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Models for the deposition of Mesozoic–Cenozoic fine-grained organic-carbon-rich sediment in the deep sea

M.A. Arthur, W.E. Dean and D.A.V. Stow

SUMMARY: The widespread occurrence of organic-carbon-rich strata ('black shales') in certain portions of Jurassic, Cretaceous and Cenozoic sequences has been well-documented from Deep Sea Drilling Project sites in the Atlantic and Pacific Oceans and from sequences, now exposed on land, originally deposited in the Tethyan ocean. These ancient black shales usually have been explained by analogy with examples of modern deep-sea sediments in which organic matter locally is preserved by (1) increasing the supply of organic matter, (2) increasing the rate of sedimentation, and/or (3) decreasing the oxygen content of the bottom water. However, detailed examination of many black shales reveals characteristics that cannot be explained by simple local models, including: their approximate coincidence in time globally; their occurrence in a variety of different environments, including open oxygenated oceans, restricted basins, deep and shallow water; their interbedding with organic-carbon-poor strata which often dominate a so-called black shale sequence; their deposition by pelagic, hemipelagic, turbiditic and other processes; and the variations in type and amount of organic matter that occur even within the same sequence.

A more complex model for the origin of black shales therefore appears most appropriate, in which the cyclic preservation of organic matter depends on the interplay of the three main variables, namely supply of organic matter, sedimentation rate, and deep-water oxygenation, each of which varies independently to some extent. The variation and relative importance of these parameters in individual basins and widespread black shale deposition in general are linked globally and temporally by changes in global sea-level, climate and related changes in oceanic circulation. An important and often overlooked factor for the supply of organic matter to deep-basin sediments is the frequency and magnitude of redepositional processes. The interplay of these variables is discussed in relation to the middle Cretaceous and Cenozoic organic-carbon-rich strata, in particular, which show marked differences in the relative importance of the different variables.

Results of analyses of Deep Sea Drilling Project (DSDP) cores show that strata containing relatively high concentrations of organic matter (more than 2% organic carbon) are common at different stratigraphic horizons and in many parts of the world ocean. These strata are often loosely described as 'black shales' although they usually consist of interbedded rocks with varying contents of carbonate and/or biogenic silica and differ mainly in colour and/or concentration of organic matter. The most common lithologies are interbedded black or dark-olive shale or claystone, and lighter greenish-grey shale or claystone (i.e. interbedded 'black' and 'green' argillaceous rocks).

The oldest occurrences of organic-carbon-rich strata in the present ocean basins are upper Jurassic (Callovian-Oxfordian) black claystones recovered at two DSDP Sites in the South Atlantic and one in the North Atlantic, and Tithonian-Neocomian deep-water marlstones and limestones that have been recovered from several DSDP sites in the North Atlantic. Black-shale deposition was widespread in the North and South Atlantic during the Barremian-Albian (100–115 my BP) and Cenomanian-Turonian (86–100 my BP) at both continental-margin and deep-basin sites (Arthur & Natland 1979; Jenkyns 1980; Weissert 1981) and locally during the Coniacian-Santonian (75–86 my BP). Cretaceous organic-carbon-rich strata in the Pacific are much more restricted in both time and space than in the Atlantic. They have been recovered from seven DSDP sites on the flanks of elevated volcanic plateaus and seamounts, mainly within very restricted stratigraphic intervals (Schlanger & Jenkyns 1976; Dean *et al.* 1981; Thiede *et al.* 1982). Organic-carbon-rich strata in the Pacific all occur within the same general middle Cretaceous time interval, but they are not strictly synchronous and occur in strata that range in age from 86–120 my BP. Middle Cretaceous strata recovered from deep basins of the Pacific generally are not rich in organic carbon, but a thin layer containing 5% organic carbon was recently recovered at the Cenomanian-Turonian boundary within a turbidite sequence in the eastern Mariana Basin (DSDP Site 585; Moberly *et al.* 1983). Organic-carbon-rich sediments of Eocene and Miocene age have been recovered at a few sites in the Atlantic (Arthur & von Rad 1979; McCave 1979a; Dean & Gardner 1982), but these strata are much more restricted in both time and space than those of Cretaceous age. In the Pacific, however, extensive organic-carbon-rich Miocene strata occur in both offshore and onshore localities.

In this paper we first provide an outline of the main modes of preserving organic matter in modern deep-sea sediments, then describe the lithology, distribution and organic-carbon content of ancient Mesozoic and Cenozoic black shales, and finally discuss the elements of models for black shale deposition. Our emphasis throughout is on deep-water and fine-grained organic-carbon-rich sediments.

Preservation of organic matter in modern deep-sea sediments

Most modern deep-sea environments are characterized by sediments that are relatively depleted in organic carbon (<0.5). At a water depth of 4000 m, less than one percent of the organic matter produced in surface waters survives to reach the sediment/water interface (Fig. 1; Suess 1980); a relatively small proportion of this organic matter is reactive in the sense of Toth & Lerman (1977). Therefore, not only are concentrations of organic matter low in the majority of deep-water sediments, but this organic matter is low in nitrogen and phosphorus, and is relatively oxidized; reactive organic matter is rapidly consumed by the benthic biomass (Müller & Suess 1979). Continued degradation of the organic

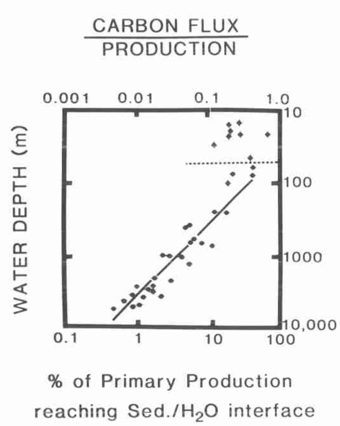


FIG. 1. Proportion of organic carbon reaching the sediment/water interface at different depths as percent of primary productivity in surface waters. Data are results of sediment trap experiments compiled by Suess (1980). The data indicate an exponential decrease of organic matter accumulation with depth, but represent the situation for an oxygenated water column only.

matter occurs during the dominantly oxic conditions of diagenesis in most pelagic deep-sea sediments (e.g. Heath *et al.* 1977; Demaison & Moore 1980).

There are three main ways of preserving relatively high concentrations of organic matter in deep-sea sediments: the first is by increasing the supply of organic matter, the second is by increasing the rate of sedimentation, and the third is by decreasing the oxygen content of the water mass overlying the sediment. These processes usually but not always act in consort in modern marine environments.

Organic matter supply

Marine organic matter is supplied to the sediments from primary biological productivity in the surface waters; terrigenous organic matter is derived from land plants and transported to the sea mainly by rivers, but some organic materials are carried from land by strong prevailing winds (charcoal, spores, pollen, etc.). Areas of increased primary productivity occur where deep (below the thermocline, i.e. from below about 150–250 metres depth) nutrient-rich waters come to the surface along an equatorial belt, along northern and southern polar divergence zones and along the western margins of the major continents (e.g. Koblentz-Mishke *et al.* 1970; Demaison & Moore 1980). Increased supplies of terrigenous organic matter are constrained by fluvial discharge and by the climate and vegetation of the drainage area; equatorial regions therefore receive more organic matter than polar or temperate seas. Arid coasts within trade wind regions may supply little terrestrially derived material.

Both marine and terrigenous organic matter can be supplied directly to deep-water sediments as a part of the pelagic 'rain' (or marine 'snow'; Shanks & Trent 1980) and by faecal pellet transport (e.g. Suess 1980; Dunbar & Berger 1981). Sinking rates of faecal pellets of up to 170 m per day and marine 'snow' of up to 95 m per day suggest that transfer of organic matter through the water column by these two processes is relatively rapid. However, sediment trap studies do suggest that much oxidation does occur in transit through oxic water masses by bacterial action or consumption by other organisms (e.g. Suess 1980; Fig. 1). In addition, and perhaps more commonly, the widespread downslope redeposition of organic-carbon-rich shelf and slope sediments is an important means of supply to deeper basinal sediments. This mode of transport includes supply from bottom nepheloid-layers as suggested by Summerhayes (1981b).

There have been few systematic studies of the organic contents of redeposited sediments along modern continental margins, but it is known that widespread redeposition of slope sediment occurs, for example, in the area of high organic productivity off north-west Africa (Embley 1976; von Rad *et al.* 1979). Slumps and slope-derived turbidites also are common along the California borderland (Nardin *et al.* 1979; Soutar *et al.* 1981), along the west coast of Mexico in the Gulf of California (Einsele & Kelts 1982), along the Peru-Chile Slope (Soutar *et al.* 1981), and in the Angola Basin off south-west Africa (Summerhayes *et al.* 1979; Embley & Morley 1980; Hay *et al.* 1982; Stow, in press). All of these regions have organic-carbon-rich slope sediments preserved as the result of upwelling and high biological productivity, and well-developed oxygen-minimum zones. More than likely, some of the sediment redeposited to the base of slope or basin floor also is rich in organic carbon. Such sediment, although redeposited, would probably retain many of the original compositional characteristics of the shallower sediment, including a high concentration of well-preserved organic matter.

Sedimentation rate

Relatively high sedimentation rates apparently aid in the preservation of organic carbon in marine sediments because, up to a point, organic carbon contents and accumulation rates in modern sediments are positively correlated with bulk sediment accumulation rates (Heath *et al.* 1977; Müller & Suess 1979; Fig. 2). Ibach (1982) suggested that this positive relationship between organic-carbon content and sedimentation rate also holds for Cenozoic and Mesozoic strata

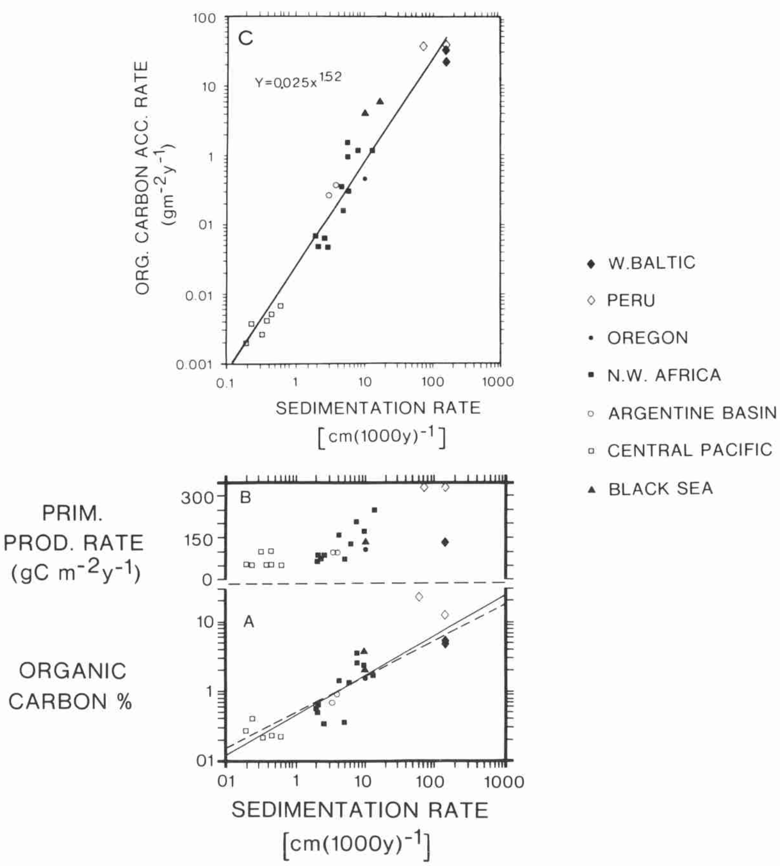


FIG. 2. Relations among bulk sedimentation rate and (A) percent organic carbon, (B) primary productivity, and (C) organic carbon accumulation rates for modern marine sediments (Müller & Suess 1979). The scatter about the best-fit regression line represents the effects of dysaerobic or anoxic conditions which further enhance preservation (e.g. Black Sea points), as well as no correction for differences in depth of deposition among other problems.

recovered at DSDP sites. The main reason for this relationship is probably that high sedimentation rates effectively bury organic matter more rapidly, removing it from zones of bioturbation, oxic decomposition, and sulphate reduction. Although organic-carbon contents and accumulation rates may be higher in high-sedimentation-rate sequences, the organic matter itself may be somewhat more poorly preserved losing a greater proportion of H, P, N than that deposited under anoxic conditions. Hydrocarbon source potential is determined not only by the organic carbon content but also the characteristics of the organic matter.

Bottom-water oxygenation

Oxygen concentrations of less than 0.5 ml/l and certainly 0.2 ml/l inhibit the activity of benthic metazoans (e.g. Rhodes & Morse 1971). Lack of bioturbation in turn limits the residence time of organic matter at the sediment/water interface and, therefore, the state of its oxic decomposition. Low-oxygen concentrations occur in both mid-water oxygen-minimum zones and in silled stratified basins, and, where lowest (i.e. <0.5 ml/l), commonly lead to enrichment of organic carbon in underlying sediments.

Most areas of modern oceans that are covered with organic-carbon-rich sediment are within well-developed oxygen-minimum zones (e.g. Demaison & Moore 1980). These zones originate partly because of the slow rate of sinking and advection of warm surface water in low latitudes (Wyrtki 1962). These warm-water masses are more poorly oxygenated than cold, deep waters that originate by sinking and advection at high latitudes because of the lower solubility of oxygen with increasing temperature (Fig. 3). Also, much of the decomposition of organic matter and concomitant consumption of dissolved oxygen takes place in the upper few hundred metres of the water column within and below the base of the photic zone (Suess 1980). In oxic environments the rate of organic matter decomposition decreases exponentially with depth (Fig. 1). Therefore, in regions of high biologic productivity, minimum concentrations of dissolved oxygen occur from about 100 m below the surface to about 1500 or 2000 m, but the extent of oxygen depletion varies (Fig. 4) depending on nutrient availability and surface water productivity, initial dissolved oxygen content of intermediate water masses, and rates of horizontal advection of oxygen. Bottom sediments within oxygen-minimum zones may preserve high concentrations of organic matter. For example, concentrations of organic carbon of up to 25% occur in sediments

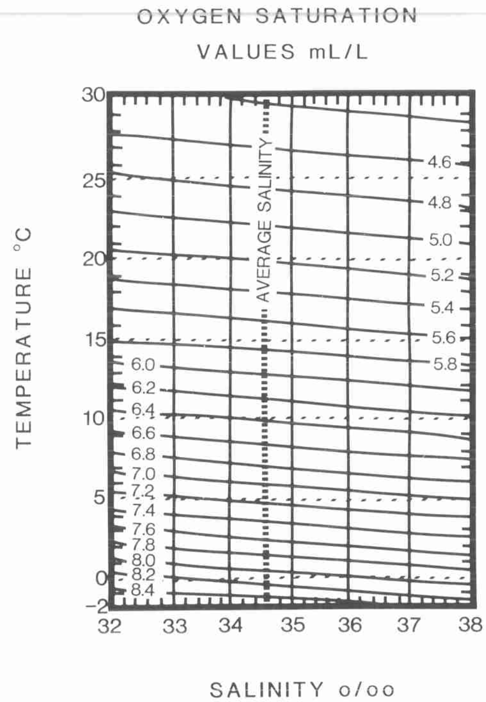


FIG. 3. Theoretical oxygen solubility in surface seawater as a function of temperature, salinity, and modern atmospheric pO_2 (after Wilde & Berry 1982).

where the oxygen-minimum zone impinges on the inner continental shelf off south-west Africa (Calvert & Price 1971). However, organic-carbon enrichment in these sediments probably is due mainly to the high productivity which is centred over a very shallow shelf (<150 m) (see later discussion).

The Black Sea is an example of a silled anoxic basin (Fig. 4) with as much as 4 to 5% organic carbon in bottom sediments (Shimkus & Trimonis 1974). The bottom water of the Black Sea is completely depleted in dissolved oxygen from 175 m to the bottom (about 2500 m). Anoxia is maintained by strong salinity stratification and relatively slow replenishment of bottom water, that is, slow replenishment in comparison to rates of oxygen consumption due to organic matter oxidation and sulphide oxidation near the point of bottom water sinking. However, the Black Sea is by no means the sluggish, stagnant basin implied by many authors; the entire deep-water mass is probably replenished every 700 yrs or less (Grasshoff 1975). Sulphate reduction occurs in the water column and results in dissolved hydrogen sulphide in the water column and a relatively high sulphur to carbon ratio in bottom sediments (Leventhal 1983). The Black Sea is a relatively

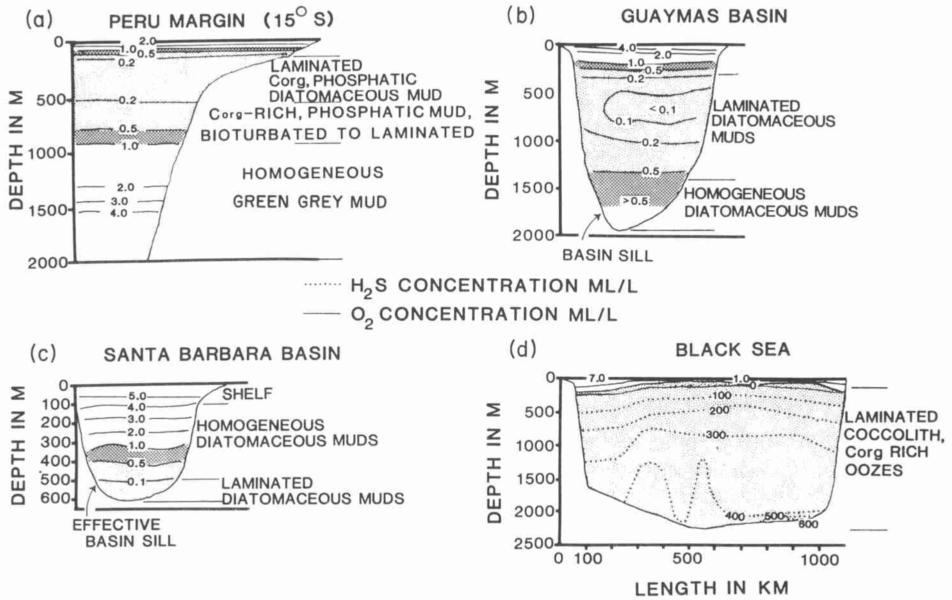


FIG. 4. Examples of modern marine settings in which dissolved oxygen is relatively depleted within intermediate or deep water masses. Shown are contours of dissolved oxygen with depth. (a) East–West section of Peru margin at 15°S (data from Burnett *et al.* 1981) showing an open-ocean oxygen-minimum zone; (b) East–West section across Guaymas Basin, Gulf of California, a narrow, semi-enclosed ocean basin with a well-developed oxygen-minimum zone; oxygen-depleted midwater masses have been advected into the basin; (c) Santa Barbara Basin off southern California (data from Emery 1960; Rittenberg *et al.* 1953); the basin is silled within the midwater oxygen-minimum zone to the west and therefore the deep waters of the basin are relatively depleted in oxygen initially; (d) Black Sea; a salinity stratified sea (data from Grasshoff 1975) that is anoxic below 175 m with sulphate reduction occurring in the water column; dotted contours below zero O_2 level are H_2S concentrations.

small silled basin with only moderate productivity in surface waters (< 100 $gC/m^2/y$) and unusual circulation conditions, but anoxia has persisted for the last 7000 yrs. About 5% of the organic carbon produced in surface waters is preserved in abyssal Black Sea sediments (Deuser 1974) as opposed to much less than 1% in most oxygenated settings. The enhanced preservation is most certainly due to anoxia and the lack of benthic metazoan activity. It is not entirely clear whether such conditions could exist continuously in a basin as large as the Cretaceous Atlantic Ocean for periods of up to 10 or 15 my.

The Mediterranean Sea is areally larger than the Black Sea, but apparently also has experienced periodic anoxic conditions since at least the Miocene (e.g. Ryan & Cita 1977; Kidd *et al.* 1978) possibly due to conditions similar to those in the Black Sea today. The Mediterranean is essentially a silled basin with respect to the Atlantic Ocean. The thin but laterally extensive sapropels (> 2% organic carbon) of the Mediterranean contain mainly marine organic carbon with maximum concentrations of about 17% (Ryan 1972; Nesteroff 1973; Kidd *et al.* 1978). Many of the later Pleistocene sapropels probably are the result of increased runoff of glacial meltwater to the Mediterranean coupled with increased surface productivity and reduced rates of sinking of bottom water (Thunell *et al.* 1977; Williams *et al.* 1978; Jenkins & Williams, in press) or possibly due to the influence of increased fluvial runoff as the result of periodically increased monsoonal rainfall in north Africa (Rossignol-Strick *et al.* 1982).

The Santa Barbara Basin is a restricted marginal basin offshore California as described by Rittenberg *et al.* (1953). Much of the basinal sediments in the Santa Barbara Basin are laminated and relatively rich in organic matter (Soutar *et al.* 1981) with contents of 3–5% organic carbon in the deepest parts of the basin and < 1% on the topographic highs (Emery 1960). The only bottom water for the basin is derived from within a well-developed oxygen-minimum zone that impinges at sill depth (Fig. 4), which, coupled with high coastal productivity, explains why oxygen depletion extends to the basin bottom.

However, the basin circulation is not 'stagnant' and deep waters may be flushed out as frequently as every 20 yrs or so (e.g. Hülsemann & Emery 1961; Soutar *et al.* 1981). Other examples of oxygen depleted basins are the thermally-stratified Cariaco Trench in the Caribbean Sea, and shallow-silled, salinity-stratified fjords, such as the Gotland Basin of the Baltic Sea and Saanich Inlet, British Columbia. However, not all silled or restricted basins are characterized by oxygen-depleted deep-water masses (e.g. the Red Sea; Grasshoff 1975).

Sedimentary indicators of productivity and levels of bottom water dissolved oxygen

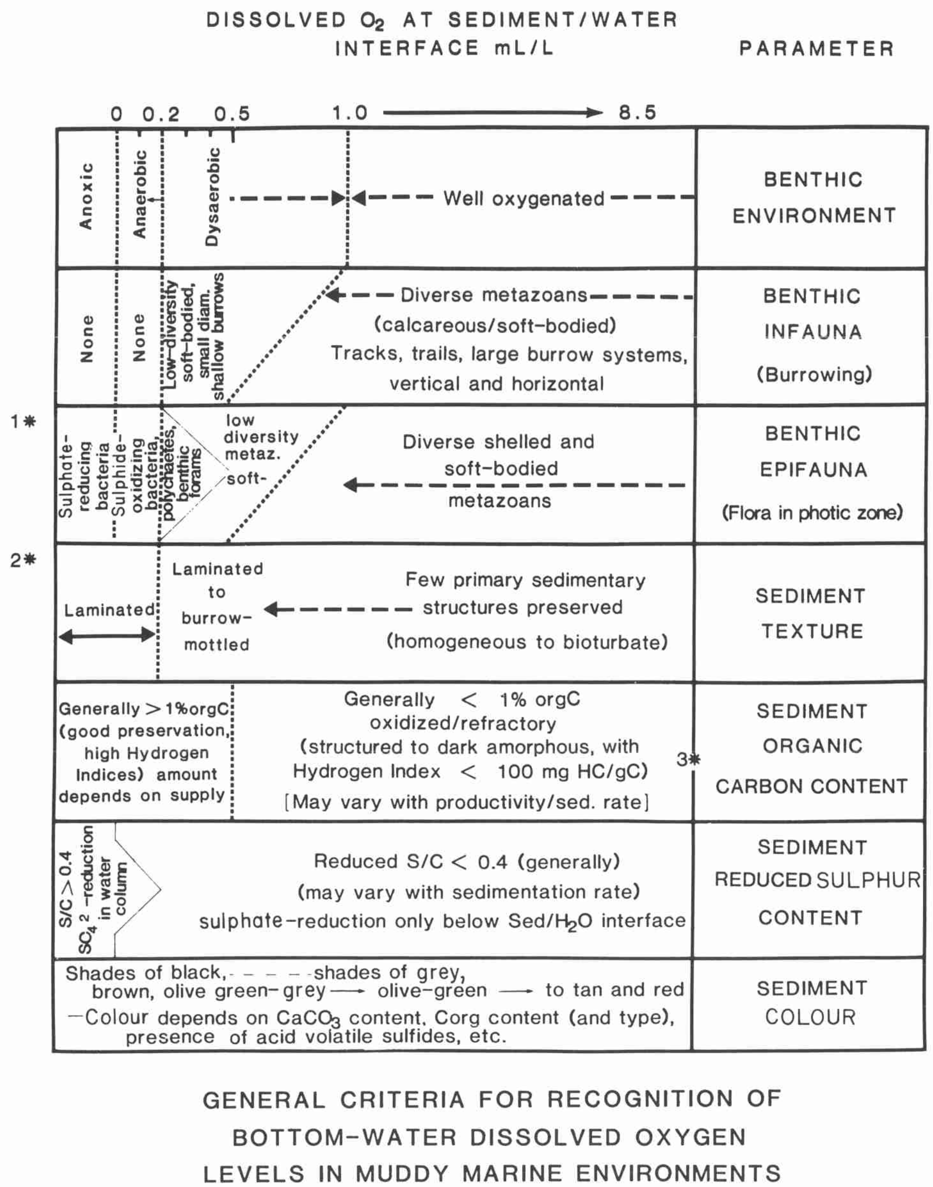
The three main factors that influence the accumulation of organic carbon: supply of organic matter; sedimentation rate; and concentration of dissolved oxygen in the water-column are commonly linked in some way in modern marine environments. It is frequently difficult to identify any one factor as the major cause of enhanced organic carbon preservation in a given ancient sedimentary environment, largely because the necessary unambiguous criteria do not exist or have not yet been formulated and tested. Primary sedimentary structures, mineralogy, biotic composition, and sediment geochemistry may, however, provide some constraints on interpretation. A basic matrix of dissolved-oxygen concentrations in bottom waters versus benthic biotic composition, primary sedimentary structures, and selected geochemical indicators in sediments is shown in Fig. 5. Such criteria are, of course, to be used with caution.

In general, bioturbated sediments occur where bottom water dissolved oxygen levels are above 0.5 ml/l. Finely laminated or nonbioturbated sediments may indicate anaerobic (<0.2 ml/l) to anoxic conditions. Many organic-carbon-rich sediments deposited under anoxic conditions are therefore finely laminated. Lamination represents preservation of the record of seasonal variation in supply of two or more components or of current-controlled deposition. For example, annual varves preserved on the slope of the Gulf of California under an intense oxygen-minimum zone represent alternation of terrigenous clay minerals supplied by run-off during the rainy season and high productivity of diatoms and organic matter during seasonal upwelling (Calvert 1966b). The small diameter of burrows and predominance of shallow horizontal versus vertical feeding traces may indicate low-oxygen conditions at the sea floor which approach but do not go below 0.2 ml/l (e.g. Savrda *et al.* 1982). Even at 0.2 ml/l it is possible to find non-metozoan organisms such as certain benthic foraminifers or serpulids (Calvert 1966b; Phleger & Soutar 1973) living on laminated sediments, and in certain settings benthic reworking of surface mud by pelagic crabs can be quite pronounced at dissolved oxygen levels of just above 0.25 ml/l (Soutar *et al.* 1981).

The control of oxygen levels on activities of larger benthic metazoans is the main way in which the concentration of dissolved oxygen affects organic carbon preservation in sediments, and probably is the main reason for better preservation of organic matter in dysaerobic to anoxic depositional environments, regardless of the rates of primary productivity. Bioturbation increases the residence time of sedimentary particles, including organic matter, and thereby increases the chance for ingestion and oxidation of organic carbon. Higher sedimentation rates (Müller & Suess 1979) act in the same manner to decrease the residence time of organic carbon in zones of oxic consumption and bacterial sulphate reduction.

The presence of pyrite or other sulphide minerals in sediments does not necessarily imply anoxic conditions at the sea floor, but only that sufficient metabolizable organic matter remained during burial to promote anoxia, sulphate reduction within the sediment and sulphide mineralization (Goldhaber & Kaplan 1974; Berner 1977, 1978). However, Leventhal (1983) suggested that a weight ratio of reduced sulphur/carbon which is consistently above 0.4 (e.g. Goldhaber & Kaplan 1974) with a positive intercept of the best fit line to the data on the sulphur axis of a S/C plot may indicate sulphate reduction within an anoxic water column as in the modern Black Sea.

Finely laminated, pyritic sediments rich in organic carbon and with high S/C ratios may indicate an anoxic depositional environment, but these criteria do not indicate the role of primary productivity in producing either the anoxic conditions or the relatively high ($>1\%$) concentration of organic carbon typical of those sediments. Large amounts of biogenic silica (mainly diatoms and radiolaria in modern sediments) commonly indicate high surface water productivity (Berger 1976), particularly where upwelling of cooler, nutrient-enriched intermediate water occurs. Other valuable criteria for evaluating productivity are relatively high accumulation rates of organic carbon and total phosphorous (biogenic debris as well as authigenic phosphate minerals; e.g. Burnett *et al.* 1980) in sediments. The data illustrated in Fig. 6 are for several modern and ancient environments compiled by Glenn & Arthur (in press; see sources of data within). The fields marked Peru and SW Africa represent regions of high primary productivity associated



1* certain species of polychaetes (e.g. Calvert, 1966) and/or calcareous or arenaceous benthic foraminifers (e.g. Phleger and Soutar, 1973) may inhabit surficial environments where benthic metazoans are excluded.

2* degree of lamination in part depends on seasonal variations in clastic input alternating with higher productivity intervals, and on sedimentation rate.

3* see text.

FIG. 5. Oxygen content at the sediment/water interface and generalized expected relationships between types of benthic organisms, primary sedimentary structures, sediment chemistry and mineralogy (sulphides), organic content and type (data and concepts from numerous sources).

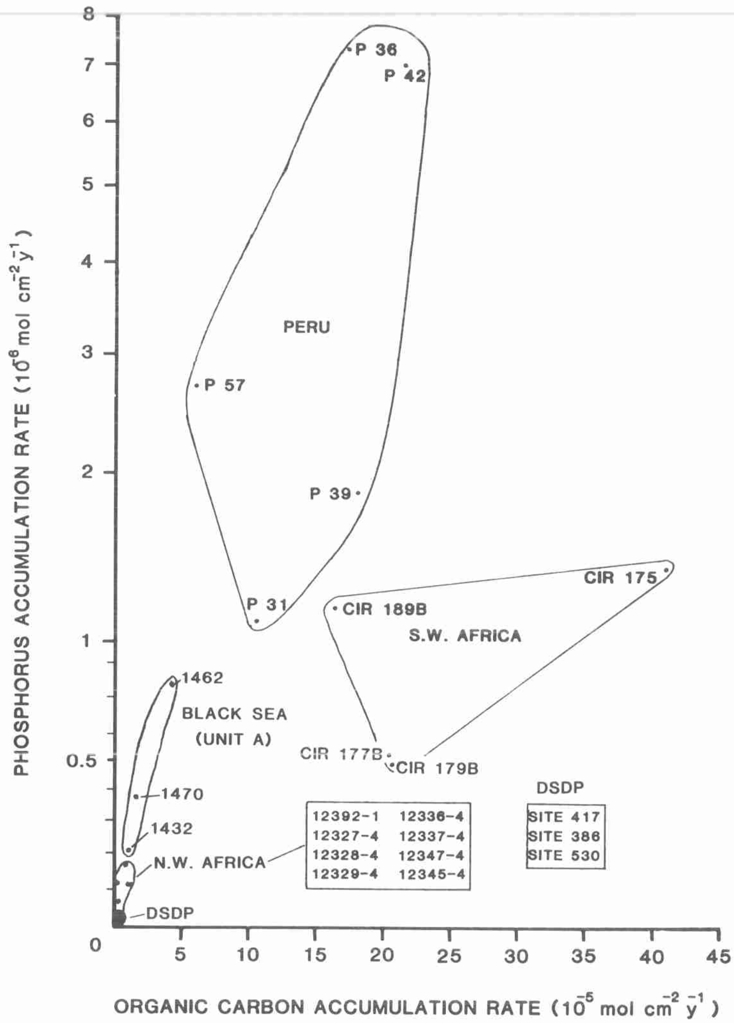


FIG. 6. Plot of total phosphorus accumulation rate *vs*. organic carbon accumulation rate for Holocene sediments from the Peruvian slope, the south-west African shelf (Namibia), the Black Sea abyssal plain and slope, the north-west African slope, and for average mid-Cretaceous 'black shales' from selected Atlantic DSDP Sites (see Glenn & Arthur, in press, for sources of data and calculations). Letters and numbers indicate individual cores.

with coastal divergences with well-developed midwater oxygen-minimum zones. Note that the organic carbon and total phosphorous accumulation rates for upwelling settings are much higher than for the Black Sea, which is characterized by an anoxic water column but moderate to low productivity. Seasonal wind-driven upwelling, high nutrient supply and high rates of organic carbon production and deposition over the relatively shallow shelf off SW Africa (e.g. Calvert & Price 1971) probably are much more important than oxygen levels in controlling the accumulation of organic carbon because bottom water oxygen levels rarely fall below 1 ml/l on either the shelf or slope (with the exception of a few local areas of <0.1 ml/l), and because the organic-carbon-rich sediments are not laminated. Lack of lamination may also be due to a lack of strong seasonal variations in supply of sediment components other than diatoms and organic carbon. On the Peruvian margin, some sediments with higher organic-carbon accumulation rates are associated with deposition under low-oxygen conditions (<0.02 ml/l) within the oxygen-minimum zone.

There are some differences in phosphorus and organic carbon accumulation rates between offshore Peru and SW Africa (Namibia). Surface water nutrient supply and productivity are nearly the same in both settings (Koblentz-Mishke *et al.* 1970; Huntsman & Barber 1977). The results shown in Fig. 6 for SW Africa all came from sediment samples from the upper 50 cm or less in cores from shallower than 134 m on the Namibian shelf; the results for Peru are from cores at depths of 186–645 m on the outer shelf or upper slope. The accumulation rates of organic carbon are not too different in either setting. The shallow depth of the Namibian examples and the high rate of supply of organic matter probably offset some of the effects of an oxygenated environment. Transfer of organic carbon to sediments is largely by faecal pellets (Bishop *et al.* 1978), and marine lipids are the only ones present off the arid coast of Namibia. Concentrations of dissolved oxygen in bottom waters off Peru are significantly lower than in bottom waters of most areas off Namibia, and this must aid in preservation of organic carbon. The higher phosphorus accumulation rates of several Peruvian cores (P-36 and P-42, Fig. 6) may indicate a supply of phosphorus in addition to that associated with organic carbon

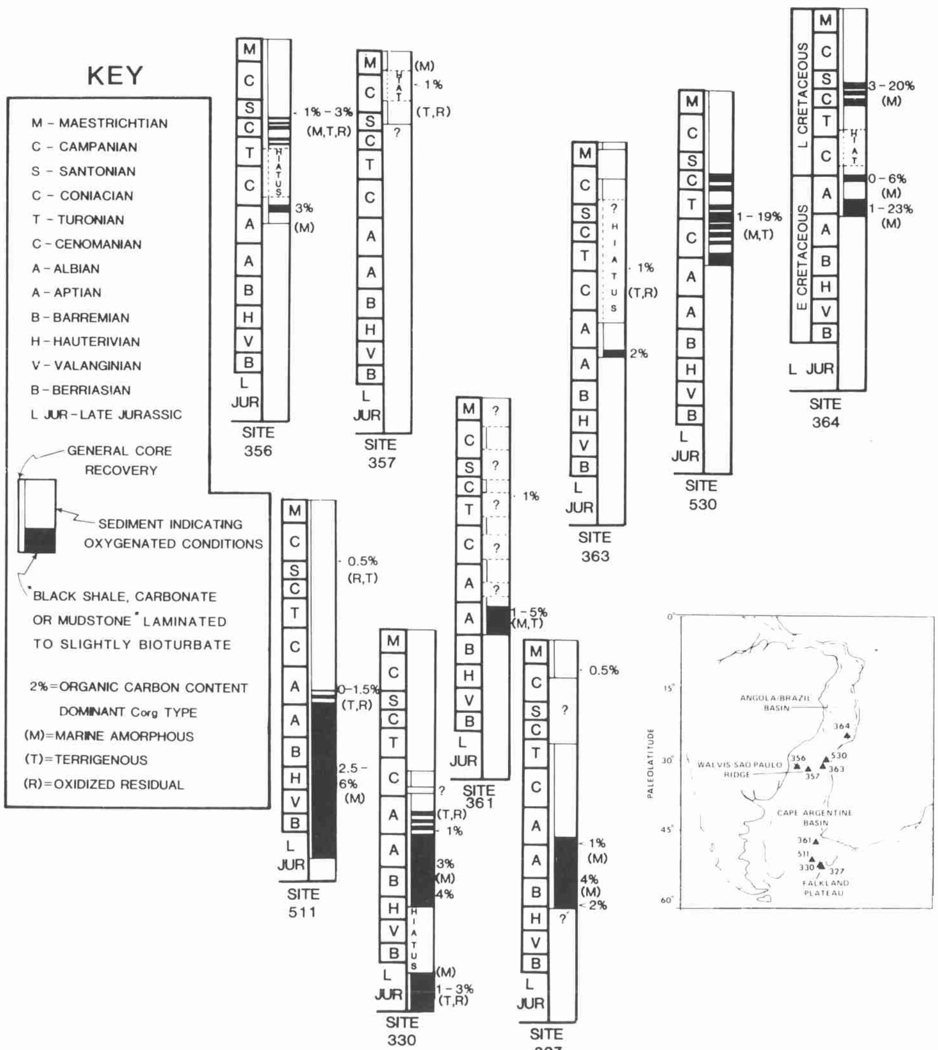


FIG. 7. Stratigraphic and core recovery columns for DSDP Sites from South Atlantic Ocean basins for the Mesozoic. Diagram illustrates intervals of 'black shale' deposition, generalized contents and types of organic matter.

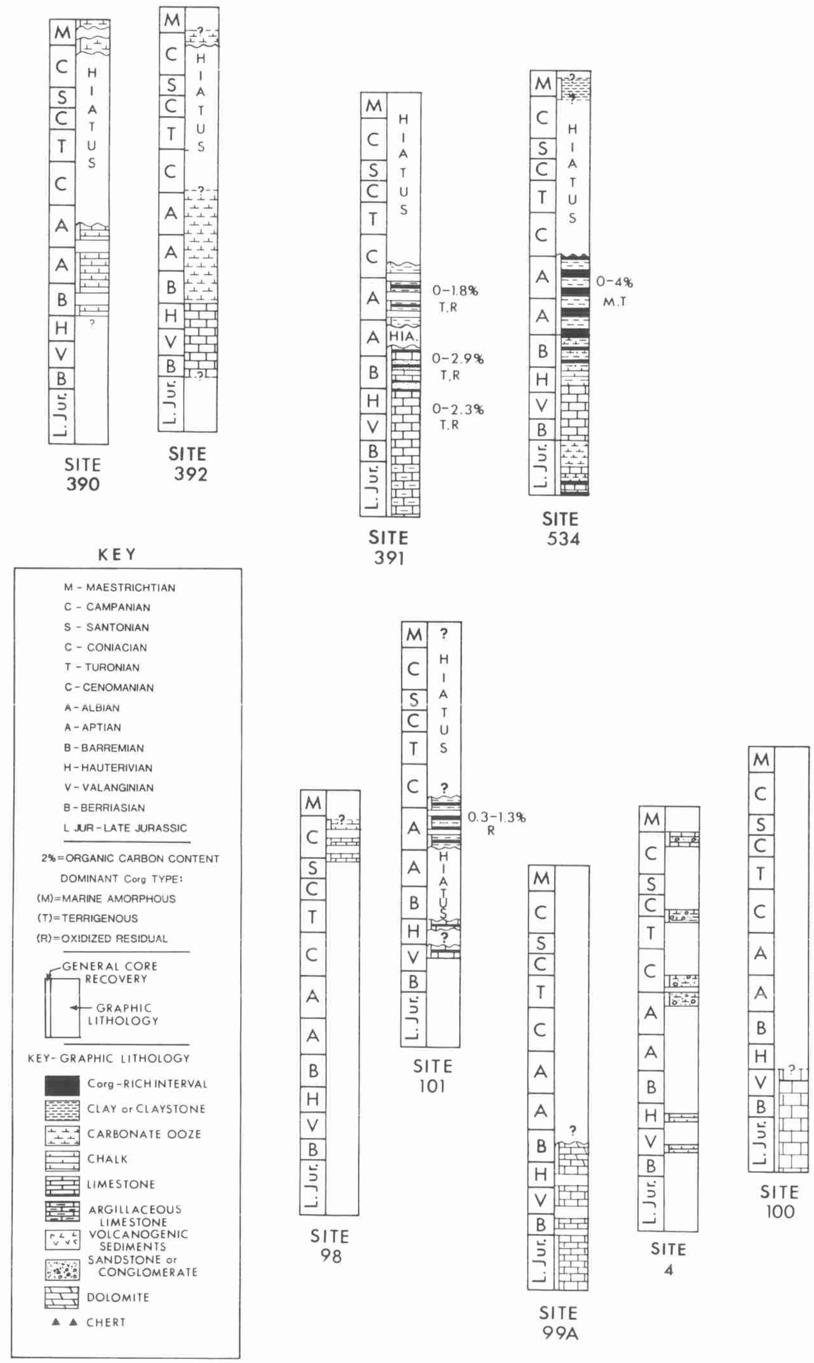
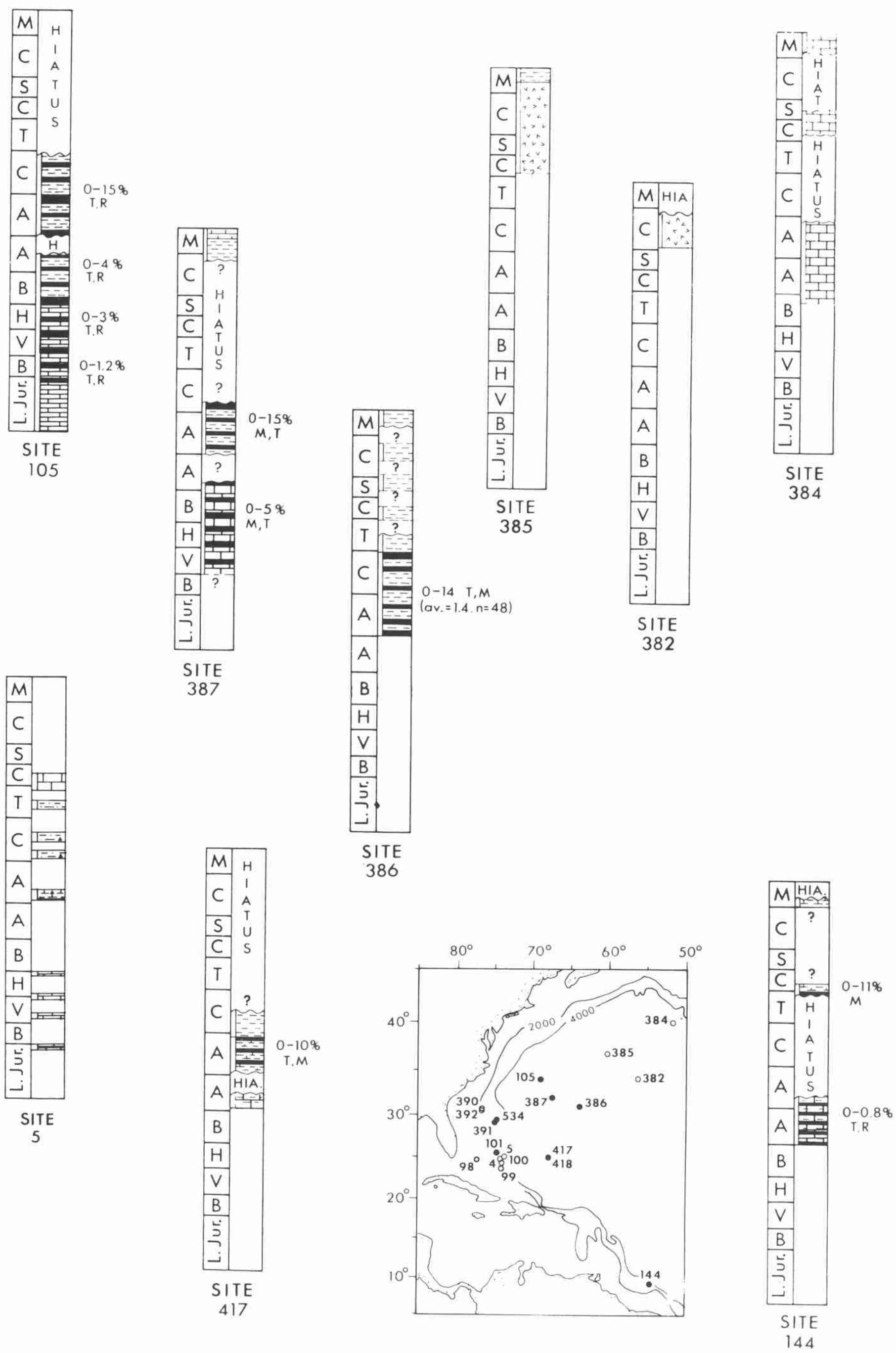


FIG. 8. Stratigraphic and core recovery columns for DSDP Sites from the western North Atlantic Ocean basin for the Mesozoic (see Key for explanation of symbols).



SITE 105
0-15% T,R
0-4% T,R
0-3% T,R
0-1.2% T,R
SITE 387
0-15% M,T
0-5% M,T
SITE 386
0-14 T,M (av.=1.4, n=48)
SITE 385
SITE 382
SITE 384
SITE 5
SITE 417
0-10% T,M
SITE 144
0-11% M
0-0.8% T,R
80° 70° 60° 50°
40° 30° 20° 10°
2000 4000
384 385 382 105 387 386 390 392 534 391 101 5 4 100 98 99 417 418 144

FIG. 8 (cont.)

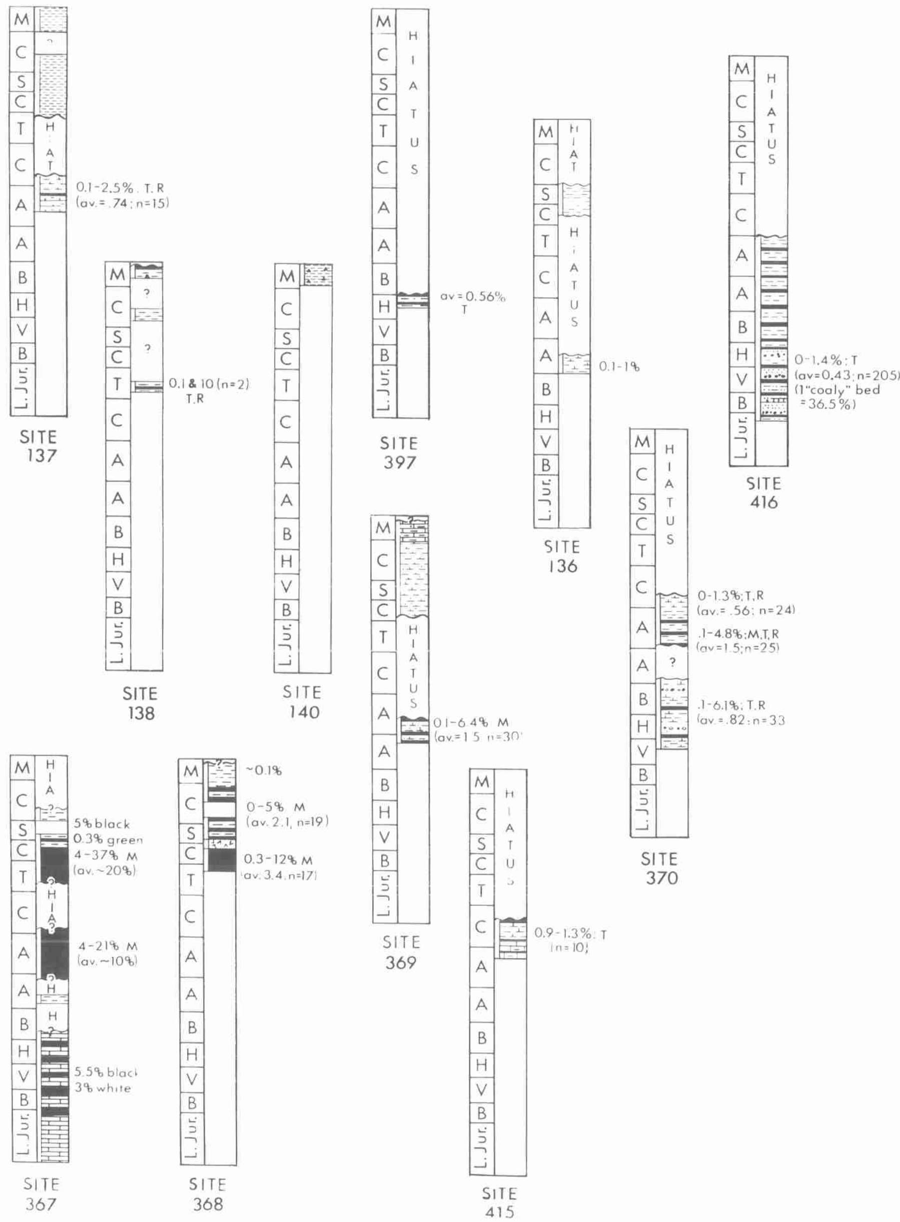


FIG. 9. Stratigraphic and core recovery columns for DSDP Sites from the eastern and northern North Atlantic Ocean basins for the Mesozoic (see Key on Fig. 8 for explanation of symbols).

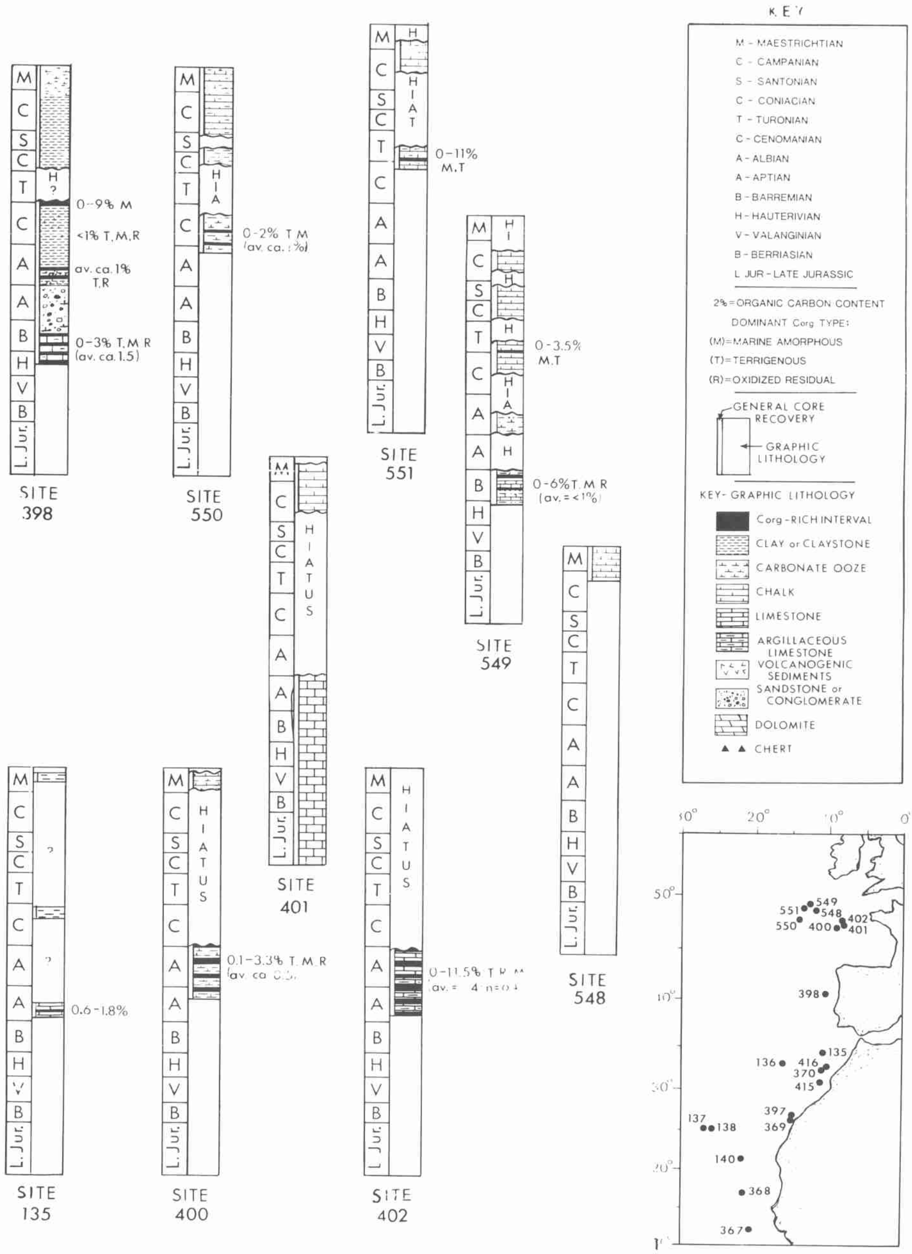

KEY
M - MAESTRICHTIAN
C - CAMPANIAN
S - SANTONIAN
C - CONIACIAN
T - TURONIAN
C - CENOMANIAN
A - ALBIAN
A - APTIAN
B - BARREMIAN
H - HAUTERIVIAN
V - VALANGINIAN
B - BERRIASIAN
L. JUR - LATE JURASSIC
2% = ORGANIC CARBON CONTENT
DOMINANT Corg TYPE:
(M) = MARINE AMORPHOUS
(T) = TERRIGENOUS
(R) = OXIDIZED RESIDUAL
GENERAL CORE RECOVERY
GRAPHIC LITHOLOGY
KEY- GRAPHIC LITHOLOGY
Corg - RICH INTERVAL
CLAY or CLAYSTONE
CARBONATE OOZE
CHALK
LIMESTONE
ARGILLACEOUS LIMESTONE
VOLCANOGENIC SEDIMENTS
SANDSTONE or CONGLOMERATE
DOLOMITE
CHERT
SITE 398
0–9% M
<1% T.M.R
av. ca. 1% T.R
0–3% T.M.R (av. ca. 1.5)
SITE 550
0-2% T.M
SITE 551
0–11% M.T
SITE 549
0-3.5% M.T
0-6% T.M.R
SITE 401
HIATUS
SITE 548
SITE 135
0.6–1.8%
SITE 400
HIATUS
0.1–3.3% T.M.R
SITE 402
HIATUS
0–11.5% T.R.M
549
551
548
550
400
402
401
398
135
136
416
370
415
397
369
137
138
140
368
367

FIG. 9 (cont.)

or other biogenic debris (e.g. fish debris) perhaps associated with detrital material which has a much higher rate of supply off Peru than off Namibia.

High-velocity currents are common in many upwelling regions (e.g. Huyer 1976; Summerhayes *et al.* 1979), and these currents may cause local winnowing or erosion of muddy sediment and organic matter and thereby reduce the rate of accumulation and/or content of organic carbon.

Primary productivity is higher (by a factor of 1.5 to 2) off NW Africa than in surface waters of the Black Sea, yet organic-carbon accumulation rates are significantly lower in sediments from the deeper NW African margin than in time-equivalent Black Sea strata. The suites of cores in each basin (Fig. 6) come from equivalent depths so that the difference in accumulation rates in large part must be related to levels of dissolved oxygen in bottom waters. Such an accumulation rate matrix (Fig. 6) when better defined and understood, may provide important constraints on interpretation of ancient 'black shales'.

Lithology and distribution of late Mesozoic deep-sea black shales

Jurassic-early Cretaceous

The first occurrences of organic-carbon-rich strata in the Atlantic Ocean basins that have been documented by DSDP drilling are of late Jurassic age. Callovian-Oxfordian black mudstones were recovered at Sites 330 and 511 on the Falkland Plateau (Fig. 7). Organic-carbon contents are as high as 4% (see review in Dean *et al.*, in press) and the organic matter is largely terrestrial or residual marine (Gilbert, in press; von der Dick *et al.* 1983). Deposition occurred under marginally oxic to occasionally anoxic water masses. Site 534 in the western North Atlantic (Fig. 8) also recovered Callovian-Oxfordian strata (Sheridan *et al.* 1982) which consist of interbedded greenish-black and brown radiolarian-rich claystone, and some interbedded grey limestone. Organic carbon contents are as high as about 4%, but most values are $<1\%$. The organic matter consists largely of terrestrial or oxidized (residual) marine material. Although anoxic conditions at the sea floor at Site 534 may have existed periodically as evidenced by finely laminated intervals, the main source of organic carbon was terrestrial (Summerhayes & Masran, 1983). The Callovian-Oxfordian has been suggested as an 'Oceanic Anoxic Event' by Jenkyns (1980) because organic-carbon-rich marine deposits of that age occur in Tethyan sequences and in other parts of Europe (Hallam & Bradshaw 1979; Morris 1979).

Interbeds of laminated dark-olive to black marlstone within bioturbated white to light-grey limestone of Tithonian to Neocomian age have been recovered from many deep-basin DSDP sites in the eastern and western basins of the North Atlantic (Dean *et al.* 1977; Jansa *et al.* 1979; Dean & Gardner 1982). The laminated dark marlstone beds generally contain more than 5% organic carbon and one bed at Site 367 in the Cape Verde Basin off north-west Africa (Fig. 9) contains 33% organic carbon (Dean *et al.* 1977). Turbidity currents were active in the deposition of at least some parts of this stratigraphic unit (Lancelot *et al.* 1972; Jansa *et al.* 1979). Jansa *et al.* (1979) suggested that this unit was deposited in an oxygenated bathyal environment, and that some of the lamination may represent reworking by bottom currents. Similar strata have been described from Tethyan localities as well (Bernoulli & Jenkyns 1974; Weissert *et al.* 1979; Arthur & Premoli Silva 1982).

The end of deposition of the Neocomian carbonates apparently was caused by a sudden rise in the carbonate compensation depth (CCD) during the Aptian (Arthur 1979; Thierstein 1979), and the remainder of the section in the North Atlantic through the Eocene is dominated by multicoloured clay-rich strata. At some DSDP sites in the North Atlantic, the Neocomian carbonates are overlain by the so-called middle Cretaceous (Aptian-Albian to Turonian) black-shale facies (Figs 8 and 9), but at some sites the two are separated by a unit of interbedded red and green claystones of late Aptian to early Albian age (Dean & Gardner 1982).

Middle Cretaceous

The middle Cretaceous black shale facies represents the main period of accumulation of organic matter in the Atlantic, and at most DSDP sites it consists of interbedded green and black clay-rich lithologies in which the black shale commonly accounts for less than 50% of the section (Figs 7–9; see reviews by Arthur 1979; Tucholke & Vogt 1979; McCave 1979b; Arthur & Natland 1979; Tissot *et al.* 1979,1980; de Graciansky *et al.* 1982; Dean & Gardner 1982; Dean *et al.*, in press). Concentrations of organic carbon in the black lithologies range from 2% to 37% and are commonly more than 5% (Lancelot, Seibold *et al.* 1977; Tissot *et al.* 1979, 1980; Summerhayes 1981b; Dean & Gardner 1982), whereas the interbedded green lithologies usually contain less than 0.5% organic carbon.

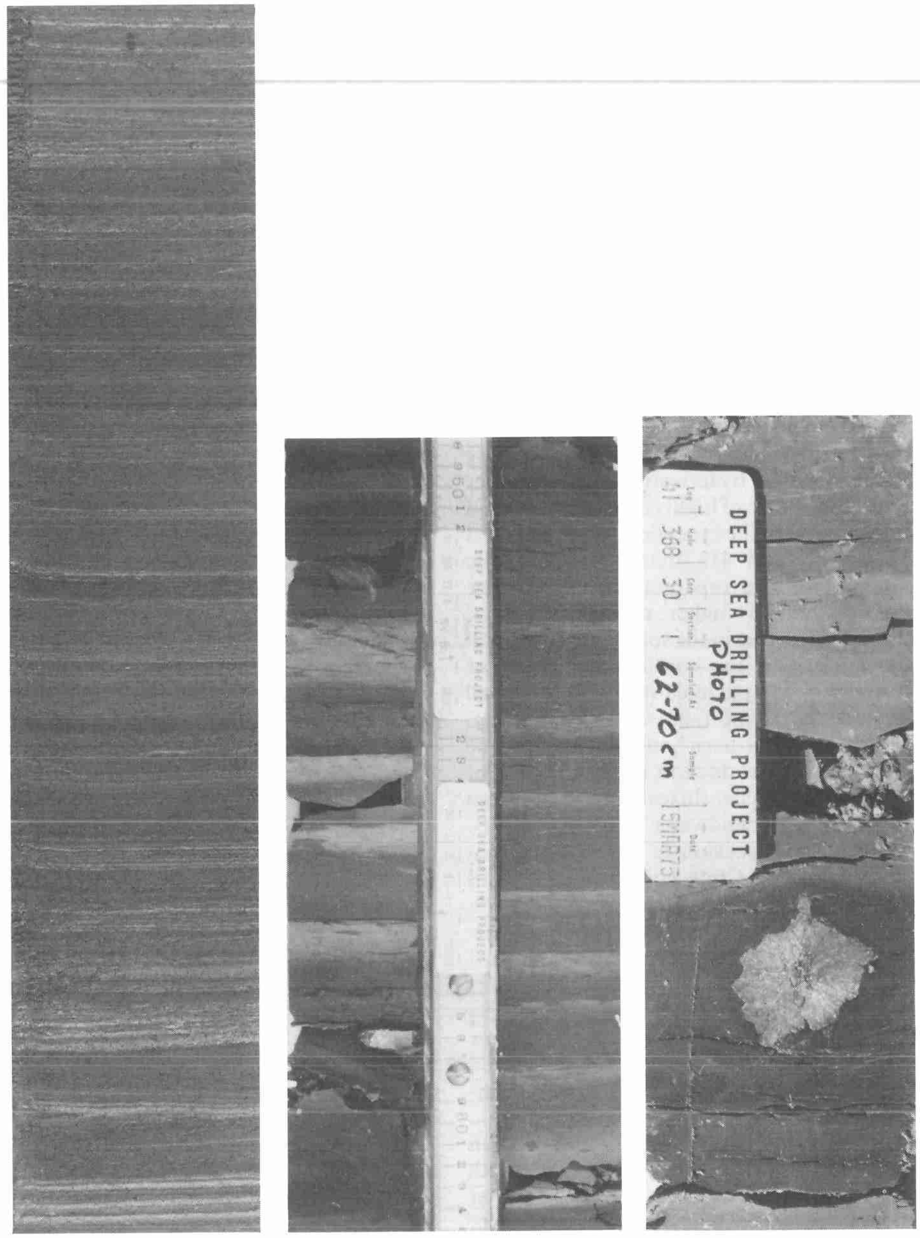
The type of organic matter in the black lithologies varies in both time and space (see Figs 7–9; Tissot *et al.* 1979, 1980; Habib 1979, 1982; de Graciansky *et al.* 1982; Summerhayes 1981b). Marine organic matter (i.e. organic matter characterized by type II kerogen of Tissot *et al.* 1974) for the most part only occurs in Aptian-Albian organic-carbon-rich strata in the South Atlantic (Fig. 7) and off the coast of north-west Africa (Fig. 9; Tissot *et al.* 1979, 1980; Summerhayes 1981b; de Graciansky *et al.* 1982). Everywhere else in the North Atlantic (Figs 8 and 9) the Aptian-Albian organic-carbon-rich strata are characterized mainly by terrestrial organic matter and some admixtures of highly degraded, residual organic matter with the exception of a few beds at Sites 386, 417, and 418 that contain marine organic matter. The predominance of marine organic matter and higher concentrations of organic carbon in sites in the south-eastern North Atlantic and along the continental margin of North Africa where palaeolatitudes were typical of trade wind circulation and coastal upwelling, and at Site 144 off northern South America, is due to higher rates of production of organic matter and consequently higher fluxes to the sea bottom and more intense midwater oxygen deficits. Numerous authors previously have pointed out this relationship for Cretaceous examples (Einsele & Wiedmann 1975; Arthur & Natland 1979; Tissot *et al.* 1979, 1980; Jenkyns 1980; Summerhayes 1981b; Parrish & Curtis 1982). In spite of these differences in sources of organic matter, most of the organic-carbon-rich strata are interbedded with bioturbated, organic-carbon-poor strata with abundant evidence of having been deposited in an oxygenated bottom water environment.

Cenomanian to possibly earliest Turonian black shales in nearly all North and South Atlantic basin sites (see Figs 7–9) contain higher amounts of organic carbon (up to 30%) much of which is amorphous marine material (Tissot *et al.* 1979, 1980; Summerhayes 1981b). Strata of this age may record an oceanwide productivity/preservation event (Schlanger & Jenkyns 1976; Arthur & Schlanger 1979; Jenkyns 1980; Scholle & Arthur 1980; Summerhayes 1981b). The Cenomanian-Turonian organic carbon burial episode is the best evidence for what is termed an 'oceanic anoxic event' (Schlanger & Jenkyns 1976) because it is both widespread and fairly brief in duration in comparison to the widespread but much longer period of deposition of Barremian-Albian 'black shales' (see later discussion).

Many middle Cretaceous black shales have very fine fissile lamination, and commonly have been interpreted as pelagic sediments that have accumulated under anoxic conditions. However, detailed examination (McCave 1979b; Stow & Dean, in press) reveals a greater variety of sedimentary structures in many intervals including horizontal silt-mud lamination, micro-cross lamination, small-scale grading and micro-bioturbation (Fig. 10). The interbedded green and red claystones commonly have a greater degree and a larger scale of bioturbation which has obliterated many of the same primary sedimentary structures as found in the black shales. It appears that both turbiditic and pelagic processes have operated throughout the deposition of most black shales sequences and are not specific to any one lithology.

In the western North Atlantic, mid-Cretaceous black shales were recovered at Sites 101 and 105 (Lancelot *et al.* 1972), 386 and 387 (McCave 1979b), 391 (Benson *et al.* 1978), and 417 and 418 (Donnelly *et al.* 1980) and 534 (Sheridan *et al.* 1982) (Fig. 8). These are interbedded with red, green and grey mudstones and marlstones and thin radiolarian-rich sand/silt layers, and commonly comprise less than 50% of the section (Figs 8 and 10). The black shales may be laminated, massive, or bioturbated and are variably enriched in organic carbon from mixed terrigenous and marine sources, although much of the marine organic carbon is degraded. Fine-grained turbidites comprise parts (mainly less than 30%) of the mid-Cretaceous section at each site, and occur in both the black shales and the associated lithologies (Fig. 10). There are thin-bedded laminated siltstone and mudstone turbidites, mostly far-travelled from the North American continental margin, as well as thicker-bedded biogenic turbidites, probably derived locally from intrabasin highs.

In the eastern North Atlantic off the Euro-African continental margin, mid-Cretaceous black shales (Fig. 9) that showed evidence of turbidite deposition were recovered at Sites 367, 368 and 370 (Dean *et al.* 1977; Dean & Gardner 1982), 397 (Cornford 1979), 398 (Arthur 1979; de Graciansky & Chenet 1979; Habib 1979), 400 and 402 (de Graciansky *et al.* 1979) and 415 (Lancelot *et al.* 1980). The black shales again form part (usually less than 50%) of a more oxic section comprising varicoloured mudstones and marlstones. Most of the turbidites recognized are also fine-grained siltstone or mudstone, although at Sites 370, 397, 398 and 402 there are interbedded coarser-grained and thicker-bedded turbidites, debris-flow deposits and slumped units (Fig. 10) including redeposited shallow-water carbonate material. There is no consistent association of

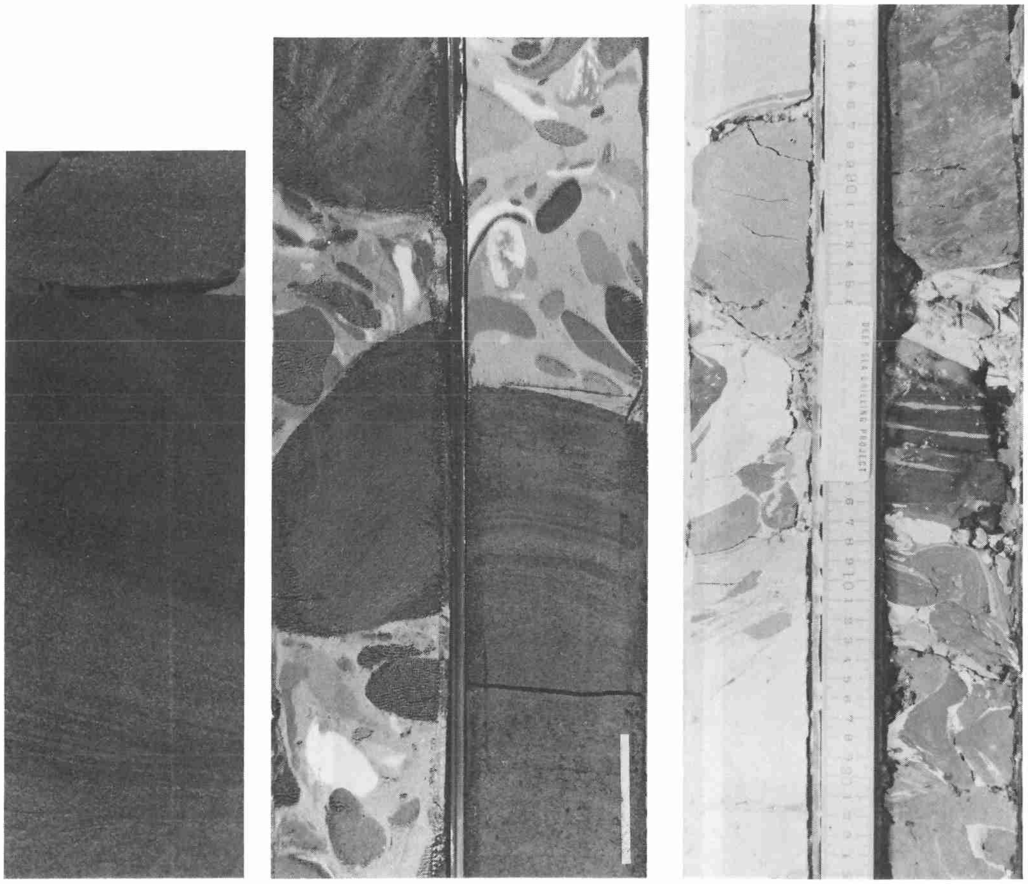


(a) (b) (c)

FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(a) 64–480, P20–3, 30–60 cm: Guaymas Basin slope, Gulf of California. Hydraulic piston core interval in finely laminated muddy diatom ooze. The white laminae are richer in diatoms, and the light-dark couplets represent annual varves. Note the even, planar nature of the laminations and general lack of irregular or cross-cutting laminations. Such continuous fine lamination is typical under an oxygen-minimum zone of relatively high sedimentation rate settings with seasonal sedimentation contrasts. Some uncomformities and apparent cross-bedding, however, were noted in these cores (see Schrader *et al.* 1981).

(b) 41–368, 37–3, 4, 47–85 cm, and (c) 41–368, 30–1, 62–70 cm: upper Palaeocene to lower Eocene, Cape Verde Rise eastern North Atlantic. Dark intervals in dark-light intercalations are mud turbidites (see Dean *et al.* 1977). Note sharp basal contacts and general lack of burrowing at base of each unit. Here the mud turbidites are black to dark green and dark olive-green and contain as much as 3% organic carbon, but usually less than 1% organic carbon. Scale is in centimetres. Photograph in (c) is a closeup of a mud turbidite containing a pyrite nodule.



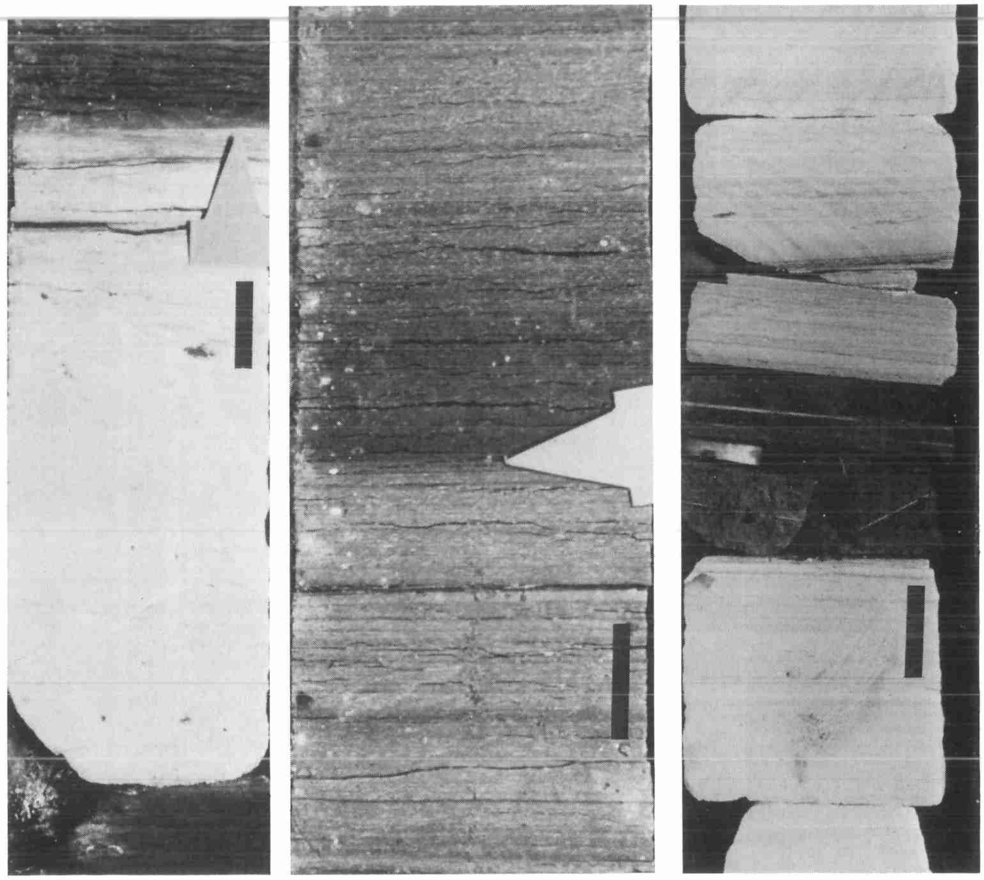
(d) (e) (f)

FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(d) 62–465A, 40–1, 88–107 cm: Hess Rise, Pacific Ocean. Laminated, olive, organic-carbon-rich limestone of late Albian age. Similar beds in sequence contain up to 9% organic carbon. The limestone contains radiolarians and organic carbon with a high hydrogen index. Although high productivity may have characterized surface waters, and bottom waters may have been anoxic, these laminated limestones probably were redeposited, as shown by cross-bedding in photo (see Dean *et al.* 1981). Bar is 3 cm.

(e) 75–530B, 8–1 and 2, 19–51 cm: Southern Angola Basin adjacent to Walvis Ridge, South Atlantic. These Pleistocene sediments contain up to 8% organic carbon, but were largely redeposited by debris-flows to deep water from highly productive shallower water settings on the nearby Walvis Ridge. Conglomeratic character of the soft mudclasts demonstrates redeposition, but some intervals are dark, fine-grained and laminated (note lower right: see Stow 1983). Similar debris-flow deposits at Site 530 also occur in the Pliocene and Miocene sections. Bar is 5 cm.

(f) 41–369A, 40–2 and 3, 81–125 cm: Continental slope off Cape Bojador, north-west Africa. This Santonian continental slope sequence of argillaceous chalks probably was deposited within an oxygen-minimum zone. Some intervals are laminated, alternating with bioturbated zones. Organic carbon contents are as high as 6.4% and average about 1.5%. Much of the sequence, like the strata illustrated here, has been redeposited by slumps and debris-flows. Scale is in centimetres.

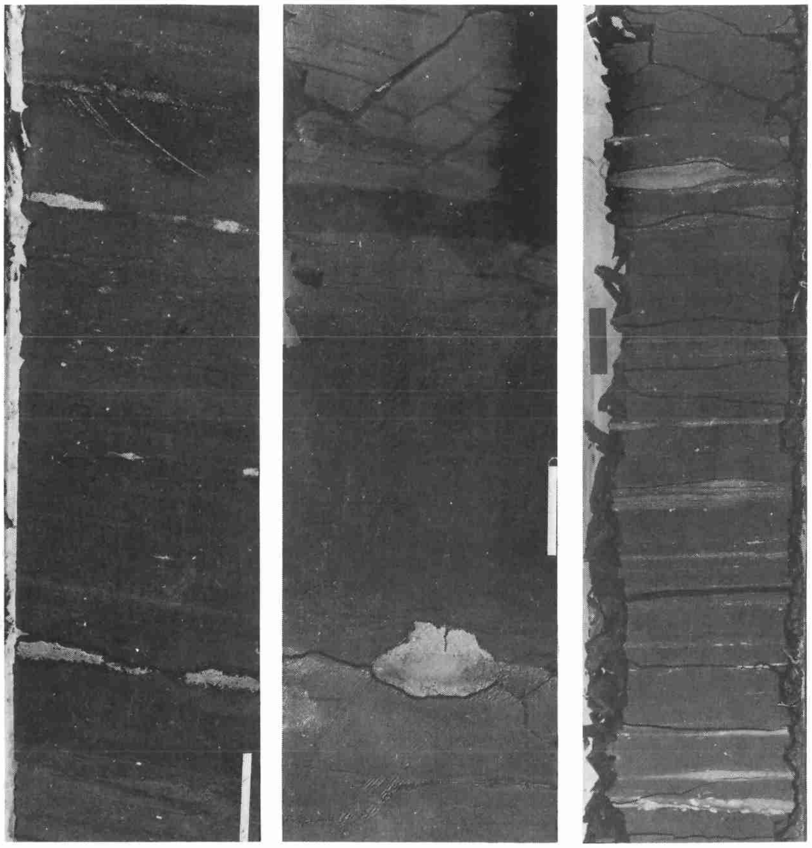


(g) (h) (i)

FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(g) 41–367, 28–3, 41–62 cm, and (h) 411–3367, 28–3, 64–79 cm: Cape Verde Basin, eastern North Atlantic. These photos are typical examples of interbedded light grey limestones and dark olive-grey to black marlstones found in lower Cretaceous (Neocomian) sequences on both sides of the Atlantic. The dark beds typically are finely laminated (note discontinuous light-coloured, coccolith-rich laminae), whereas adjacent light grey limestones are highly bioturbated. The laminated marlstones typically contain up to 3 or 4% organic carbon of marine origin. Such sequences constitute the primary evidence that the deep Atlantic was at least periodically anoxic at or above the sea floor. Bars in both photographs are 2 cm.

(i) 43–387, 49–5, 100–119 cm: western North Atlantic basin. Cycles of Neocomian marlstone-limestone similar to those in (g) and (h), but in this sequence the light-coloured limestones also are finely laminated. However, very faint burrow mottles do occur in the limestones. Although the laminations may indicate deposition under anoxic conditions, they may also be due to the action of bottom currents. High sedimentation rates, therefore, may have limited benthic activity. Bar is 2 cm.



(j) (k) (l)

FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(j) 75530-A, 98–2, 85–105 cm and (k) 75–530A, 104–3, 90–105 cm: Angola Basin adjacent to Walvis Ridge, South Atlantic. Examples of Albian black shale or mudstone facies which are homogeneous to faintly laminated and contain up to 15% organic carbon, largely of marine derivation. Numerous thin dark layers with sharp basal contacts and bioturbated upper parts are evident in photograph (j). These may represent thin redeposited units (perhaps from adjacent Walvis Ridge). A thin bioturbated zone in photograph (k) passes through the middle of the black unit suggesting that this unit may be composed of more than one 'event'. Note also pyrite (light-coloured) at boundaries of oxidized and reduced lithologies in both photographs. Bars in both photographs are 2 cm.

(l) 47B-398D, 77–3, 42–67 cm: Vigo Seamount, north-east Atlantic. Middle Albian black mudstone unit containing about 1% organic carbon, mainly of terrestrial or residual origin. Note the light-coloured layers which contain concentrations of calcareous nannofossils and siderite. Much of this unit may have been redeposited (see Arthur 1979). Bar is 2 cm.

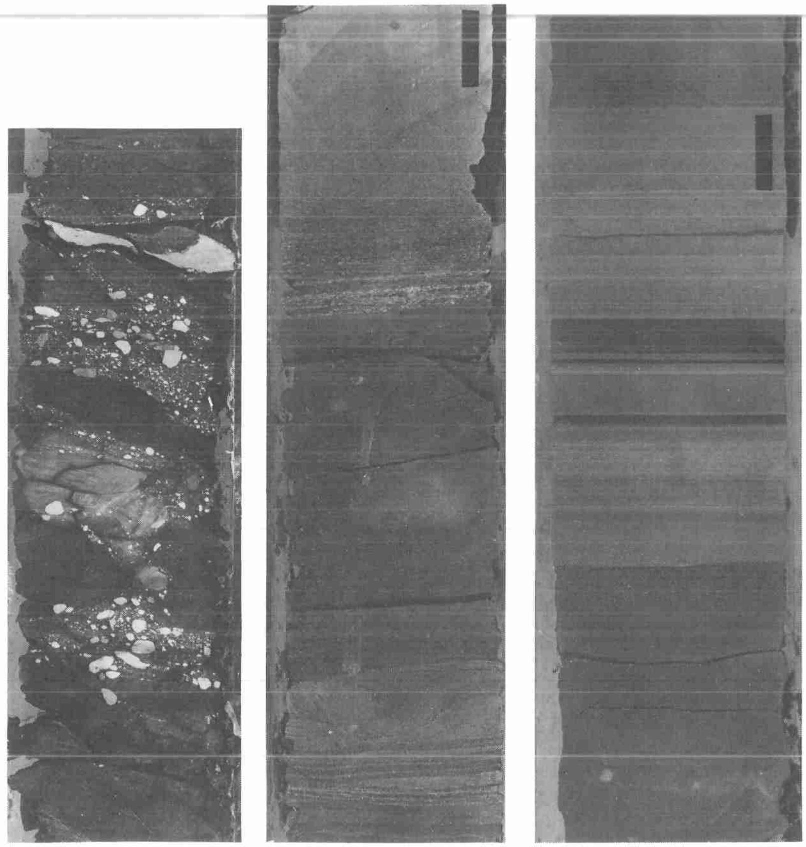
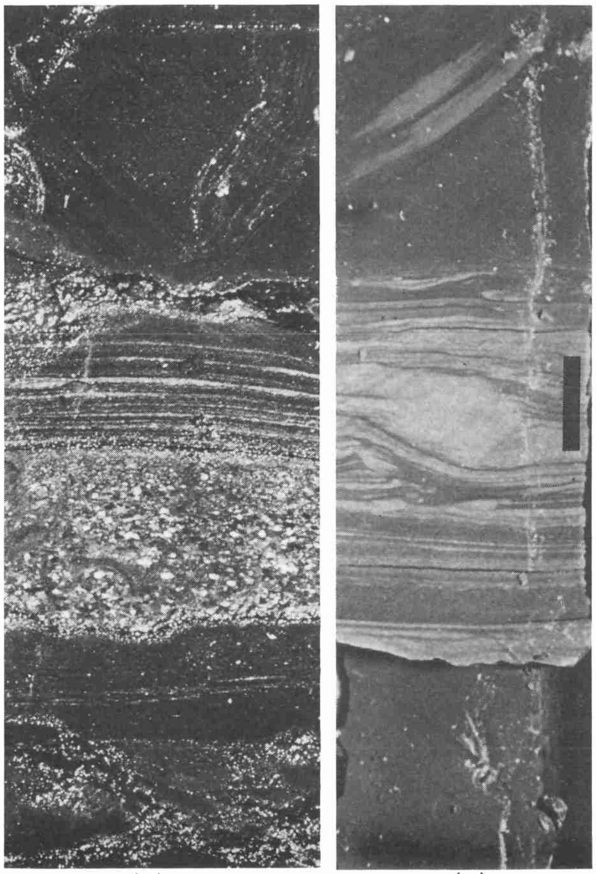


FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(m) 47B–398D, 116–1, 97–118 cm: Vigo Seamount, north-east Atlantic. This unit of upper Aptian black to dark grey mudstone is similar to (l) above, except that there is abundant evidence of redeposition of terrigenous sand and carbonate material, including large slump units just above. Bar is 2 cm.

(n) 47B–398D, 123–4, 0–24 cm: Vigo Seamount, north-east Atlantic. Dark grey to black homogeneous to laminated mudstone and calcareous mudstone of early Aptian age contains up to 2.5% organic carbon, largely of terrestrial and/or residual marine origin. Terrigenous sand/silt turbidites are common, and it is likely that much, if not all of this 'black shale' unit was redeposited. Bar is 2 cm.

(o) 47B–398D, 127–4, 84–105 cm: Vigo Seamount, north-east Atlantic. Black mudstones of late Barremian to early Aptian age in this interval contain up to 2% organic carbon, mainly of terrestrial origin. Note the thin-bedded, fine-grained turbidities that are totally unbioturbated. Fine laminations are exceptionally well-preserved. Either bottom waters were anoxic or high sedimentation rates inhibited a benthic fauna. Bar is 2 cm.



(p) (q)

FIG. 10. Examples of redeposited sediments and sedimentary structures associated with 'black shale' facies in DSDP drillholes.

(p) 41–370, 35–5, 5–16 cm: Moroccan Basin, eastern North Atlantic. Dark intervals within Barremian turbidites contain an average of about 1% organic carbon, mainly of terrestrial origin. Redeposited intervals dominate the entire sequence recovered in the Moroccan Basin. The turbidite shown here contains glauconite, dolomitized ooids and other shelf carbonate debris. Bar is 1 cm.

(q) 41–370, 41–3, 28–46 cm: Moroccan Basin, eastern North Atlantic. Barremian terrigenous silt/sand turbidite similar to photograph (n). Bar is 2 cm.

turbidites with any one particular facies. For example, at Site 398 thin black mudstone turbidites grade upwards into biogenic-rich pelagites, whereas at Site 400 biogenic calcilutite turbidites grade up into black shales, parts of which may represent pelagic sedimentation.

Of the eight DSDP Sites in the South Atlantic (Fig. 7; see Dean *et al.*, in press, for review) that have recovered mid-Cretaceous black shales, at least three show clear evidence of interbedded turbidites. At Site 361 in the Cape Basin (Natland 1978; Arthur & Natland 1979) thick sandstone and pebbly sandstone turbidites and slumps occur within a sequence of green, grey and black mudstones. The coarse-grained turbidites commonly are rich in terrestrial organic debris and grade upward into thinly laminated black shales some of which have more marine organic matter. Both pelagite and turbidite interpretations of the black shales at Site 361 are possible. The black shales and associated facies in the Angola Basin at Sites 364 (Bolli *et al.* 1978) and 530 (Stow & Dean, in press) have a much more distal or basinal character. Black shale beds comprise from 5 to 50% of the section at Site 530, and occur as numerous thin beds with variable organic carbon content that is dominantly of marine origin (Meyers *et al.*, in press). Very fine-grained, thin-bedded turbidites comprise between 5 and 20% of the section at Site 530 and occur both in the black shales and in the associated facies.

The few occurrences of middle Cretaceous organic-carbon-rich strata in the Pacific, except those on southern Hess Rise (Sites 465 and 466) appear to have been caused by short-lived events of oxygen depletion that were associated with volcanic activity and downslope sediment redeposition (Dean *et al.* 1981; Thiede *et al.* 1982). These strata appear to be the result of a coincidence of local tectonic factors acting to produce and preserve organic matter at times that were not strictly synchronous but were within the same general middle Cretaceous time period as Atlantic-Tethyan black shale occurrences. The organic-carbon-rich strata on Hess Rise also are of the same general age and are associated with volcanic activity and sediment redeposition but cover a longer time span (late Albian-early Turonian) as Hess Rise crossed under the highly productive equatorial divergence. At all Pacific sites, organic-carbon-rich strata probably accumulated on the flanks of volcanic edifaces within oxygen-minimum zones in relatively shallow water (several hundred metres).

An increase in organic carbon concentrations also occurs within Coniacian-Santonian sequences at DSDP Sites in the eastern South Atlantic (Fig. 7; the so-called sapropels of Ryan & Cita 1977; McCoy & Zimmerman 1977), the eastern North Atlantic and the Caribbean (Arthur & Natland 1979). Many of these organic-carbon-rich beds may be the result of higher rates of supply of organic matter related to upwelling and/or enhanced preservation under marginally oxic conditions. Redeposition from adjacent continental margins, however, cannot be ruled out as a cause. Organic-carbon-rich strata at Sites 369 and 144 may have been deposited within expanded and intensified oxygen-minimum zones associated with coastal divergences and upwelling.

Cenozoic

Eocene sediments in both the eastern and western basins of the North Atlantic are relatively enriched in organic carbon. Dean *et al.* (1977) and Dean & Gardner (1982) described rhythmically interbedded black and green radiolarian clay layers of Eocene age from Site 367 in the Cape Verde Basin in the eastern North Atlantic (Fig. 10). The black clay layers contain up to 4% organic carbon of marine origin. These beds commonly have sharp basal contacts with silt stringers and bioturbated tops, and are best interpreted as mud turbidites. McCave (1979a) proposed a similar interpretation for dark siliceous Eocene mudstones in the western North Atlantic (Site 386) that contain up to 2% organic carbon.

More than 1000 m of lower to middle Miocene sediment at Site 397 on the Cape Bajador upper continental rise was redeposited from the outer shelf and upper continental slope along the continental margin of north-west Africa (Arthur & von Rad 1979; Arthur *et al.* 1979). Dark-coloured clays at Site 397, interpreted as mud turbidites, and the matrix and clasts of thick debris-flow deposits contain up to 7% organic carbon of marine origin (Cornford 1979) (Fig. 10). Benthic foraminifera and sedimentary structures (Arthur & von Rad 1979) all indicate downslope redeposition of these sediments.

Organic-carbon-rich sediments in the south-eastern Atlantic off south-west Africa (Calvert & Price 1971) extend over a wide area to depths of at least 1330 m, and over a time-span of late Miocene to Recent (Site 362, Bolli *et al.* 1978; Site 532, Hay *et al.* 1982; Gardner *et al.*, in press). At Site 532 the concentration of organic carbon of marine origin is as much as 8% in the dark green and olive siliceous clays of late Pliocene to early Pleistocene age, and less in the interbedded calcareous oozes, marls and muds (Gardner *et al.*, in press; Meyers *et al.*, in press). All the lithologies are pelagic or hemipelagic in origin, are thoroughly bioturbated and accumulated relatively rapidly at rates of 2.5 to 6 cm/1000 yrs under an oxygen-minimum zone.

The late Miocene to Recent section at Site 530 in the Angola Basin, immediately north of the Walvis Ridge in 4650 m water depth also is high in organic carbon with a maximum of 6% in lower Pleistocene sediments (Meyers *et al.*, in press). The organic-carbon-rich facies corresponds to that on Walvis Ridge except that they are all redeposited as thick-bedded mud-ooze turbidites and debrites (Fig. 10; Stow, in press). The Walvis Ridge and Angola Basin sediments are mostly well-bioturbated, non-laminated and dark greenish-olive in colour, but with subsequent burial to greater depth they may come to more closely resemble the Mesozoic black shales in appearance.

Neogene black shales or sapropels are common in the Mediterranean and Black Seas, and they have been interpreted, in part, as being of turbidite origin (e.g. Sites 378 and 380, Hsü *et al.* 1978; Ross *et al.* 1978). DSDP drilling off the coast of southern California and in the Gulf of California recovered relatively organic-carbon-rich sequences of Miocene to Recent age (e.g. Schrader *et al.* 1980; Summerhayes 1981b; Gilbert & Summerhayes 1981, 1982). Most of these sequences contain marine organic matter or mixtures of marine and terrestrial organic matter, and represent deposition associated with upwelling and preservation of organic matter under intense oxygen-minimum zones. They are younger analogues or offshore equivalents of the Miocene Monterey Formation (Pisciotto & Garrison 1981).

Discussion

By analogy with modern examples of organic-carbon-rich sediments we can assume that three main variables were significant for the preservation of organic matter in ancient black shales. These were:

(1) variation in the supply of both marine and terrigenous organic matter from surface productivity (pelagic settling, faecal pellets), fluvial discharge, redepositional processes and settling from slope and bottom nepheloid-layers;
(2) variation in the rate of sedimentation and hence rate of burial of organic matter;
(3) variation in bottom water oxygenation within an oxygen-minimum zone or throughout the basin as a result of oceanographic and climatic conditions.

Various authors have usually stressed the particular significance of one or another of these variables from study of a black shale sequence in a particular locality or region, and have, to some extent, presented models that were too generalized or simplified.

Basin deoxygenation and expansion and intensification of an oxygen-minimum layer have been proposed to explain the accumulation of the middle Cretaceous organic-carbon-rich strata (e.g. Schlanger & Jenkyns 1976; Ryan & Cita 1977; Fischer & Arthur 1977; Thiede & van Andel 1977; Arthur & Schlanger 1979). Both models require very low to zero concentrations of dissolved oxygen in part of the water column as a result of reduced advection of oxygenated water and/or increased supply of organic matter. Both models imply that reducing conditions in the sediments (and therefore the increased degree of preservation of organic matter) are largely the result of anoxic or near-anoxic conditions in the overlying waters. On the other hand, the importance of supply factors, namely increased productivity, terrigenous runoff and/or redeposition of organic carbon by turbidity currents, have been stressed by Dean *et al.* (1977), Cornford (1979), de Graciansky *et al.* (1979), Habib (1979, 1982), Welte *et al.* (1979) and Natland (1978) for more specific areas off north-west Europe and north-west and south-west Africa. Settling of fine-grained terrigenous organic matter from possible bottom nepheloid-layers also has been proposed as a mechanism of transporting organic carbon to deep-sea environments (Summerhayes 1981b).

However, it seems clear to us that there are important differences between black shales of different ages and places, and even between different beds within the same sequence. No simple model therefore will be adequate to explain what is in fact a very complex depositional system, and we must appeal to a more complex interplay of controls each of which may have varied independently to some extent.

The chief characteristics of ancient deep-water black shales that must be explained by any model are the following:

(1) the approximate coincidence in time of the main periods of black shale accumulation in many ocean basins within the Jurassic, Cretaceous and Cenozoic;
(2) the range of environments in which black shales were deposited, from open ocean to restricted basin, and from hundreds of metres to a few kilometres water depth;
(3) the formation of many black shales under poorly-oxygenated waters as evidenced by bioturbation characteristics in Cretaceous sequences, and in fully-oxygenated ocean

basins such as those deposited during the Eocene and Miocene;

(4) the interbedding of organic-carbon-rich facies with organic-carbon-poor facies, in cycles with irregular periodicities that average 20 000 to 140 000 yrs, (Dean *et al.* 1977; McCave 1979b; Arthur 1979; Dean *et al.*, in press), and with a variable order of occurrence of the different facies (McCave 1979b; Stow & Dean, in press);

(5) the variation in lithologies that are enriched in organic-carbon, including mudstones, marlstones, limestones, sandstones and diatomites, all of which have been loosely referred to as 'black shales' (e.g. See Ibach 1982);

(6) the range of processes that have been responsible for the deposition of different black shales, including pelagic, hemipelagic (e.g. settling from nepheloid-layer) turbiditic and other mass-flow processes, and the fact that no *one* process is uniquely associated with any *one* of the associated facies (Stow & Dean, in press);

(7) the different types of organic matter and amounts of total organic carbon in black shales even within the same sequence.

Factors in the origin of Cretaceous black shales

To explain these characteristics for the middle Cretaceous black shales that are so widespread throughout the Atlantic and Tethyan ocean as well as on isolated plateaus and seamounts in the Pacific, we must consider the worldwide conditions of climate and ocean circulation that probably were influential.

Global climate during the middle Cretaceous was warm, eustatic sea-levels were generally high, spreading rates were fast, pelagic sediments in the world ocean were accumulating rapidly, and oceanic surface and bottom water temperatures were high (Douglas & Savin 1975; Fischer & Arthur 1977; Berger 1979; Brass *et al.* 1982). Increased surface water temperatures would have had two main effects: first, thermohaline deep-water circulation, driven today by the sinking of cold, oxygen-rich surface waters in high latitudes, would have been less efficient and driven mainly by salinity differences; and second, warm, saline water, which would contain lower initial concentrations of dissolved oxygen (Fig. 3), and was created on restricted, evaporative low latitude shelves and in semi-isolated basins would become the main deep-water mass in the ocean (Roth 1978; Arthur & Natland, 1979; Wilde & Berry 1982; Brass *et al.* 1982). It is not clear what the relative rates of deep-water overturn would be under such conditions, although Brass *et al.* (1982) argued that turnover rates could be as high or higher than those of today, and Barron & Washington (1982) suggested that oceanic circulation on the whole was not 'sluggish' as is commonly assumed because of warmer surface and bottom waters during the Cretaceous. In any case, bottom water circulation was sufficient to supply some oxygen to maintain oxidizing conditions in the deep basins of the Pacific but oxygen supply was inefficient enough to permit depletion of dissolved oxygen over a broad range of midwater depths in areas of high productivity of organic matter, possibly because of lower initial dissolved oxygen contents in intermediate and deep-water masses due to the higher temperature and salinity. Even in the Angola Basin, which was highly restricted during the middle Cretaceous, circulation was sufficient to provide oxygenated bottom waters most of the time.

The accumulation of organic-carbon-rich sediments at many places in the world ocean at times that were not always strictly synchronous undoubtedly is the result of a coincidence of several factors acting to produce and preserve organic matter. During the middle Cretaceous, much of the world ocean may have been so poised that relatively small changes in the flux of organic matter and/or rates of deep-water circulation at any one place may have caused anoxia or near-anoxia within midwater oxygen-minimum zones and possibly throughout much of the bottom water mass under extreme conditions in more tectonically restricted basins.

An expanded and intensified oxygen-minimum zone during much of the early and middle Cretaceous would explain the excellent preservation of organic carbon, and the increase in accumulation rate of organic carbon over much larger areas in continental slope environments, on flanks of high-standing features on the sea floor, and in deep-sea settings largely by redeposition. It is not inconceivable that intensified oxygen-minimum zones periodically extended as deep as 2500 to 3000 m; this would help to explain the presence of upper Aptian-lower Albian black shales containing well-preserved organic carbon on or near the crest of the Mid-Atlantic Ridge in the North Atlantic (Site 386, Tissot *et al.* 1980; Summerhayes 1981b; Sites 417 and 418, Deroo *et al.* 1980) where it is unlikely that continental-margin-derived turbidites could have been the supply of the organic matter. Oxygen-minimum zones probably were the most expanded and intensified in areas of pronounced upwelling and high productivity, such as off north-west Africa (e.g. Berger & von Rad 1972; Arthur & Natland 1979; Thierstein 1979; Fig. 9), which would explain the higher overall contents, accumulation rates, and more marine character of organic matter in mid-Cretaceous sediments from the eastern North Atlantic basin. This explanation also probably accounts for equatorial Pacific mid-Cretaceous black shales that accumulated on the flanks of high-standing seamounts and plateaus. In the south-east Angola Basin, burrows commonly become smaller and less abundant going from a green claystone to an overlying black shale bed (DSDP Site 530; Stow & Dean, in press), which suggests that periodic deep-basin oxygen-deficient conditions probably occurred there.

However, although bottom water anoxia or near anoxia may have aided in the preservation of organic matter it was not necessarily the only cause for accumulation of organic-carbon-rich strata. An increase in the relative amount of organic debris deposited was equally important. At different sites, there were differences in the relative amounts of organic debris derived from areas of increased productivity within the basin and preserved in sediments under oxygen-minima on continental margins and organic debris from areas with increased terrigenous supply. In many places, organic matter was redeposited to basinal sites by turbidity currents; in other places, there was relatively rapid accumulation of pelagic organic matter. The main variables that may have

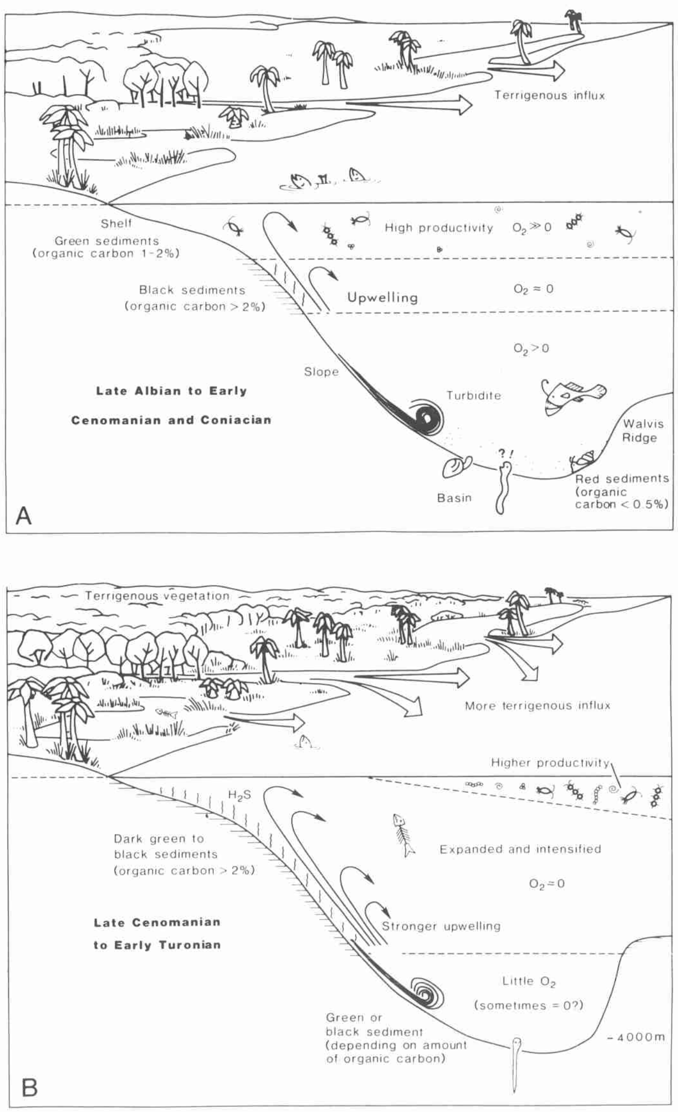


FIG. 11. Cartoons illustrating the interplay between organic carbon fluxes from terrigenous sources and from surface water productivity, and changes in their relative importance in sedimentary deposits as the result of changes in climate, sea-level, upwelling and oceanic fertility, and deep-water oxygen contents. Also emphasized are pathways for organic matter deposition, including by downslope movement (e.g. turbidites) from river mouths (terrigenous organic matter) or from within oxygen-minimum zones (well-preserved marine organic matter), pelagic settling (e.g. faecal pellets and/or marine snow), and settling from nepheloid-layers (both marine and terrigenous organic matter). (A) Late Albian-early Cenomanian or Coniacian-Santonian in North and South Atlantic: low to moderate productivity with 'normal' mid-water oxygen-minimum zone, oxygenated (i.e. > 1.0 ml/l) deep-water masses. (B) Late Cenomanian-early Turonian or possibly middle Aptian-middle Albian in South and North Atlantic. Higher rate of supply of organic carbon and greater extent of oxygen depletion in deep-water masses (after Dean *et al.*, in press).

operated during a period of time when less organic matter was accumulating and during a period of time when more organic matter was accumulating are summarized in Fig. 11.

The widespread distribution of black shale and related organic-carbon-rich facies occurred within relatively restricted time periods during the Cretaceous. The term 'Oceanic Anoxic Events' (OAE) has been applied to these periods (Schlanger & Jenkyns 1976). Unfortunately, this term has been construed by others to indicate that anoxic conditions existed throughout the water column in all ocean basins at precisely the same time. The original intent of the term was to point out that there were periods of time (Barremian-Albian, Cenomanian-Turonian and possibly Coniacian-Santonian) when organic-carbon-rich marine facies were much more widespread in epicontinental seas and open ocean basins than they are today. Although single black shale beds probably cannot be correlated from basin to basin, packages of organic-carbon-rich facies, no matter what their amount and type of organic matter, generally are time equivalent. This emphasizes the possibility that the occurrences of organic-carbon-rich facies in different basins might be linked in some way, even though they may occur in different lithologies and depositional settings.

We favour fluctuations in sea-level and climate as the dominant causes for variations in amount of organic matter supplied and preserved (e.g. Fischer & Arthur 1977). Higher global sea-level and warm, equable climates probably influenced rainfall, production of terrestrial vegetation and runoff to the oceans, thereby increasing the flux of terrigenous sediment and organic matter to ocean basins, particularly the relatively narrow North and South Atlantic Oceans during the Hauterivian through Albian (e.g. Sites 398; 400; 402; Habib 1979). The rapid rate of supply of terrigenous organic matter to a basin that was already poorly-oxygenated (but not necessarily anoxic) probably led to enhanced oxygen deficits in some basins, and allowed preservation of what marine organic matter was produced and supplied to basin deeps. Locally, particularly in shallow epicontinental seas and restricted ocean basins, high runoff periodically may have led to less saline surface waters and stable stratification of the water column, which, in turn, produced deeper water oxygen deficits (e.g. Ryan and Cita 1977; Thierstein & Berger 1978; Arthur & Natland 1979). This would be particularly true if the supply of terrigenous organic matter was high, or if autochthonous productivity increased, perhaps stimulated by nutrients supplied by runoff.

McCave (1979b) and Hochuli & Kelts (1980) suggested that periodic (every 50 ky or less) high productivity intervals could have produced the interbedded organic-carbon-rich and organic-carbon-poor facies so typical of the Cretaceous deep-water black shale sequences. In general, however, sea surface productivity during the Cretaceous probably was low overall in comparison to that of today (Roth 1978; Berger 1979; Thierstein 1979), and only in certain regions (e.g. off N.W. Africa) can high sea surface productivity be called upon to produce Hauterivian-Albian deep-water organic-carbon-rich facies.

Marine organic matter appears to be more prevalent in strata of mid-Aptian to early Albian age, and of Cenomanian age in many Atlantic-Tethyan sequences. This might indicate periods of overall higher sea surface fertility and intensification of midwater oxygen minima related to higher sea-level stands coupled with more rapid production of warm, saline bottom water (Southam *et al.* 1982; Brass *et al.* 1982) or some other mechanism (Tucholke & Vogt 1979; Summerhayes 1981b).

A high sedimentation rate may have been a factor in the preservation of organic matter in some regions (e.g. Ibach 1982), but some of the lowest sedimentation rate intervals, which occur during higher sea-level stands (e.g. de Graciansky *et al.* 1981), have the greatest enrichment of organic matter. Figure 12 illustrates a compilation of organic-carbon contents (core averages) for the western North Atlantic Cretaceous sequences plotted against interval sedimentation rates (compare with Fig. 2, after Müller & Suess 1979). The dotted line represents the approximate Müller & Suess (1979) relationship for sediments from modern marine environments corrected for compaction of the older Cretaceous strata. The scatter of points is high, and many of the points lie above the line, suggesting that the rate of organic carbon supply and/or preservation under low-oxygen conditions was at times more important than bulk sedimentation rate in causing enhanced organic carbon accumulation in Cretaceous ocean basins. Even so, most Cretaceous deep-water organic carbon accumulation rates were, on average, much slower than modern shallower water organic carbon accumulation rates (Fig. 6).

Factors in the origin of Cenozoic organic-carbon-rich strata

The oceanographic and climatic conditions under which the Cenozoic black shales were deposited in the Atlantic are very different from those in the Cretaceous. The ocean basin was wide and unrestricted by the Eocene so that basin floor deoxygenation is not likely and there are no character-

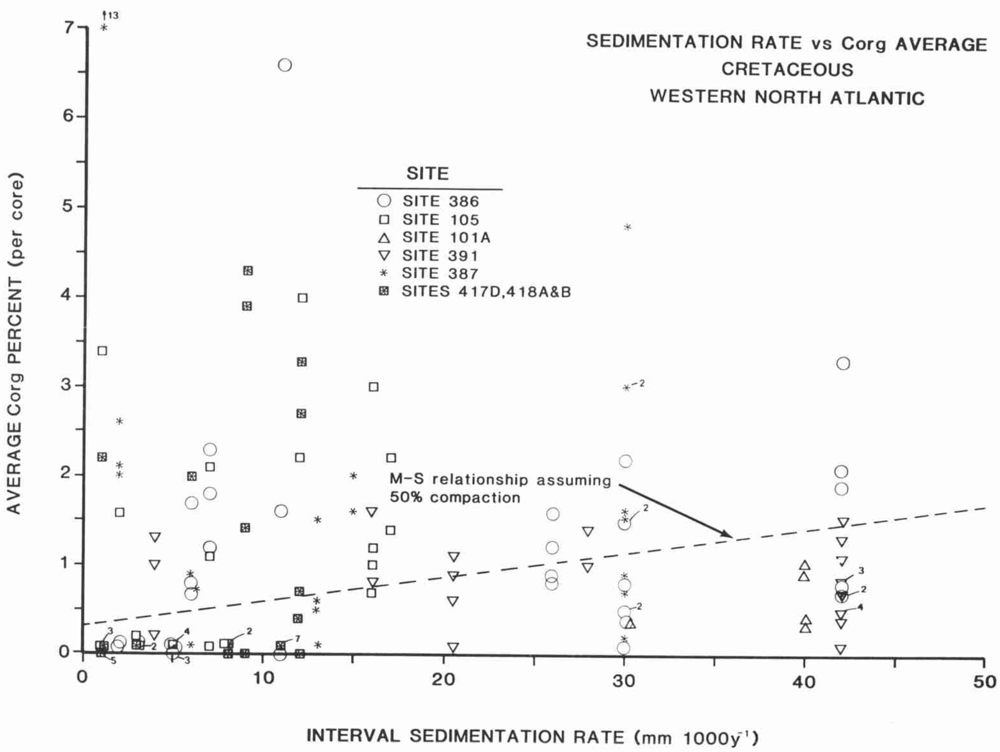


FIG. 12. Plot of interval sedimentation rate (mm 1000 y^{-1}) versus average organic carbon content by core (or core section) in DSDP Sites from the western North Atlantic. The points represent compilation of data from the Hauterivian through Cenomanian of Sites 386, 387, 101, 105, and 417, 418 (data in Arthur & Dean, in press). The dashed line is the 'best-fit' regression line of Müller & Suess (1979) for recent marine sediments adjusted for 50% compaction of Cretaceous strata. Note large scatter of points (see text for discussion).

istics of the organic-carbon-rich facies that would suggest that such an event occurred. Global climate had begun to deteriorate near the beginning of the Tertiary so that by the late Eocene there was a fairly marked temperature differential between the poles and equator. There is considerable evidence now to infer that modern deep thermohaline circulation began at this time, with the formation of cold dense water at high southern latitudes and its northward flow into the Atlantic as Antarctic Bottom Water (Berggren & Hollister 1977; Kennett 1977; Tucholke & Vogt 1979; Shor & Poore 1978; Moore *et al.* 1978). Initially upwelling of older nutrient-rich waters forced by the deep circulation changes would have led to increased productivity, particularly along the continental margins, and hence to an increased supply of organic matter to the sediments. Redeposition of this material to the deeper ocean basins (e.g. at DSDP Sites 367, 370 and 386) was most likely by turbidity currents and other mass-flow processes. In contrast to the above possibility, Brass *et al.* (1982) suggested that the production of warm saline bottom water was linked to eustatic sea-level rise and transgression of shelf areas. An early to middle Eocene sea-level highstand may have increased production of bottom water and increased deep-water turnover rates thereby stimulating biologic productivity, which could have led to intensification of oxygen-minimum zones (Southam *et al.* 1982). Kelts & Arthur (1981; and references within) further suggested that the widespread Eocene cherts of the deep North Atlantic ('Horizon A') may be explained by increased productivity of siliceous organisms on the margins and subsequent downslope redeposition.

The middle to late Miocene to Pliocene

organic-carbon-rich facies (mainly biogenic siliceous sediments) in the Atlantic (Diester-Haas & Schrader 1978; Gardner *et al.*, in press) and Pacific (Ingle 1981; Summerhayes 1981a) also appear to be related to vigorous ocean circulation, upwelling and prolific diatom productivity. This was in response to the continued deterioration of global climate and mid-Miocene build-up of the Antarctic ice cap (Savin 1977; Kennett 1977). Antarctic Bottom Water, Arctic Bottom Water and Norwegian Sea Overflow Water all provided deep vigorous circulation at this time (e.g. Shor & Poore 1978). Organic-carbon-rich sediments accumulated under upwelling zones in the open, oxygenated north-east and south-east Atlantic, and all around the margin of the north Pacific which apparently was more fertile than the Atlantic because of an estuarine-type circulation (Berger 1970). A widespread period of tectonism in the circum-Pacific led to the formation of silled marginal basins which became deoxygenated where the midwater oxygen-minimum zone impinged on the bottom at sill depth (Ingle 1981). Lowered sea-level also may have helped to lower the base of the oxygen-minimum zone (Summerhayes 1981a), and coastal divergences may have moved seaward, off the shelf edge, and provided a more direct source of organic matter to deeper oceanic settings (e.g. Gardner *et al.*, in press).

Conclusions

From the above discussion it is apparent that the origin of interbedded more- and less-reduced lithologies with variable amounts of organic matter and variable amounts of pelagic, hemipelagic, and terrestrial sediment is complex and probably is not due to any one simple process although events in different ocean basins may be linked by a common factor, such as changes in global sea-level. The main variables are supply of organic matter from land and from surface water productivity, sedimentation rate, and oxygen concentration in bottom waters. These variables are greatly influenced by climatic, oceanographic, geographic and tectonic factors. The supply of organic matter in turn determines the thickness and intensity of a midwater oxygen-minimum zone. Surface water productivity is determined, at least in part, by the intensity of upwelling of nutrient-rich water from within the upper part of the oxygen-minimum zone. Another important factor in the accumulation of deep-water black shales is the frequency and magnitude of sediment redepositional events.

The middle Cretaceous black shales owe their origin to an interplay of these three main variables. At this time, the oceans were poised at relatively low oxygen levels which periodically tipped in favour of organic matter preservation, largely due to increased supply of organic matter from surface productivity or terrigenous input, commonly aided by downslope resedimentation. By contrast, the Cenozoic oceans witnessed an increasingly vigorous thermohaline circulation, possibly higher overall fertility, widespread upwelling and enhanced productivity. Organic-matter was preserved both in restricted marginal basins and in open ocean sites; redeposition was locally important.

Each region and even each bed within a black shale sequence is likely to have its own particular combination of factors that led to organic-matter preservation, so that generalized models may not always be applicable.

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