1970) and from ancient sequences (Mutti & Ricci Lucchi 1972); resedimented conglomerates were recognized as different from sandy turbidites but equally important (Walker 1975); similarly, the characteristics of silt and mud turbidites have only recently been detailed (Piper 1978; Stow & Shanmugam 1980; Stow *et al.* 1984).

The Deep-Sea Drilling Project, initiated in 1968, has seen nearly one hundred different voyages by the drillship *Glomar Challenger* throughout the world's oceans. These, together with many other marine and land expeditions, have clearly contributed enormously, not only to the study of deep-sea clastics but also to the related studies of pelagic and authigenic sediments, ocean history, tectonics and ocean-margin development.

Processes

For clastic sedimentary particles to accumulate in the deep sea they must undergo erosion, from land or from the seafloor, transportation and deposition. Biogenic material may be similarly eroded, but much is synthesized directly in the oceans and simply transported to and deposited

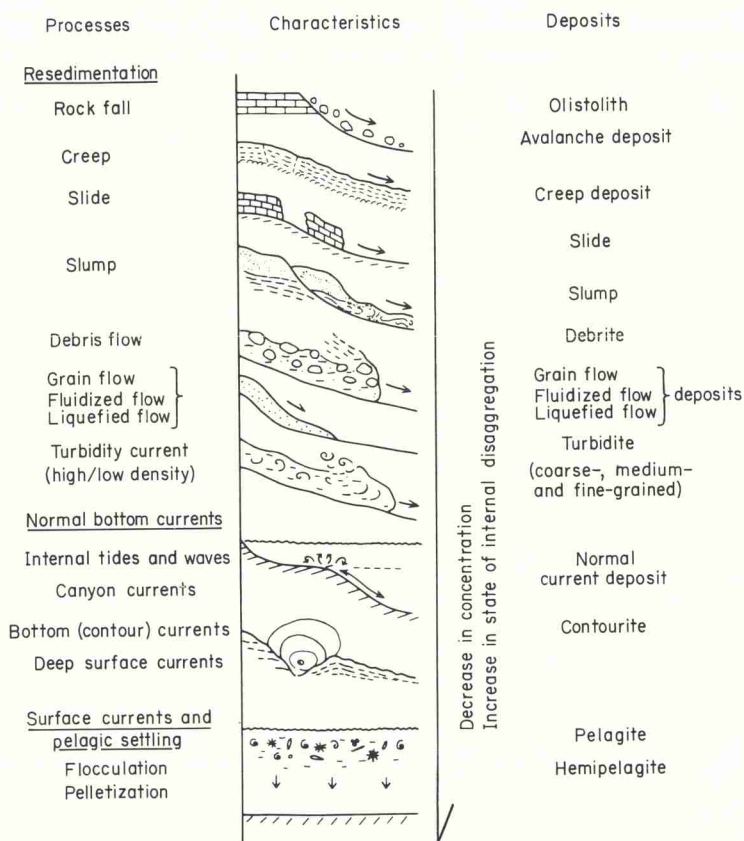


FIG. 2. Schematic representation of the main transport and depositional processes in the deep sea, illustrating the concept of a process continuum.

on the seafloor. Authigenic minerals grow *in situ* at or near the sediment-water interface, although they too may be subsequently reworked. There are three main groups of processes that operate on both terrigenous and biogenic material in the deep sea (Fig. 2): resedimentation processes, movement by normal bottom currents and transport by surface currents/pelagic settling.

(1) Resedimentation

Resedimentation processes (synonymous with mass gravity transport) are all those processes that move sediment downslope over the seafloor from shallower to deeper water and that are driven by gravitational forces (Fig. 2) (Middleton & Hampton 1976; Saxov & Nieuwenhuis 1982; Hein 1982; Hill *et al.* 1984; Gorsline 1984).

Rock falls are sudden, rapid freefall events that occur only on steep slopes of faulted or

carbonate margins or in the heads of submarine canyons. They are initiated by undercutting or erosion, earthquakes and other shock events. Single displaced clasts or olistoliths (up to tens of metres in size) and avalanche deposits bounce, roll and slide downslope for several tens or hundreds of metres before coming to rest.

Sediment creep is a process of slow strain due to constant load-induced stress. It is a barely perceptible but probably widespread phenomenon on many slopes, both gentle and steep, depending on the physical properties of the sediment. It appears to propagate along an internal décollement zone, and may be a precursor to slide or slump failure.

Slides and slumps involve complete sediment failure, downslope displacement and remoulding. They are widespread on slopes of all gradients over about 0.5° and range in size from less than 1 m³ to over 100 km³. They comprise mobile shear zones along which the sediment

mass moves. Coherent slide blocks may be less easily recognized than contorted slump units in ancient sediments. They are triggered by various shock events or may develop progressively from sediment creep.

Debris flows are slurry-like flows of sand to boulder-size clasts supported by their own buoyancy in a muddy matrix. They can be large or small (several to tens of metres thick), occur commonly on slopes greater than $1-2^\circ$, and advance slowly downslope either continuously or intermittently for distances up to tens of kilometres.

Grain flows are characterized by grain to grain collisions and a resulting dispersive-pressure support mechanism. They require slopes in excess of about 18° and so are probably very localized and relatively small events on steep slopes such as in the heads of submarine canyons. They are not a process of long-distance transport in the deep sea.

Liquefied or fluidized flows are related processes in which the grains settle rapidly through upward-moving pore fluids. Liquefied flows may move rapidly down slopes in excess of $2-3^\circ$ for a short distance only, before 'freezing' and deposition occurs. These flows rarely occur alone as a separate process in the deep sea, but are common during the final stages of deposition from high concentration turbidity currents.

Turbidity currents are very widespread throughout the deep sea over variable slopes and are of various types. High-density ($50-250 \text{ g l}^{-1}$) and low-density ($0.025-2.5 \text{ g l}^{-1}$) currents have been identified, as well as storm-generated flows, turbid-layer flows and mid-water density flows. Flow thickness may be from less than 10 to over 500 m, and flow velocities from $10-50 \text{ cm s}^{-1}$ for low-density currents to over 25 m s^{-1} (70 km h^{-1}) for high-density currents. In all cases the flow is sustained and sediment supported by a process of turbulent auto-suspension. Turbidity currents can develop either from slumps and debris flows by mixing with seawater, or more directly from sediment suspensions caused by storm stirring, rivers in flood and melting glaciers.

These various flows are simply end-members of a process continuum in which all intergradations exist. A single resedimentation event may be initiated by seismic tremors or sediment overloading on a slope causing slumping which, by mixing with seawater, evolves through a debris flow into a turbidity current. A rock fall may trigger a sand avalanche or grain flow that deposits rapidly through a phase of liquefied or fluidized flow. Storm stirring at the shelf break

or across the shelf may generate a low-density turbidity current, and it appears that major river floods or glacial melting may initiate or augment this same process.

That a process continuum exists is emphasized by the progressive decrease in flow concentration and increase in the state of internal disaggregation from rock falls (solid) through to low-density turbidity currents. The maximum distance of transport is a few tens of metres for rock falls and probably similar for grain flows and liquefied flows, a few kilometres for slides and slumps, 50-100 km for debris flows and several 1000 km for turbidity currents.

(2) Normal bottom currents

The second group of processes (Fig. 2) are all those deep currents that actively erode, transport and deposit sediment on the seafloor but that are not driven by gravity, and may therefore flow alongslope and upslope as well as downslope (Hollister & Heezen 1972; Heezen 1977; Shepard 1973; Shepard *et al.* 1979; Stow & Lovell 1979; Stow 1982).

Internal tides and waves are widespread in the upper few hundreds of metres, at the thermocline and at other density discontinuities. Their effects on the bottom sediments are particularly apparent at the shelf break, in submarine canyons or in deep narrow passages.

Canyon currents have now been recorded from very many submarine canyons throughout the oceans even at depths in excess of 4000 m. They commonly flow alternately up and down the canyon, with a tidal periodicity in the deeper parts but with a higher frequency of flow reversal 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. Typical velocities range up to 30 cm s^{-1} .

Bottom (contour) currents are formed by cooling and sinking of surface water at high latitudes and the deep, slow, thermohaline circulation of these polar water masses throughout the world's oceans. Highly saline but warm water also flows out of the Mediterranean as an intermediate level contour current. Current intensification is caused by flow restriction through narrow passages and on the western margins of basins by the Coriolis force. These stronger currents commonly attain velocities of $10-20 \text{ cm s}^{-1}$ (rarely over 100 cm s^{-1}), but are highly variable in direction, velocity and periodicity, and advance with a slow eddy-like progression. They may be associated with well-developed nepheloid layers.

Surface currents driven by the winds may also impinge on the seafloor at very great depths (several kilometres), such as the deep Gulf-Stream gyres of the North Atlantic or the deep Kuroshio current off Japan.

It is worth emphasizing again that, for these flow types as for the resedimentation processes, there are no rigorous boundaries between flows. Thus, internal tides and waves can cause or contribute to canyon currents, and these latter grade progressively into swifter flowing, more concentrated low-density turbidity currents. Deep, surface currents and thermohaline bottom currents are practically indistinguishable at the seafloor and, where associated with intensified nepheloid layers, grade into the lowest-density large-scale turbidity currents.

Mostly, however, the concentration ($0.025\text{--}2.5\text{ mg l}^{-1}$) and velocity (average $0.05\text{--}0.2\text{ m s}^{-1}$) of the normal bottom currents are significantly lower than those of the various resedimentation flows. Their range of thickness (up to 2000 m) and distance of transport (up to thousands of kilometres) is variable. Such currents are clearly capable of profoundly affecting the bottom sediments and morphology, and so it is not surprising that we are discovering more and more facies and features in the deep sea that display 'fluvial-like' characteristics.

(3) Surface currents and pelagic settling

The third important depositional process in the deep sea, slow pelagic settling through the water column (Hsu & Jenkyns 1974; Jenkyns 1978; Gorsline 1984), can be considered one extreme end-member of the process continuum (Fig. 2). 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. Terrigenous elements (clays, very fine-grained quartz, volcanic dust, etc.) are transported to the open ocean in variable amounts by surface currents, winds and floating ice.

Vertical settling of the finest particles is extremely slow ($10^{-4}\text{--}10^{-6}\text{ m s}^{-1}$), although much of the material settles more quickly ($10^{-2}\text{--}10^{-3}\text{ m s}^{-1}$) as flocs and faecal pellets (McCave 1984). As it settles, the material is subject to dissolution of calcareous and siliceous tests, oxidation of organic matter and lateral transport by bottom and turbidity currents.

Facies

The most important features used to define

different deep-sea facies are: grain size and other textural attributes, sand/mud ratio, bed thickness and geometry, internal organization of beds, biogenic and 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 or environment.

However, with at least ten distinct depositional processes, a large range of environments, and sediments ranging from huge boulders to the finest clays, there are clearly a very large number of possible facies in the deep sea. More than fifty facies have been identified for clastic sediments alone (Mutti & Ricci Lucchi 1972, 1975; Walker & Mutti 1973; Carter 1975) (Fig. 3), although this degree of subdivision is clearly unnecessary for most purposes other than very detailed sedimentological interpretation.

A rather more simple grouping into seven major facies classes, based largely on grain size differences, is also shown in Fig. 3. We can identify, in both modern and ancient sediments, the following classes:

- A gravels and pebbly sands (conglomerates and pebbly sandstones),
- B sands (sandstones),
- C sand-mud units (sandstone-mudstone units),
- D silts and silt-mud units (siltstones and siltstone-mudstone units),
- E muds (mudstones),
- F chaotic mixed-grade deposits (rockfalls, slumps, debrites),
- G oozes and arls (limestones, marlstones, cherts, etc.). ('Arl' is a term introduced by Dean *et al.* (1984) for muddy oozes and biogenic muds.)

These are modified and updated from those originally proposed by Mutti & Ricci Lucchi (1972, 1975).

It is sometimes convenient to subdivide the first five of these classes (A–E) further into disorganized and organized facies groups. The disorganized groups contain such facies as thick-bedded, massive, 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 contain 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; contorted and disturbed

Class	Group	Facies	Facies				
			1	2	3	4	5
A Gravels + pebbly sands	A1 Disorganized grvl + p.sst						
	A2 Organized p.sst						
B Sands	B1 Disorganized						
	B2 Organized						
C Sand - mud units	C1 Disorganized						
	C2 Organized						
D Silts + silt-mud units	D1 Disorganized						
	D2 Organized						
E Muds	E1 Disorganized						
	E2 Organized						
F Chaotic mixed-grade units	F1 Isolated displaced clasts						
	F2 Contorted + disturbed beds						
	F3 Muddy gravel + pebbly mud						
G Oozes + 'Arls'	G1 Ooze						
	G2 Arl						

FIG. 3. The main classes and groups of sediment facies recognized in the deep sea (modified after Mutti & Ricci Lucchi 1972; Rupke 1978). Over fifty distinct facies have been identified and these are illustrated schematically. See text for further discussion.

slumps and slide masses; and pebbly muds or muddy gravels. Finally, facies class G comprises the purely biogenic (pelagic) sediments, the calcareous and siliceous oozes, and the mixed muddy oozes and biogenic muds that are so common in the deep sea. Dean *et al.* (1984) have called this latter group of sediments 'arls', adding 'sarl' (siliceous biogenic mud) and 'smarl' (calcareous and siliceous biogenic mud) to the commonly used and convenient term, marl.

Facies models

Most of the large number of separate facies shown in Fig. 3 can now be interpreted in terms of depositional process by reference to one of the facies models for resedimented, normal

current deposited and pelagic sediments outlined in Figs 4 and 5. These facies models show schematically the idealized sequences and sedimentary characteristics of sediments deposited by single events or particular processes. Actual examples of some of these facies from both the recent and ancient record are shown in Figs 6 and 7. Facies models are not shown for the isolated displaced clasts, sediment creep deposits, gravel and coarse sand lags, dune-bedded sands and black shale facies, as more work is required on each of these groups before an adequate synthesis can be attempted.

Slumps (facies group F2) can involve any lithology and be very thick (tens to hundreds of metres) or very thin (a few centimetres). Laminae and beds are rolled, contorted or rotated, but sedimentary structures will often

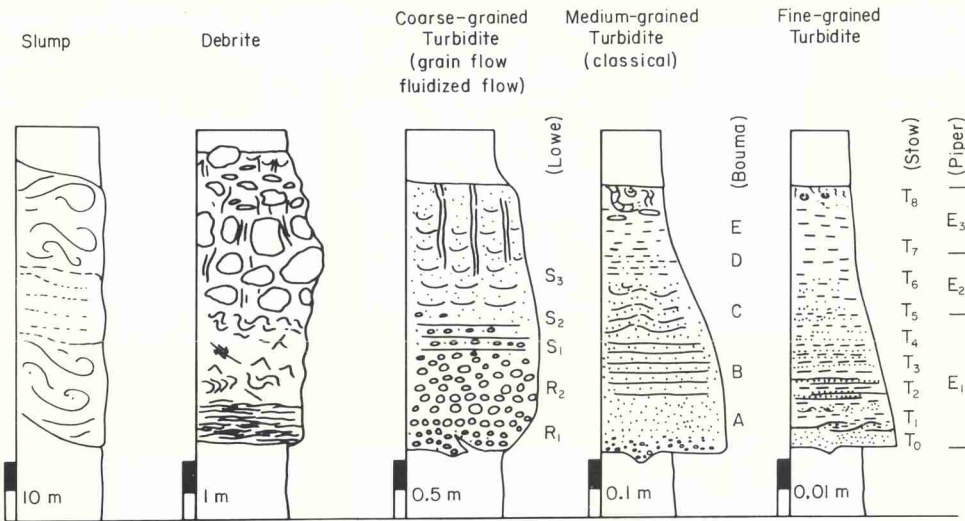


FIG. 4. Models of resedimented facies, for slumps, debrites and turbidites, showing the idealized structural sequences for debrites and for coarse-, medium- and fine-grained turbidites. The scale bars give only an indication of typical unit thicknesses, which may vary widely in practice. Grain size increases to the right for each column.

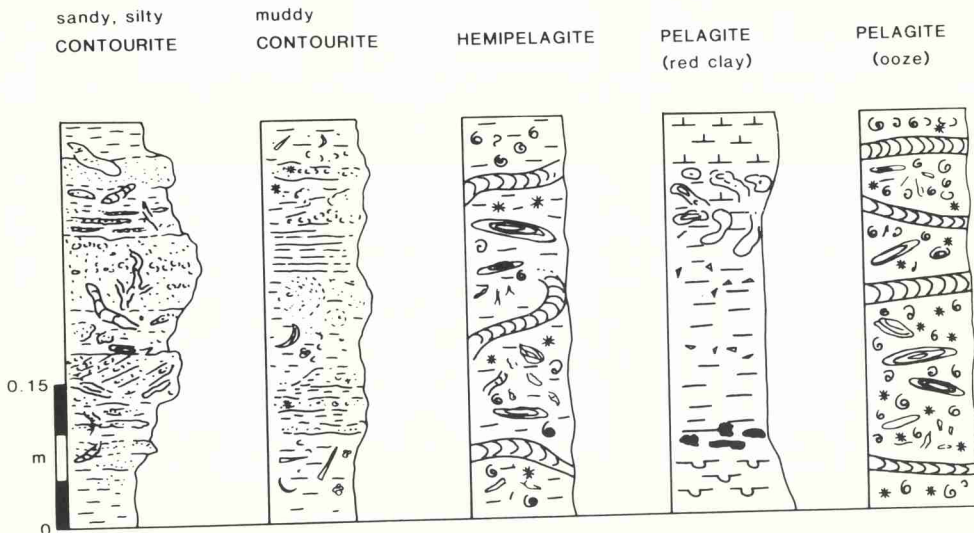


FIG. 5. Normal sedimentation facies models for contourites, hemipelagites and pelagites. Grain size increases to the right for each column.

show the way-up of the beds and the facing direction will determine the sense of the slope. No standard vertical sequence has been identified (Moore 1961; Morgenstern 1967; Lewis 1971; Saxov & Nieuwenhuis 1982).

Debrites (facies group F3) also involve mixed lithologies, with hard pebbles and boulders or soft mudstone clasts set in a muddy matrix, and

vary in thickness up to several tens of m. They may be disorganized or minimally organized with a basal zone of lensoid (?shear) lamination, a middle zone of high-angle faults and slump folds capped by convolute lamination, and an upper clast-rich zone that can show slight positive grading, water-escape pipe and dish structures and some horizontal alignment of

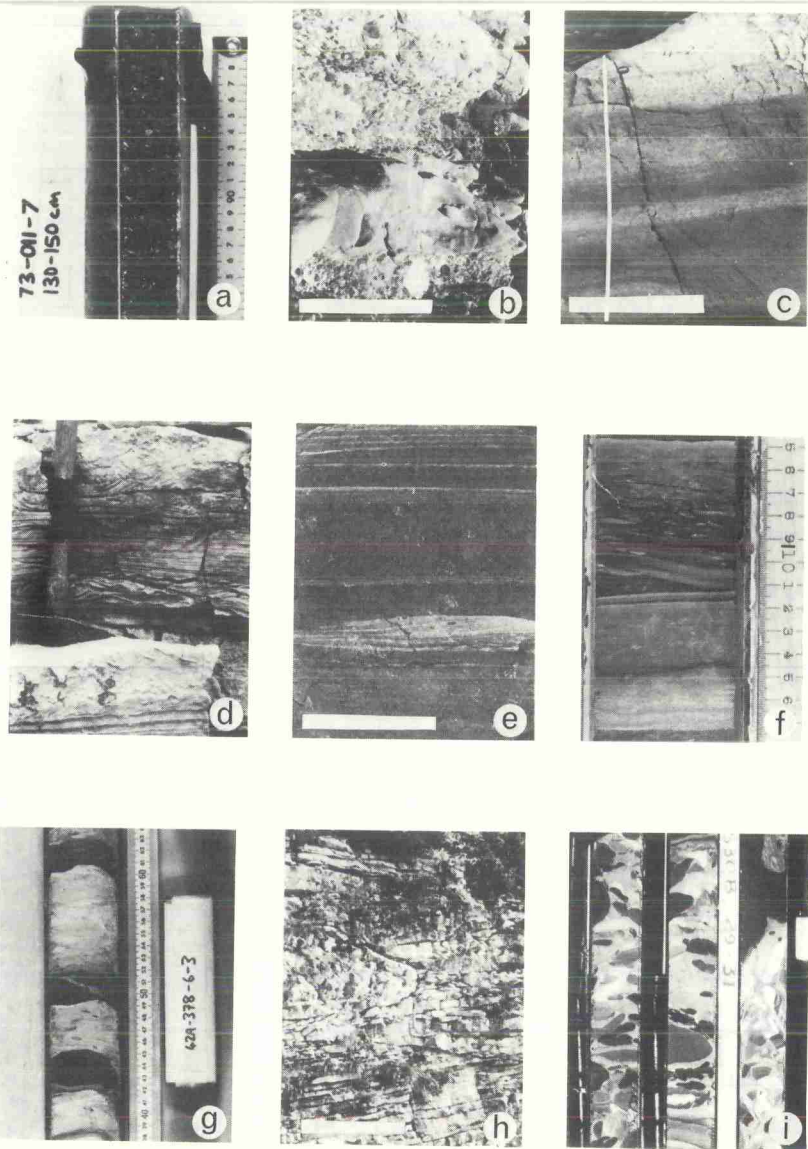


FIG. 6. Selected photographs illustrating resedimented facies characteristics: (a) part of thick coarse-grained, graded, gravel turbidite (?Holocene) from Laurentian Fan channel, western North Atlantic (Facies A3.3); (b) interbedded disorganised conglomerates and pebbly sandstones from the Upper Cretaceous of Pigeon Point, California (Facies A1/A2.1 and 2) (scale bar = 1 m); (c) parallel-laminated, thick-bedded, coarse-grained sandstone turbidite, with deep mud-scour at top surface, from the Ordovician Halifax Formation, Nova Scotia (Facies B2.1) (scale bar = 1 m); (d) typical medium-bedded, Bouma BCDE turbidite from the Tertiary of the central Apennines, Italy (Facies C2.2); (e) thin-bedded, fine-grained silt-mud turbidites from the Ordovician, Halifax Formation, Nova Scotia (Facies D2.2 and 3) (scale bar = 3 cm); (f) thin-bedded, fine-grained silt-mud turbidites from DSDP Site 530A, late Cretaceous, SE Angola Basin (Facies D2.2 and 3); (g) thin-bedded, graded mud turbidites (dark) within a pelagic ooze sequence from DSDP Site 378, Pliocene, Aegean Sea (Facies E2.3); (h) slumped beds from the Cretaceous-Tertiary Scaglia Rossa Formation, Umbro-Marchean Apennines, Italy (Facies F2.2) (Scale bar = 1 m); (i) detail of thick-bedded marl-ooze debrites from DSDP Site 530B, Plio-Quaternary, SE Angola Basin (Facies F3.1) (core width = 7 cm). Individual graded units (turbidites) shown by arrows where possible.

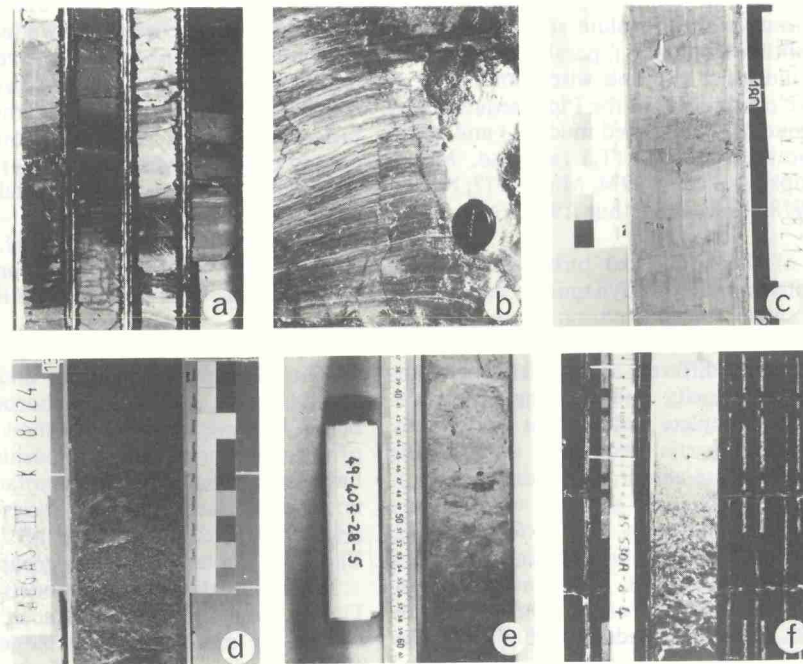


FIG. 7. Selected photographs illustrating normal sediment facies characteristics: (a) interbedded black organic-rich fissile mudstones (Facies E2.4), and green calcareous hemipelagic mudstones (marls) (Facies G2.1) from DSDP Site 530A mid-Cretaceous, SE Angola Basin (core width = 7 cm); (b) finely laminated diatomaceous pelagic ooze from the Miocene Monterey Formation California (Facies G1.2); (c) homogeneous muddy contourite (Facies E1.2) from Faro Drift, Gulf of Cadiz. (d) fine sandy contourite (lower) (Facies C1.2) overlain by mottled silt and mud contourites (Facies D1.3) and muddy contourite (upper) (Facies E1.2) from Faro Drift, Gulf of Cadiz (core width = 10 cm); (e) homogeneous muddy contourite (Facies E1.2) and irregularly laminated silty contourites (Facies D1.3) from DSDP Site 407, NE Atlantic; (f) interbedded, bioturbated, pelagic calcareous ooze (light) (Facies G1.1) and siliceous ooze (dark) Facies G1.2) DSDP Site 530A, Pliocene, SE Angola Basin.

clasts. (Abbate *et al.* 1970; Middleton & Hampton 1976; Naylor 1980; Thornton 1984).

The *coarse-grained* turbidite model is a composite including many of the facies in facies classes A and B which are each represented by one of the divisions of the idealized sequence ($R_1, 2, S_{1, 2, 3}$, Lowe 1982). Deposition can be by grain flow, fluidized- or liquified-flow mechanisms often following long-distance transport by turbidity currents. The lower part of the sequence can comprise either gravel or sand or any gradation of pebbly sand between the two. Thus there is a sharp, scoured base, a negatively graded lower division, an intermediate massive, stratified, graded-stratified division and an upper division with dish and pipe structures. The top is commonly sharp and flat (Walker 1975, 1978; Carter 1975; Middleton & Hampton 1976; Lowe 1979, 1982; Hein 1982).

The *medium-grained turbidite* model is the

classical Bouma (1962) sequence and represents most of the facies class C and parts of B and D (there being some overlap between the three turbidite models). The five divisions are well-known: massive to graded sand (A), parallel-laminated sand (B), cross-laminated to convolute sand (C), parallel-laminated fine sand and silt (D) and massive to bioturbated mud (E) (Bouma 1962; Kuenen 1964; Walker & Mutti 1973; Hesse 1975).

The *fine-grained turbidite* model, representing much of facies classes D and E, was developed to facilitate description and interpretation of the resedimented muds and silts not adequately covered by Bouma's E division. Thus, Piper (1978) recognized a graded silt-laminated mud division (E_1), a graded mud (E_2) and an ungraded mud (E_3). Stow & Shanmugam (1980) recognized an idealized vertical sequence of silt laminae structures passing up from the

graded laminated unit including fading ripples (T_0), to mud with convolute silt laminae (T_1), low-amplitude ripples (T_2), parallel distinct (T_3), parallel indistinct (T_4) and wispy laminae (T_5). These are overlain, as in the Piper sequence, by graded mud (T_6), ungraded mud (T_7) and a thin micro-bioturbated zone (T_8) (see also, Moore 1969; Rupke & Stanley 1974; Mutti 1977; Nelson *et al.* 1978; Kelts & Arthur 1981; Stow *et al.* 1984).

Each of these idealized turbidite sequences can be interpreted hydrodynamically as resulting from a single re-sedimentation event that deposited progressively finer grades of sediment and gave rise to different sedimentary structures as the flow velocity and carrying power decreased. 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. 3; for example, deposition of top-absent classical turbidites (Pouma divisions A, AB, ABC) 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_2E_3 , Stow divisions T_{678}) give massive and graded mud turbidites (facies E1.1, E2.1, E2.2 and E2.3); and so on.

Different facies can thus be related to different parts of the idealized sequences and hence to a particular type of flow and to a stage in the evolution of that flow. This information then leads to an interpretation of the bathymetric, environmental or other factors controlling the location and occurrence of different flow types.

Two *contourite* models are shown in Fig. 5, one for sandy contourites and one for muddy contourites (Stow & Lovell 1979; Stow 1982). These represent the sediments deposited from bottom currents in the large elongate contourite drifts of the ocean basins, and also interbedded with other facies on continental rises. *Muddy contourites* appear homogeneous and thoroughly bioturbated, although rare parallel or wavy lamination, pockets of coarser material, irregular sharp contacts between silts and mud, and certain textural characteristics are evidence that deposition has been current-influenced. *Sandy contourites* (also coarse silt grade) occur in thin to medium irregular layers or beds with sharp or diffuse contacts, rare internal lamination, and very common bioturbation and burrows. Compositionally, both types generally contain planktonic and benthonic biogenic material mixed with terrigenous material, often broken or iron-stained, although other composi-

tional types are also known. A complete gradation exists between the two facies, and they commonly occur together in an irregular vertical 'sequence' that shows negative grading from a muddy through mottled silt and mud to a sandy facies and then a positive grading back to a muddy facies (Faugères *et al.* 1984). Parts of this 'sequence' occur as the individual facies E1.2, D1.3 and C1.2 (Fig. 3).

Reworking and winnowing by bottom currents results in *coarse sand* and *gravel-lag contourite* facies, shown as facies B1.2 and A1.3 in Fig. 3.

Hemipelagites are compositionally very similar to muddy contourites and 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 & Enos 1977; Hill 1981). These are the 'arls' of facies group G2.

Two contrasting *pelagite* models are shown in Fig. 5, one for oozes (facies group G1) comprising more than 70% biogenic material, and one for red clays (facies E1.3) that commonly have less than 10% biogenics. Neither of these are considered in detail here (but see Jenkyns 1978, Hoffert 1980, Thiede *et al.* 1981 and Leggett this volume).

Environments

Distinctive vertical sequences and horizontal distributions of facies and the occurrence of particular depositional processes are characteristic of specific environments in the deep sea. Amidst the great variability of the marine realm we can identify three fundamentally different environments, each with their own mixture of sedimentary, tectonic and morphological features. These are slope aprons, submarine fans and basin plains. Slope aprons probably account for the greatest volume of sediments, basin plains are areally most extensive, whereas submarine fans, having early attracted the geologists' attention, have been the most thoroughly studied to date. Nevertheless, it is now possible to begin to construct generalized environmental models for each of these different regions (e.g. Pickering 1983).

(1) 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 where they are taken here to include the con-

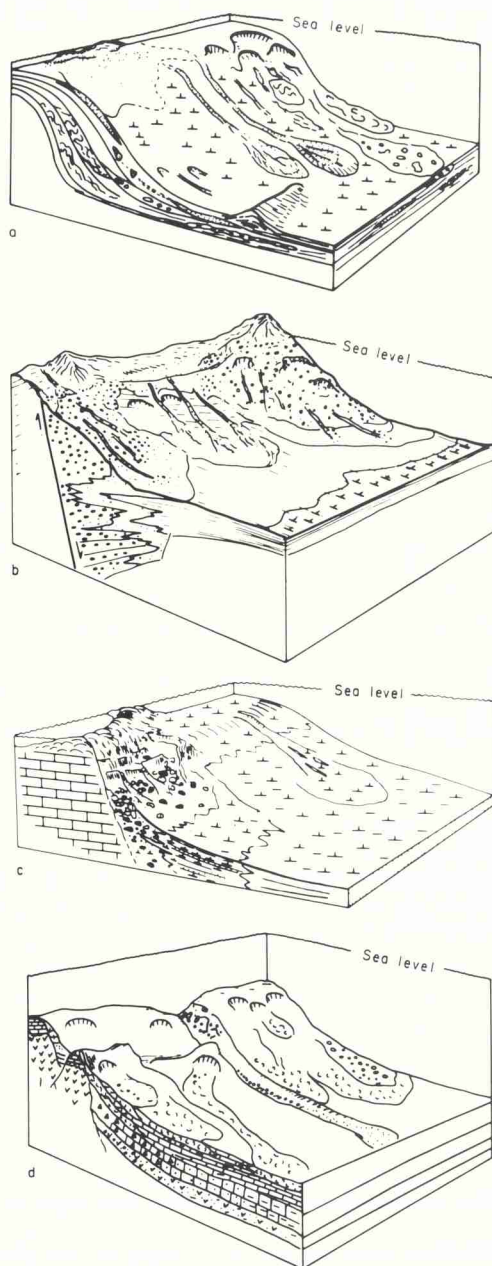


FIG. 8. Sedimentary environment models for submarine slope aprons showing schematic distribution of facies and morphological elements. No fixed scale applies: slope widths vary between 1 and 500 km; slope gradients 1–7°. (a) normal (clastic), (b) faulted, (c) carbonate, and (d) ridge-flank.

tinental rise. Broad areas of slope also occur on the flanks of mid-ocean ridges and surrounding isolated seamounts or plateaus.

They 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, slump and slide scars, irregular slump and debris flow masses, small straight or slightly sinuous channels and gulleys, more complex dendritic canyons, isolated lobes, mounds and drifts, and broad areas of smooth or current-moulded surface.

Although at least ten different types of slope apron can be distinguished on the basis of the morpho-tectonic setting (Emery 1977; Bouma *et al.* 1978; McIlreath & James 1978; Doyle & Pilkey 1979), there are only four major types that are significantly different in terms of their sedimentological characteristics. These are: normal (clastic), faulted, carbonate and ridge-flank slope aprons (Fig. 8).

Normal (clastic) slope aprons (Fig. 8a) range from mainly constructional, with a relatively smooth convex-concave profile built upwards and outwards by slope progradation, to mainly destructional, with erosion (slumping, sliding, etc.) on the face of the slope causing a steeper and more irregular profile to be developed. In the former, lower-energy case the slope surface tends to be smooth or current-moulded, whereas in the latter, higher-energy case the slope is often gullied and slump-scarred, with sediment lobes, debris-flow masses and slump blocks at the foot of the slope. Large canyons and channels may cut across the slope at intervals, and elongate contourite drifts be constructed near the base-of-slope.

The distribution of facies is highly irregular and dominated by finer-grained sediments (silts, muds, oozes and arls). Coarser-grained sediments (sands and gravels) occur above the mudline, as sand-spillover sheets, in the axes of channels and in base-of-slope lobes. Slides, slumps and debrites are common throughout.

Many of the continental slopes and rises throughout the world fit into this category, as well as the slope-aprons around marginal seas and shelf basins. A particularly well-studied modern example is the Nova Scotian slope-rise system off eastern Canada (Stanley *et al.* 1972; Piper 1975; King & Young 1977; Stow 1978; Hill 1981). Ancient equivalents include the Lower Palaeozoic Meguma Group onshore Nova Scotia (Schenk *et al.* 1980; Stow *et al.* 1984), and

the Carboniferous–Permian Sweetwater slope in the subsurface of Texas where oil and gas are found in small channel and lobe sandstone reservoirs.

Faulted slope aprons (Fig. 8b) are those that form across an active synsedimentary fault margin. They commonly have relatively steep portions alternating with flatter perched basins forming a stepped profile. There is an abrupt change of profile at the base-of-slope to a flat basin floor and little development of a lower slope or rise. A thick wedge of sediment accumulates in a narrow trough at the foot of the slope as a result of continued down-faulting.

Steeper portions of the slope may be bare of sediment or have a relatively thin veneer, with frequent slump-scars, slump masses and short-lived shallow channels. A roughly slope-parallel arrangement of coarse to fine grained facies may be developed, often built up from numerous small fans or lobes that overlap along the length of the slope. However, much lateral facies variability occurs as a result of the non-uniform, periodic nature of fault activity.

Faulted slope-aprons develop mainly along active compressional and strike-slip margins, as well as young rifting margins such as parts of the Red Sea rift system. Modern examples include the Gulf of Guinea slope (Delteil *et al.* 1974) and the western margin of the Tyrrhenian Sea (Wezel *et al.* 1981). Several ancient examples are known from the later Jurassic early rifting phase around the North Atlantic margins; these include the Wollaston Foreland Group of eastern Greenland (Surlyk 1978), and the Brae oilfield and related subsurface systems of the northern North Sea (Stow *et al.* 1982).

Carbonate slope aprons (Fig. 8c) have either a steep off-reef profile, often with stepped portions of submarine cliffs, or a gentle convex-concave off-platform profile. These are the bypass and depositional margins identified by McIlreath & James (1978). Sediments are thin or absent on the steeper parts which are fringed by a periplatform, calcirudite talus wedge that grades rapidly downslope into calcarenites, calcilutites and pelagic-hemipelagic limestones. Channels and canyons locally funnel coarser sediments into deeper water. The gentler slopes have a more irregular facies distribution of finer-grained calciturbidites, bottom current and pelagic deposits. Slumps, debris flow masses, channels and lobes further complicate the arrangement of facies.

Such slope aprons are found surrounding coral reefs and carbonate platforms throughout the lower latitude regions of the world. The best studied modern examples in terms of sediments

are those around the Bahamas (Mullins & Neumann 1979), and the Belize barrier and atoll reefs (James & Ginsburg 1979). Many ancient carbonate slope-apron systems have been described, two well-known examples being the Cambro-Ordovician Hales Limestone Formation of central Nevada (Cook & Taylor 1977), and the Cow Head Breccia of the same age in western Newfoundland with spectacular giant limestone clasts in slope-apron megabreccias and debrites (Hubert *et al.* 1977).

Ridge-flank slope aprons are the very low-gradient, highly dissected slopes of oceanic ridges, linear island chains, oceanic plateaus and isolated seamounts. They have a distinctive irregular to concave profile with ocean-crust basement highs, perched basins and transverse fracture zone valleys. Sediment cover is usually thin and irregular, comprising both biogenic (calcareous or siliceous) and volcanigenic material in slump, debrite, turbidite and pelagite facies.

Modern examples are summarized by Kelts & Arthur (1981) from many DSDP sites. More detailed studies are given for the Walvis Ridge slope-apron in the South Atlantic (Stow 1984), and the Mid-Atlantic Ridge flank in the vicinity of Gibbs Fracture Zone (Faugères *et al.* 1982). Ancient examples are best known from the sediments associated with ophiolites, although these are not often well preserved (Nisbet & Price 1974).

(2) Submarine fans

Submarine fans are distinctive constructional features at the foot of slopes, both in small shelf or marginal basins and in the large ocean basins. Unlike slope aprons, which are continuous 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 also very variable in size, from a radius of little more than 1 km to a length of more than 2000 km, and have gradients similar to those of slopes, decreasing from the upper to lower fan region. They comprise one or more feeder channels or canyons, tributary and distributary channels, abandoned half-filled channels, slump and slide scars and blocks, debris flow masses, broad channel levees, lobes built up at the ends of channels and distributaries, and relatively smooth or current-moulded interchannel and inter-lobe areas.

A number of different fan models have been developed over the past 15 years (Normark 1970, 1978; Nelson & Nilsen 1974; Mutti & Ricci

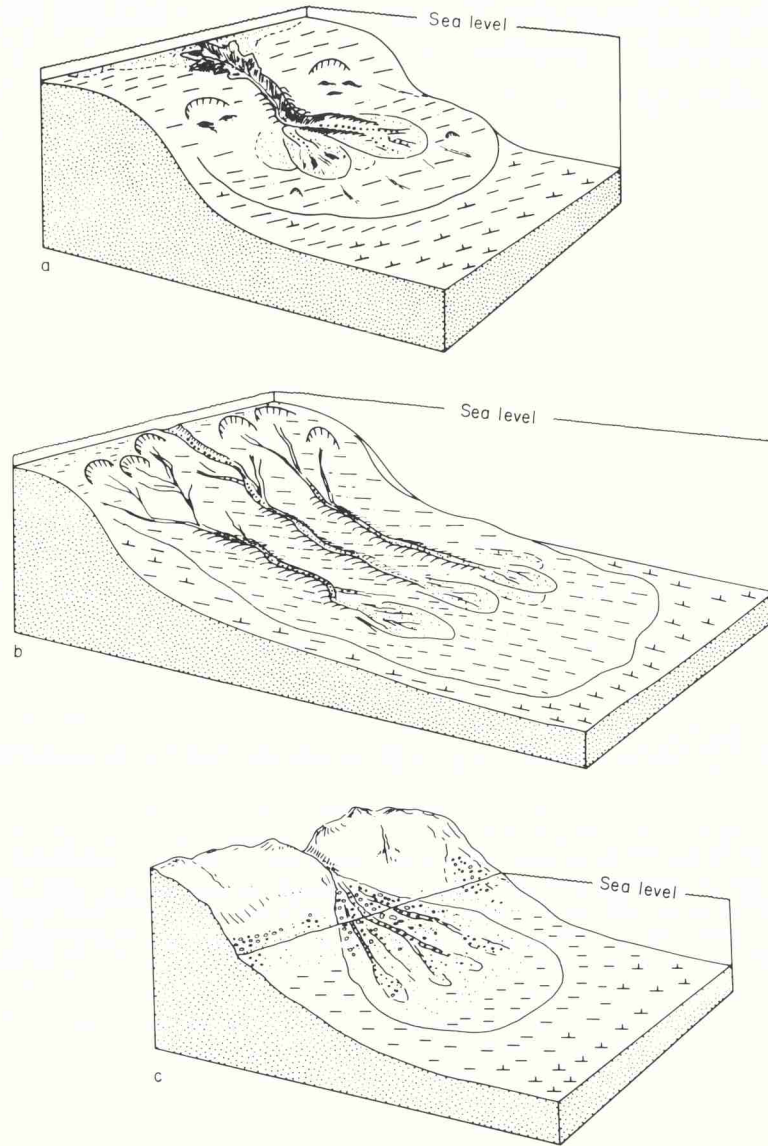


FIG. 9. Sedimentary environment models for submarine fans showing schematic distribution of facies and morphological elements. No fixed scale applies: fan radius normally <math>< 150\text{ km}</math> for A, <math>< 1500\text{ km}</math> for B and <math>< 15\text{ km}</math> for C; steepest parts of upper fans rarely exceed

Lucchi 1972, 1975; Walker 1978; Stow 1981; Howell & Normark 1982) from studies of both modern and ancient systems. However, there appear to be two principal end-member types developed in deeper water (Fig. 9a, b), with all possible gradations between the two (Stow *et al.* 1984), and a third shallow-water type (Fig. 9c). All these types may be substantially modified by morphological confinement and syndimentary

tectonics (Scott & Tillman 1981).

Radial fans (also called 'sandy', 'low-efficiency', 'canyon-fed', 'small', 'restricted-basin') have a true fan-like shape arranged concentrically about a small apex, a single feeder channel or canyon, a concave-convex-concave longitudinal profile, and a hummocky mid-fan region characterized by distributary channels and lobes. There is both an elongate and

concentric distribution of coarse-to fine-grained facies. Slumps, slides and debris occur in the lower slope, upper fan and channel margin areas. Coarse-grained turbidites are mainly confined to channels and lobes, whereas fine-grained turbidites occur throughout together with hemipelagic and pelagic facies.

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, 1978). The Carboniferous Pesaguero Fan of northern Spain (Rupke 1977) and some of the Upper Miocene Stevens Sandstone fans that provide hydrocarbon reservoirs in subsurface California, (Scott & Tillman 1981) may be ancient equivalents.

Elongate fans (also called 'muddy', 'high-efficiency', 'delta-fed', 'large', 'open-basin') are longitudinally extended perpendicular to the margin, commonly with a broad head region, two or more main feeder channels, a complex tributary-distributary system, an irregular-concave-smooth profile, and large terminal lobes constructed at the ends of channels on to the lower fan region. The sediment distribution is more elongate than concentric, and there is usually a high proportion of mud to sand. The various resedimented and pelagic facies are otherwise similar to those of radial fans.

Modern examples include the giant (3000 km long) Bengal fan (Curry & Moore 1971) and the Indus fan (Jipa & Kidd 1974) amongst others. In reality, there is a complete gradation between the different fan types. Stow *et al.* (1984) plot a number of modern fans on a triangular diagram with the apices as radial fans, elongate fans and normal slope-apron systems. Ancient examples probably include the Eocene Butano Sandstone system of California (Nelson & Nilsen 1974), and the Precambrian Kongsfjord Formation of northern Norway (Pickering 1982).

Fan deltas (also called 'short-headed delta-front fans') develop as the subaqueous part of alluvial fans that prograde from highlands directly into a standing body of water (lake or sea). They are mostly relatively small, pear-shaped in outline and with an ephemeral system of shallow braided channels radiating down-slope from the fan head. Sediments are mainly coarse-grained in the upper regions and in the channels, with muddier sediments on the levees, interchannel and more distal areas.

Modern fan deltas have been reviewed by Westcott & Ethridge (1980), well-studied examples being the Yallahs fan-delta off south-east Jamaica, and a number of fan-deltas along the south-east coast of Alaska (Boothroyd &

Nummedal, 1978). Ancient fan-deltas are also well-known from rocks of all ages around the world. Westcott & Ethridge (1983) review these briefly, and document in more detail an example from the Wagwater Trough of east central Jamaica.

(3) Basin plains

Basin plains are the flattest and deepest of our three deep-sea environments, and may vary widely in their areal extent up to the size of the major ocean basins, and in their depth to the deep floors of submarine trenches. They generally have a very gentle relief and merge gradually or more abruptly with the surrounding slopes and isolated seamounts or other basement highs.

Their main morphological elements include the extreme distal portions of submarine fans, channels and lobes, isolated intra-basinal channels, ridges and drifts, structurally controlled grabens, morphologically restricted passages, and very large areas of smooth or current-modified seafloor.

Several different basin classifications have been proposed using the criteria of composition, depth, restrictedness, fill geometry or morphotectonic setting (Dott & Shaver 1974; Hesse & Butt 1976; Gorsline 1978; Ballance & Reading 1980; Pilkey *et al.* 1980). From the point of view of their sedimentary characteristics, the most important controls are the basin size, the sediment supply and source area, and tectonic activity. On this basis, we recognize two end member basin plain types, those that are undersupplied and those that are oversupplied, and a complete gradation between the two.

Undersupplied basin plains (Fig. 10a) are commonly large, open basins far from land in a low-relief relatively stable tectonic setting. They have a low sediment supply to area ratio, and fill very slowly with simple sediment drape or current drift forms. The facies are dominantly fine-grained, terrigenous and biogenic, and overall thicknesses rarely exceed about 1 km.

Most of the major abyssal plains of the ocean basins are of this kind, including the Sohm and Hatteras abyssal plains in the North Atlantic (Horn *et al.* 1971, 1972; Pilkey *et al.* 1980). Although these Atlantic plains receive an enormous volume of terrigenous material derived from very large drainage basins, their very great size maintains their undersupplied character.

Oversupplied basin plains (Fig. 10b), by contrast, are mostly rather small, often partly confined and located in tectonically active areas.

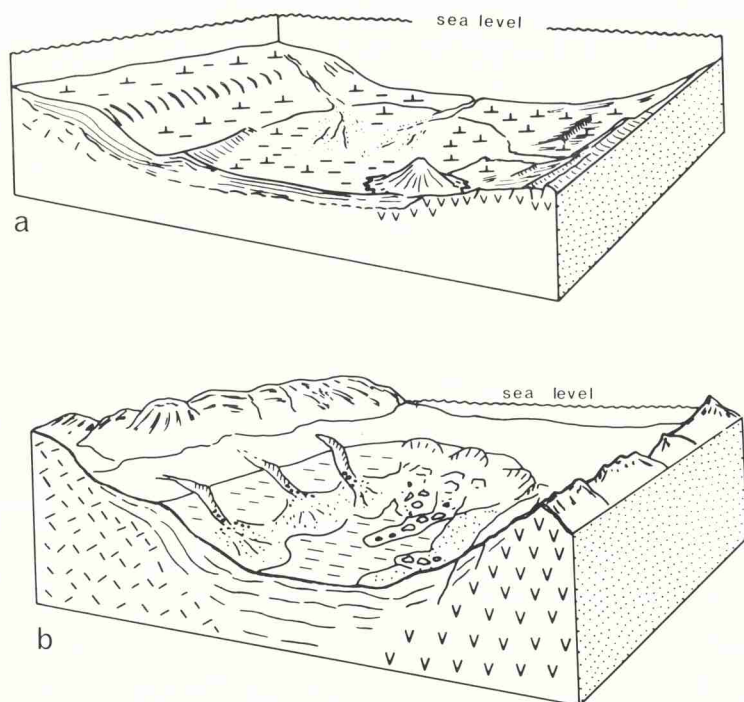


FIG. 10. Sedimentary environment models for submarine basin plains showing schematic distribution of facies and morphological elements. No fixed scale applies: basin plain areas vary from $<200 \text{ km}^2$ to $>100,000 \text{ km}^2$, gradients flat to very low. (A) Undersupplied, (B) oversupplied.

One or more of the basin margins may be a zone of active synsedimentary faulting, and sediment supply is large compared with the basin floor area. The basins fill up rapidly with progradational, mounded and onlap fill geometries and an overlapping, somewhat chaotic distribution of coarse and fine-grained facies is developed.

Modern basin plains of this type occur in regions of active compressive or strike-slip tectonics, and are particularly well known from the Californian borderland (Gorsline *et al.* 1984) and the New Zealand continental margin (Spörl 1980). The classical ancient examples are the Oligo–Miocene periadriatic foredeep basin plains of Italy (Ricci Lucchi 1975, 1978).

(4) Morphological elements and vertical sequences

The first steps in interpreting the palaeoenvironment from an ancient rock series are to identify the facies correctly and then to assess, as far as exposure and structural complexity will allow, the vertical arrangement and horizontal distribution of those facies. We are just begin-

ning to recognize particular vertical sequences and horizontal associations that we believe are characteristic of specific morphological elements in the deep sea (e.g. Rupke 1978; Walker 1978).

Some of these sequences are shown schematically in Fig. 11. Several types of canyon or channel-fill are recognized: a blocky, massive coarse-grained fill of canyon or proximal channel; more regular fining-upward sequences of mid-slope or mid-fan channels; packets of sands deposited in distributary-type channels; and a blocky to fining-upward mud-dominated channel-fill. Regular coarsening-upward sequences appear typical of more proximal sandy (suprafan) lobes and probably also of proximal muddy lobes, whereas distal (terminal) sandy and muddy lobes are commonly more symmetrical. Other mounds on the seafloor 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. Irregular sequences also characterize levees, interchannel, smooth slope and several basinal environments. The main differences

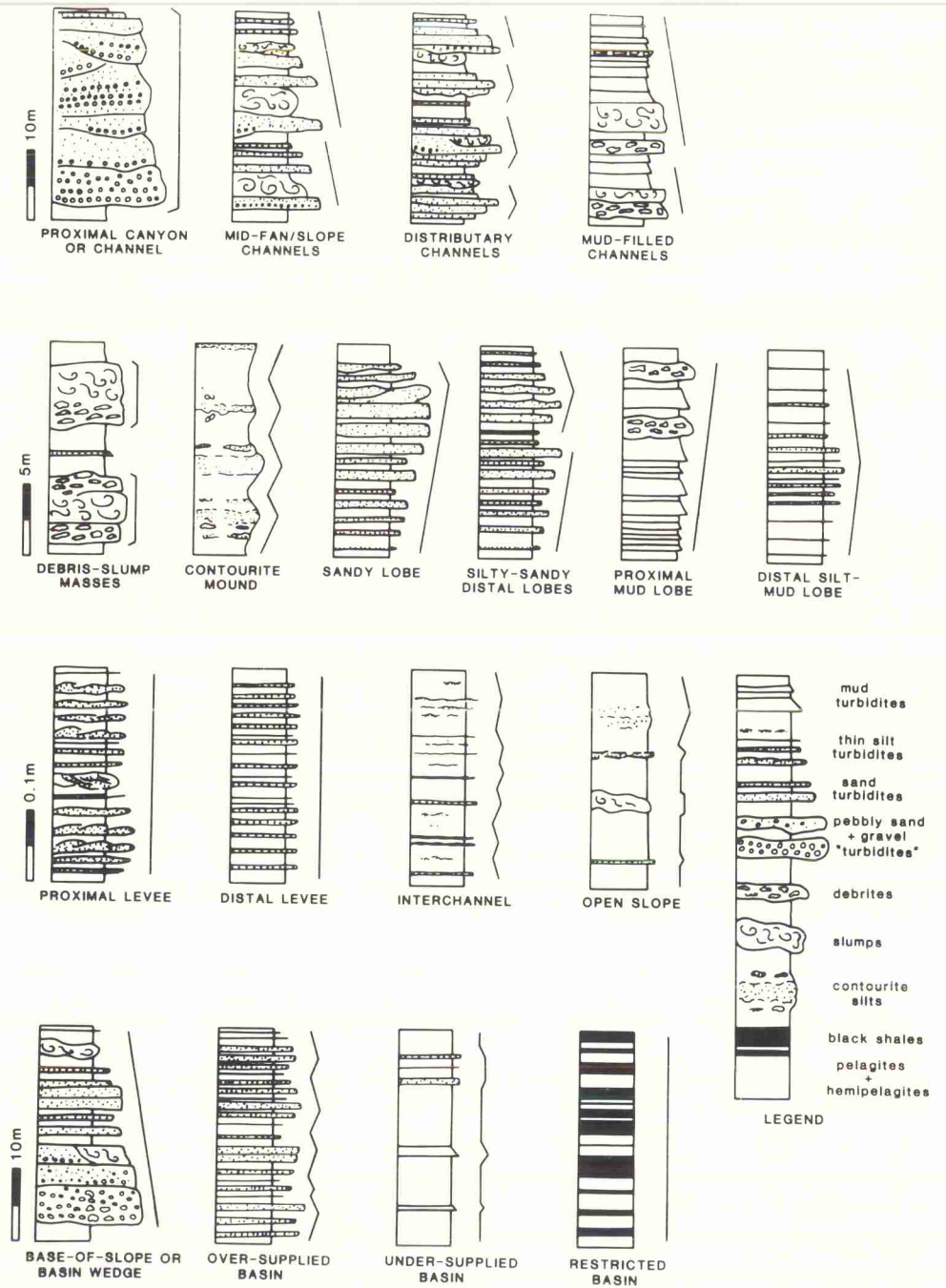


FIG. 11. Typical vertical sequences of turbidites and associated sediments from various morphological elements in the different deep-sea environments. Fining-upward, coarsening-upward, blocky, symmetrical and irregular sequence types are indicated by the lines to the right of lithological columns.

between these settings are 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 over-supplied basins.

These are generalized sequences associated with specific elements in the deep sea. They do show variations of scale, sedimentary materials and regularity and it is not always easy to make a definitive interpretation. It is still more difficult to reconstruct the larger palaeoenvironment as this requires the vertical or lateral association of several of the required elements. It must also be remembered that isolated channel, lobe or other sequences may develop in any of the major environments.

Controls and rates

A whole new dimension to our understanding of the deep sea is added when we consider the factors that control the development of different processes, facies or environments (Stow *et al.* 1982, 1984; Howell & Normark 1982), the rates at which these controls operate (Howell & Von Huene 1980; Blatt *et al.* 1980; Stow *et al.* 1984), and also the material budgets involved (Gorsline 1978, 1984). Three primary controls, as well as a number of secondary controls, on deep-sea sedimentation can be identified: (1) sediment

type and supply, (2) tectonic setting and activity, and (3) sea level. These controls are not, of course, entirely independent; for example, tectonic factors influence sediment supply or local sea-level changes, and so on (Fig. 12).

(1) Sedimentary controls

One of the key sedimentary variables is, clearly, the type of sediment available for deposition or redeposition. The most important types are clastics, including gravel, sand and mud grade terrigenous materials, and biogenics, including skeletal or reefal platform debris, calcareous and siliceous oozes. Locally, evaporites, volcanoclastics, organic-carbon-rich and other sediments are also significant. The volume and rate at which sediments are made available for deposition is another important variable. For example, major rivers can provide a large and rapid supply, glaciers a somewhat lesser supply, and pelagic settling a relatively low supply of sediment. Finally, the number and position of input points exerts a considerable influence over the morphology developed.

The secondary factors that influence sediment type and supply are also illustrated schematically in Fig. 12. Rates of accumulation and denudation, together with the periodicity or frequency of significant sedimentary aspects are shown in Fig. 13.

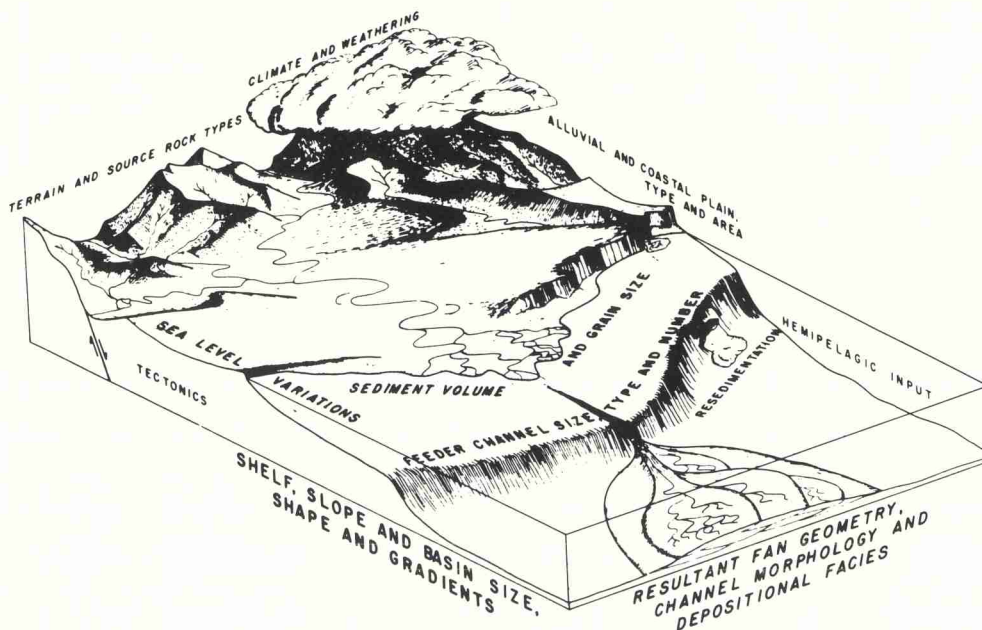


FIG. 12. Schematic diagram of the major controls on sedimentation in the deep sea (from Stow *et al.*, 1982, 1984).

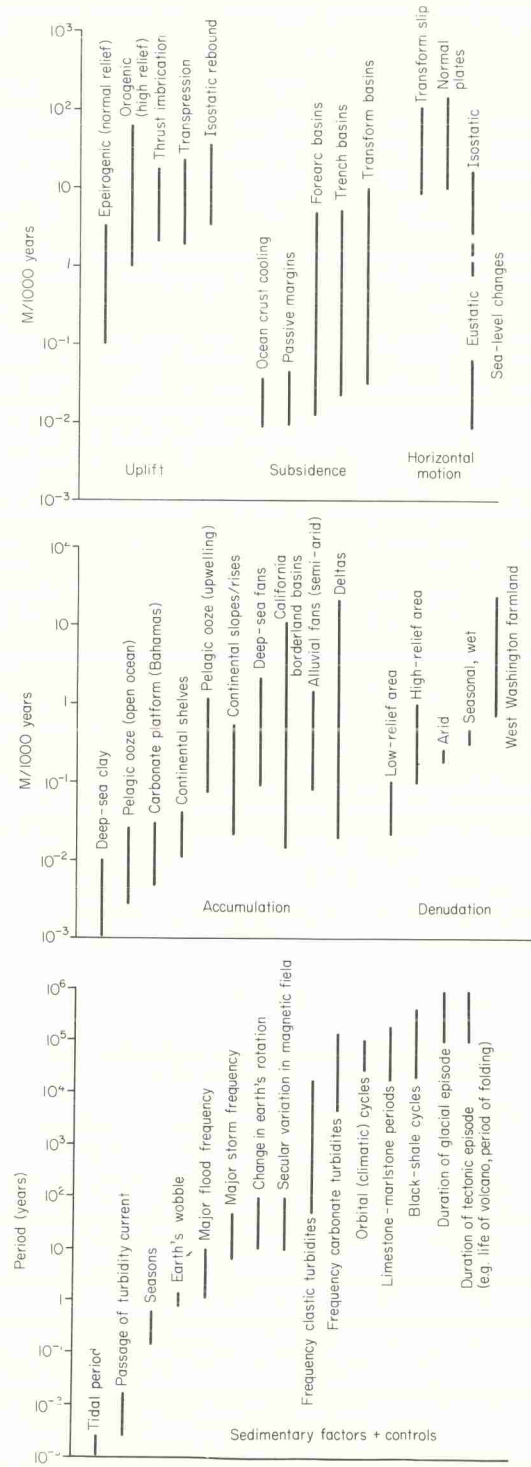


FIG. 13. Rates (in m 1000 yr⁻¹) of tectonic and sedimentary processes and periods (in years) of sedimentary factors that influence sedimentation in the deep sea.

(2) Tectonics

The broad tectonic settings in which the various deep-water systems can develop include: mature passive margins (American or African types); immature or rifting passive margins; convergent margins with arc or trench systems; transform margins; marginal seas and back-arc basins; intraoceanic basins on the flanks of ridges and seamounts; and intracratonic basins on continental shelves and within continents.

These tectonic settings exert a first-order control on the types of system developed by affecting the regional stress regime, rates of uplift and denudation, drainage patterns, coastal plain and shelf widths, 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. This degree of tectonic activity varies temporally and spatially within and between the main tectonic settings, but is most pronounced in convergent, transform and young passive margins.

Secondary factors involved in tectonic activity include the rates of horizontal and vertical motion, the maturity of the margin, and the relationship of a particular setting to neighbouring plates. The rates of motion can be critical. If deposition rates are slower than tectonic rates, growth will be controlled by tectonism rather than by sedimentary factors such as fluctuating gradients and migrating channels, distributaries, and terminal lobes. Some of these rates are indicated in Fig. 13.

(3) Sea level

Fluctuation in sea level not only affects the nearshore realm of sedimentation, but also profoundly influences deep-sea depositional and re-sedimentation patterns. Shoreline sources such as rivers or littoral drift cells either may have direct access to basin slopes during periods of low sea level or indirect access through paralic and continental shelf environments during periods of high sea level.

Sea level changes may be global (eustatic) or regional in nature. Eustatic fluctuations occur as a result of a change in the total volume of ocean basins or a change in the volume of seawater. The volume of ocean basins is affected by four secondary factors (Pitman 1979): sediment input, continental collision and subduction, growth of seamount chains, swelling and shrinking of mid-ocean ridge systems.

Mostly, the rates of sea-level change effected (Fig. 13) appear too small in comparison with tectonic rates of uplift and subsidence to have any appreciable effect on active margins. However, prolonged eustatic changes, due to changes in ocean-ridge volume for example, acting in consort with passive margin subsidence will have a significant effect on sea level along the more stable coastlines. And also, the locking-up of very large amounts of water in expanded polar ice sheets during glacial periods and its release during warm climatic epochs, can cause enormous sea-level fluctuations of up to 10 m 1000 yr⁻¹.

Regional fluctuations in sea level as a result of local tectonic and isostatic factors can be very much greater than most eustatic changes (Fig. 13), as evidenced by the rates of tectonic processes discussed in the preceding section. They are not always readily distinguished from and may completely mask the eustatic change. However, various attempts have been made to chronicle the eustatic as distinct from local sea level fluctuations through time, and the currently popular world wide sea-level curve is that of Vail & Mitchum (1979). Clearly, during periods of very high stands the deep-sea sedimentation regime is likely to contrast markedly with that during very low stands.

Economic aspects

The demonstrable importance of deep-sea sediments as a source of great economic potential has certainly provided a new impetus to their study, as well as a new focus for much research. The materials they contain that have so far attracted the most attention are: oil and gas, ferromanganese nodules, metalliferous sediments and ores, and the uranium series elements of black shales. Most of these are in the pelagic and associated facies discussed by Leggett (this volume); I will only briefly consider the potential hydrocarbon resources in this paper.

There have been numerous important hydrocarbon discoveries in deep-sea sediments of Devonian to Tertiary age (Table 1) in the U.S.A. (Walker 1978) in the North Sea (Woodland 1975; Illing & Hobson 1981) and off the Brazilian coast (Tofelli & Barros 1979). There are also vast new areas of petroleum potential beneath present-day continental margins, where canyon, fan or other reservoirs may become economically viable in the near future (Yarborough *et al.* 1977; Mattick *et al.* 1978; Wilde *et al.* 1978).

These reservoirs are mostly found in the

TABLE 1. *Hydrocarbon reservoirs in submarine fans and associated sequences*

Author	Area	Other details	Reference
Barbat	1958 Los Angeles Basin, California		In: L. G. Weeks, (ed.) <i>Habitat of Oil</i>
Sullwold	1960 Los Angeles Basin, California	Tarzana Fan, late Miocene	<i>AAPGB</i> , 44 , 433-57.
Yerkes <i>et al.</i>	1965 Los Angeles Basin, California		<i>USGS Prof. Pap.</i> , 420A,
Mayuga	1970 Los Angeles Basin, California	Wilmington Oilfield	<i>AAPG Mem.</i> 14 , 158-84
Gardet	1971 Los Angeles Basin, California		<i>AAPG Mem.</i> 15 , 298-308
Weser	1978 Los Angeles and Ventura Basins		<i>AAPG Stud. Geol.</i> 7 , 227-42
Nagle & Parker	1971 Ventura Basin, California		<i>AAPG Mem.</i> 15 , 254-97
Hsü	1977 Ventura Basin, California	Ventura Field, lower Pliocene turbidites	
Sullwold	1961 Great Valley, California	Stevens Sandstone, Bakersfield arch, Miocene	<i>AAPGB</i> 61 , 137-68
Martin	1963 Great Valley, California	Rosedale Channel, late Miocene	In: Peterson & Osmond (eds)
Seiden	1975 Great Valley, California	Asphalt, San Joaquin Valley, Miocene	<i>Geometry of Sandstone Bodies</i>
Sanem & Soddard	1965 Great Valley, California	Stevens Sand, San Joaquin Valley, Miocene	<i>AAPGB</i> , 47 , 441-56
Callaway	1968 Great Valley, California	W. San Joaquin Valley, Miocene	<i>Pac. Sect. Am. Ass. Petrol. Geol.</i> 40th AGM
Bazeley	1972 Great Valley, California	San Emido Nose Field, Miocene	<i>AAPGB</i> , 49 , 1089
Maier <i>et al.</i>	1975 Great Valley, California	Naval Petroleum Reserve, Miocene	<i>AAPG-SEPM Pac. Sect. Guidebk</i> , 43rd AGM
MacPherson	1978 Great Valley, California	San Joaquin Valley, Bakersfield Arch, Miocene fans	<i>AAPG Mem.</i> 16 , 297-316
Edmonson	1965 Sacramento Valley, California		<i>USGS Prof. Pap.</i> 912 , 109 pp.
Dickas & Payne	1967 Sacramento Valley, California	Upper Palaeocene Channel	<i>AAPGB</i> , 62 , 2243-74
Weagant	1972 Sacramento Valley, California	Grimes Gas Field	<i>San Joaquin geol. Soc. select. Pap.</i> 3 , 36-61
Paine	1966 Louisiana and Gulf of Mexico		<i>AAPGB</i> 51 , 873-82
Sabate	1968 Louisiana and Gulf and Mexico	Pleistocene oil and gas fields	<i>AAGP Mem.</i> 16 , 428-39
Benson	1971 Louisiana and Gulf and Mexico	Oligocene	<i>Trans. Gulf-Cst Ass. geol. Socs</i> 16 , 261-74
Hoyt	1959 Texas		<i>Trans. Gulf-Cst Ass. geol. Socs</i> 18 , 373-86
			<i>Trans. Gulf-Cst Ass. geol. Socs</i> 21 , 1-14
			<i>Trans. Gulf-Cst Ass. geol. Socs</i> 9 , 41-50

- Galloway & Brown 1973 Texas
 Bloomer 1977 Texas
 Dixon 1972 Pennsylvania
- Vedros & Visser 1978 Arkona Basin, Oklahoma
- Jacka *et al.* 1968 Delaware
- Thomas *et al.* 1974 North Sea
 Parker 1975 North Sea
- Fowler 1975 North Sea
- Walmsley 1975 North Sea
 De'ath & Schuyleman 1981 North Sea
- Carman & Young 1981 North Sea
- Heritier *et al.* 1981 North Sea
- Kessler *et al.* 1980 North Sea
 Skjold 1980 North Sea
 Stow *et al.* 1980 North Sea
- Stow *et al.* 1982 North Sea
- Tofelli & Barros 1979 Brazil continental margin
- Ojeda 1982 Brazil continental margin
 Casnedi 1983 Italy, Periadriatic basins
- Carboniferous
 Bradford Oilfield
 Red Oak Sandstone, Carboniferous
 Bell Canyon Formation
- Forties Field, Palaeocene
 Tertiary sands
- Montrose Field, Palaeocene
- Forties Field, Palaeocene
 Magnus Field, Jurassic
- Forties Field, Palaeocene
- Frigg Field, Eocene
- Cod Field, Palaeocene
 Balder Field, Palaeocene
 Brae Field, Jurassic
- Brae Field, Jurassic
- Various fields
 Various fields
 Pliocene Cellino Formation
- AAPGB* 57, 1185–218
AAPGB 61, 344–59
Mem. Am. Ass. Petrol. Geol. Eastern Section Field Trip Gdbk
In: Stanley & Kelling (eds) Sedimentation in Canyons, Fans and Trench
Pac. Sect. Gdbk Am. Ass. Petrol. Geol. Soc. econ. Paleont. Mineral., Tulsa, Permian Basin Section Guide I
AAPGB 58, 396–406
In: A. W. Woodland (ed.) 447–52
Petroleum and the Continental Shelf of N. W. Europe, 467–76
Shelf of N. W. Europe, 467–52
In: Illing & Hobson (eds) Petroleum geology of the Continental Shelf of N. W. Europe
In: Illing & Hobson (eds) Petroleum geology of the Continental Shelf of N. W. Europe
In: Illing & Hobson (eds) Petroleum geology of the Continental Shelf of N. W. Europe
NPF, Geilo Volume
NPF, Geilo Volume
Mem. Am. Ass. Petrol. Geol. mtg, Houston, 1980
J. Petrol. Geol. 5, 129–68; 6, 103–4
Ist Cony. Brasil. Petrol. 1:11, 51–61
AAPGB 66, 732–49
AAPGB 67, 359–70

coarser-grained facies (sands and gravels) of canyons, channels, lobes, slope wedges and over-supplied basins. Such reservoir facies, however, are commonly associated with muddy deposits that may be sufficiently enriched in marine or terrigenous-derived organic matter to provide an adequate source of oil or gas after burial and thermal maturation. Furthermore, the particular organization of potential reservoir and source facies in certain deep-sea settings (fans for example) provides the possibility of suitable stratigraphic or structural-stratigraphic traps forming with time.

Future directions

If the next three decades are as productive as the past three, then we will cover a very large amount of ground in deep-sea studies. But just what ground will that be, where will it lead and how will we carry out the research? These questions are addressed briefly in this concluding section.

(1) Areas of study

Although we may have now reached at least some consensus in the five areas discussed in this review paper, there remains much room for improvement to, and refinement of, the working models outlined here. In particular, the following developments may be anticipated: development of rigorous physical models of processes and of the critical conditions for erosion, transport and deposition; emphasis on the effects of normal currents in the deep sea, particularly on the growth of contourite drifts and fluvial-like morphologies; work on fine-grained facies, dune-bedded sands and black shales; elaboration of the slope and basin environmental models, work on the shelf-slope interface, and further emphasis on characterizing individual morphological elements; further exploration of sedimentary controls, rates and budgets and a more critical understanding of their effects; and determination of the relationships between source-rock organofacies and lithofacies, the effects of diagenesis on reservoir rocks, and the geometry of reservoir bodies.

Other quite different avenues of study will doubtless also be pursued, particularly in fields where we are almost but not quite at the synthesis and model-construction stage. These include: the petrography and geochemistry of both coarse- and fine-grained deep-sea facies, and the development of post-depositional diagenetic models; the influence of organisms on sediments in terms of primary productivity,

organic matter content and particle coatings, erosion and suspension, and bioturbation; and the long-awaited marriage between studies of ancient and modern sediments.

(2) Methods of research

In many ways, even with the advent of deep-ocean drilling technology and the voyages of the *Glomar Challenger*, our studies of deep-sea sediments have remained relatively simple. I suspect that the next three decades will bring immense sophistication to both the methodology and technology we use. Again, I am guessing as well as being selective when suggesting that the advances listed below will prove the most constructive: technological developments in, for example, marine-based observation of smaller and smaller sea-bottom features approaching the scale of observations on land, and laboratory-based studies at the ultramicroscopic scale; the use of actualistic experiments on the seafloor (or in lakes, perhaps), as well as of actual observations on the water column and recently deposited sediments; specific programmes designed to test sedimentological models using closely spaced drill sites or sea to land drill transects, and to examine in detail particular areas of the seafloor using a multidisciplinary approach; the computer handling of increasingly large data-sets as well as the systemization of data produced in different countries and different laboratories; a communications breakthrough that overcomes the many barriers between laboratories, countries and languages, and that renders accessible large volumes of information that are otherwise comparatively lost.

The deep-sea sedimentological community is already grappling with these areas and methods of study. It will perhaps not be too long now before we can claim to have entered the 'Age of Order'.

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