Deep-water fine-grained sediments: facies models

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SUMMARY: Based on a large amount of published data and stimulated by the papers and discussion at the International Workshop on Fine-Grained Sediments held in Halifax, Canada in August 1982, we have attempted a synthesis of deep-water fine-grained sediment facies. Three main facies groups related to depositional processes can be identified: turbidites, contourites and pelagites/hemipelagites. There is a continuum between the different processes and hence a continuum between facies. Nevertheless, it is possible to define several distinct facies models within each of these groups on the basis of sedimentary structures, texture and composition, and to provisionally interpret these in terms of depositional hydrodynamics. Patterns of horizontal and vertical facies distribution can be related to depositional subenvironments. There is much variability within and departure from the facies models we propose, and many interesting and problematic areas of research remain in the quest for better understanding of deep-water fine-grained sediments.

Only five years ago it was fair to say that fine-grained sediments were grossly understudied by comparison with, for example, sands and gravels. However, in the last few years, a large amount of data has appeared on both modern and ancient clays, muds and silts and their biogenic equivalents (e.g. Potter *et al.* 1980). A variety of fine-grained facies types have been recognized on the basis of size, grade and internal organization of beds (e.g. Piper 1978; Stow 1984a; in press), and new questions have emerged regarding their origin, processes and environments of deposition.

However, there is still need for a general synthesis or organization of these data into composite facies models. Such models may then serve as a framework to guide description, a standard against which to compare new information, a predictive tool in new geological settings and as a basis for hydrodynamic interpretation (Walker 1975). There have been several recent attempts to construct facies models for deep sea sediments, including turbidite muds and silts (Piper 1978; Stow & Shanmugam 1980), 'homogeneous' muds (Stanley 1981), carbonate turbidites (Hesse 1975; Stow, Wezel *et al.*, this volume), muddy contourites (Stow & Lowell 1979; Stow 1982; Faugères *et al.* 1984; Gonthier *et al.* this volume), hemipelagities and pelagites (Hoffert 1980; Einsele 1982; Jenkyns, in press; Thornton, this volume; Hill, this volume). There have been similar attempts to collect together the diverse data on more shallow-water fine-grained sediments (e.g. tidal deposits, Ginsburg 1975; deltaic deposits, Coleman 1976; shelf deposits, Shepard *et al.* 1960): these are not considered further here.

In this paper we attempt, therefore, to collate information on deep-water fine-grained sediments from our own studies, from a growing body of literature, and from the papers and discussion at the International Workshop on Fine-Grained Sediments (Halifax, Canada, 1982) from which this volume has been edited. We recognize and describe various facies models within three main facies groups: fine-grained turbidites, muddy contourites, and pelagites/hemipelagites; and briefly discuss their interpretation in terms of depositional processes. We then consider the distribution of these facies in different environmental settings and some of the post-depositional changes that they undergo. Our discussion highlights some of the interesting and problematic areas of interpretation.

There is by no means a concensus view on the description and interpretation of fine-grained sediments. Many readers will feel we are premature in defining some of the facies models and are guilty of generalization in the discussion of their horizontal and vertical distribution in the deep sea. However, we believe that some degree of synthesis and simplification is necessary and that the conclusions are useful. Certainly, they should be the subject of rigorous scrutiny by future research.

Processes and facies

There appears to be a continuum of processes operating in the deep sea (e.g. Walker 1978; Stow 1984a). These include resedimentation (mass gravity) processes, normal bottom currents and pelagic settling (Fig. 1). There is close interaction and overlap between these processes both during transport and during the final stages of deposition. Any single resedimentation event, for example, may be initiated by the slumping of unstable slope sediments and then, by mixing with seawater, evolve into a debris flow, a high concentration turbidity current and, finally, a low-concent-

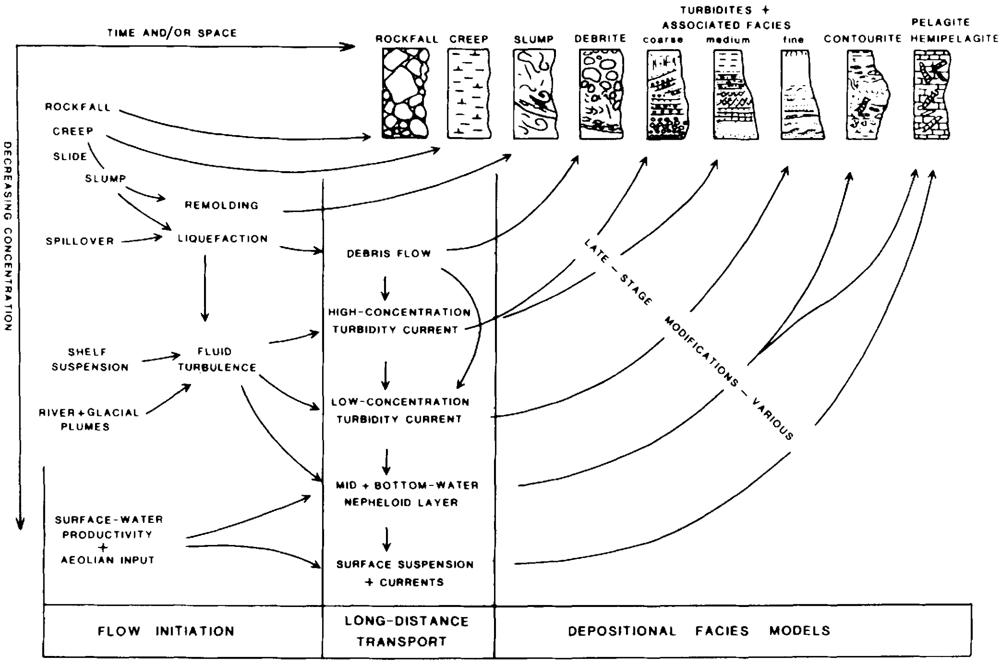

FIG. 1. Diagram showing the main processes of flow initiation and long distance transport in the deep sea and the depositional facies models related to each distinct process. In nature, there is a continuum of processes and facies. (Modified after Walker 1978).

ration turbidity current. Deflection of the low-velocity dilute tail of a turbidity current by a regional contour-following bottom current will cause the downslope flow to grade imperceptibly into an alongslope bottom current. Pelagic sediment settling vertically through the water column in the open ocean may be deflected slightly by the action of very weak bottom currents.

The mechanisms involved in erosion, transport and deposition of fine-grained sediments within this process continuum are discussed more fully in Section 1 of this volume (papers by McCave, Gorsline, Kranck, Eittreim, all this volume). We recognize that there is also a facies continuum in the deep sea but that it is nevertheless possible to distinguish facies types that are the result of a distinct turbidity current, bottom current and pelagic processes. Our facies models are, therefore, closely related to the depositional process (Fig. 1). We have not attempted to include facies models for slumps, sediment creep deposits or debris-flow deposits ('debrites'), although these commonly involve fine-grained sediments (but see Naylor 1980, 1981; Thornton, this volume; Stow 1984 a,b, in press.).

Several other facies types have been proposed in the literature, resulting from processes such as turbid layer flows (Moore 1969; Stanley and Maldonado 1981), suspension cascading (McCave 1972), unifite flows (Stanley 1981), nepheloid layers (Biscaye & Eittreim 1977), canyon currents (Drake *et al.* 1978) and so on. These do not, however, differ significantly either in character or in the depositional process from the broader categories that we suggest and do not appear valid to us as distinct facies. They do serve to indicate the range of mechanisms that may exist within our general process groups.

Black shales may be deposited by a variety of processes and so are not considered as a separate facies in this context. However, several of the papers in this volume describe organic-rich sediments of different types and deposited by different processes (e.g. Crevello *et al.*; Isaacs; Anastasakis & Stanley; Thickpenny, all this volume). Arthur *et al.* (also this volume) have attempted to synthesize a more generalized model for the black shale sedimentation. Neither do we include a detailed discussion of the acoustic character of deep-sea sediments and the recognition of acoustic facies: useful summaries are given by Damuth (1975, 1978), Jacobi (1982) and Nardin *et al.* (1979).

Fine-grained turbidites

Fine-grained turbidites are made up of material dominantly in the silt and clay size grades (i.e. over 50% less than 63 μm grain size). They are widespread in the deep sea and, volumetrically, the most important of our facies groups (Piper 1978). They occur as very thin to very thick beds that have been deposited rapidly (from a few hours to a few days) from a single resedimentation event. The chief criteria that can be used to distinguish them from other facies in the deep sea which they may resemble, include:

(1) a regular vertical sequence of sedimentary structures commonly associated with a positive grading;
(2) the presence of sedimentary structures indicating rapid deposition, with bioturbation restricted to the tops of beds;
(3) compositional, textural or other features which indicate that they are exotic to their depositional environment.

With high resolution acoustic profiling systems, fine-grained turbidites show good penetration and appear well stratified, in contrast to the hard, often irregular, reflective bottom found in coarser turbidites. Fine-grained turbidites occur in a variety of environments that can be distinguished in seismic reflection profiles, such as levees and ponded basin plains. They may pass laterally into coarser turbidites that often fill channels and appear as discontinuous strong reflectors in seismic reflection profiles.

In Bouma's (1962) classical sequence for sand-mud turbidites, all fine material was classed as a featureless E division. Later work distinguished between turbiditic and pelagic mud (Kuenen 1964; Van der Lingen 1969) and Piper (1978) further subdivided the turbidite mud into E1, E2 and E3 structural divisions. Fine-grained turbidites that occur alone are, in some cases, the distal equivalents of thicker-bedded, coarser-grained turbidites, but they may also occur without the proximal sand-mud 'parents'. On the basis of grain size, internal organization and composition we recognize four distinctive facies within this group: silt turbidites, mud turbidites, biogenic turbidites and disorganized turbidites. Each of these can best be described in terms of a separate facies model.

Silt turbidites

In distal turbidite environments, silt beds (> 70% silt-sized particles) are more abundant than sands and commonly occur as thin or medium-bedded turbidites. Early work demonstrated a change from sands to silts moving basinwards from fans to abyssal plains (e.g. Horn *et al.* 1971; Normark & Piper 1972), and there have since been many descriptions of silt turbidites from the deep sea. Well-documented examples include those of the Aleutian Trench (Piper 1973), the Indus fan (Jipa & Kidd 1974), the Nile cone (Maldonado & Stanley 1976), the Antarctic continental rise (Piper & Brisco 1975), the Zaire fan (van Weering & van Iperen, this volume) and the south-east Angola Basin (Stow 1984c). Similarly, many ancient slope, fan and basin plain successions comprise interbedded siltstone turbidites and mudstones (e.g. Lundegard *et al.* 1980; Pickering, this volume; Stow, Wezel *et al.*, this volume).

Facies model (Fig. 2)

Silt turbidites commonly exhibit the same suite of structures (Piper 1978; Kelts & Arthur 1981) as the thicker-bedded, classical sandy turbidites described by Bouma (1962). A *complete* sequence from top to bottom would be (Fig. 2):

F hemipelagic or pelagic sediment, biogenic, bioturbated
E mud, graded, commonly bioturbated
D fine silt, parallel-laminated alternating silt and minor clay, often with synsedimentary deformational structures, graded
C medium silt, cross-laminated, rarely convolute, graded

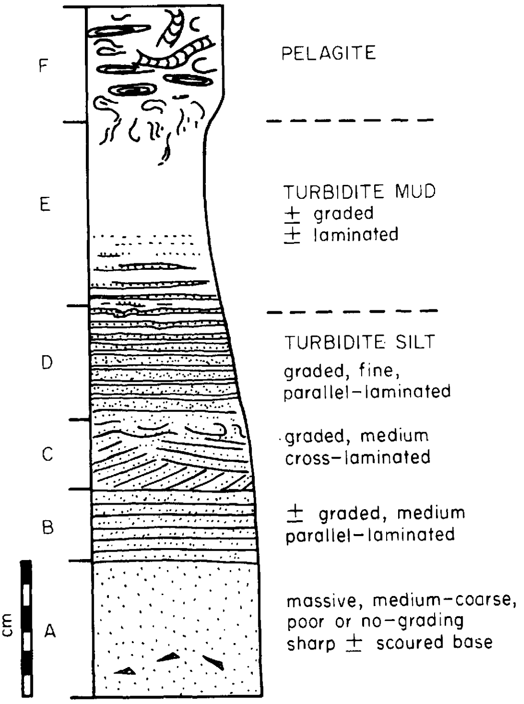

FIG. 2. Silt turbidite facies model. Structural divisions follow Bouma (1962) as modified by van der Lingen (1969)

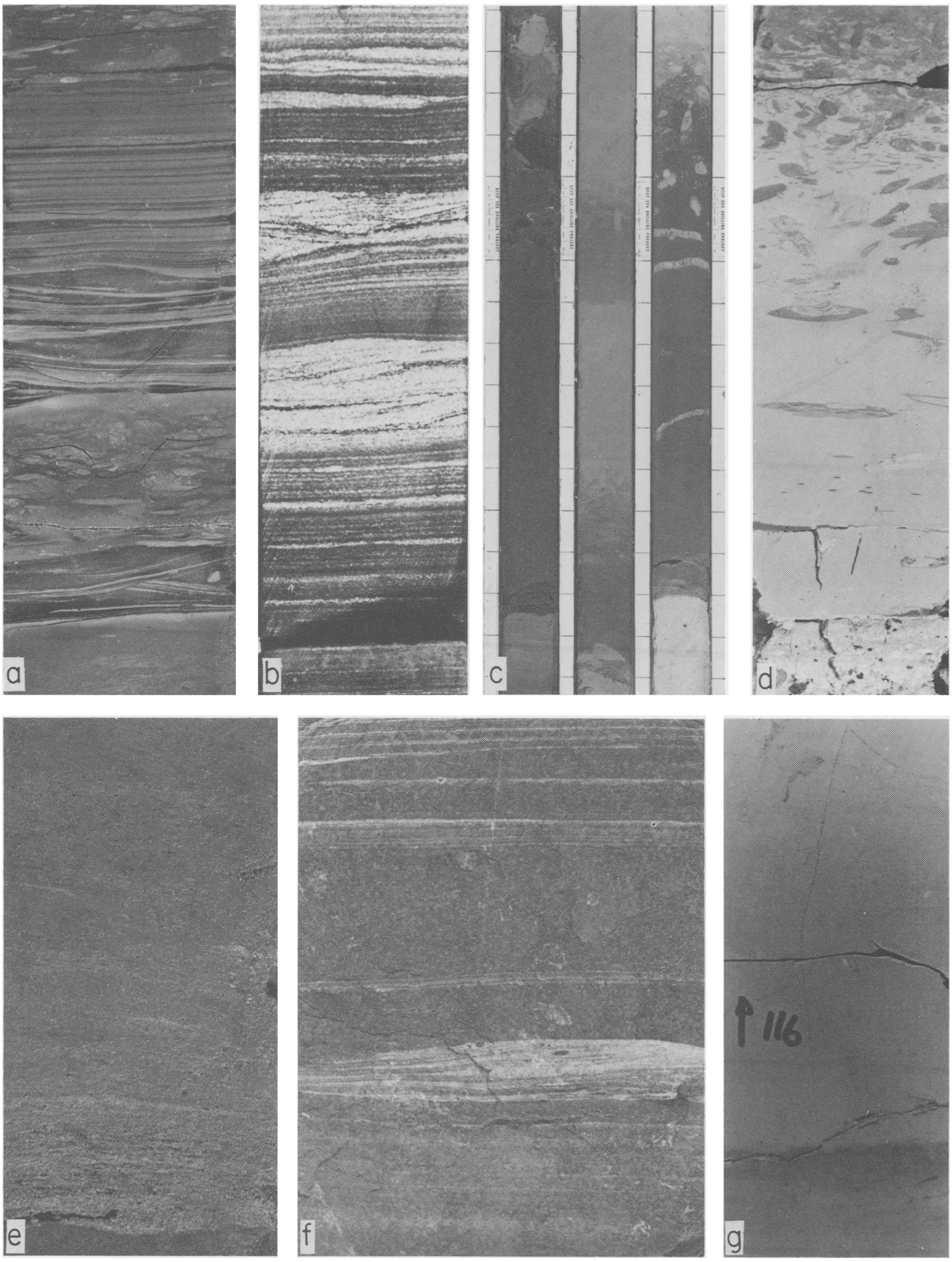

FIG. 3. Photographs of fine-grained turbidite facies. Width of sections approximately 7 cm. (a) Silt turbidites, upper Cretaceous, DSDP Site 530A, Angola Basin. (b) Silt and mud turbidites, upper Jurassic, Brae slope apron, North Sea. (c) Mud and ooze turbidites (3 core sections), Plio-Pleistocene, DSDP Site 530B, Angola Basin. (d) Calcirudite–calcilutite turbidite, upper Cretaceous, DSDP Site 530A, Angola Basin. (e) Disorganized silty-mud turbidite, Pleistocene, DSDP Site 530B, Angola Basin. (f) Silt laminated mud turbidites, Cambro–Ordovician, Halifax Formation, Nova Scotia (g) Calcarenite–calcilutite turbidite, upper Cretaceous, Scaglia Rossa Formation, Italy.

B medium silt, parallel-laminated, graded
A medium-coarse silt or sandy silt, massive, poor or no grading, some floating clasts, minor scouring at base

The interpretation and origin of this sequence, in terms of flow decline during deposition from a single turbidity current event, we suggest is similar to that proposed for the Bouma sequence in sandy turbidites (Harms & Fahnestock 1965). However, the complete sequence is very rarely found (e.g. Fig. 3). More commonly, base-cut-out sequences occur comprising CDE and DE divisions (graded sorted silts and laminated silts respectively of Piper, 1978) in beds from 1 to 10 cm in thickness. Thicker-bedded and coarser silt top-cut-out (AB and B) or mid-cut-out (AE) sequences occur more rarely (the ungraded massive silts of Piper, 1978).

Mud turbidites

Deep-sea terrigenous successions are commonly dominated by muds and, in many settings, between 50% and 80% of this is of turbidite origin (e.g. Hesse 1975; Piper 1978). The features that characterize mud turbidites are subtle and frequently overlooked, especially when the turbidites are interbedded within a thick monotonous hemipelagic mud sequence. However, clear examples have been described from both the east and west Mediterranean (Rupke & Stanley 1974; Bartolini *et al.* 1975; Got, this volume), the Cascadia Channel (Griggs & Kulm 1970) and Astoria fan (Nelson 1976) in the eastern Pacific, the Angola Basin (Stow 1984c), the Northwest Atlantic Mid-Ocean Channel (Chough & Hesse 1980), the Laurentian Fan (Stow 1981) and the Cap Ferret fan in the north-east Atlantic (Cremer 1981, 1983).

In ancient sequences exposed on land, weathering and fracturing often make it impossible to discern characteristic structures in mudstones and shales. Where preservation has been good and rocks have been suitably smoothed, as on wave-cut platforms, the same features as described for modern mud turbidites may be even more clearly demonstrated. Examples range from the Precambrian slope facies of northern Norway (Pickering 1982a, this volume), through successions from the Cambro-Ordovician (Piper 1972a; Stow, Wezel, *et al.*, this volume), Silurian (Piper 1972b), Devonian-Carboniferous (Hall & Stanley 1973; Lundegard *et al.* 1980), Triassic (Hicks 1981) and Cretaceous (Ingersoll 1977) to the well-known Cenozoic turbidite formations of the Italian Apennines (e.g. Mutti 1977; Mutti *et al.* 1978; Ricci Lucchi 1978, 1981).

Facies model

From this large amount of data it is possible to synthesize an ideal facies model for mud turbidites (Fig. 4). Piper (1978) proposed the subdivision of Bouma's (1962) E division in to three parts, giving from top to bottom:

F hemipelagic or pelagic sediment
E3 ungraded mud
E2 graded mud
E1 laminated mud
D laminated sand and silt

Apart from the topmost E3 division, mud turbidites commonly show slight but distinct positive grading in both grain size and composition. Textural grading is best observed as a progressive decrease in maximum or mean size through successive silt laminae (E1 division) (Piper 1972b), or as a decrease in silt to clay ratio through the E2 division. A wide range of compositional grading has been observed, including an upwards increase in micas, various clay mineral species and organic carbon, and upward decrease in heavy minerals, quartz and foraminifera (Rupke & Stanley 1975). Carbonate is known to show either an increase or decrease depending on the type and grain size of the carbonate and associated components. Colour changes commonly mirror the textural and compositional grading and provide the simplest method for visual identification of mud turbidites.

Stow (1977) and Stow & Shanmugam (1980) have shown that there is a further hierarchy of structures within the graded laminated mud and silt of Piper's scheme, many of which can be helpful as diagnostic criteria in describing mud turbidites. The *complete* sequence from top to bottom is as follows (Fig. 4):

P pelagite or hemipelagite, bioturbated
T_8 turbidite (± part pelagite), microbioturbated
T_7 ungraded mud, occasionally with silt pseudonodules
T_6 graded mud, often with dispersed silt lenses
T_5 wispy convolute silt laminae in mud
T_4 indistinct, discontinuous silt laminae in mud
T_3 thin, regular, continuous parallel silt laminae in mud
T_2 thin, irregular, slightly lenticular silt laminae in mud, often with low-amplitude climbing ripples
T_1 thick mud layer, often with thin convolute silt laminae
T_0 thick, basal, lenticular, silt lamina, often with fading-ripple top, microlaminated interior and scoured, load-cast base

The complete sequence is interpreted as a single

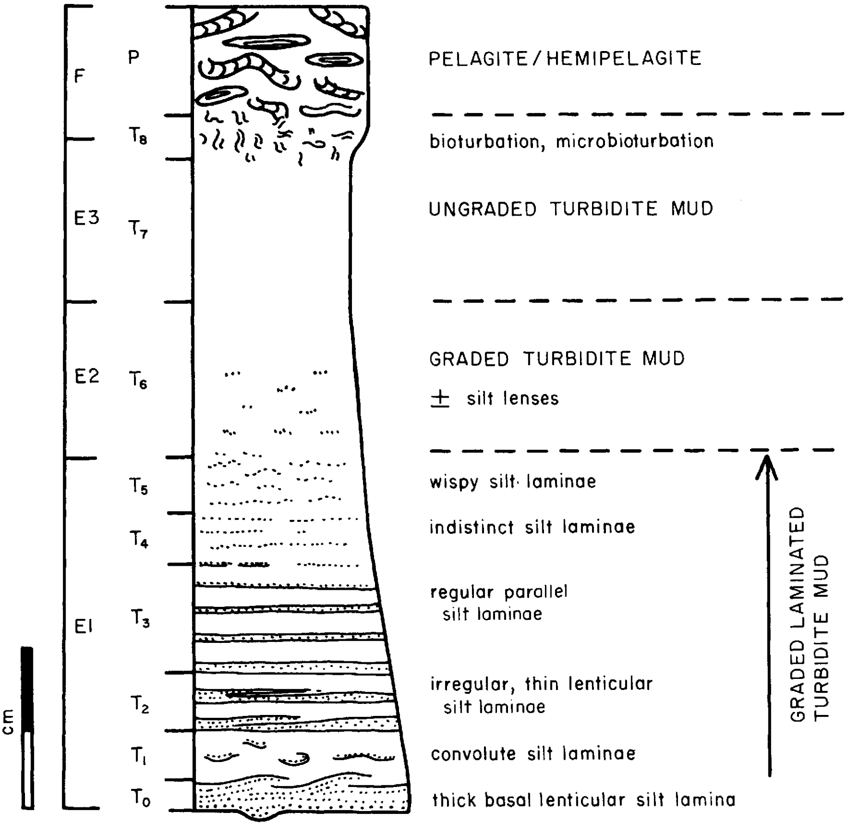

FIG. 4. Mud turbidite facies model. Structural divisions after Piper (1978) and Stow (1977).

depositional unit from a large fine-grained turbidity current (e.g. Stow & Bowen 1980). Internal lamination in the basal layer (T_0) and migrating ripple lamination indicate periods of tractional movement during deposition of the coarser silt grains. The silt ripples with muddy troughs (fading ripples) are deposited from turbidity currents containing a high proportion of clay-sized material. This commonly also gives rise to a thick mud layer (T_1), with convolute silt laminae formed either by loading into the soft mud or by incipient ripple development. At a lower current velocity and with less silt available the very thin laminae of low-amplitude, long wavelength ripples (T_2) are formed. Continued waning of the current velocity as the flow passes results in the overall positive grading observed (T_3-T_4) in which the alternation of silt and mud laminae is believed to be due to the depositional sorting of silt grains from clay flocs caused by increased shear in the bottom boundary layer (Stow & Bowen 1978, 1980; but see Hesse & Chough 1980). The more homogeneous mud unit (T_7) at the top of the sequence comprises the finest silt and clay which was not so effectively sorted into distinct laminae because of the lack of silt and the very fine grain size.

Variability

Although a standard structural sequence can be identified, a wide range of variations is possible and, as with the Bouma sequence for sandy turbidites, the complete set of divisions is rarely present in any one bed (Figs 3 and 5). In many cases the beds are relatively thin (< 10 cm), averaging 2–5 cm, and comprise only the upper parts of the sequence (E_2E_3 or T_4-T_8, base-cut-out units), or lower parts of the sequence E1 or T_0-T_4, top-cut-out units). In the extreme base-cut-out case we have a mud or clay turbidite with no trace of silt lamination, whereas the top-cut-out case is gradational to a silt turbidite. In other cases, only the middle parts of the sequence occur (E_2 or T_2-T_5), or the middle parts are cut out to give an E_1E_3 (T_{078}) bed analogous to a Bouma AE turbidite. Repetition of these different bed types in any one area can lead to the development of a

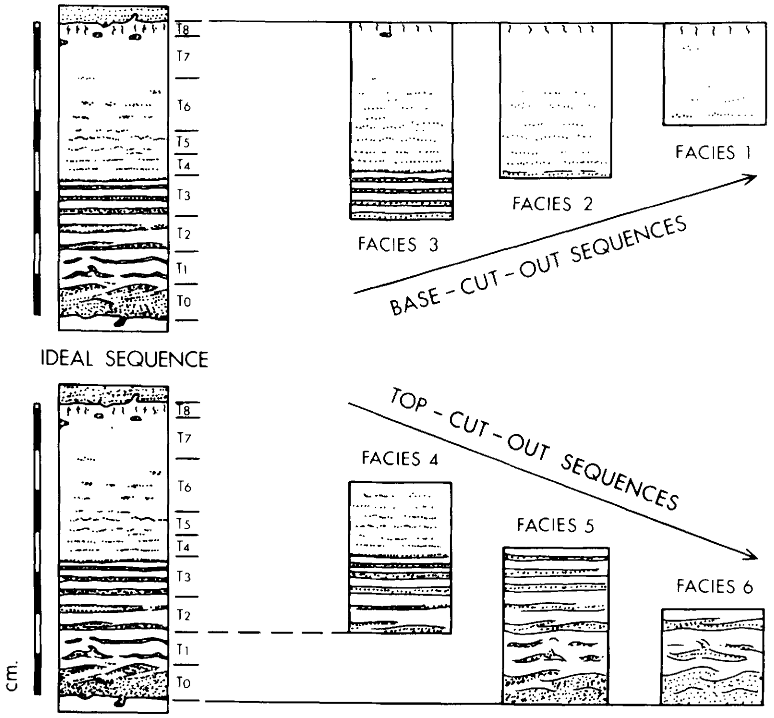

FIG. 5. Examples of variability in mud turbidites: typical base-cut-out and top-cut-out sequences (after Stow, Alam & Piper, this volume).

distinctive facies, such as thick lenticularly laminated silts and thin parallel-silt-laminated mud, in which individual turbidite units may not be readily distinguished. Thin-bedded turbidites of these various kinds have been widely reported from both modern and ancient successions (e.g. Piper 1978; Nelson *et al.* 1978; Stow & Shanmugam 1980; Kelts & Arthur 1981; Hill 1981 and this volume; Chough, this volume; van Weering & van Iperen, this volume).

In some cases, the thick basal laminae of a graded laminated unit is preceded by one or two thinner, finer grained silt and mud laminae. These appear to be precursors of the main depositional event. In other cases, there is variation in the sequence of T_0 to T_8 divisions through a single unit, perhaps related to an instability in the flow conditions.

Both medium and thick-bedded mud turbidites have also been described, commonly 10–50 cm thick but ranging up to several metres (Fig. 3) (e.g. Rupke & Stanley 1974; Piper 1978; Blandpied & Stanley 1980; Stanley 1981; Stow 1984c). These show the same sequence of structures (E_1-E_3, T_0-T_8) but each division is vertically expanded. They probably result either from large muddy turbidity currents derived entirely from fine-grained sediments (e.g. from a slump on a muddy upper slope), or from the ponding of a muddy turbidite tail in an enclosed basin (Wezel 1973; Bowen *et al.* 1984). Stanley (1983) suggests that slow (hemipelagic) settling from detached turbidity currents in a well stratified basin is an important process in the accumulation of beds that he describes as unifites. Similar slow settling from thick dilute turbidity currents may be important elsewhere (e.g. Stow & Bowen 1980). However, there are generally no diagnostic features that allow recognition of depositional processes involving turbidity current detachment. We recognize that the upper parts of mud turbidites may have been deposited by a variety of turbidity-current-related processes.

Biogenic turbidites

Biogenic pelagic sediments are very widespread in the open ocean and, where terrigenous input is minimal, along the continental margin. In areas of topographic relief and/or tectonic activity, such as mid-ocean ridges, seamounts and other submarine highs, resedimentation of pelagic siliceous and calcareous oozes occurs via slumping, debris flows and turbidity currents (e.g. Kelts & Arthur 1981). Off carbonate platforms and reef margins, resedimented biogenic material derived from shallow water is also common. Calcirudites, calcarenites, calcidebrites and carbonate slump deposits have been described from many slope and basinal systems (e.g. McIlreath & James 1979).

As with terrigenous mud turbidites, fine-grained carbonate turbidites are not always readily distinguished from associated pelagic and hemipelagic facies, except where deposition has occurred below the carbonate compensation depth (Hesse 1975). Several authors besides Hesse have encountered similar problems in making clear facies distinctions (e.g. Wilson 1969; Carrasco 1977; Cook & Taylor 1977; Enos 1977; Reinhardt 1977; Homewood & Winkler 1977; Stow, Wezel *et al.* this volume; Faugeres *et al.* this volume; Heath & Mullins, this volume). In other cases, more definitive interpretations of carbonate turbidites have been made for a variety of Palaeozoic to Recent limestones (e.g. Van Andel & Komar 1969; Thomson & Thomasson 1969; Davies 1977; Anatra *et al.* 1980; Kennedy 1980; Kelts & Arthur 1981; Faugeres *et al.* 1982).

Siliceous turbidites rich in diatoms, radiolarians, sponge spicules and other siliceous organisms are less well known than the carbonate equivalents. However, they have been encountered at several DSDP sites in the North Atlantic (Beall & Fisher 1969; Peterson *et al.* 1970; Kagami 1979; McCave 1979) and in the Gulf of California (Curray *et al.* 1980). A few ancient examples have also been documented from the Mediterranean area (Nisbet & Price 1974; Kalin *et al.* 1979), Japan (Imoto & Fukutomi 1975) and North America (Folk & McBride 1978).

There is not always a clear compositional distinction between the different biogenic turbidites so that all mixtures of carbonate, silica and clay materials can occur. In addition to relatively pure carbonate and siliceous ooze turbidites, Stow (1984c) describes marl, 'sarl' (siliceous-clayey) and 'smarl' (siliceous-calcareous-clayey) biogenic turbidites from the south-east Angola Basin (terminology of Dean *et al.* 1984).

Facies model

Because there are close similarities between the numerous descriptions of calcareous and siliceous fine-grained turbidites and because many of them comprise various admixtures of these components, it seems appropriate at this stage to propose a unified facies model for biogenic turbidites irrespective of composition. We have used a modified Piper (1978) sequence for mud turbidites, including an E/F division to emphasize the extremely gradational transition from turbidite to pelagite that commonly occurs (Fig. 6).

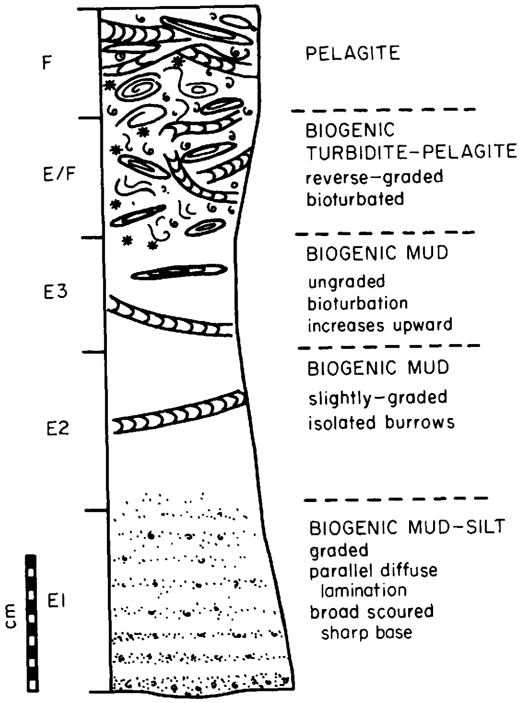

FIG. 6. Biogenic turbidite facies model. Structural divisions introduced here, modified from Piper (1978) and Stow, Wezel *et al.*, (this volume).

The detailed structural divisions of the Stow (1977) sequence have not yet been recognized in biogenic turbidites, whereas the twofold unifite-hemipelagite division of Stanley (1981) appears too simple. The complete sequence from top to bottom is:

F hemipelagic or pelagic sediment, bioturbated
E/F mixed turbidite and pelagite, reverse-graded, extensively bioturbated
E3 ungraded, very fine-grained, very homogeneous, isolated burrows
E2 slightly graded, fine-grained to very fine-grained, otherwise homogeneous, rare isolated burrows
E1 graded, laminated, laminae horizontal and indistinct or diffuse, alternation of coarser (biogenic) and finer (clay-biogenic) material, thicker basal lamina of sandy-silty biogenic debris with sharp scoured base.

Textural grading is most evident through the graded laminated division (E1), but is very slight through the E2 and absent in the E3 divisions, which are commonly very fine-grained. The background pelagic or hemipelagic sediment, comprising large unbroken forams, diatoms and other planktonic organisms, is often coarser grained than the E2-E3 turbidite divisions. The gradual upward transition from turbidite to pelagite, therefore, shows a marked reverse grading.

Where the components are of mixed types, compositional grading is the most striking. There has been little experimental work on the relative hydraulic equivalence of different biogenic particles (e.g. Berger & Piper 1972) but empirical data from descriptions of biogenic turbidites suggests that forams and shell debris alternate with terrigeneous clay-rich laminae in the lower divisions, whereas diatoms and the finest carbonate material (e.g. nannofossils) are concentrated towards the top.

This sequence of structures and associated grading can be interpreted in terms of deposition from a waning turbidity current, in the same way as for terrigenous mud turbidites. The coarsest material deposits first as a thin layer at the base. Shear-sorting together with a waning current velocity results in the graded laminated division (Piper 1972b; Stow & Bowen 1978, 1980), in which the rather diffuse lamination indicates a poor separation of silt-sized biogenics from clay-sized material. This may be, in part, a function of the floc composition, with floc strength less than in clay-rich sediment. The absence of structures and poor or absent grading in the E2-E3 divisions, the extended transitional division, E/F, and the extent of burrowing into the turbidite, all appear to suggest that there is an even less clear distinction between the processes responsible for deposition of the turbidite and pelagite facies in biogenic than in terrigenous systems (Heath & Mullins; Crevello *et al.*, both this volume). Stow, Wezel *et al.* (this volume) suggest that fine-grained 'pure' carbonate carried into a basin by a turbidity current is more likely to disperse into the water column than the equivalent terrigenous material, perhaps because it is less prone to forming strongly-bound flocs. Thus, the fine-grained low-concentration tails of carbonate-charged turbidity currents mix with surface and mid-water suspensions derived *in situ* or from the basin margins and settle slowly through the water column. The biogenic turbidite will therefore grade imperceptibly upwards into pelagite.

Whether or not a similar reasoning applies to siliceous biogenic turbidites is not certain. However, it seems less common that both the background pelagic sedimentation and the resedimenting turbiditic material are of equivalent siliceous composition, so that there is mostly a marked compositional distinction between the pelagite and turbidite facies.

Variability

Examples of variability of biogenic turbidites are illustrated in Fig. 3. Thick-bedded (commonly > 1 m), fine-grained, biogenic turbidites showing the complete structural sequence (E1-E/F) occur in two main settings. They are found as relatively proximal deposits on platform slopes and ridge flanks where the source material is entirely fine-grained (e.g. Stow 1984c). Thinner-bedded base-cut-out turbidites (E2-E/F, E3-E/F and E/F) occur more distally (e.g. Bartolini *et al.* 1975). They also occur as more distal deposits in small ponded basins (e.g. Wezel 1973; Stanley 1981) where the original turbidity current may have already deposited its coarser load.

There are many examples of thick fine-grained caps to even thicker (up to 10 m) coarse-grained biogenic debrites and turbidites. Complete gradational sequences do occur (e.g. Faugères *et al.* 1982; Gonthier *et al.* 1982; Stow 1984b,c), but more commonly the basal units are variously missing so that a thin to thick fine-grained bed overlies a much reduced biogenic sandy or gravelly layer. The contact between the two is often sharp or apparently erosive.

The other main variable besides structural sequence and bed thickness is composition. Carbonate mud turbidities may be rich in nannofossils (e.g. Kennedy 1980), resedimented periplatform ooze of mixed composition (e.g. Crevello &

Schlager 1980; Heath & Mullins, this volume), or mixed planktonics and indeterminate lime mud (Stow, Wezel *et al.*, this volume). Every gradation of mixed carbonate-clastic mud turbidites occurs (see examples in Cook & Enos 1977; McIlreath & James 1978), with the characteristic features of the facies changing accordingly. Gradations between siliceous and calcareous biogenic turbidites are also well-known (e.g. Kelts & Arthur 1981; Stow 1984c). Almost pure diatomaceous turbidites have been described from the Gulf of California (Curray *et al.* 1979), radiolarian chert turbidites from Greece (Nisbet & Price 1974), and mixed siliceous turbidites from many parts of the world (e.g. Kelts & Arthur 1981; Chough, this volume).

Disorganized turbidites

Disorganized turbidites (Fig. 7) show only poor diffuse positive grading with bioturbation, if present, concentrated towards the tops of beds. Otherwise they are structureless. They may be dominantly silt grade, clay grade or mixed and have variable admixtures of biogenic and terrigenous material. However, separation into distinct laminae of different grain size or composition has not occurred and the sorting is therefore poor. They have a sharp, often scoured base and a gradational, indistinct top, occurring in beds from 5 cm to several metres in thickness.

Disorganized fine-grained turbidites of this sort are relatively widespread, occurring both as the dominant turbidite type in an area and as isolated examples within a suite of more organized thin-bedded turbidites (Fig. 3). However, their interpretation is not always clear. In some cases they may be a part of one of the other turbidite sequences, such as the massive silts and ungraded muds described by Piper (1978). In other cases they may result from the ponding of large turbidity currents in small basins (e.g. Wezel 1973; Stanley 1981). In proximal slope or shelf-edge settings, disorganized turbidites may be deposited by turbidity currents that have not matured sufficiently during flow to allow some internal sorting of their loads (e.g. Ballance *et al.*, this volume; Hill, this volume). In more distal settings, where the associated turbidites are well-organized (e.g. van Weering & van Iperen, this volume), very rapid deposition perhaps due to a local topographic effect may cause deposition of isolated disorganized beds. Chough & Hesse (1980) interpret such thin poorly-sorted layers, intercalated with thinly laminated fine-grained turbidites on the levees of the Northwest Atlantic Mid-Ocean Channel, as resulting from spill-over of the head rather than body of a turbidity current.

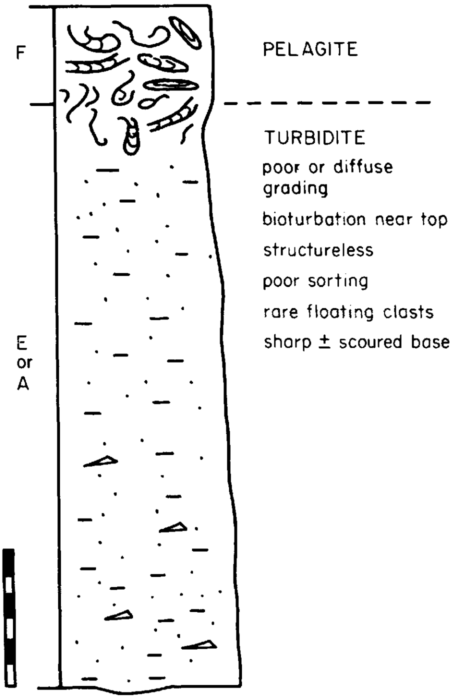

FIG. 7. Disorganized turbidite facies model (this paper).

Contourites

Contourites are a volumetrically significant facies in the present-day deep sea, where they can occur in enormous sediment drifts that are equivalent in size to some larger submarine fans, but they have proved extremely difficult to identify in the ancient record. This is perhaps because they have a low preservation potential, since almost all are developed on oceanic crust. Contourites are sediments deposited or reworked by currents that are persistent in time and space and flow mainly along-slope in relatively deep water. Deposition is controlled by bottom currents driven by thermohaline circulation or, more rarely, by wind-forced surface currents, rather than by gravity-driven turbidity currents. Johnson *et al.* (1980) and Halfman & Johnson (this volume) have extended the definition of the facies to include lacustrine contourites from Lake Superior.

The chief sedimentary criteria that can be used to distinguish contourites from other deep sea facies include:

(1) an irregular vertical arrangement of facies types and structures with both negative and

positive grading, but no regular structural sequence;

(2) evidence for more or less continuous bioturbation that has kept pace with deposition, but with the 'ghosts' of current-induced structures remaining;

(3) compositional, textural or other features that indicate a combined *in situ* and exotic origin.

For present-day examples there are a range of non-sedimentary criteria that provide important additional evidence for a contourite interpretation. These include the measurement of bottom currents and nepheloid layers, the identification of large and small-scale current-induced morphological features, and certain seismic and stratigraphic characteristics. The acoustic character of contourites using high-resolution profiling systems is dependent on grain size variation and on the presence of intermediate-scale bedforms. In seismic reflection profiles, both acoustically transparent and well-laminated contourite drifts have been described (e.g. Scrutton & Stow 1984). The nature of drift growth and the positive accumulation features (large-scale mud waves, sediment drifts) are diagnostic.

There are other bottom currents in the deep sea, such as the up and down-canyon currents measured by Shepard *et al.* (1979), and these may deposit sediments with very similar characteristics to the contourites we describe. There have been very few descriptions of sediments that can be related clearly to bottom currents of this type, so we have not considered them further in this paper.

According to Heezen *et al.* (1966) and Hollister & Heezen (1972), the thin laminated silt and fine sands on the continental rise off eastern North America are the typical contourite facies. Subsequent work on this margin has shown that the distinction between contourites and interbedded facies is not so clear cut (Stanley *et al.* 1979; Stow 1979a; Shor *et al.*, this volume). Piper & Brisco (1975) describe a different set of characteristics for silty contourites on the Antarctic margin. Based largely on evidence from coring and drilling the major sediment drifts in the North Atlantic, Stow & Lovell (1979) and Stow (1982) have characterized two main contourite facies:

(1) muddy contourites, that result from deposition by bottom currents; and

(2) sandy contourites, that result from a combination of deposition and reworking by bottom currents.

To these we might add a third, less common facies type:

(3) gravel-lag contourites, that result from erosion and winnowing of fines by powerful bottom currents (Carter & Schafer 1983).

Faugères *et al.* (1984) and Gonthier *et al.* (this volume), from a study of the Faro Drift contourites in the Gulf of Cadiz, point out that there is in fact a continuum of contourite facies from the very fine clay grade to the coarse sand and gravel grades. They recognize an additional intermediate mottled silt and mud contourite facies as well as a silty-sandy facies (rather than a strictly sandy facies).

We describe two contourite facies models in this paper, one for muddy contourites and one for silty or fine sandy contourites, and then discuss their occurrence in irregular vertical sequences.

Muddy contourites

Muddy contourites include a range of deposits at the fine end of the grain size spectrum, and appear to be the dominant contourite facies in the deep sea, making up over 75% of the main contourite drifts. They have been described from some 15 separate drifts in the North Atlantic (see reviews by Stow & Lovell 1979; Stow & Holbrook, this volume) and show a consistent set of sedimentary characteristics (Fig. 8) that are similar in many ways to hemipelagic and pelagic sediments. There have been no specific identifications of muddy contourites in ancient rocks, although several authors have interpreted ancient slope-rise sequences where they believe bottom currents to have been active (e.g. Bein & Weiler 1976; Anketell & Lovell 1976).

On first appearance the facies is monotonous, homogeneous and structureless, but closer inspection reveals important structural features. There are no very distinct beds, but both positive and negative grading between more clayey and more silty-sandy sediment is commonly present over a few centimetres to a few tens of centimetres. The contacts between clay-rich and silt-rich parts are mostly gradational, less commonly sharp and erosive. Primary lamination of either silt or clay is rare, and may be distinct and planar or indistinct and wavy. Where not laminated, the muds are thoroughly bioturbated. The silt fraction is often concentrated in irregular pockets and lenses that show some horizontal alignment, but that appear to have been broken up by bioturbation (Faugères *et al.* 1984).

Texturally, these mud contourites are silty-clays with up to about 10 or 15% sand fraction and a mean size that varies from less than 5 μm to about 40 μm. They are mostly poorly sorted. The shape of the grain size curves (Rivière 1977) varies from logarithmic with slight hyperbolic tendency for the finer-grained samples to parabolic for the more silty samples (Gonthier *et al.*, this volume).

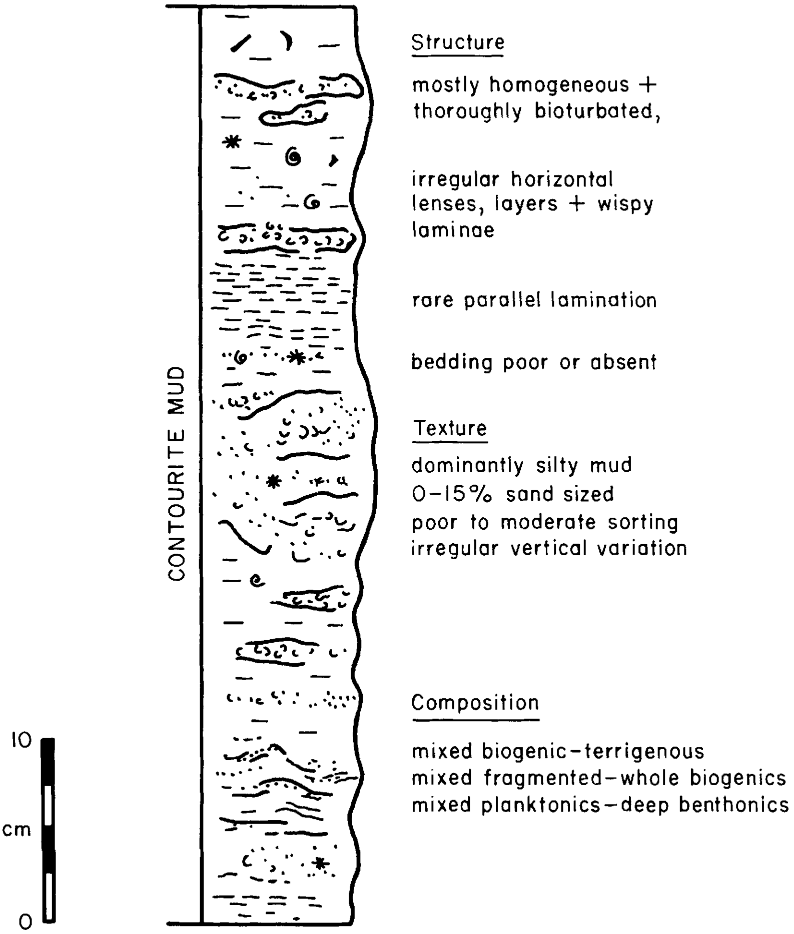

FIG. 8. Muddy contourite facies model (after Stow & Lovell 1978; Stow 1982).

In terms of composition, many muddy contourites are mixtures of biogenic and terrigenous material in variable proportions depending on the distance from land, source of supply and so on (Laughton *et al.* 1972; McCave *et al.* 1980; Stow & Holbrook, this volume). The biogenic fraction includes calcareous and siliceous planktonic and benthonic forms. Some are broken and iron-stained. The benthonics are usually *in situ* deeper-water forms rather than shallow-water derived species. The terrigenous fraction is commonly dominated by fine quartz and clays. Other contourites may have a more or less pure pelagic composition (Shor & Poore 1978; Kidd, pers. comm. 1983) and some appear to be dominated by volcanogenic material from mid-ocean ridges and seamounts (Taylor *et al.* 1975; Von Stackelberg *et al.* 1979).

Silty-sandy contourites

Silt and fine sand contourites appear to be less abundant in the present deep sea as distinct beds. Many of the muddy contourites described from contourite drifts contain intervals that are more sandy and silty, some of which might be termed sandy muddy silts. In other cases, cleaner sandy and silty beds have been documented (e.g. McCave *et al.* 1980; Faugères *et al.* 1979, 1984; Gonthier *et al.*, this volume). The origin of the sharply-defined silt laminae and cross-laminated fine sands on the continental rise off North America remains enigmatic (e.g. Hollister & Heezen 1972; Stow 1979; Shor *et al.*, this volume), and so we have not included the features of these deposits in our synthesis of a silty-sandy contourite facies model (Fig. 9). Most of the ancient

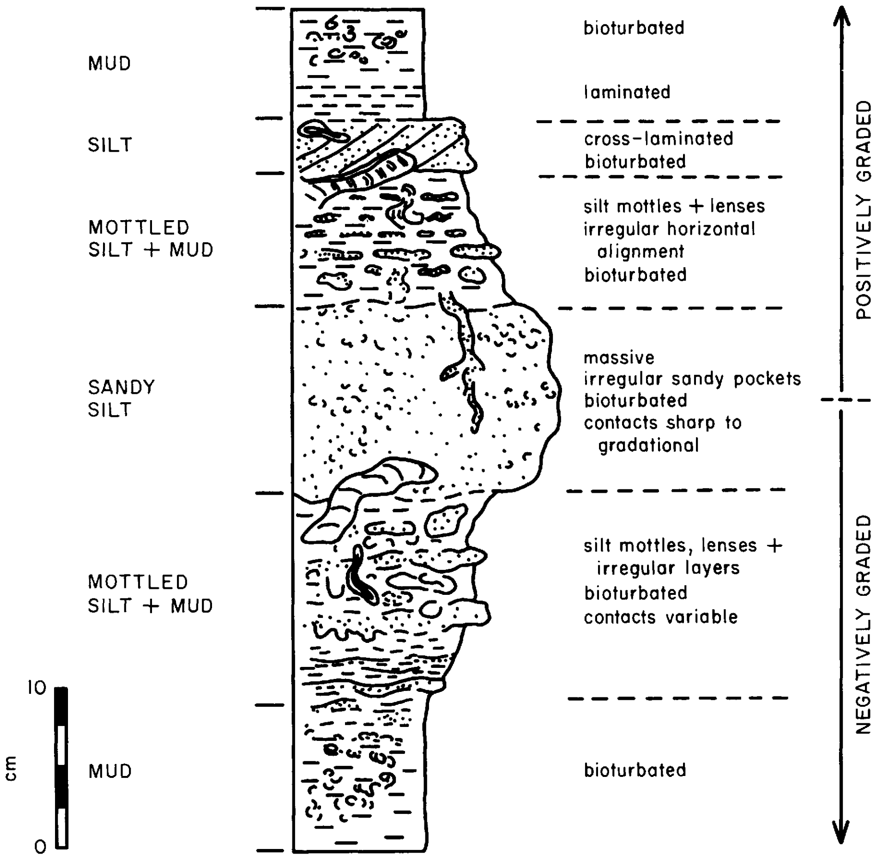

FIG. 9. Typical arrangement of contourite facies in negatively to positively graded couplets (after Gonthier *et al.*, this volume; Faugères *et al.*, 1984).

sandy contourites that have been described have been identified on the basis of the Heezen-Hollister criteria (see reviews in Stow & Lovell 1979; Lovell & Stow 1981), and may in fact not be contourites.

Silty or sandy contourites occur in irregular beds from 1 cm to about 20 cm in thickness. The top and bottom contacts of beds may be sharp and relatively flat, erosional or completely gradational. In many cases they display no primary structures apart from irregular concentrations of coarser material and slight positive or negative grading. In other cases, primary cross-lamination and parallel lamination has been preserved, but no consistent sequence of structures has been observed. Secondary bioturbation, continuous throughout the beds, is the most common feature, ranging from large distinct burrows to small irregular mottling (e.g. Chough & Hesse, in press).

Medium to coarse-grained silts with up to 40% sand and less than about 10% clay are the most common, but fine sandy contourites are also known. They are mostly moderately well-sorted but with a distinct fine tail evident on grain size curves. According to Gonthier *et al.* (this volume) the shapes of the curves tends towards parabolic or to a combination of hyperbolic (coarse tail) and parabolic (fine tail).

Compositionally, silty-sand contourites are similar to muddy contourites in comprising mixed biogenic-terrigenous material (Faugères *et al.* 1984). The larger biogenic grains are commonly fragmented and iron-stained, and there is clearly less clay material. More or less pure foraminiferal contourite sands have also been described (Faugères *et al.* 1979; Shor *et al.* 1980).

Contourite 'sequence'

There is no regular sequence of structures within contourites as there is in the various turbidite models described. However, a distinctive feature of contourite successions appears to be the presence of both negatively-graded sequences in which the grain size increases, and positively-graded sequences in which the grain size decreases upwards (Gradstein *et al.* 1982; Faugères *et al.* 1984; Gonthier *et al.*, this volume). Both sequences may be from about 10 to 100 cm thick and occur separately or as a combined negative-positive unit (Fig. 9) which shows, from top to bottom:

— homogeneous mud facies
— mottled silt and mud facies
— mottled facies with silt layers
— silt-sand facies
— mottled facies with silt layers
— mottled silt and mud facies
— homogeneous mud facies

There is considerable variability in this sequence as it is observed in different contourite successions, particularly in terms of its thickness, its completeness and its symmetry. Such variations in grain size of the facies and in the associated structural and compositional characteristics can be interpreted in several different ways. Faugères *et al.* (1984) relate the Faro Drift sequences to long-term variation in velocity of the transporting current. A complete negative-positive sequence represents a gradual increase, a maximum and then gradual decrease in the average current velocity at a given site. The time-scale for deposition of the sequence would be of the order, say of 1000 to 30 000 yrs, and the mean velocities might vary between about 5 and 25 cm/s (Gonthier *et al.*, this volume). Alternatively, such sequences might reflect variation in the grain size and/or biogenic content of material supplied to the system.

We emphasize, again, both the subtlety of the sedimentary features observed in contourites and their variability (Fig. 10a–c). On the one hand they are similar to those of some indistinct fine-grained turbidites, and on the other hand they are almost indistinguishable from hemipelagite characteristics. This fact underlines the existence of a process-continuum in which flow velocity, concentration and frequency can all vary, together with sediment supply.

Pelagites and hemipelagites

The third major facies group of the deep sea comprises the pelagic and hemipelagic sediments (Hsu & Jenkyns 1974; Jenkyns 1978, in press; Einsele & Seilacher 1982). These are widespread throughout the world's oceans and widely recognized in both modern and ancient successions. They have been deposited primarily by slow settling through the water column in the absence of any *substantial* bottom current or turbidity current activity. However, many of the processes proposed for the accumulation of hemipelagites involve current-induced slow advection of suspended sediment (Drake *et al.* 1978).

Hemipelagic sediments accumulate on continental margins and in other settings not far removed from terrigenous sediment sources. They comprise a mixture of indigenous biogenic material and silt and clay size terrigenous detritus. They accumulate slowly and are thus intensely bioturbated. True pelagic sediments accumulate in the open ocean and comprise principally skeletal parts of plankton with some admixture of very fine silt and clay, much of which has reached the open ocean by aeolian transport. The proportion of terrigenous material may be increased by dissolution of the biogenic components. Bioturbation is ubiquitous except in anoxic basins.

Textural and compositional criteria can be used to distinguish four types of hemipelagite/pelagite facies (modified after Berger 1974):

(1) pelagic ooze, with $>75\%$ biogenics;
(2) muddy pelagic ooze ('arl'), with 25–75% biogenics and a terrigenous component predominantly of clay;
(3) pelagic clay, with $<25\%$ biogenics and $>60\%$ clay in the terrigenous fraction; and
(4) hemipelagites, with $>5\%$ biogenics and a terrigenous component with $>40\%$ silt.

Other minor facies include the purely chemogenic sediments that are composed almost entirely of authigenic minerals, such as ferromanganese nodules and phosphorites. These are not considered further here.

The chief distinguishing features of pelagites and hemipelagites include:

(1) evidence for low or very low rates of sedimentation and continuous bioturbation (except in anoxic basins);
(2) no primary sedimentary structures or other evidence of current-controlled deposition;
(3) a mainly uniform composition within any one succession, that may show a regular cyclicity related to climatic or other controls;
(4) a variable biogenic component mainly of planktonic tests, a very fine grained, often far-travelled, terrigenous component, and commonly a significant authigenic component.

Seismic reflection profiling shows a more or less uniform, acoustically transparent sediment drape over bottom irregularities although there may be a thickening in basins (Moore 1969). Stratigraphically, pelagites and hemipelagites commonly show continuous deposition, although accumulation rates may be as low as one metre per million years and hiatuses may be present.

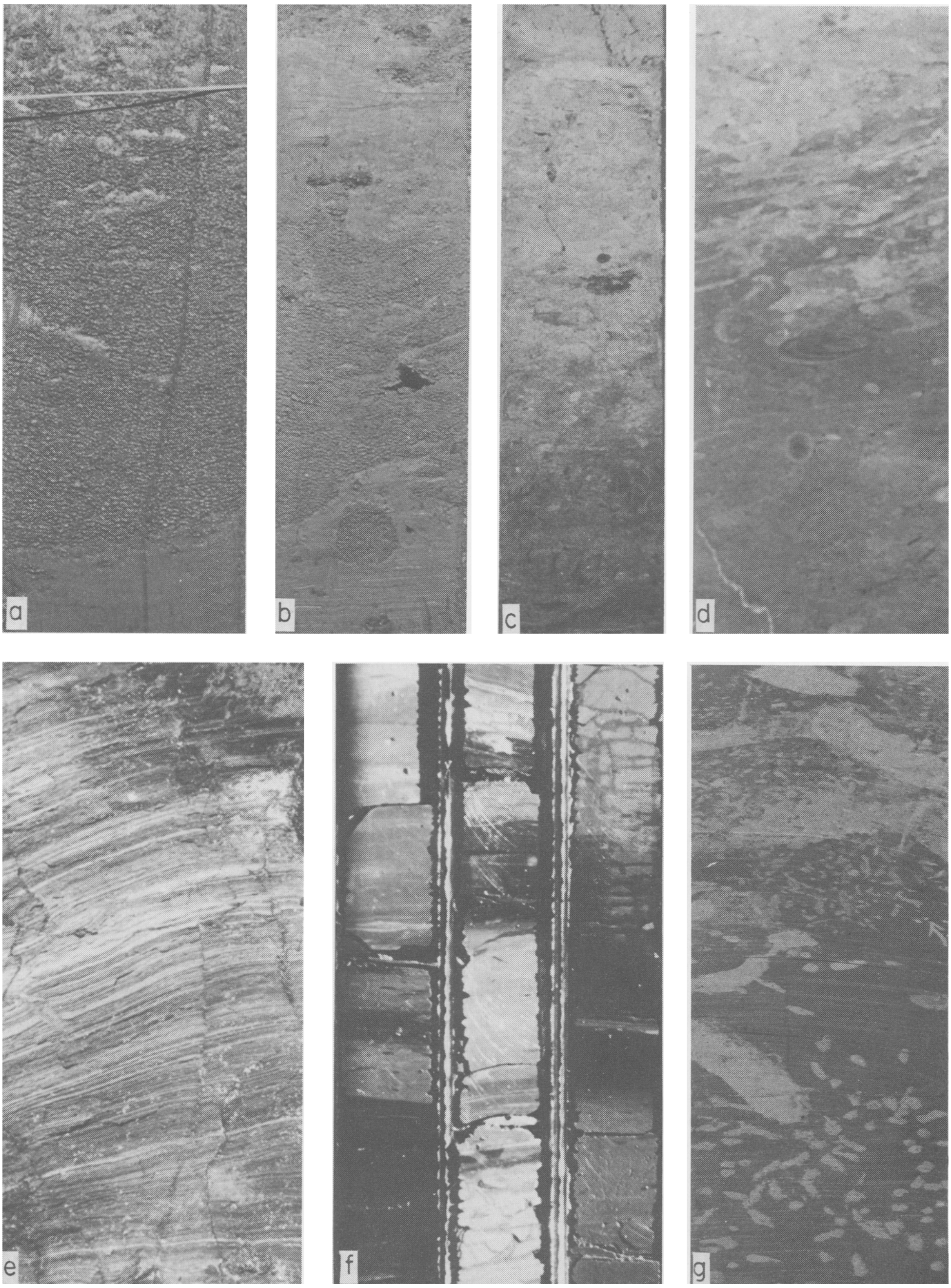

FIG. 10. Photographs of contourite, pelagite and hemipelagite facies. Section widths about 7cm. (a) Silty contourite, Pleistocene, Faro Drift, Gulf of Cadiz. (b) Mottled silt and mud contourite, Pleistocene, Faro Drift, Gulf of Cadiz. (c) Muddy contourite, Plio-Pleistocene, Blake–Bahama Outer Ridge, western North Atlantic. (d) Bioturbated pelagite–hemipelagite, upper Cretaceous, Scaglia Rossa Formation, Italy. (e) Organic-carbon-rich siliceous pelagite (diatomite), Miocene, Monterey Formation, California. (f) Alternating black shade and marlstone, mainly hemipelagic (part turbiditic), Cretaceous, DSDP Site 530A, Angola Basin. (g) Bioturbated siliceous/calcareous pelagite, Plio-Pleistocene, DSDP Site 532, Walvis Ridge.

Pelagic ooze

Pelagic oozes are most typical of the open ocean basins far from a terrigenous source. They have been the subject of much study over the past 125 years and hence their sedimentary characteristics are well known. Some important syntheses are to be found in the volumes by Hsu & Jenkyns (1974) and Cook & Enos (1977), and in papers by Arrhenius (1963) and Jenkyns (1978 and in press). Specific examples in this volume where pelagic sediments are described as part of the succession, include those of the western and central North Atlantic (Robertson), the north-western Mediterranean (Monaco & Mear), the Miocene of southern Turkey (Hayward) and the Cretaceous–Tertiary of the Apennines (Stow, Wezel *et al.*).

Facies model (Fig. 11)

One of the chief characteristics of pelagic oozes is their very slow rate of accumulation, commonly from less than 1 mm to 10 mm/1000 yrs, although this can be an order of magnitude higher under zones of upwelling. They are usually, therefore, thoroughly homogenized by bioturbation, and without any primary current-induced structures. A variety of burrow types may be preserved with different assemblages or ichnofacies dependent on different environmental factors, such as water depth, grain size, sedimentation rate and redox conditions (Seilacher 1967; Werner & Wetzel 1982; Wetzel, this volume). Some of the main diagnostic trace fossils are *Zoophycos, Chondrites, Planolites, Scolicia, Trichichnus, Teichichnus* and *Lophoctenium*. Several tiers of trace fossil assemblages are commonly superimposed on one another (Werner & Wetzel 1982).

The grain size of pelagic oozes is largely dependent on the composition of the biogenic fraction. Coccolith plates are very small (clay size), whereas some foram-rich or diatom-rich oozes may have a mean grain size that is in the silt range. The terrigenous component is mainly clay-sized. Full grain size analyses, however, are rarely carried out on pelagic oozes as the hydraulic equivalence of biogenic particles is not well-known and hence interpretation of the grain size distribution would be difficult.

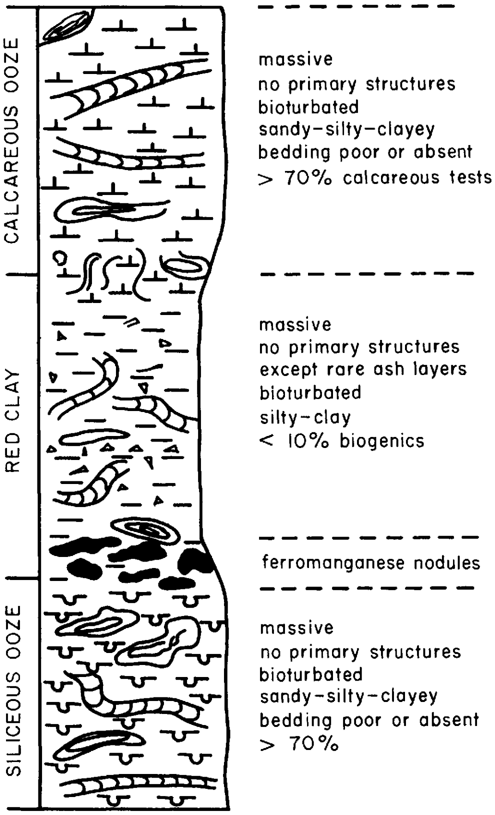

FIG. 11. Pelagite facies models. Schematically interbedded calcareous ooze, red clay and siliceous ooze, with ferromanganese nodules developed at ooze-clay contact.

Pelagic oozes are composed dominantly (>75%) of the tests of planktonic organisms, either calcareous (coccoliths, forams, pteropods) or siliceous (radiolarians, diatoms, silicoflagellates) or a mixture of both (Berger 1974). The other components (Lisitzin 1972) can include very fine-grained terrigenous material (principally quartz, feldspars and clays), volcanogenic debris (palagonite and derived clay minerals), authigenic minerals (such as phosphate, barite, zeolites, ferromanganese nodules and coatings), and rare extraterrestrial material. Under normally oxic conditions the organic-carbon content is extremely low, but under anoxic conditions pelagic black shales can contain over 20% organic carbon (Isaacs 1981; this volume; Arthur *et al.*, this volume).

The characteristics, composition and distribution of the different types of pelagic ooze are dependent on a number of interacting variables which we will not discuss at length (but see Berger 1970; Lisitzin 1972; Broecker 1974). These include: the water depth and the corresponding carbonate compensation depth; the source and supply of terrigenous and volcanogenic material; surface water productivity and the supply of biogenic material; surface currents and bottom circulation patterns; climate and basin physiography; and physiochemical conditions. Pelagic sediments are perhaps more affected by this range of different controls than are turbidites or contourites because of their two principal attributes: (1) they are composed mainly of biogenic $CaCO_3$ and/or SiO_2 both of which are soluble in sea water; and (2) they settle relatively slowly through the water column and are buried only slowly, and hence are exposed to external factors for a relatively long time period, either in the water column and/or on the sea-floor. The actual processes of settling as single grains or as larger flocs and pellets are discussed at more length in papers by Gorsline, McCave, and Eittreim (all this volume).

Muddy pelagic ooze

There is a continuum of facies from pelagic ooze with > 75% biogenic material to pelagic clay with < 25% biogenics. Muddy pelagic ooze (or 'arl', terminology of Dean *et al.* 1984) is a relatively common intermediate sediment type, with characteristics intermediate between an ooze and a clay. It differs from true hemipelagic sediment in having a dominantly clay-sized rather than silt-sized terrigenous component, and in being an open-ocean rather than continental margin facies.

Pelagic clays

Pelagic clays, also known as red clays, brown clays and abyssal clays, accumulate in the deepest, and most remote parts of the ocean basins. They are particularly well represented in parts of the Pacific Ocean, resulting from dissolution of biogenics and aeolian transport of dust. Their sedimentary characteristics are relatively well known (e.g. Arrhenius 1963; Griffin *et al.* 1968; Hsu & Jenkyns 1974; Jenkyns 1978; Hoffert 1980). Ancient examples have also been documented (e.g. Audley-Charles 1965).

Facies model (Fig. 11)

Pelagic clays have one of the lowest sedimentation rates of all the pelagic/hemipelagic sediments, commonly less than 1 mm/1000 yrs but ranging up to about 7.5 mm/1000 yrs. The rate of growth of ferromanganese nodules and crusts, however, is one or two orders of magnitude lower. As with pelagic oozes, pelagic clays are usually well-oxygenated and thoroughly bioturbated. The trace fossil assemblages are those adapted to the deepest water, and finest grain size.

Texturally, they are very fine-grained, clay or fine-silt sized, and poorly to moderately well sorted with an even distribution of grain sizes over a small size range.

Clay minerals are the dominant components, whereas quartz, feldspar and other terrigenous materials are very minor (Arrhenius 1963). Authigenic components include the zeolites, ferromanganese minerals (goethite, micronodules) together with some clays and feldspars. Biogenic material is often very scarce, but a complete gradation exists with muddy pelagic oozes (> 25% biogenics). Volcanogenic material, essentially palagonite, is present in very variable amounts. In certain environments especially close to mid-ocean ridges or immediately overlying ocean floor basalts, pelagic clays can be highly enriched in a variety of metals and trace elements.

Hoffert (1980) has identified four types of pelagic clay based on slight differences in their mineralogical and chemical compositions. These are associated with (1) siliceous oozes, (2) calcareous oozes, (3) volcanogenic material, and (4) none of the above, but with a mixed composition. It is by alteration and dissolution of the different biogenic or volcanogenic components of these other oceanic sediments, together with authigenesis and diagenesis of new minerals, that pelagic clays are formed in a process somewhat analogous to pedogenesis on land.

Hemipelagites

Hemipelagic sediments are more typical of marginal oceanic settings where there is a ready supply of terrigenous material. In published descriptions of sediment sequences they are frequently referred to as the background, normal, ubiquitous or interbedded facies, and are often not described in any great detail although they may comprise the greater part of a given sequence. At high latitudes, for example in the Arctic Ocean and Baffin Bay, ice-rafting is a major contributor to fine-grained hemipelagic sediments.

Amongst the large volume of literature that simply refer to hemipelagic sediments in this vein there are, however, some authors who have documented the facies characteristics in rather more detail. Particularly useful descriptions of modern hemipelagites include those of Rupke (1975) and Stanley & Maldonado (1979) from the western and eastern Mediterranean respectively, Stanley *et al.* (1972) and Hill (1981) from Nova Scotian margin, Moore (1974) and Kolla *et al.* (1980) from the Indian Ocean, and various contributions by Gorsline and colleagues (e.g. Gorsline 1978, 1981) from the basins of the California Borderland. In this volume there are a number of papers dealing, in part, with hemipelagic facies in the modern deep sea (e.g. Monaco & Mear, Chough, Krissek, Thornton, Gorsline *et* *al.*, Faugères *et al.*, Isaacs, and Auffret *et al.*; all this volume).

Although more difficult to identify in ancient rocks there have been a number of papers describing hemipelagites from inferred slope and basinal settings (e.g. Hesse 1975; Piper *et al.* 1976; Ingersoll 1978; Hicks 1981; Pickering 1982b). Papers in this volume by Ballance *et al.*, Bourrouilh & Gorsline, and Pickering also describe ancient hemipelagites.

Facies model (Fig. 12)

We are therefore able to summarize the chief sedimentary characteristics of hemipelagites (Fig. 12). They are commonly homogeneous and structureless, with bedding poorly defined or absent except when it has been accentuated by burial and diagenesis (see below). There are no primary current-induced sedimentary structures such as lamination, ripples or erosional contacts, although a depositional lamination may be preserved under anoxic conditions (e.g. Isaacs, this volume; Thornton, this volume).

Under normal oxic conditions, however, bioturbation is ubiquitous and thorough, often resulting in a completely homogenized sediment with a mottled aspect. Burrow traces are also often preserved, with the same major trace fossil assemblages represented as for the pelagic oozes described above. These similarly depend on water depth, grain size, sedimentation rate and redox state (Werner & Wetzel 1982; Wetzel, this volume). Iron sulphide filaments (Mycelia) and mottles are also a common feature of hemipelagic sediments.

Texturally, hemipelagic sediments are silty clays with 1 to 15% of sand of mainly biogenic origin. They are poorly sorted and show no systematic grading apart from that associated with the compositional cyclicity discussed below. The shape of the grain size cumulative curves (Rivière 1977) appears almost uniform or logarithmic, although there may be irregularities due to a typical biogenic input or rare exotic terrigenous material (e.g. ice-rafted debris).

Apart from the mixed biogenic-terrigenous aspect of hemipelagites, it is difficult to generalize

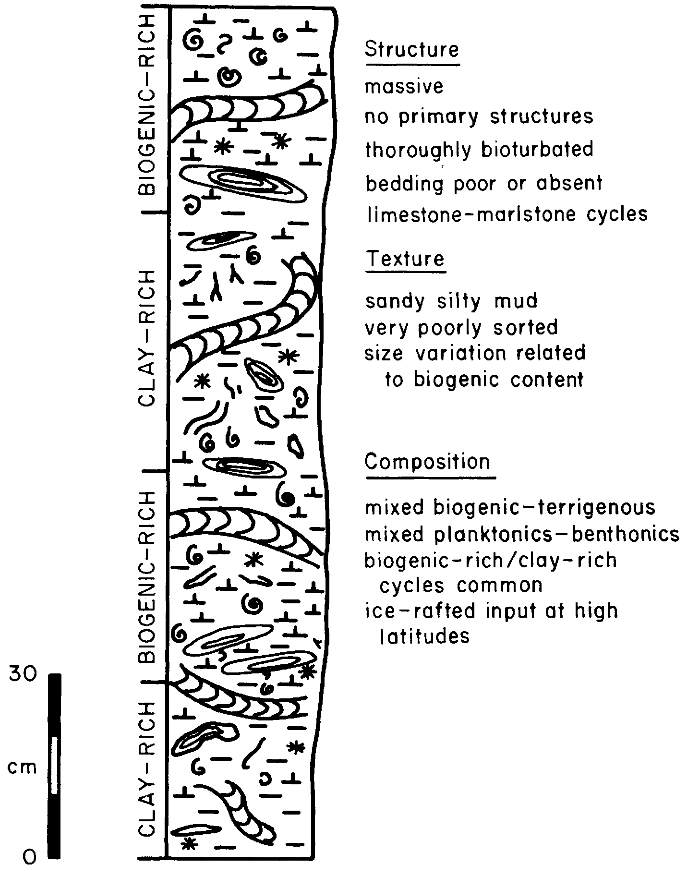

FIG. 12. Hemipelagite facies model. Rhythmic alternation of biogenic-rich and clay-rich intervals.

further about their composition. The biogenic input may be slight or dominant, calcareous or siliceous depending on surface productivity, carbonate compensation depth, etc. It is commonly of mixed planktonic and *in situ* (deep-water) benthonic origin. The terrigenous components are mostly uniform in any given area and show very gradual compositional trends towards the source area. Their nature can, of course, be highly varied depending on tectonic, climatic and other factors, and the sediments may also contain a minor to significant fraction of far-travelled (wind-blown, ice rafted, etc.) terrigenous debris.

Seismic-reflection profiles of hemipelagic sediments frequently show a draped morphology over bathymetric irregularities suggesting that they were deposited by vertical settling through the water column. Observations of oceanographic processes and of regional patterns of sediment distribution, however, suggest that other rather more complex processes may also have an effect on hemipelagic deposition (e.g. Moore 1969; Damuth & Kumar 1975; Drake *et al.* 1978; Stanley & Kelling 1978; Karl *et al.* 1983; Hill & Bowen 1983). In particular, slow lateral displacement of fine suspended sediments in mid-water and bottom water nepheloid layers; sediment dispersion by up and down-canyon normal currents; resuspension and slow diffusion at the shelf break; and sediment creep over gentle slopes all appear to be important processes. These are discussed at more length in papers by Gorsline, McCave, Eittreim, Thornton and Gorsline *et al.* (all this volume).

Hemipelagite–pelagite cycles

Examples of some hemipelagic and pelagic sediments are shown in Fig. 10d–g. An important feature of many, but not all, hemipelagic and pelagic successions is a cyclic alternation of biogenic-rich (pelagite) and clay-rich (hemipelagite or pelagic clay) beds (Fig. 12) (e.g. Dean *et al.* 1981; Einsele 1982; Einsele & Seilacher 1982; Stow, Wezel *et al.*, this volume). Most of the cyclic sequences that have been described from both modern and ancient successions are limestone–marl cycles with the primary variation being the relative proportions of carbonate and clay, although similar cycles are also known from biogenic siliceous sequences. Other compositional (e.g. organic carbon, trace elements) and textural (grain size) characteristics may mirror the limestone–marl cycle or may vary independently. There is a close correspondence worldwide in the order of magnitude of both Quaternary–Recent and older cycles. They are commonly between 10 and 100 cm thick, with a sedimentation rate of 0.5 to 3 cm/1000 yrs and a periodicity of 20 000 to 100 000 yrs.

Possible causes for cyclic variation in the relative abundance of $CaCO_3$ or biogenic silica include: (1) variation in the rate of production of $CaCO_3$ or SiO_2 by planktonic organisms; (2) variation in the rate of input of $CaCO_3$ or SiO_2 via turbidity currents; (3) variation in the amount of dissolution of $CaCO_3$ or SiO_2 either shortly after sedimentation or at depth after burial; and (4) dilution of the biogenic component by variation in the non-carbonate (terrigenous) input. All of these processes have been shown to be important in different cases. The close relationship of the periodicity to the earth's orbital changes and climatic variation suggest that climate is most often the ultimate control, with sea-level changes and fluctuation in the carbonate compensation depth also being important (e.g. Fischer & Arthur 1977).

There are rather fewer examples of similar cycles in siliceous biogenic sediments (e.g. Garrison & Fischer 1969; Barrett 1982). Some of these probably have a similar origin to the limestone–marl cycles, but others were more clearly controlled by turbidity current input of the siliceous material (e.g. Nisbet & Price 1974; Folk & McBride 1978).

Environments and facies distribution

The various fine-grained facies outlined above are not distributed randomly through the deep sea but occur preferentially in certain environments. They are commonly closely associated and interbedded with each other and with the range of coarser grained facies also found in deep water. Based largely on their occurrence at the present day, we outline here some aspects both of their generalized distribution on a global scale and of their more detailed horizontal distribution within slope, fan and basin plain settings. We then describe their occurrence in characteristic vertical sequences in the ancient record. Our intention here is to highlight only the main aspects of facies distribution and not to attempt rigorous documentation. We believe that more detailed consideration will be a fruitful area for further research.

Global distribution

The principal environments of turbidite deposition (both coarse-grained and fine-grained) are the very large areas of continental slopes and rises and oceanic abyssal plains (Fig. 13). Also impor-

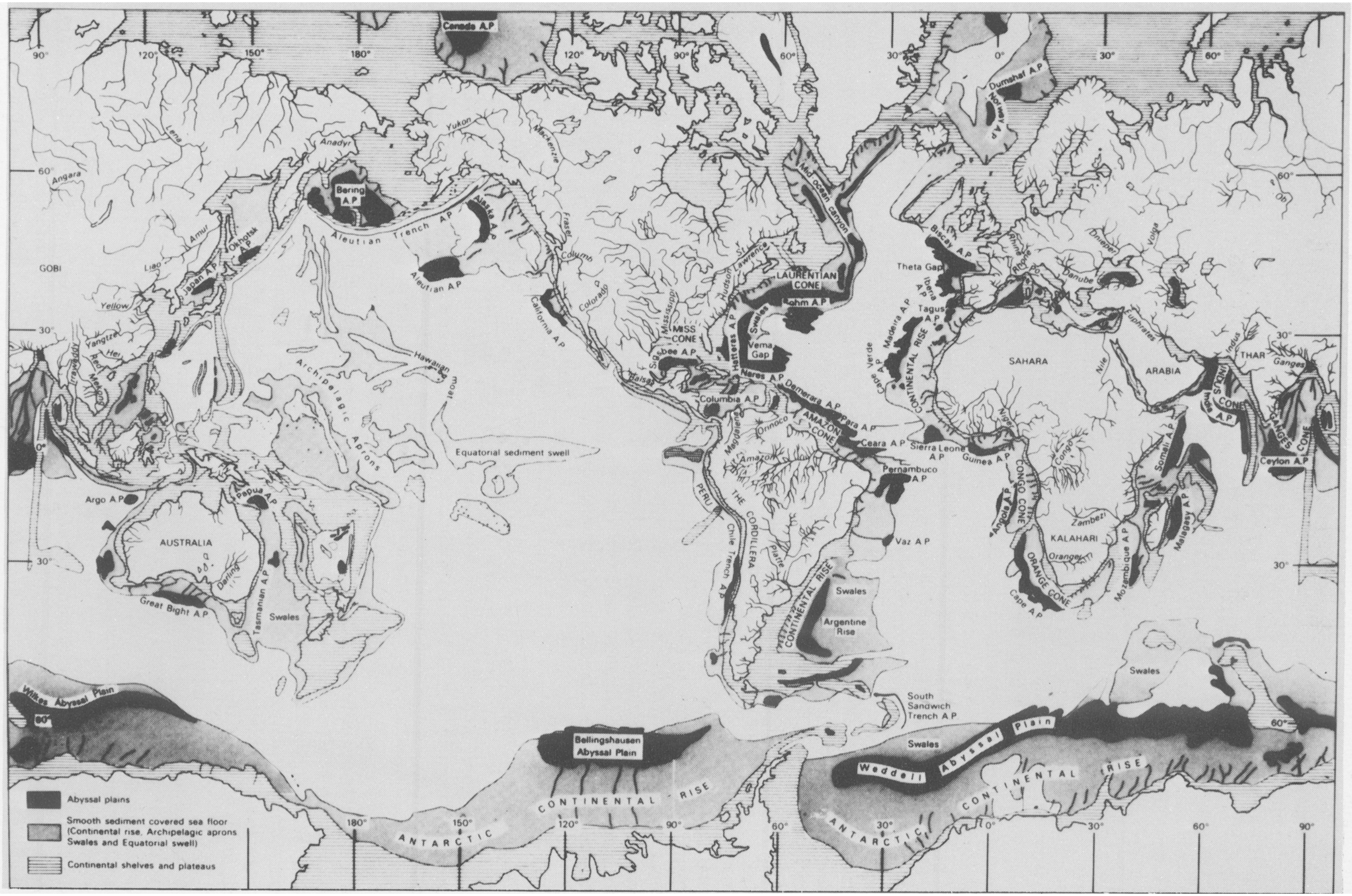

FIG. 13. World map showing principal areas of major turbidite sedimentation: continental slopes and rises, submarine fans and abyssal plains. Main terrestrial drainage pattern also indicated. (From Heezen & Hollister 1971).

tant are the slopes and floors of smaller marginal seas and land-locked basins. Turbidites are dominant, therefore, around the margins of the continents and particularly those trailing-edge margins that receive most sediment from continental drainage systems and glacial erosion, including much of the Atlantic Ocean, north Indian Ocean, Mediterranean Sea and the circum-Antarctic margin (Inman & Nordstrom 1971). Large amounts of sediment draining from south-east Asia is mostly trapped in back-arc basins or on broad continental shelves of the western Pacific. There is generally less sediment supplied to the coasts along collision margins, although the tectonically unstable and metastable slopes in these areas are important sites of resedimentation via slumping, debris flows and muddy turbidity currents (e.g. Gorsline *et al.*, this volume).

Silt turbidites are associated with channels and other conduits across the slope, and are commonly fed to the channel distributary regions on the lower slope (e.g. mid and lower fan lobes) and to more proximal parts of abyssal plains (Piper 1978). Mud turbidites are more widespread, occurring in levee and interchannel areas of the slope and throughout the basin plain. Biogenic turbidites are restricted to regions off carbonate banks and reefs, upwelling zones or seamounts and oceanic ridges, where a biogenic source is available. The distribution of disorganized turbidites is less clearly understood; they occur both in proximal and distal regions of turbidite sedimentation.

The chief environments of contourite deposition are closely related to the deep-water thermohaline circulation pattern of the ocean basins and, in particular, to the higher velocity bottom currents that occur principally on the western margins of basins or by the acceleration of flow through restricted passageways. In the North Atlantic, for example (Fig. 14), a number of large sediment drifts can be identified that are made up almost entirely of muddy and silty contourites, and of pelagic or muddy pelagic oozes that have been moulded by bottom currents during deposition. These occur as isolated elongate mounds both in the middle of the ocean and parallel to or projecting from the continental rise, and as irregular dome-shaped mounds near seamounts and other topographic highs. Coarser-grained sand and gravel lag contourites occur in association with muddy contourites in the Straits of Gibraltar and in the western Labrador Sea. On the continental rise of eastern North America contourites are closely interbedded with fine-grained turbidite and hemipelagite facies.

Pelagic and hemipelagic sediments are interbedded to varying degrees with both turbidite and contourite facies wherever they occur. Elsewhere in the deep sea they are the principal facies. Calcareous pelagic oozes are dominant in the Atlantic Ocean and over the shallower mid-ocean ridge parts of the other oceans. Siliceous pelagic oozes occur mainly in two high latitude circum-polar bands, a Pacific equatorial band and under upwelling zones on the western boundaries of the continents. Pelagic clays are confined to the very deepest and central parts of the oceans, in particular the Pacific. Hemipelagites are most abundant near the continents, except where masked by a major input of resedimented facies. At very high latitudes, there is an ice-rafted component within the other facies which, if dominant, gives an ice-rafted hemipelagite facies.

Slope aprons, submarine fans and basin plains: horizontal distribution

Slope aprons are morphologically heterogeneous and have a correspondingly complex and irregular distribution of facies (Fig. 15) (Bouma *et al.* 1978; Doyle & Pilkey 1979; Stow 1984a). On normal terrigenous-supplied slopes there is commonly a mud-line (Stanley & Wear 1978) that separates the shallow, higher-energy, sandy shelf facies from the mainly fine-grained slope sediments. Some sand spillover occur along the shelf break as well as the funnelling of coarser sediments down canyons or gullies to isolated depositional lobes. Areas of sediment creep and slumping give rise to resedimented slump and debrite masses and, in some cases, to mud turbidites. On the open slope there is an interbedding of and gradation between fine-grained turbidites, contourites, hemipelagites and pelagites, with the turbidites being more common near channels, the contourites occurring in areas of bottom current activity, and the pelagites increasing in abundance distally.

Very rapid progradation occurs off deltas, particularly at times of lowered sea-level. Fluctuations in river discharge may produce graded beds resembling turbidites (see discussion in Stow, Alam & Piper, this volume). Sediment facies include both turbidites and hemipelagites accumulated from suspension fall-out. Both may be little bioturbated because of high rates of deposition. Similar facies occur in high latitudes off active ice margins.

At the base of slopes there may be either a smooth gradual transition to the basin plain facies or areas of more positive construction. Isolated channel-fed lobes comprise silt and fine sand turbidites (Piper 1978), slump/debris-flow-fed lobes comprise mud turbidites (Stow 1984c),

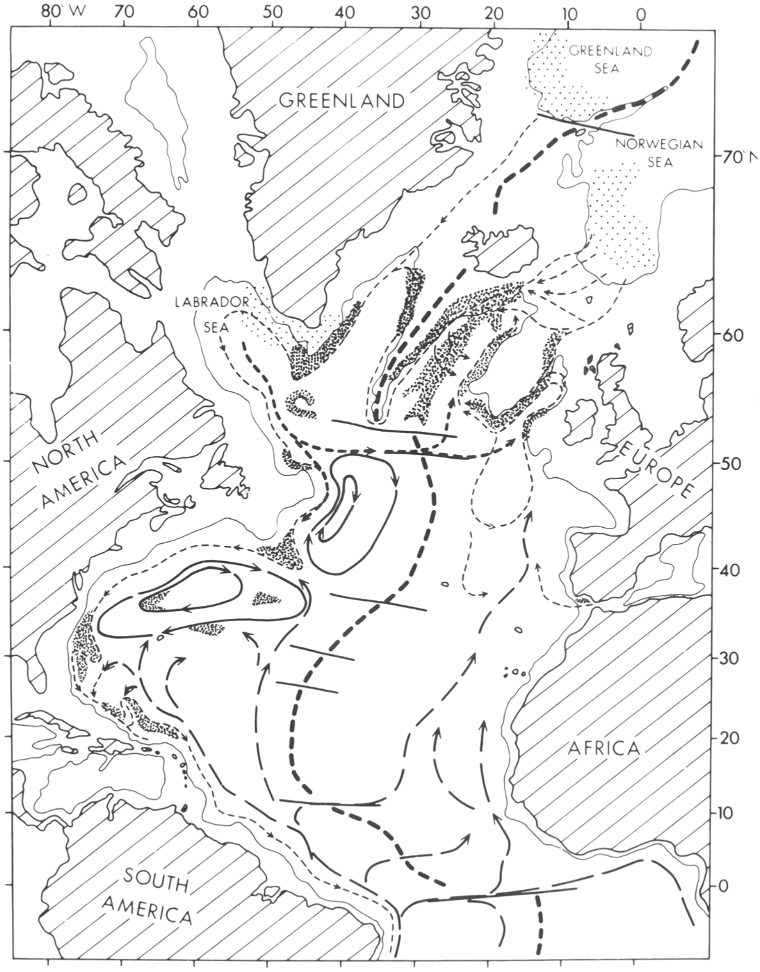

FIG. 14. Present-day deep-water circulation in the North Atlantic. Areas of bottom-water formation shown in wide stipple. Main contourite drifts shown in close stipple. Contour at 2000 m; mid-ocean ridge in heavy dashes. (After Stow 1982).

and contourite mounds or drifts may be constructed parallel to the slope contours. Along-slope trends of sediment characteristics are observed in these drift deposits (Gonthier *et al.*, this volume). Other papers in this volume that discuss facies distribution on modern clastic slopes include those by Hill, McGregor *et al.*, Ballance *et al.*, Krissek, Got, and Gorsline *et al* (all this volume).

Variations on these facies distributions occur on slopes that are controlled by active faulting or diapirism. Resedimented and pelagic biogenic facies are dominant on carbonate slopes (e.g. McIlreath & James 1978; Mullins & Neumann 1979; Heath & Mullins, this volume; Faugères *et al.*, this volume) and on the flanks of oceanic ridges and seamounts (e.g. Gonthier *et al.* 1982).

The distribution of facies on *submarine fans* has received rather more attention in the past than that on slopes (e.g. Mutti & Ricci Lucchi 1972; Walker & Mutti 1973; Walker 1978). Several important papers have addressed specifically the question of distribution of the fine-grained facies (Piper 1978; Nelsen *et al.* 1978; Stow 1981; Cremer 1981, 1983) so that a relatively clear picture has emerged (Fig. 16). Van Weering & van Iperen (this volume) and Hayward (this volume) also deal specifically with fine-grained facies on deep-sea fans.

There is both an elongate and concentric distribution of coarse to fine-grained facies on most fans. Slumps and slides are mostly confined to the slope, upper fan and channel margins. Debrites may be more widespread (Nardin *et al.*

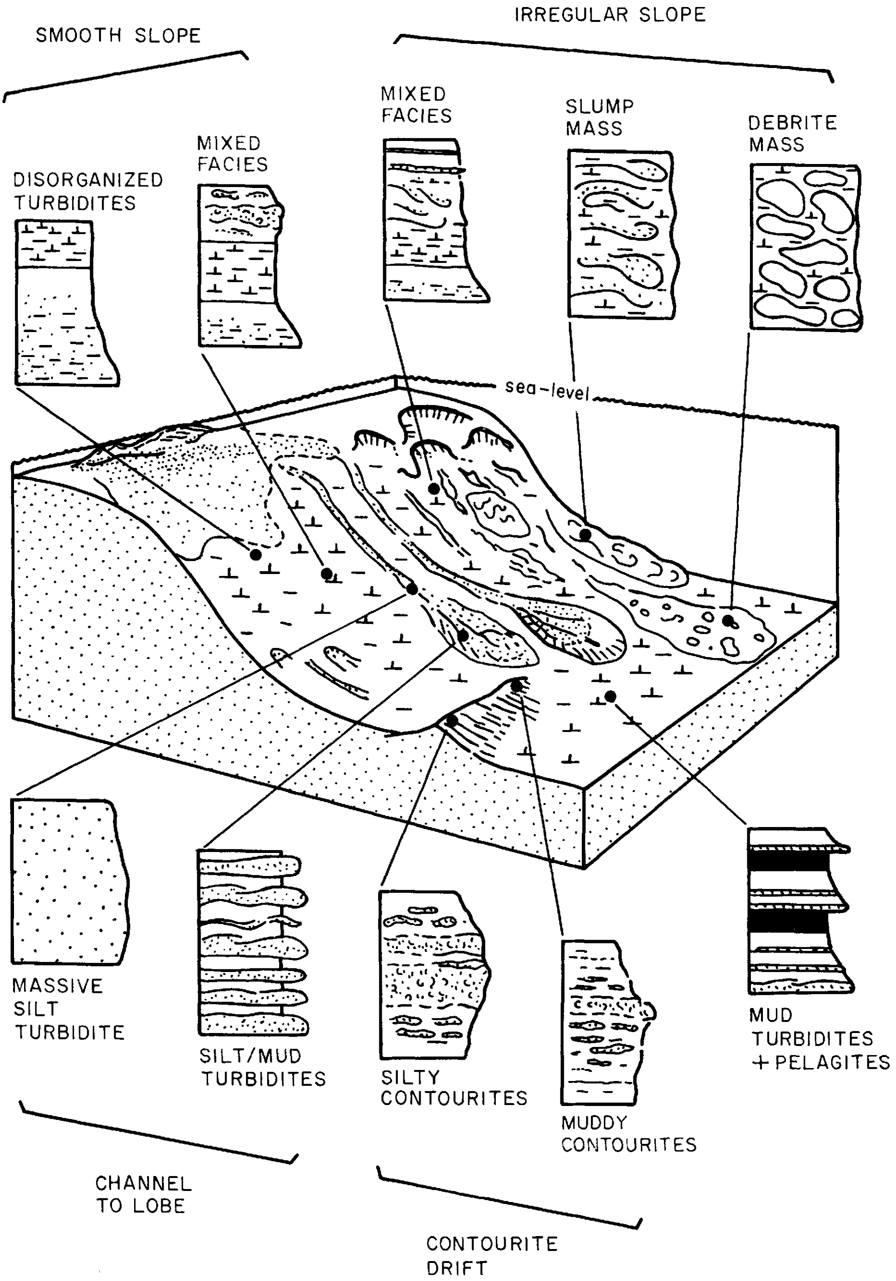

FIG. 15. Schematic distribution of fine-grained sediment facies across a typical slope and rise. Note irregular distribution of mixed facies types.

1979). Coarse-grained turbidites are transported through the channel system and deposited along their length and on the sandy lobes that spread out at their terminations. Fine-grained turbidites are transported either down channels and then laterally by overflow onto the levees and interchannel areas, or as thick unconfined low-density flows (Bowen *et al.* 1984). They commonly show both a down-fan and away-from-channel evolution of textural, structural and compositional features. Hemipelagites, pelagites and, in some cases, contourites are interbedded with the resedimented facies in areas of lower energy.

Basin plains are the ultimate trap for deep-sea

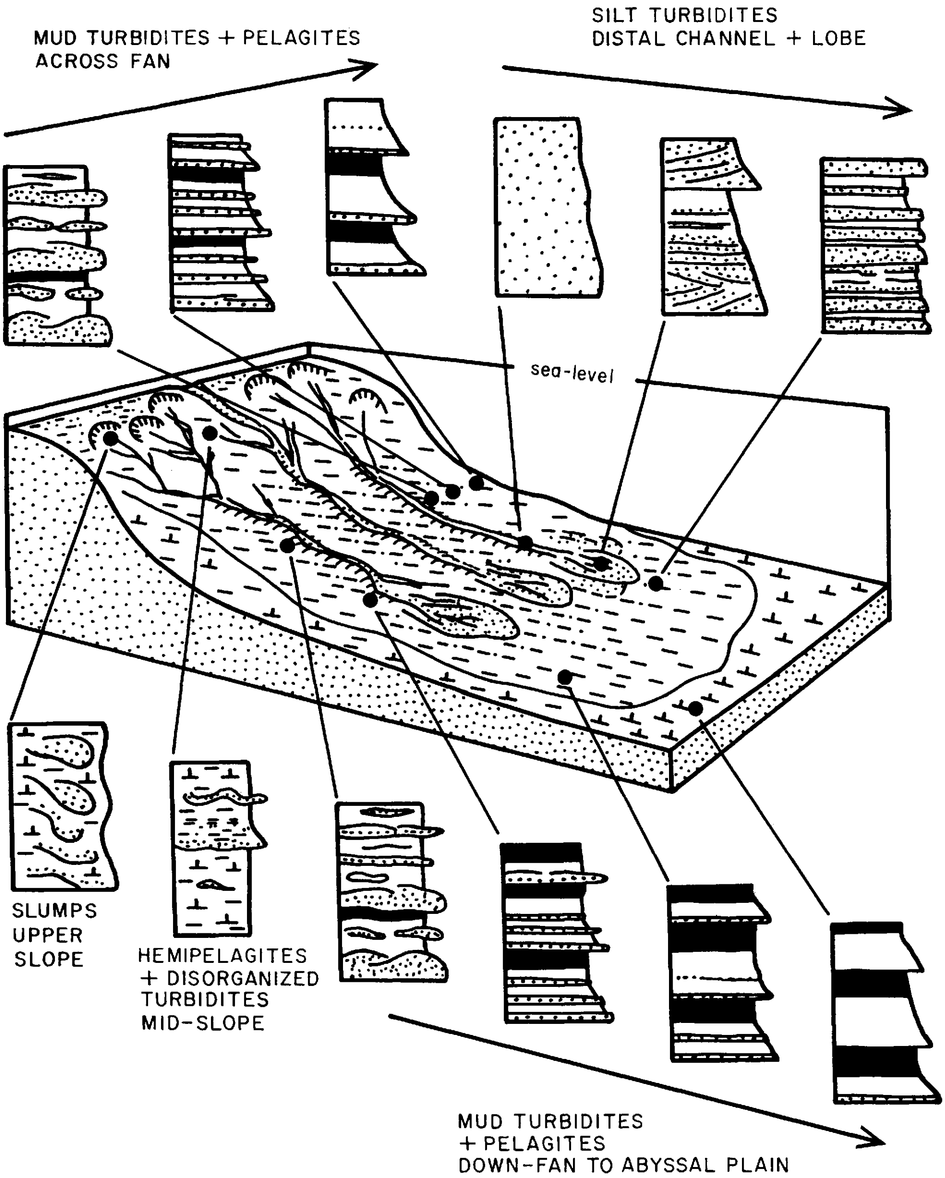

FIG. 16. Schematic distribution of fine-grained sediment facies across a typical large deep-sea fan. Note systematic down-fan and across-fan distribution of mainly turbiditic facies.

sediments that have been eroded and transported by currents or that have settled slowly through the water column (Gorsline 1978; Pilkey *et al.* 1980; Stow, 1984a). They vary widely in their areal extent, depth and shape, including for example, the tiny plains of slope basins and the major oceanic abyssal plains. A single basin plain may be fed from several sources, including channels and fans, the surrounding slopes and surface waters. All sediment facies types are represented in basin plains (Fig. 17), depending very much on morphological, tectonic, sedimentary and sea-level controls, and the fine-grained facies commonly predominate. Both centripetal and longitudinal facies distributions are observed; turbidites, contourites and hemipelagites/pelagites are often intimately interbedded. Several papers in this volume outline specific examples of basinal sedimentation, including those by Auffret *et al.*, Got, Chough, Thornton, Gorsline *et al.*, Robertson and Crevello *et al.* (all this volume).

Where a basin is fully enclosed and circulation

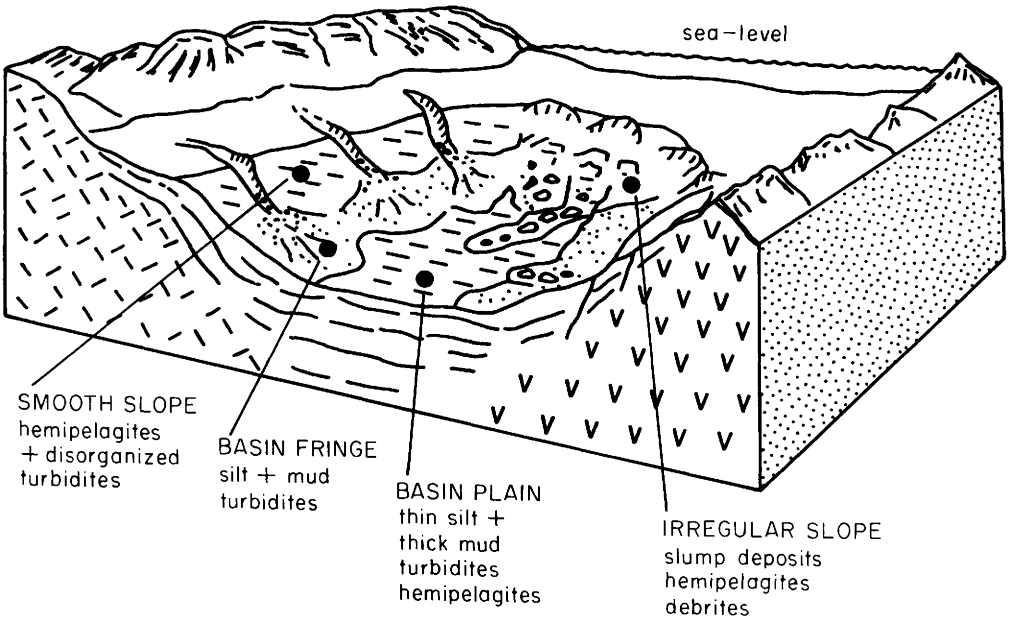

FIG. 17. Schematic distribution of fine-grained sediment facies within a relatively small enclosed marginal basin. Note irregular to concentric distribution of mixed facies types.

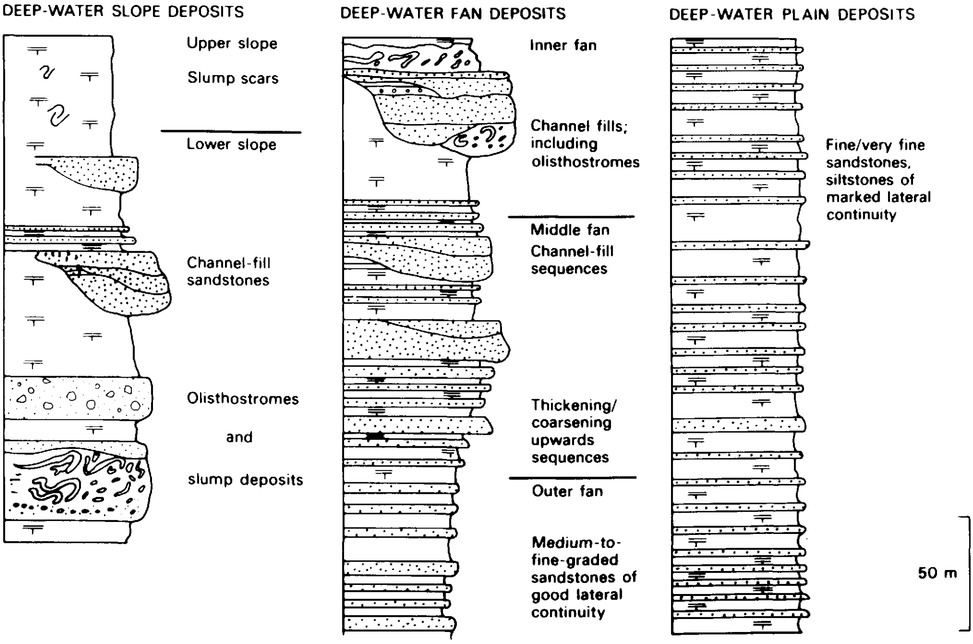

FIG. 18. Typical facies associations for coarse-grained and fine-grained sediments in slope, fan and basin plain environments. (After Mutti & Ricci Lucchi 1972).

restricted, periodic anoxic bottom conditions can lead to the accumulation of organic-rich sediments (black shale facies) (e.g. Arthur *et al.*, this volume).

Facies associations & vertical sequences

In ancient record it is often possible to recognize slope, fan and basin plain palaeoenvironments on the basis of the vertical and/or lateral association of the characteristic facies outlined above. Thus, Mutti & Ricci Lucchi (1972) proposed facies associations for each of these three main environments (Fig. 18). Finer-grained facies make up a large part of these facies associations.

Within these associations there are commonly smaller scale vertical sequences of beds that appear characteristic of different subenvironments or morphological elements in the deep sea (Fig. 19). For example, fining-upwards sequences

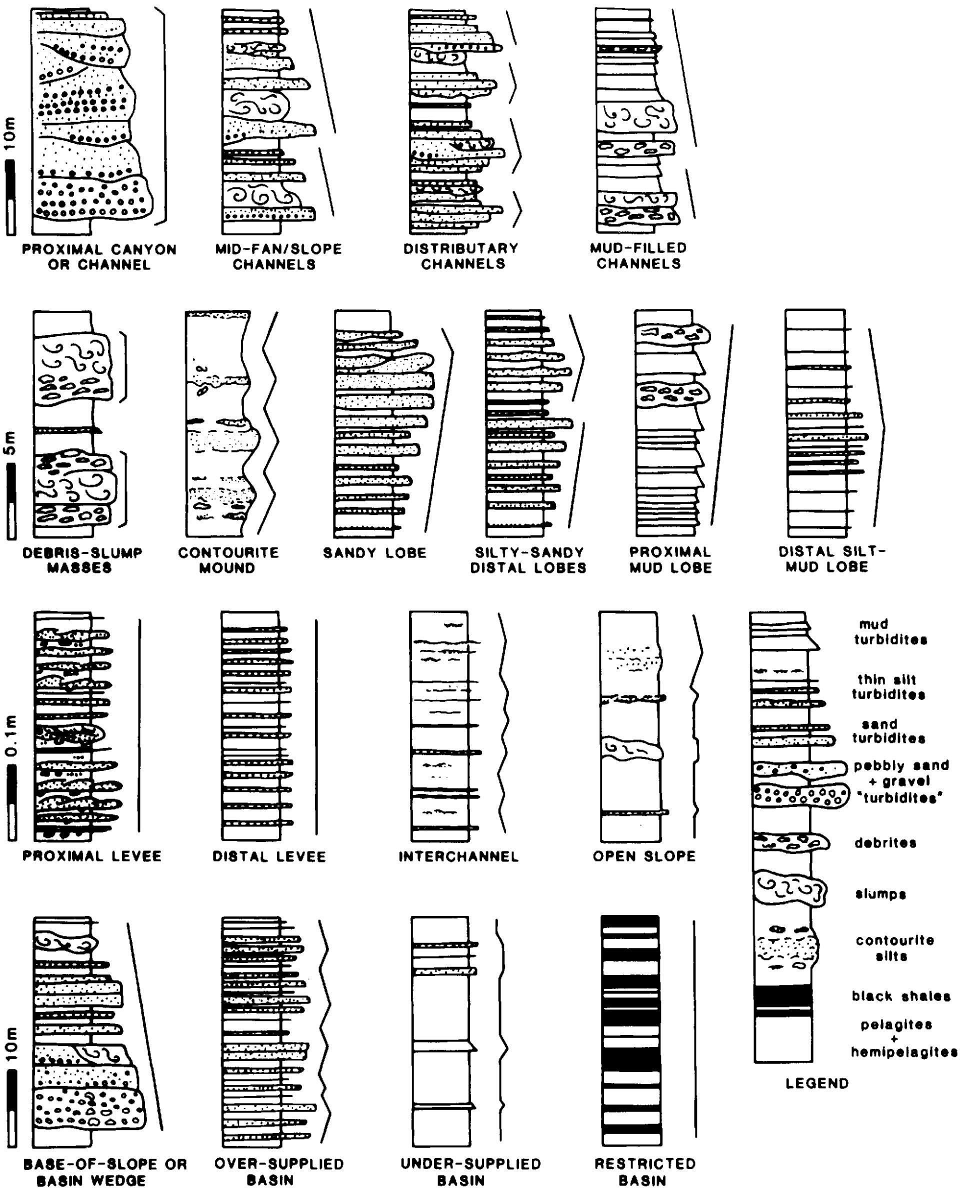

FIG. 19. Typical vertical facies sequences for various deep-water sub-environments (or morphological elements). Note variation in scale and in grading characteristics of the different sequences. (From Stow 1984a).

of turbidites are often interpreted as representing channel- or canyon-fill deposits, whereas coarsening-upwards sequences are taken as indicative of lobe deposits (Ricci Lucchi 1975; Rupke 1977; Walker 1978). Vertical sequences of predominantly fine-grained facies have been described by Shanmugam (1980); Stow *et al.* (1982); and Stow (1984a) among others. In this volume, papers by Hill; Pickering; Stow, Alam & Piper; Bourrouilh & Gorsline; Hayward; and Isaacs (all this volume) give further specific examples of sequences of a variety of facies in different environments.

Canyons and channels can be filled by muddy sediments rather than coarse-grained facies, although much of the fill comprises slumps and debrites as well as mud turbidites and hemipelagites in a rather chaotic vertical sequence (e.g. Coleman *et al.* 1983). Stow (1984c) has interpreted thickening-upward followed by thinning upward sequences of biogenic-mud turbidites and debrites from the south-east Angola Basin as representing progradational–regradational submarine fan facies. The more distal parts of fan lobes or finer-grained terminal lobes on large muddy fans appear to show symmetrical vertical sequences of silt and mud turbidites (e.g. Pickering 1982b).

Irregular or relatively uniform sequences of interbedded turbidites, contourites and hemipelagites/pelagites appear to be more characteristic of levees, interchannel, open slope and some basinal environments. These commonly differ in the relative proportions of the main facies present. More purely contourite sequences occur through contourite drifts (e.g. Gonthier *et al.*, this volume; Stow & Holbrook, this volume), and these apparently show an irregular alternation of sandy, silty and muddy contourite facies, although very little data are yet available. Thick sequences of regularly-spaced pelagite–hemipelagite alternations characterize many low energy slope and basinal environments.

Discussion

In this section we highlight briefly some interesting and problematical areas, mostly concerning detailed characteristics of the deep-water fine-grained facies outlined above. These are some of the areas of current research and where future advances are likely to be made.

Silt-mud lamination

Silt-laminated muds are a very common deep-sea facies and occur repeatedly in the geological record. However, their mode of emplacement and the process of lamina formation is not fully understood, and many different mechanisms have been proposed (see reviews by Stow & Bowen 1978, 1980). In many, perhaps most, cases, the lamination is current-induced. Velocity fluctuation within a single current (Lombard 1963), reflection of a turbidity current from the walls of a small basin (Van Andel & Komar 1969), the quasi-cyclic bursting process in boundary layers (Hesse & Chough 1980), and a series of small distinct flows or suspension clouds (Dzulynski & Radomski 1955) have all been suggested as the prime cause of such lamination. Perhaps more widely applicable are the processes of congregational sorting described by Piper (1972b), and of shear sorting of clay flocs from silt grains during final deposition through the base of the bottom boundary layer (Stow & Bowen 1978). This latter mechanism has received some support from experimental work with mud-water suspensions (see Kranck, this volume).

Although these kinds of processes would apply equally to bottom currents, the association of abundant laminae with bottom currents as proposed by Hollister & Heezen (1972) has not subsequently been confirmed (see e.g. Hill, this volume; Shor *et al.*, this volume). The laminae that have been described from contourites of sediment drifts are irregular or wispy and discontinuous (Piper & Brisco 1975; Stow 1982; Gonthier *et al.* this volume). Those of turbidites are more regular, distinct and continuous (e.g. Hill; Stow, Piper & Alam; Pickering; all this volume).

In anoxic environments that cannot support a burrowing infauna or epifauna, pelagic and hemipelagic sediments can also preserve a primary lamination. This lamination is not current-induced but results from periodic changes in the type of sediments supplied. The laminae are thus more truly varves, though not necessarily of seasonal periodicity (e.g. Isaacs, this volume; Robertson, this volume). Arthur *et al.* (this volume) show that lamination in black shales is in some cases turbiditic, in other cases pelagic and in other cases is better described as a fissility rather than true silt-mud alternation.

Mud fabric & fissility

The fabric of sands and gravels has long been used as a diagnostic of depositional process and flow direction and silt fabric in fine-grained sediments can be used in a similar manner (Piper 1972a,b; Stow 1979b). However, there has been very much less work of an equivalent nature on the fabric of muds and shales (see reviews by

Moon 1972; Bennett *et al.* 1977; O'Brien 1981; Moon & Hurst, this volume).

O'Brien (1981) and Moon & Hurst (this volume) agree that the development of fissility in shales is associated with an original orientation of clay flakes parallel to bedding. However, it remains uncertain whether the parallel orientation results from deposition of clay in the dispersed state as a result of an organic or inorganic deflocculant, or from deposition of oriented floc domains and subsequent mechanical reorientation on burial. Moon & Hurst (this volume) further discuss the importance of shale fissility in the generation and primary migration of hydrocarbons from organic-rich black shales.

Geotechnical properties

The study of sea floor soil mechanics has lagged behind that of its terrestrial counterpart, but has recently gained great impetus from the offshore oil, gas and minerals industries. In an important synthesis, Bennett *et al.* (1977) review the work on clay fabric and methodology and begin to relate different clay fabric types to selected geotechnical properties, depth of burial and laboratory consolidation loads. They suggest that for smectite or illite-rich muddy sediments, low void ratios result from high density packing of oriented clay particles, whereas high void ratios develop in sediments having non-oriented chain-like domains of clay particles.

The possible relationship of geotechnical properties to particular fine-grained facies deposited by different deep-sea processes has been investigated by Hein & Gorsline (1981) for muddy sediments in various borderland basins off California (see also Gorsline *et al.*, this volume). In a study of various hemipelagic sediments from the Guatemalan margin, including the slope, trench and Cocos oceanic plate environments, Faas (1982 and this volume) identifies marked regional trends in geotechnical properties. He notes high plasticity in the upper slope mudflow and debris flow deposits, together with high organic-carbon content, and a decrease downslope to lowest values in the trench-fill turbidites. The siliceous hemipelagites of the Cocos plate also have a relatively high plasticity index and are relatively overcompacted compared to oozes.

Bioturbation

Trace fossil and bioturbation assemblages (ichnofacies) are very common in fine-grained sediments throughout the deep sea and are useful for sedimentological and environmental interpretation (Seilacher 1967, 1978; Crimes & Harper 1977; Werner & Wetzel 1982; Wetzel 1982 and this volume). Various attempts have been made to summarize the distribution and character of ichnofacies assemblages in relation to bathymetry, sediment type, ecological stress and biotic diversity and density (e.g. Potter *et al.*, 1980, p. 44). Whereas early work suggested a *Zoophycos* slope assemblage and a *Nereites* basinal assemblage, the situation is now seen to be more complex with at least five different assemblages found in the deep sea (Werner & Wetzel 1982).

Although there is not a simple relationship between sediment facies and ichnofacies, it appears that in many cases the three main facies groups do have a characteristic suite of burrows and/or degree of bioturbation. Pelagites and hemipelagites are very thoroughly bioturbated and may show a wide range of burrow types with several superimposed tiers being present. Contourites show almost complete bioturbation but insufficient to have destroyed all primary structures. Different burrow types characterize the more muddy and more silty contourite facies. Turbidites tend to be markedly less bioturbated with a concentration of burrows towards the tops of individual beds and a still more restricted diversity of burrow types. Superimposed tiers of burrows may also be present where there has been sufficient time between successive flows. Turbidity currents can apparently introduce an exotic shallow-water burrowing fauna into a deep-water setting (Wetzel, this volume).

Ichnofacies assemblages are also proving to be an important characteristic for understanding the depositional environment and origin of black shales. In particular, the nature of the burrows in the interbedded non organic-rich facies and in the transition zone to black shales appear related to bottom water oxygenation (Byers 1977; Ekdale 1980; Stow & Dean 1984).

Bioturbation is presumably absent from many black shales because of the anoxic bottom conditions during sedimentation. However, there are many examples of apparently oxic environments and sediments that still lack bioturbation. The other factors that deter burrowing organisms are still poorly known.

Textures

The analysis of grain size and other textural attributes of fine-grained sediments can provide very useful information on the processes and mechanics of deposition (e.g. Piper 1973; Rivière 1977; McCave, this volume; Kranck, this volume). However, textural analyses are more difficult to perform on consolidated material, and hence comparison of ancient and modern sediments is difficult.

There is commonly a clear textural distinction between current-deposited and pelagic-settled sediments (e.g. Passega 1964). This can be seen as a sorting parameter, on a plot of coarser one percentile against median diameter, or in terms of the shape of the cumulative grain-size distribution curve. Using Rivière's (1977) terminology, the grain-size curves tend towards logarithmic for pelagic/hemipelagic sediments, hyperbolic (coarse tail) for muddy contourites and parabolic (fine tail) for turbidites and silty or fine sandy contourites. The types of grain-size sequence or grading are also diagnostic of the different facies: distinct positive grading and positive grading through grouped silt laminae are characteristic of turbidites; more irregular, alternating positive-negative grading is characteristic of contourites; and negative grading is most pronounced through the turbidite-pelagite transition in biogenic turbidites.

Grain size as analysed in the laboratory differs significantly from grain size during deposition in two respects: (1) the clay fraction is analysed in the dispersed state but usually deposited as flocs of various sizes; and (2) the correspondence between grain size and hydraulic equivalence of biogenic and terrigenous material is poorly known. The measurement of grain size of suspended sediment in sea water allows greater insight into the depositional state of fine-grained material (e.g. McCave, this volume; Eittreim, this volume). Kranck (this volume) attempts to distinguish between grain-size populations that have been deposited as flocs and those that have settled in the dispersed state. There has been little serious attempt to determine the hydraulic equivalence of biogenic material (but see Berger & Piper 1972).

Composition and colour

There are many different aspects to the composition of fine-grained sediments, some of which have received much attention and others less attention (see, for example, Potter *et al.* 1980; Tissot & Welte 1978). The biogenic content is clearly important for both biostratigraphic and palaeoecological information. Clay mineralogy has been much used in palaeoenvironmental and provenance studies. Inorganic geochemistry is beginning to be used more for provenance study. Organic geochemistry is vital for our understanding of hydrocarbon source-rock deposition and potential. What seems to be lacking most in our understanding and interpretation of compositional characteristics is better use of a combined approach involving a number of separate compositional studies on the same suite of rocks.

Although in individual cases compositional criteria are extremely valuable in distinguishing between turbidite, contourite and pelagite facies (e.g. Piper 1978; Stow & Lovell 1979; Spears & Amin 1981), such criteria are not readily generalized as so much depends on the local source and supply, competition from other components, and so on.

The colour of fine-grained sediments is closely related to composition and has been a topic of long-standing interest, but one in which relatively little advance has been made in our understanding of the origin of colour and hence in using it as a diagnostic feature (but see McBride 1974; Pettijohn 1975; Potter *et al.* 1980). Potter *et al.* (1980) present a useful synthesis of what we know so far and have attempted to quantify the relationship of shale colour to both carbon content and the oxidation state of iron.

Diagenesis

Of prime importance to all studies of ancient fine-grained sediments is the question of diagenesis. The twin processes of mechanical and chemical diagenesis can so profoundly alter the structural, textural, compositional, colour and fabric characteristics of deep-sea facies, that their use as criteria for distinction and for interpreting depositional history must necessarily change.

The clay-rich beds in a pelagite–hemipelagite sequence can be reduced to thin shale partings by carbonate dissolution (e.g. Fischer & Arthur 1977). Primary sedimentary structures can be destroyed and deformational structures introduced during compaction (e.g. Rieke & Chilingarian 1974). Clay minerals undergo progressive alteration with increased burial pressures and temperatures (e.g. Perry & Hower 1970). Organic carbon is subject to progressive maturation into hydrocarbons with increasing temperature (e.g. Tissot & Welte 1978). Fluids released and mobilized during these various processes migrate through the formation and cause further chemical changes (e.g. Neglia 1979).

These and numerous other effects of diagenesis we do not discuss here, and were not discussed at any length at the Fine-Grained Sediments Workshop in Halifax. We would like simply to emphasize the importance of considering diagenetic history in any study of fine-grained sediments in the deep sea. One of the key areas of research in the near future must surely be the development of diagenetic models, for both fine-grained and coarse-grained sediments, to complement the facies models we have outlined in this paper.

Conclusions

(1) The wealth of data that exists and the depths of our knowledge on fine-grained sediments has increased dramatically in the past five to ten years. The holding of a workshop on just the deeper water fine-grained sediments and the editing of this volume has enabled us to take stock of where we have reached and where we should be going in this field of research. In this paper we have attempted a synthesis of facies models and have drawn heavily on other papers in the volume.

(2) We identify three broad facies groups in the deep sea: turbidites, contourites and pelagites–hemipelagites. These are related primarily to the processes of deposition by, respectively, gravity-driven turbidity currents, thermohaline/wind-driven normal bottom currents, and vertical settling through the water column. Within each of these groups we describe several distinct facies models:

turbidites
  (1) silt turbidites
  (2) mud turbidites
  (3) biogenic turbidites
  (4) disorganized turbidites
contourites
  (1) muddy contourites
  (2) silty/fine sandy contourites
pelagites–hemipelagites
  (1) pelagic oozes
  (2) muddy pelagic oozes
  (3) pelagic clays
  (4) hemipelagites

These facies models, based on a large amount of observational data on both modern and ancient sediments, are intended as a summary of principal sedimentary characteristics and as a basis for interpretation of the depositional processes involved. Neither the complete sets of characteristics nor the full sequences of sedimentary structures will necessarily be observed in every section examined and we have tried to show, to some extent, the variability to be expected in each case.

(3) We have also given a brief summary of the hydrodynamic interpretation for each facies model. However, it should be emphasized that there appears to be a continuum between the various processes that operate in the deep sea and hence there will be a continuum of resultant facies deposited.

(4) The different fine-grained facies show certain characteristic patterns of occurrence both horizontally over the surface of the sea floor and vertically in boreholes or in ancient rock successions. They occur interbedded with and adjacent to each other and to coarser-grained facies. These patterns of distribution, which we have briefly described, can be used to help identify depositional palaeoenvironments.

(5) Finally, in discussion, we attempt to do little more than highlight some of the interesting and problematical areas of current research in fine-grained deep-water sediments. We mention, in particular, certain aspects of facies characteristics including: silt-mud lamination, fabric and fissility, geotechnical properties, bioturbation, textures, composition and colour, and diagenesis.

ACKNOWLEDGEMENTS: In a synthesis paper of this kind we have many people to thank for their helpful and stimulating discussion. In particular, we thank all the participants in the International Workshop on Fine-Grained Sediments held at Halifax, Canada in 1982. DAVS acknowledges financial support from the Natural Environment Research Council, UK, and technical and secretarial support at the Grant Institute of Geology. We are very grateful for the helpful comments of our reviewers, Ali Aksu, Donn Gorsline, Phil Hill and Julian Holbrook.

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