MARINE BLACK SHALES:
Depositional Mechanisms and Environments of Ancient Deposits

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INTRODUCTION

The well-mixed oxygenated ocean of today seems not to be the model for the past 600 m.y.

(Degens & Stoffers 1976)

Organic-carbon-rich strata or "black shales," including dark gray to black, laminated, carbonaceous mudrocks characterized by impoverished benthonic faunas, or devoid of metazoan life, have long intrigued geologists, both because of their widespread distribution at certain times in the past and their early recognition as potential hydrocarbon source rocks. One of the major obsessions of many early workers, to the mid-1900s, was the application of uniformitarian principles to depositional models for black shales. Thus, although it was recognized that some black shale units were unusually widespread, many early workers sought less extensive, modern analogues for environments of formation of dark-colored, rela-
tively organic-carbon (OC)-rich muds. In fact, the main debate focused on shallow- vs deep-water origins; most workers assumed that “black shale” deposition required a supply of organic matter and conditions conducive for preservation of that organic material, including depletion of dissolved oxygen in waters overlaying the sediment/water interface. However, the “barred (silled) basin” depositional model dominated most interpretations of ancient black shales.

With the advent of the Deep Sea Drilling Project in 1968, a wealth of new data became available from the previously unsampled ocean basins. This led to the realization that the deep-sea record of the past 150 million years (m.y.) also contains several widespread “black shale” sequences and, more importantly, that OC-rich shales of Cretaceous age appear to be confined to particular stratigraphic horizons. This discovery brought the realization that local basinal bathymetry could not be the entire explanation for many black shale units. Some workers concluded that certain widespread black shales were deposited globally and synchronously, perhaps even within relatively narrow (<1 m.y.) time envelopes. These time envelopes, during which the global ocean conditions were propitious for the deposition of OC-rich sediments (but not implying global total anoxia of deep-water masses), were termed “Oceanic Anoxic Events” (OAEs). Ocean drilling and the consequent OAE concept, whether correct or not, renewed interest in black shales in general and promoted considerable debate as to the mechanisms that control black shale deposition.

Controversy continues in regard to the origin of modern OC-rich sediments as well as the origin of ancient widespread black shales. Disagreement on the origin of modern OC-rich sediments centers on the relative importance of primary productivity (OC flux) vs possible preservational mechanisms—water-column anoxia and sedimentation rate in particular. The continuing debate about ancient black shales stems from the “no-analogue” global environment that exists at present and the insufficient stratigraphic, sedimentological, and geochemical data for black shales for resolving accumulation rates and definitive indicators of productivity and anoxia, among other parameters.

In this paper we outline the salient features and inferred depositional regimes for black shales in the stratigraphic record and provide guidelines for further study. Rather than review all of the important studies that have contributed to our knowledge of black shales, we have instead attempted to synthesize data and concepts while providing examples for further reference. We have included a discussion of OC preservation in modern marine environments as a prelude to interpretation of the ancient record and a discussion of proposed biotic and geochemical indicators of degree of oxygenation in the water column.

FUNDAMENTAL PROPERTIES OF BLACK SHALES

Composition

Over the past century the term “black shale” has been applied to a considerable range of related facies interpreted to represent quite varied depositional settings and paleoenvironmental conditions. Thus, it is useful to first establish a clear definition of “black shale” in terms of sedimentologic and geochemical parameters before addressing depositional mechanisms.

In this paper we confine our discussion of black shale to marine mudrocks (see Ingram 1953, Dunbar & Rogers 1957, Blatt et al 1972, Lundegard & Samuels 1980, Spears 1980 for definitions of mudrocks) that are rich in organic matter (as defined below); true color can vary from medium gray to olive-brown to black. Although 10% laminated is a defining characteristic of most black shales, there are numerous OC-rich rocks that fall under the classification of (nonlaminated) mudstones or claystones. In addition to the dominant sediment type (mud or clay), mudrocks may contain various constituents, including calcium carbonate (CaCO₃) and biogenic SiO₂, and there are a number of appropriate classification schemes (e.g. Hattin 1975, Pratt 1984, Dean et al 1984). But perhaps the most important constituent that differentiates black shales from other mudrocks is organic matter content, which is the main cause of the dark color in ancient black shale facies (e.g. Trask & Patnode 1942).

Coloration of shales is influenced by amount, type, and maturity of OC, all of which can vary significantly. Sediments containing more than a few percent of immature amorphous organic matter are typically more brown than black; strata containing only 1–2% highly oxidized or thermally mature OC are commonly black. CaCO₃ contents in excess of 25% can also impart a lighter color to relatively OC-rich mudrocks. The black coloration of shales also is attributed at times to the presence of very fine-grained iron monosulfide minerals (grietite and/or mackinawite; Berner 1984), but these monosulfides are not stable over long (geologic) periods of time and the black coloration in monosulfide-bearing sediments fades rapidly (within hours) as they are exposed to oxygen. Huyck (1990) has discussed some of the problems and variations in defining “black shale” on the basis of OC contents. Most problems revolve around determining the OC content of “average” shale, which has commonly been taken as 0.65% OC (Vine & Tourtelot 1970). As an operational definition marine “black shales” should contain at least 1% organic matter (e.g. Huyck 1990)—typically a mixed assemblage including material derived from higher plants, algae, and phytoplankton (Cook et al 1981). Proximity to terrestrial sources of organic matter and marine productivity are the pri-
mary controls on the type of organic matter that becomes incorporated in black shale deposits as outlined below. This enhanced OC content sets these mudrocks apart from others; however, one must also keep in mind that many ancient black shales that once contained >1% OC may now have <1% OC because of OC loss during thermal maturation or metamorphism (e.g. Raiswell & Berner 1987). Thus the definition of “black shale” by organic content will always be somewhat arbitrary.

A number of workers have referred to “black shales” as “sapropels.” It is thus useful to examine briefly the genetic implication of “sapropel” because, in many cases, the term has been misapplied (e.g. Tyson 1986). Many workers apply the operational definition used by Kidd et al (1978) for Mediterranean strata which set an arbitrary lower limit of >2% OC for sapropel and use “sapropelic” to refer to sediments with OC contents between 0.5–1.9%. Anastasakis & Stanley (1984) applied the term “episapropel” to those units that could be determined to have been redeposited on the basis of sedimentary structures. However, although some workers used “sapropel” as a modifier for OC-rich sediments (e.g. Grabau 1913), the original definition of the term includes the inference that “sapropel” is a sediment rich in algal organic matter that was formed under reducing conditions in a “stagnant” water body (e.g. Pontonié 1937, Wasmund 1930). Following the recommendation of Tyson (1986), the term “sapropel” should be reserved only for descriptions of strata that meet the latter criteria; otherwise the term has no more utility than “black shale (clay or mud).”

Lamination and Sedimentary Structures

Lamination is a fundamental distinguishing characteristic of black shales and has long been a topic of interest (e.g. Bradley 1931). Lamination in mudrocks is defined as a parallel arrangement of layers <10 mm thick that results from a regular alternation in fabric, grain size, and/or color (Lundegard & Samuels 1980). Although many classic black shales are finely laminated, not all are, either because original laminae were disturbed by bioturbation or by other physical processes, such as winnowing and redeposition.

Laminae couplets typify sediments in modern environments of formation of OC-rich muds where oxygen concentrations inhibit benthic metazoan activity and where there is pronounced seasonality in productivity or sediment supply. Examples are the Black Sea (e.g. Degens & Ross 1974) and the Gulf of California (e.g. Calvert 1964). One of the most common types of lamination found in Cretaceous and younger rocks are light-dark alternations composed of laminae rich in calcareous material (commonly coccoliths) and clay/organic matter, respectively. Although a number of recent studies have suggested that such laminations result from seasonal phytoplankton blooms (Ross et al 1970, Thomsen 1989, Hattin 1975), evidence from other ancient black shale deposits indicates that there are a wide range of different types of laminae that appear macroscopically similar, but do not reflect a seasonal depositional cycle (Sageman 1991), possibly because sedimentation rates are too slow. In cases such as the Cretaceous Greenhorn Formation of the U. S. Western Interior Basin it appears that individual light-colored laminae represent the amalgamation of numerous annual productivity events, separated by darker clay and organic-rich laminae that probably reflect more rapidly and intermittently deposited mud. Fine lamination in other Cretaceous calcareous black shales of the western North Atlantic has been attributed to bottom-current activity, albeit in an oxygen-deficient basin (e.g. Robertson 1984, Dean & Arthur 1989). Williams & Reimers (1989) attributed lamination in the Miocene Monterey Formation of California, and by inference other OC-rich deposits of upwelling zones associated with oxygen-minimum zones, to periodic (perhaps seasonal) colonization of the sediment/water interface by sulide-oxidizing bacterial mats.

Other features common to epicontinental black shales include a variety of event beds and horizons that indicate significant condensation and/or erosion. Whereas examples from the Western Interior Cretaceous include skeletal limestone beds and lenses composed of winnowed biogenic debris (Pratt 1984, Hattin 1986, Sageman 1991), the Devonian black shales of New York State contain similarly interpreted pyritic lag deposits (Baird & Brett 1986). Black shale sequences may also be typified by redeposited beds relatively rich in organic matter as suggested for many “episapropels” of Quaternary age in the Mediterranean (e.g. Anastasakis & Stanley 1984) and Cretaceous black mudstones of the deeper marginal North Atlantic (Arthur et al 1984, Degens et al 1986). Many of these redeposited OC-rich beds have sharp basal contacts and are homogeneous to coarsely laminated.

HISTORICAL PERSPECTIVE

In Pettijohn’s (1957, pp. 622–26), review of the “Black Shale Facies” (euxinic) he suggested, “That the pyritic black shales were deposited under anoxic conditions is unquestioned. Whether, however, the basin of accumulation was shallow or deep and whether it was landlocked or freely connected with the sea or even a stagnant area in the open sea has been much debated.” With this statement he essentially outlined the major sources of disagreement as to the origin of black shales then and now. His cogent summary of the black shale facies followed years of speculation
and assertion by a number of influential workers. For example, Grabau (1913) argued that widely distributed marine black shales (his “sapro-pelargillite”) were deposited in coastal lagoons or marginal epicontinental seas on extensive mud flats exposed at low tide. Grabau knew of the Black Sea because Pompeckj (1901) had interpreted the Jurassic Posidonienschiefer (Europe) and the Devonian Portage (eastern U.S.) black shales, respectively, as having originated in Black Sea–like environments (see also Schuchert 1915), but he viewed this analogy as doubtful, suggesting only that this model might apply to the Permian Kupferschiefer of Europe. Subsequently, Strom (1939) summarized his work on organic matter deposition in Norwegian fjords, which probably unintentionally reinforced the barred basin concept. Fleming & Revelle (1939), on the other hand, presented a thorough and insightful treatment of the role of physical oceanographic processes on dissolved oxygen distribution in the water column and even discussed the origin of mid-water oxygen-minimum zones (OMZ); they outlined causes of stable stratification, including the susceptibility of tropical environments to thermal stratification, but, again, particularly emphasized the role of sill depth or basin thresholds in controlling the rate of renewal of bottom water, citing California borderland basins and the Black Sea as examples where stable stratification and/or sills limited exchange of bottom waters and promoted oxygen deficiency. In contrast, Twenhofel (1939) apparently saw a more open-marine origin for “bituminous black shales” stating that “…bituminous black shales accumulated in general, in less stagnant and more open water, but waters that did not have good circulation.” Nonetheless, he favored a “Black Sea type” model for Paleozoic laminated black shale deposits of northwest Europe and the U.S. Appalachian basins. He suggested water depths between 300 and 600 feet, however, for the euxinic water masses, arguing that deposition above wave base would have been difficult as a result of oxygen mixing downward periodically.

At the time of publication of Pettijohn’s treatise there was little appreciation of the possible role of primary productivity, and the emphasis clearly focused on reducing environments. However, this subject, a source of disagreement among present workers (e.g. Pedersen & Calvert 1990 vs Demaison & Moore 1980), was not completely ignored. For example, Goldman (1924) explicitly suggested the importance of rate of supply of organic matter with or without oxygen depletion in stating that, “In shallow bays that have no threshold, but which are rich in vegetation and have only a feeble deep water circulation, the bottom deposits often consist of black muds high in hydrogen sulﬁde and containing no animal life. The waters themselves, however, will usually contain dissolved oxygen. The moderate stagnation in these cases is due to the large supply of decomposable organic matter rather than the lack of water exchange.” This concept was echoed by Fleming & Revelle (1939) and was implicit in Trask’s (1932) discussion of sedimentary organic matter contents, but really was not entertained as a major theme in the origin of black shales until more recently (e.g. Parrish 1982, Parrish & Curtis 1982, Waples 1983). More complete models for black shale deposition, such as that for midcontinent U.S. Pennsylvania black shales outlined by Heckel (1977), suggested that both high primary productivity (upwelling) and oxygen depletion in deeper water masses (preservation) played a role in producing relatively OC-rich strata.

Like Ruedemann (1935, also Hard 1931), who argued for “toxic” bottom-water conditions and against the concept of small embayments or lagoons as primary sites for black shale deposition in their work on Devonian black shales of the Appalachian region, Pettijohn (1957) also summarized black shale facies as being widespread at certain times and characterized by a slow rate of accumulation. He advocated the idea that many black shales represent deposition in “starved” basin phases (following the terminology of Adams et al 1951), which now would be recognized as relatively condensed zones related to transgressive episodes. Nonetheless, Pettijohn followed the prevailing view that OC-rich strata were the result of more local factors, such as deposition in “barred” or tectonically isolated basins (e.g. Trask 1932, Woolnough 1937, Caspers 1957), and suggested the Black Sea as the “type euxinic basin,” thus acknowledging a possible role for density stratification of water masses and dissolved oxygen concentration (“strong reducing environment”) in OC preservation, but in a restricted basin context.

That variations in global oceanic circulation likely were involved in the formation of the black shales was recognized by van der Gracht (1931) who argued that during non-glacial periods the equatorward circulation of deep, cold oxygen-rich water would cease and that the bottom water of large oceanic basins would become anoxic. This concept built on the ideas of Chamberlin (1906), who suggested reversal of circulation and warmer, more saline deep waters during warm climatic epochs. Stroom (1936) also recognized that more widespread episodes of “black sediment” deposition might require more than local causal factors. Thus, long before the 1970s it was conceived that the “Black Shale Facies” represented rather unique but widespread paleoceanographic conditions that occurred episodically. One common theme of nearly all black shale studies links black shales to marine transgression (e.g. Hallam & Bradshaw 1979, Loutit et al 1988) — a cause which has been invoked for modern occurrences of OC-rich sediments in silled basins as well (e.g. Middelburg et al 1991).

A minor revolution in how the community viewed the origin of black
shales came as a result of the Deep Sea Drilling Project (DSDP). Cretaceous black shales were recovered in a number of DSDP sites in all ocean basins (Ryan & Cita 1977, Thiede & van Andel 1977, Fischer & Arthur 1977). The recognition that these organic-rich units were equivalent to black shales exposed on a number of continents led to the “OAE” concept, which was originally defined by Schlanger & Jenkyns (1976) to signify time envelopes during which black shale deposition was particularly prevalent. It has become apparent that this terminology may be misleading in the sense that it does not distinguish between events with long and those with short durations. The term “anoxic event” subsequently has been used very loosely in describing almost any relatively OC-rich bed, including, for example, the Quaternary sapropels of the Mediterranean Sea. With respect to the Cretaceous, it is quite clear that the Cenomanian-Turonian OAE was, geologically speaking, a very brief episode (<1 m.y.) recorded nearly globally. By comparison, the Aptian-Albian OAE, as defined by Schlanger & Jenkyns (1976) encompasses nearly 20 m.y. An additional problem is that the term OAE can be construed as indicating that the global ocean was anoxic (widespread “euxinic” conditions), but this sort of genetic implication should be avoided because of the multitude of possible ways in which black shales may have formed, even during one discrete episode.

MODERN ENVIRONMENTS AND ORGANIC CARBON BURIAL

Modern Analogues for Ancient Environments

Interpretation of ancient black shale deposits has been strongly influenced by observations of modern environments where OC-rich muds presently accumulate. Depositional analogues for marine black shales most commonly cited include large landlocked or silled basins, such as the Black Sea and the Baltic Sea, as well as open-ocean environments such as coastal upwelling zones (e.g. Peru, Namibia, NW margin of the Indian Ocean, California borderland basins, and the Gulf of California) affected by impingement of the midwater oxygen-minimum zone. More localized environments of organic matter burial include tropical embayments, coastal swamps, large estuaries, fjords, and brine-stratified slope basins. The physical characteristics of these different environments have been extensively reviewed by Rhoads & Morse (1971), Demaison & Moore (1980), and Arthur et al (1984) among others, and are summarized in Figure 1.

Modern analogues for black shale-forming environments are classified based on relative water depth, and the nature and position of the redox boundary and its affect on benthic communities (Figure 1). The categories include: 1. deep, enclosed, strongly stratified basins, characterized by a

**Figure 1.** A summary of the major modern marine environments of organic matter accumulation showing: (a) idealized basin morphology and bathymetry; water mass distribution; dominant position of redox boundary (Rdb); and predominant marine biota; (b) characteristic environments; and (c) key differences for each environment.
redox boundary high in the water column and long-term anoxia throughout much of the basin; 2. deep, unenclosed basins in which dysoxic conditions predominate and the redox boundary normally remains at some depth within the substrate, but may fluctuate up and down; 3. shallow, stratified basins that are subject to frequent mixing by storms in which dysoxic to oxic conditions predominate and the redox boundary normally remains at some depth within the substrate, but may fluctuate above and below the sediment/water interface (SWI); and 4. shallow-water areas in which the water column is dysoxic to fully oxic, but the sediments are highly OC-rich and reducing. The redox boundary is commonly coincident with the SWI.

The Black Sea is the best example of a large, deep, stratified basin characterized by an anoxic marine water column and finely laminated, OC-rich sediments (Figure 1). It has played an influential role in the interpretation of ancient black shales (i.e. Rhoads & Morse 1971). However, conditions in the Black Sea are unique and represent one extreme in a spectrum of oxygen-deficient environments. Black Sea waters below about 150 meters are completely anoxic and contain dissolved H₂S; they are devoid of metazoan life. Yet, recent studies of a wide range of ancient black shales have revealed a common history of repeated benthic colonization events, suggesting relatively dynamic, and largely dysoxic benthic environments (e.g. Kaufman 1981, Savrda & Bottjer 1986, Sageman 1989, Wignall 1990). Thus, the late Holocene Black Sea is best viewed as an analogue for sedimentary and geochemical processes under an anoxic water column (euxinic conditions), which, through geologic history, probably occurred more commonly in the benthic zone, than throughout the water column of relatively deep basins.

Oceanic OMZs result from the oxidation of sinking pelagic organic matter and the long residence time and slow rate of exchange between intermediate-depth waters and better oxygenated surface waters (Demaison & Moore 1980). They are particularly intense and well-developed on the margins of ocean basins, where wind-driven upwelling and higher nutrient levels enhance surface-water productivity. In some cases, OMZs impinge upon structurally complex continental margins, such as on the California borderland, where a series of graben-type basins preserve significant thicknesses of OC-rich sediment (Figure 1). These basins are silled below about 500 m depth and contain predominantly dysoxic waters with the main redox boundary at shallow depth within the substrate (Savrda et al 1984). Biotic characteristics of such basins reveal surprisingly high benthic community diversity and biomass from within the dysoxic zone, with up to ca. 300 species reported in some cases (Jumars 1976). The highest species diversity and biomass are among polychaete worms (up to 150 species; Hartman & Barnard 1958), which can bioturbate the sediment at oxygen levels as low as 0.3 ml/l (Thompson et al 1985). In addition, a wide range of taxa comprise less abundant elements in these communities, including arthropods, echinoderms, coelenterates, infaunal protobranch bivalves, nematodes, and benthic foraminifera (Hartman & Barnard 1958, Hülsemann & Emery 1961, Calvert 1964, Nichols 1976, Jumars 1976, Smith & Hamilton 1983, Savrda et al 1984, Thompson et al 1985). Though relatively OC-rich sediments are commonly bioturbated, laminated intervals with up to 4% relatively hydrogen-rich OC occur (Hülsemann & Emery 1961, Summerhayes 1987), suggesting significant spatial and/or temporal variability in the position of the redox boundary.

In upwelling zones the export of abundant pelagic organic matter to the benthic zone, due to high productivity, results in extensive metazoan and bacterial decomposition and consumption of dissolved oxygen (dysoxic to anoxic). The Peru outer shelf to upper slope region is characterized by a well-developed OMZ (e.g. Thiede & Suess 1983, Suess, von Huene, et al 1990, Froelich et al 1988). Sediment OC contents are high relative to the California borderland, due to enhanced productivity [up to 22% TOC (Total Organic Carbon) with HI (Hydrogen index for hydrocarbons per g OC) to 600] and lower sedimentation rates; however, the biotic characteristics of these basins are almost identical to those of the California borderland (Rosenberg et al 1983, Arnitz et al 1991). Despite evidence of pervasive colonization of the sediment by infauna (polychaetes dominant; Arnitz et al 1991), intervals of undisturbed laminae described from cores suggest spatial and/or temporal variability in the position of the redox boundary. Mats of sulfide-oxidizing bacteria (e.g. Thioploca) that occur in patchy distributions on the surface of these OC-rich sediments (Galarraga 1977, Arnitz et al 1991) may play a role in determining redox boundary dynamics.

Oxygen deficiency and accumulation of OC-rich mud has been described from large estuaries (e.g. Chesapeake Bay), shallow basins with high freshwater input (e.g. the Baltic Sea), and fjords, attributed mainly to seasonal intensification of salinity stratification (Goldman 1924, Ström 1936, Leppekosski 1969, Theede 1973), or in tropical bays (e.g. Kau Bay, Jellyfish Lake, Palau, Cariaco Trench, Venezuela) as the result of thermal and/or salinity stratification (Orem et al 1991, Middelburg et al 1991, Richards & Redfield 1954). At such times, bottom waters become dysoxic, and the substrate is characterized by black, anoxic muds saturated with H₂S, particularly in topographically low areas. Organic content of the sediment typically varies from less than 1 to 7% TOC (but Jellyfish Lake, Palau has up to 46% TOC; Orem et al 1991) and organic matter may be mostly terrestrial in origin, resulting in low hydrogen indices. The benthic fauna
is typically dominated by polychaetes, although a number of low-oxygen tolerant bivalves also have been described (Leppakoski 1969, Theede 1973). Such environments are quite dynamic, with frequent variations in the position of the redox boundary due to seasonal or longer-term changes in biologic production, stratification, and benthic oxygen levels. Another rather local environment for OC preservation occurs in small, restricted slope basins associated with regions of salt tectonics and salt outcrop (e.g., Orca Basin, Gulf of Mexico, Mediterranean basins; see Cita 1991 for review). These small basins are not likely to produce major preserved black shales in the record but are interesting geochemically.

Certain coastal areas where fine-grained sediments are deposited, such as large bays, lagoons, tidal flats, and intertidal swamps, are also sites of deposition of black, H2S-containing, OC-rich muds (Twenhofel 1915, Jorgensen & Revsbech 1985, Rosenberg 1977, Martens & Klump 1984). The distribution of organic-rich sediment in such areas is extremely patchy. The OC content of the sediment is highly variable and likely to be dominated by terrestrial components, resulting in low hydrogen richness. Prevalence of mixing in these shallow areas results in mostly oxic to slightly dysoxic conditions in the water column. However, a sharp redox boundary commonly develops at the SWI due to the rapid uptake of oxygen by bacteria in the OC-rich sediment. In some cases, development of a diffusive boundary layer creates a physical microboundary to oxygen exchange; such a boundary may or may not be associated with a conspicuous microbial mat of sulfur-oxidizing bacteria (such as Beggiatona) on the sediment surface (e.g. Jorgensen & Revsbech 1985). In general, the major faunal constituents of shallow-water OC-rich muds are quite similar to other modern low-oxygen environments (where polychaetes and infaunal deposit-feeding bivalves dominate). However, these muds also contain a unique group of highly specialized bivalves that survive within toxic, H2S-containing muds by virtue of chemosymbiosis with sulfur-oxidizing bacteria. These taxa, including species of the families Lucinidae, Thyasiridae, and Solemyidae, evolved from infaunal suspension-feeding bivalves and developed morphologic and metabolic adaptations for survival in oxygen-deficient reducing environments (Cavanaugh 1985, Dando & Southward 1986).

In summary, most modern environments of OC burial are dominantly dysoxic, but relatively dynamic, with pronounced variability of water column dissolved oxygen concentrations, and at least periodic excursions of the anoxic zone above the SWI. This temporal and spatial heterogeneity in the position of the redox boundary exerts a strong control on the development of benthic communities (soft-bodied infauna vs shelly epifauna), and thus the preservation of primary laminae. Infaunal deposit feeding is the dominant trophic strategy in modern low-oxygen environments while epifaunal suspension-feeding organisms are rare (i.e. Rhoads & Young 1970).

Sulfide-oxidizing bacterial mats, common in modern low-oxygen environments with OC-rich substrates (Spies & Davis 1979, Jannasch 1984, Tuttle 1985), may help to stabilize the redox boundary at the SWI, and thus probably play a significant role in the ecology of most low-oxygen benthic communities. Under conditions of a relatively stable redox boundary at the SWI specialized epifaunal or infaunal suspension feeders have come to dominate the benthic community; where the boundary is more dynamic the associated macrofauna tend to be opportunistic colonizers.

Because modern environments of organic-rich sedimentation include examples of either relatively deep basins (types 1 and 2, Figure 1) or relatively shallow environments (types 3 and 4, Figure 1), it is not surprising that one of the longest running geological debates has focused on the relative depth of black shale deposition. However, many epicontinental black shales—the typical subjects of deep vs shallow debates (e.g. Twenhofel 1915, Ruedemann 1935, Trask 1951, Conant & Swanson 1961)—have characteristics that are inconsistent with those of most modern low-oxygen environments. For example: 1. many black shale units are widely distributed and temporally long ranging within epicontinental basins; 2. many intervals are predominantly laminated, rather than predominantly bioturbated; 3. their constituent faunas are dominated by low diversity assemblages of epifaunal suspension feeders, rather than diverse infaunal deposit feeders; and 4. they are characterized by common event communities indicating a dynamic benthic environment (e.g. Kauflman & Sageman 1990, Sageman et al 1991). Increasingly geologists are seeking to understand the processes responsible for accumulation of OC-rich sediments in modern marine environments, independent of factors such as basin depth, in an effort to develop more actualistic depositional models for ancient black shales.

**Processes: Productivity vs Preservation**

Because OC flux to the seafloor is an important determinant in the origin of black shales, we outline briefly the major controls on OC production and initial accumulation on the seafloor. Organic carbon is usually a minor constituent (<0.5 wt.%) of marine sediments; however, its burial over vast areas of the seafloor plays a very important role in biogeochemical cycles. The main sources of OC to marine sediments are terrestrial organic matter transported to the ocean via rivers (e.g. Hedges & Parker 1976, Hedges et al 1986, Meybeck 1982, Berner 1982), and to a lesser extent by winds (e.g. Cachier et al 1986), and photosynthetic production of marine
phytoplanktic organic matter with a minor contribution from organisms higher in the food chain. The amount of marine OC fixed in near-surface waters by photosynthesis is dependent, in part, on nutrient availability in surface waters. These principles have been reviewed in Berger et al (1989) and Jahnke (1990) and are summarized briefly here. In central pelagic regions of large oceanic gyres, where primary productivity and surface water nutrient concentrations are generally low (so-called oligotrophic regions), regenerated production (i.e., using nutrients largely recycled from the surface biomass) accounts for most of the primary productivity, and only a very small portion of the OC is exported to deep waters. In highly productive regions, new production (i.e., nutrients supplied from the base of the thermocline) accounts for a more significant fraction of the primary productivity, and a larger portion of the OC is transported from the photic zone. This flux is concentrated near the continental margins and is reflected in sediment OC concentrations (e.g., Premuzic et al 1982, Jahnke et al 1990, Reimers et al 1992).

A global average of less than 20% of the organic matter produced in oceanic surface waters escapes consumption/oxidation within the photic zone (e.g., Müller & Suess 1979, Berger et al 1989) to fall through the main oceanic thermocline. Of the material that escapes the upper 100 m of the water column, 75 to 85% is decomposed within the upper 500 to 1000 m and only 3 to 5% of the total primary productivity is transported below about 1000 m. In addition, greater than 90% of this organic matter that reaches the seafloor is decomposed, primarily by oxygen-based metabolism of organisms (e.g., Emerson & Hedges 1988), so that sediment OC burial fluxes provide minimum estimates of the original organic particle flux to the deep sea. Thus, a small fraction (<0.5% globally of the primary production) of the OC reaching the sediments is buried (e.g., Berner 1982). Previous studies (e.g., Suess 1980) suggest that, on a regional and local basis, the rate of OC burial reflects primary productivity, in spite of the many processes that diminish OC fluxes to sediments. Such a relationship between sedimentary OC accumulation and primary productivity might allow inferences about surface productivity in the past from sediment accumulation (Müller & Suess 1979, Sarastein, et al 1988, Stein 1986).

While primary production (photosynthesis) in surface waters and river input control the amount of OC initially supplied to the seafloor, many factors modify the concentration of OC buried in the sediments (e.g., Jahnke et al 1990). For example, because OC may be modified by microbial oxidation and metazoan consumption in the water column and at the SWI (e.g., Aller & Mackin 1984), the OC fluxes (and OC concentrations, see Figure 2) may depend on depth-dependent oxygen concentrations. In areas where the midwater OMZ intersects the seafloor along continental margins, many workers have argued that OC is preferentially preserved in surface sediments because of the low dissolved-oxygen concentration (e.g., Richards & Redfield 1954, Demaison & Moore 1980, but see Calvert et al 1992a for opposite view).
The rate of OC burial and sedimentation rate covary in all marine sediments (Toth & Lerman 1977, Heath et al 1977, Müller & Suess 1979). A recent summary of available data from modern marine environments (Betts & Holland 1991) attributes increasing OC burial efficiency (the percentage of OC that reaches the SWI, survives degradation, and is ultimately buried) almost entirely to a sedimentation rate effect at linear sedimentation rates below 60 cm/ky (Figure 2; see also Henrichs & Reeburgh 1987). Higher sedimentation rates apparently reduce the residence time of OC at the SWI and thus remove it from the zone of oxidative benthic metabolism, particularly in moderate- to well-oxygenated regions where large benthic metazoans (see below) churn the sediment. In addition, pumping of dissolved oxygen below the SWI which could be utilized to further degrade OC and/or mitigate buildup of inhibitory metabolites (e.g. Lee 1992) is eliminated by the exclusion of large infaunal organisms in dysoxic to anoxic settings (e.g. Aller 1982, Berner & Westrich 1985).

OC distribution may also be affected by the grain size and mineralogy of the accompanying sediments. Some organic compounds are sorbed onto clays (Morris & Calvert 1977) and/or onto calcium carbonate grain surfaces ( Müller & Suess 1979), thereby increasing OC preservation in the fine fraction. Fine-grained sediments have been found to be higher in OC than other lithologies in similar environments (e.g. Emery & Uchupi 1972, Bordovsky 1965). Of course, this relationship may be the result of depositional energy and higher OC fluxes to the seafloor over the continental slope where finer grained sediments preferentially accumulate. However, recent work (Keil et al 1993) suggests that organic matter forms a coating which has a fairly uniform thickness on clastic grains, resulting in a strong relationship between decreasing grain size, increasing grain surface area, and OC concentration. Nonetheless, this association cannot be the main control on black shale occurrence because low-OC clay-rich strata dominate the record.

Substantial debate centers on whether marine sediments and sedimentary rocks containing substantial amounts of organic matter originated because of unusual chemical conditions in the water column overlying the seafloor (enhanced preservation of organic matter as the result of oxygen depletion) or because of an unusually high flux of OC to the seafloor with rate of burial exceeding rate of oxidation. In the latter case the high OC fluxes are commonly viewed as resulting from an increase in available nutrients at the sea surface due to upwelling or other circulation phenomena. Bulk sedimentation rate clearly is a factor in preservation of organic matter. However, the effects of variations in primary productivity (increased carbon flux) and perhaps of enhanced preservation in oxygen-deficient environments on OC accumulation in sediments are probably also important (e.g. Emerson 1985, Emerson & Hedges 1988), yet these have not been calibrated satisfactorily in modern depositional environments. On the basis of data in Figure 2, it appears that a distinction between anoxic ( euxinic) and anoxic depositional environments at low to moderate primary productivity (<150 g C m$^{-2}$ y$^{-1}$) is possible at sedimentation rates below 5 cm/ky (Stein 1986). At sedimentation rates above 10 cm/ky, it is difficult to distinguish the relative effects of oxygen concentration from productivity. However, other global data synthescise has found no significant effect of anoxic conditions (at or above the SWI) on OC preservation (Figure 2; e.g. Henrichs & Reeburgh 1987, Canfield 1989, Betts & Holland 1991). For example, burial efficiencies in oxic Dabob Bay (Washington) and nearby euxinic Saanich Inlet (British Columbia) are indistinguishable and the SWI is an important site of degradation in both settings (Cowie & Hedges 1992). Lee (1992) also found that the intrinsic rates of oxic and anoxic degradation are not significantly different (see also Canfield 1989); she proposed, however, that the higher content of lipid-rich OC observed for many sediments deposited in euxinic environments could be the result of buildup of anaerobic bacterial mass that would normally be grazed in oxic environments (see Harvey et al 1986).

The lack of difference in OC degradation rates in oxic vs anoxic settings appears to be confirmed by recent studies of sediments deposited in high-productivity OMZ settings. In the Gulf of California (Calvert et al 1992) there appears to be no significant difference in OC concentration and pyrolysis hydrogen index between laminated and bioturbated Recent muds; likewise, there is no evidence of enhanced OC preservation within a modern well-developed OMZ on the Oman margin (Pedersen et al 1992). OC accumulation rate data for the Black Sea (Calvert et al 1991, Arthur et al 1993) also indicate no strong difference between productivity-normalized Holocene OC accumulation rates in that deep euxinic basin as compared to oxic open-ocean settings characterized by similar depths and sedimentation rates, although the Black Sea data fall at the high end of the range. In fact, it is inadvisable to use accumulation rates from the Black Sea as a comparison because of apparent focusing of organic matter sedimentation in deep water relative to the margins (Arthur et al 1993). These results are vexing in light of the many workers who have attributed relatively high OC concentrations and evidence of lack of degradation to the effects of low dissolved oxygen concentrations in the depositional milieu (Demaision & Moore 1980). Are productivity and sedimentation rates the only control on OC accumulation in marine environments?
SELECTED TECHNIQUES IN BLACK SHALE STUDIES

Chemical Analyses of Organic Carbon-Rich Sediments: Clues to Origin?

Here we present approaches, using geochemical analyses of OC-rich strata and paleobiological analysis of black shale faunas, to the study of ancient black shales. In each case these should be grounded within detailed stratigraphic and sedimentologic frameworks. We believe that the integration of these different approaches holds great promise for addressing many of the questions facing black shale workers today. Geochemical studies of black shale sequences may provide important information on the causes and consequences of enhanced organic matter burial in ancient environments. Yet none of the potential techniques proposed to date constitute foolproof means of inferring absolute dissolved oxygen concentrations or primary productivities. For example, several workers have suggested that concentrations (and accumulation rates) of Ba in sediments signal the influence of biologic productivity in determining sedimentary organic carbon fluxes (e.g. Brumsack 1986). Although highly correlated with OC flux, the barium is most likely tied up in carbonate phases, highly insoluble barite crystals, and nodules. However, because of subsequent bacterial sulfate reduction and decreasing pore-water sulfate concentrations, barite can be solubilized to migrate within the sediment column. Because of this process, barite in modern OC-rich sediments tends to reprecipitate at or near interfaces representing reduced sedimentation rates or oxygenation events (Brumsack 1986). Thus, drawing conclusions from Ba concentration patterns alone might be misleading. Below we outline several commonly cited geochemical indicators of conditions promoting black shale deposition and briefly discuss their potential and the problems with their interpretation. These are: 1. simple organic geochemical indicators, such as the pyrolysis hydrogen index and applications of carbon isotope data (e.g. Hollander et al 1993), that provide evidence of the degree and conditions of preservation of organic matter; 2. the reduced sulfur to organic carbon (S/C) ratio (e.g. Leventhal 1983, Berner & Raiswell 1983, Berner 1984) and “degree of pyritization” (e.g. Raiswell et al 1988) that have been suggested as a means for discriminating between euxinic conditions as opposed to pore-water anoxia; and 3. ratios of concentrations of transition metals such as V and Ni that may provide evidence for anoxia in the overlying water column (euxinic basin) (e.g. Lewan & Maynard 1982) and the utility of trace metals in general.

PYROLYSIS HYDROGEN INDEX AND SIMPLE INDICES OF PRESERVATION/TYP

In addition to the problems in discriminating oxic from anoxic depositional environments using OC concentrations or accumulation rates, one of the major differences between anoxic and oxic depositional environments is the apparent better preservation of lipid-rich organic matter in anoxic settings. Although a detailed discussion of the plethora of organic geochemical techniques available is beyond the scope of this review, below we examine one of the more readily applied techniques: the pyrolysis of organic matter by Rock-Eval pyrolysis, which rapidly provides estimates of hydrocarbon type, hydrocarbon-generating potential, and degree of preservation of organic matter in sediments and sedimentary rocks (Espitalié et al 1985). This technique has commonly been applied to the study of modern and ancient OC-rich strata, and, with care, can be relatively easily interpreted in terms of both preservation and source of organic matter (Peters 1986, Katz 1983).

For example, laminated Unit II sediments (2–8 ka) in deep-water sites in the Black Sea are characterized by HI values > 400 mgHC/gOC (Figure 3a). Laminated Unit II sediments from shallow-water, nearshore areas are characterized by more variable HI values of 150 to 500 mgHC/gOC. Values of HI < 150 mgHC/gOC might indicate either highly oxidized organic matter of planktonic origin or primarily terrestrial derived higher plant material. Mixing of terrestrial material (HI of approximately 100 mgHC/gOC) and well-preserved algal organic matter (HI > 600 mgHC/gOC) in various proportions will lead to HI values between these end-member values. Likewise, partial oxidation of marine authochothous organic matter can produce HI values lower than 600 as well as higher oxygen indices. In modern marine environments it is unusual to find HI values > 400 mgHC/gOC in primarily oxygenated environments, even under highly productive regimes with somewhat oxygen-depleted bottom waters such as the Peru margin (Figure 3b) or in the Gulf of California (Calvert et al 1992), where values in surface sediments underlying dysoxic waters are entirely below 300 mgHC/gOC. Note that there is no clear correspondence between OC concentration and HI or OI for Peru sediments in Figure 3b. Consistently high HI values of 400–600 are found mainly in environments where organic matter is preserved under euxinic conditions and where there is no significant reworked material. In the upper part of Unit I in the deepest Black Sea, deposited under euxinic conditions, HI values range between 250 and 550 (Arthur et al 1993); thus, HI values of < 400 mgHC/gOC do not necessarily indicate deposition under oxic conditions. Pyrolysis data are sensitive to both source and preservation, and interpretation should be aided by other types of information on kerogen type, such as visual kerogen analysis and/or carbon isotopic data on organic carbon, among others.

Hollander et al (1993) proposed a method for discriminating between
laminated black shales produced by high productivity, with concomitant euxinic conditions, and those formed in low-productivity euxinic environments. This method utilizes the difference ($\Delta \delta^{13}C$) between $\delta^{13}C_{oc}$ (organic C) and $\delta^{13}C_{cc}$ (carbonate C), with the constraint that the OC be entirely marine derived, plotted against the pyrolysis HI, which is assumed to reflect degree of preservation of the marine OC as a function of dissolved oxygen availability. The method relies on the dependency of $\delta^{13}C_{oc}$ on the concentration of dissolved, undissociated CO$_2$ [CO$_2$-aq] in surface waters, which has been shown to increase with increasing productivity and decreasing [CO$_2$-aq] in modern lacustrine (e.g. Hollander & McKenzie 1991) and marine (e.g. Freeman & Hayes 1992) environments. Normalization to carbonate-C isotopic values provides a correction for possible variations in the carbon isotopic composition of total dissolved carbon.

When applied to the Upper Jurassic Kimmeridge Clay (UK) (Hollander et al 1993) $\Delta \delta^{13}C$ decreased with increasing HI and this trend was interpreted as indicating that higher biologic productivity was the main factor in producing the most OC-rich, lipid-rich units. The Lower Jurassic Toarcian Shale of France exhibited the opposite trend—one which suggested that euxinic conditions were most responsible for increasing preservation of OC.

One must apply caution when using this approach for several reasons. First, the carbonate carbon isotope values used as a baseline are commonly altered diagenetically in black shales (e.g. Irwin et al 1977), rendering them more depleted in $\delta^{13}C_{cc}$ and therefore reducing the apparent $\Delta \delta^{13}C$. Secondly, changes in the dominant marine phytoplankton occur across black shale cycles (e.g. Prauss & Riegel 1989, for a Toarcian example), and sufficient data have not yet been collected to constrain possible species effects on $\delta^{13}C$.

**Sulfur to Carbon Ratios and Degree of Pyritization** Following the proposal of Leventhal (1983), the reduced S/C ratio in sediments has commonly been used to infer euxinic conditions (e.g. Beier & Hayes 1989). Leventhal (1983) originally suggested that a positive intercept on the sulfur axis of a sulfur-carbon crossplot and S/C ratios higher than the average of about 0.4 for oxic marine environments (Goldhaber & Kaplan 1974, Berner & Raiswell 1984) would denote deposition in an euxinic environment. Some subsequent workers have stretched this definition to include any set of values with higher S/C ratios. However, care must be taken in interpreting such data because of the role that availability of iron plays in fixing reduced sulfur in sediments (e.g. Raiswell & Berner 1985, Calvert & Karlin 1991). Many black shales are reactive-iron limited, i.e. Fe that can be reduced, solubilized, and made available for pyrite precipitation, particularly if they contain relatively high amounts of OC, calcium carbonate, or biogenic silica (Raiswell et al 1988, Dean & Arthur 1989). Thus, total reduced sulfur need not be high in sediment deposited in euxinic environments. Likewise, sediments deposited in an oxic environment which
are rich in reactive iron and have sufficient but low concentrations of reactive OC can have a high S/C ratio. Raiswell et al (1988) suggested that the “degree of pyritization” (DOP) is a more reliable index of euxinic versus anoxic deposition; DOP is the ratio of iron in sulfide minerals to total reactive iron (Fe-sulfides + remaining reactive Fe as determined by boiling with concentrated HCl). High proportions of sulfidized iron (high DOP) tend to occur in sediments deposited in euxinic environments, such as the Black Sea (e.g. Calvert & Karlin 1991). Raiswell et al (1988) suggested the following guidelines: DOP < 0.45 indicates oxic bottom water, for 0.45 < DOP < 0.75 one has “restricted” bottom-water conditions, and DOP > 0.75 results from “inhospitable” (euxinic) bottom-water conditions. They also suggest that for the sediment under analysis to be useful, it should contain < 65% skeletal material.

Figures 4b and 4c show Fe-S-C ternary diagrams (see Figure 4a and Dean & Arthur 1989, for construction and principles of interpretation) for the Black Sea and Peru margin, respectively. Peru margin samples (Figure 4c) come from either oxic sites (group 1), Fe-poor phosphorites (group 2), or dysoxic sites (all others along dashed line-3). The oxic (bioturbated with 1–9% OC) samples fall along the line representing S/C of 0.4, typical of oxic marine environments, whereas the dysoxic samples fall in an array along a line representing relatively constant S/Fe, regardless of OC content. The intercept on the Fe-S axis is at an S/Fe ratio of about 0.27. Because the theoretical S/Fe ratio is 0.54 for a sample where all Fe and S reside in pyrite, the plot indicates that only about 50% of the Fe is sulfidized. Although not a direct indicator of DOP, this rapid plotting technique has good utility in environmental determination. For example, DOP averaging 50% was determined for other Peru margin samples by Mossman et al (1992). The highest DOP values for Peru margin sediments would rate an assignment to the “restricted” facies using the above criteria. A Black Sea data set (Figure 4b; Calvert & Karlin 1991) allows discrimination between oxic shelf/slope (group 1) and euxinic marginal to deep-basin (group 2) environments. In this case, the intercept for the euxinic samples along a relatively constant S/Fe ratio represented by the dashed line is about 0.38 or about 70% of the Fe sulfidized. Calvert & Karlin (1991) independently determined a mean DOP of about 73% (from a range of 59–85%) for these samples. Using the criteria of Raiswell et al (1988), these samples would fall on the transition between “restricted” and “inhospitable.” The oxic samples had DOP values between 3 and 55%. Figure 4d shows an application to ancient black shales using data for the Pennsylvanian Stark Shale (Desborough et al 1990). Marginally oxic (group 1) vs anoxic (euxinic; group 2) clusters, based on other geochemical and fabric criteria, are clearly distinguished. For the samples thought to be deposited under euxinic conditions, the degree of Fe sulfidization and apparent DOP is high (about 87%). Therefore, although DOP is certainly one potential indicator of depositional conditions, it is not foolproof, and should not be used solely to differentiate between euxinic and marginally oxic conditions.

**TRACE METALS** Many ancient black shales are characterized by enrich-
ments in certain trace metals, particularly Cd, Ag, Mo, Zn, V, Cu, Ni, and U (e.g. Vine & Tourtelot 1970, Brumsack 1986, Holland 1979). Such enrichments are quite variable in magnitude and a number of possible causes for this enrichment exist. For example, it is often assumed for sequences of intercalated OC-rich and OC-poor beds that mobilization and migration of redox-sensitive trace elements from relatively OC-poor beds to OC-rich beds can account for enrichments. For many elements (Ag, As, Cd, Cr, Mo, Sb, U, V, and Zn) a sedimentary sequence more than 10 times as thick as the black shale layer has to be leached of its trace metal content with complete transfer to the black shale bed in order to account for enrichments (Brumsack 1986); in general, this mechanism does not explain relatively thick sequences of trace-metal enriched black shale. Other mechanisms, outlined below, include (a) biocenconcentration and transfer with particulate flux to the seafloor and/or sorption on biogenic and other particles in an oxic water column; (b) diffusion from bottom waters into sediments and fixation in authigenic phases associated with organic matter degradation by anaerobic bacteria; and (c) particulate flux associated with a water-column oxic/anoxic interface. The latter two mechanisms are explored briefly below as they bear on environmental interpretation of black shales.

In addition to the "direct" contribution from the particulate flux, OC-rich sediments receive trace metals during or shortly after deposition. Oxygen is consumed and disappears at or near the SWI, and ensuing bacterial sulfate reduction causes the generation of hydrogen sulfide. Metals like As, Mo, U, and V are sensitive to changes in redox potential and the presence of hydrogen sulfide. These elements may diffuse from seawater (particularly eucinic water column; see below) into the sediments where they are bound as sulfides (As, Mo) or possibly bound to organic matter (U, V) in their reduced states (Veeh 1967, Brumsack 1986). Because primarily molecular diffusion is involved, the magnitude of enrichment of such elements is likely inversely proportional to the sediment accumulation rate (Anderson et al 1989).

Under eucinic conditions, such as in the present Black Sea or Cariaco Trench, the nutrient/trace metal patterns differ from those observed in oxic waters. Many trace metals undergo a dramatic solubility decrease at the O₂/H₂S–boundary (e.g. Spencer et al 1972, Jacobs et al 1985 1987) and are removed from the water column, usually by precipitation as sulfides, and finally buried in the sediments. Again, this trace element enrichment may be more noticeable when sedimentation rates and corresponding dilution by terrigenous detritus are low. Therefore, sediments deposited under an anoxic water column, with low terrigenous input, and consequent high-OC preservation should exhibit a strong trace metal signal. By contrast the trace metal content of sediments that were deposited in an environment with enhanced surface-water biological productivity and concomitant high sedimentation rates should be comparable to those of modern upwelling sediments, regardless of the presence of an anoxic water column.

It appears then that trace element enrichment in black shales can occur in a variety of ways, and that there is no perfect discriminator of the conditions under which a given black shale was deposited. However, at least two trace-element-based techniques may be of use, when combined with other data: Th/U and V/(V + Ni).

The Th/U ratio is sensitive to the relative inputs of Th-bearing clay minerals and U that is fixed in sediments during early diagenesis in eucinic environments. Zelt (1985), Wignall & Myers (1988), and Doveton (1991) have applied this technique in studies of sedimentary sequences using spectral gamma logs (either hand-held or wireline), and suggest that low Th/U reflects low redox conditions. Soluble uranium in its oxidized state must be reduced and fixed as tetravalent U in the sediments of eucinic basins as opposed to its unreactive behavior in oxic marine basins (Anderson et al 1989). The precipitation of U in the sediments drives diffusion of U into pore waters, thus further enriching U in these OC-rich sediments. Because Th is scavenged by particles and supplied in the elatic fraction, high rates of sedimentation, even in a eucinic basin, will tend to cause high Th/U ratios. Only at relatively low sedimentation rates will the Th/U ratio be a sensitive indicator of a eucinic water column. For example, Mangini & Dominik (1979) demonstrated that the U/OC ratio is inversely proportional to the sedimentation rate in Mediterranean sapropels. This pattern has been confirmed for deep basin sediments of late Holocene age in the Black Sea (Anderson et al 1989). Again, some care must be exercised in applying this technique.

Vanadium is apparently more effectively fixed in sediments in association with organic compounds (e.g. porphyrin compounds) when competing metals, particularly Ni, are removed from anoxic pore waters as sulfide complexes (Lewan & Maynard 1982). Vanadium enrichments in OC-rich sediments appear to be enhanced by high concentrations of dissolved sulfide, especially in the overlying water column (causing V reduction), and slow sedimentation rates, which allow sufficient diffusion of V (reduced by sulfide) into sediment pore waters (e.g. Breit & Wanty 1991) where it can bind with porphyrin complexes. Increasing V enrichment over Ni may therefore indicate prevalence of sulfidic (eucinic) conditions and variations in the V/(V + Ni) ratio would indicate relative changes in oxygenation in the water column. The efficacy of this technique, however, is reduced where high sedimentation rates prevail (both limiting V diffusion and diluting trace metals overall) and/or where there is no
source of sufficiently well preserved organic matter with attendant geoporphyrins, regardless of the presence or absence of euxinic conditions.

Examples from two modern OC-rich environments are shown in Figure 5. Holocene Black Sea sapropels (Unit II; Arthur et al 1994) are enriched in trace metals (e.g. Calvert 1990, Brumsack 1989), including V and Ni. Figure 5a is a plot of V/(V + Ni) vs %TOC for Unit II samples of laminated sediments from gravity cores taken from shallow to deep water depths (200–2250 m) in the Black Sea, all from below the oxic/anoxic interface. In these sediments, OC concentrations increase toward deeper water and lower sedimentation rates. V/(V + Ni) values are moderately high for all samples (because all were from euxinic environment) and increase slightly with increasing TOC, probably because of more effective diffusion into sediments at slower sedimentation rates (the black squares are from a site anomalously enriched in Ni and therefore do not lie on the same trend). The Peru margin data (Figure 5b) show no trend with increasing TOC, and the highest as well as lowest values of V/(V + Ni) occur in oxic slope sites (Bx2, 5, and GS6 are oxic, the remainder dysoxic; Froelich et al 1988); it is likely that V is sorbed onto oxyhydroxides in the surface sediments there. Because there are no true euxinic conditions in the Peru OMZ, there is little opportunity for V reduction in the water column. Nonetheless, samples from Bx6 have TOC and V/(V + Ni) values in the middle range of euxinic Black Sea sediments. V/(V + Ni) values for ancient black shales also exhibit little relationship to TOC, as in the case of the Pennsylvania Stark Shale (Figure 6b), and it is likely that this is because TOC concentrations are not strictly a function of euxinic conditions. On the other hand, there is a strong correspondence between V/(V + Ni) values and DOP in the Stark Shale (Figure 6a), suggesting that higher V/(V + Ni) values signal more strongly euxinic conditions of deposition. Again, some care must be taken in applying this technique.

Biofacies Analysis and Paleo-Oxygenation History of Black Shales

Due to excellent preservation, which is a common characteristic of fossil assemblages within organic-rich facies (Canfield & Raiswell 1991a,b), there is a robust record of paleobiologic data from the study of black shales. As a result of early investigations of Devonian, Pennsylvanian, Jurassic, and Cretaceous deposits in the United States and Europe, for example, extensive species lists were compiled (e.g. Ruedemann 1935, Hauf 1921, Einsele & Mosebach 1955). In most cases, however, fossil occurrences were not recorded in detailed stratigraphic sequence, and little taphonomic information is available. Moreover, due to the prevailing interpretation of black shale environments as “stagnant” anoxic seas, black shale faunas were largely believed to have been transported into these environments from shallower areas or from life habits as epi- or pseudoplankton. Yet, throughout the literature there are references to the persistent and widespread occurrence of small, thin-shelled taxa that are normally benthonic, as well as to shell horizons containing abundant specimens of a single or few species that suggest brief episodes of opportunistic benthic colonization.

Following a review of trends in faunal assemblages and associated biogenic sediment fabrics of modern low-oxygen environments, Rhoads & Morse (1971) suggested a model for the interpretation of oxygen-deficient biofacies. In this model, three distinct biofacies were recognized: anoxic, dysaerobic, and aerobic. Each biofacies was defined by specific faunas and characteristic sediment fabrics that were correlated to specific ranges of dissolved oxygen (Figure 7a). This scheme was the first to provide a means of reconstructing paleo-oxygenation histories in OC-rich facies, and has persisted as the central framework for interpretation of oxygen-restricted black shale faunas. Refinement of the model, resulting from increasingly detailed studies of both modern and ancient low-oxygen faunas (e.g. Byers 1977; Morris 1980; Savrda et al 1984, 1991; Thompson et al 1985; Savrda & Botteri 1986, 1987; Wignall & Myers 1988; Brett et al 1989; Sageman 1989, 1991; Kaufman & Sageman 1990; Wignall 1990; Sageman et al 1991; Wignall & Hallam 1991) has included the recognition of 1. finer subdivisions of dysaerobic biofacies, 2. patterns of dominance among infaunal vs epifaunal taxa, and 3. trends in low-oxygen community structure, including trophic group dominance, species richness, and community equitability. These improvements have made possible even more detailed paleoenvironmental reconstructions of OC-rich facies.

Figure 7b shows an example of a current black shale biofacies model, derived mainly from detailed study of the Late Cenomanian Hartland Shale Member of the Greenhorn Formation in the Western Interior basin (Sageman 1989, 1991; Kaufman & Sageman 1990; Sageman et al 1991). It illustrates a number of the revisions that have been made to the Rhoads & Morse (1971) scheme. In the Hartland Shale Model:

1. Two main biofacies categories are identified—one dominantly infaunal, the other dominantly epifaunal.
2. Seven subdivisions (biofacies levels) are recognized, spanning the range from anaerobic to aerobic.
3. Each biofacies is defined by characteristic taxa with specific life habits and trophic strategies, characteristic sediment fabrics related to the level of infaunal colonization, relatively well-constrained levels of species richness and equitability, and in some cases, recurring ecological relationships between component taxa, such as commensalistic host-epibiont interactions.
4. The biofacies levels are generally cumulative upward with respect to both taxa and trophic strategies (specific taxa, or members of the same trophic group in any given level may occur at all higher levels), reflecting increasingly favorable conditions for benthic habitation.

**Infaunal Biofacies** Predominantly infaunal biofacies are described from strata with greater than 90% of the total fauna represented by sediment feeding or dwelling ichnotaxa, and/or body fossils of organisms that lived infaunally. Infaunal biofacies level 1 (Figure 7b) reflects anoxic conditions both above and below the SW1, with laminated sediments, the highest levels of preserved TOC, and no evidence of benthic metazoan life. At biofacies level 2, microborrowing—described from the Greenhorn Formation by Pratt (1984) as the slight disruption of primary laminae by burrows less than 1 mm in diameter—probably indicates the presence of meiofaunal organisms (such as nematodes) in the uppermost layers of the sediment. Level 3 refers to small deposit-feeding pioneer ichnotaxa, such as Planolites or Chondrites, which are widely recognized as the traces of the most low-oxygen tolerant macroinvertebrate infauna (e.g. Bromley & Ekdale 1984, Savrda & Bottjer 1986, Ekdale & Mason 1988). Level 4 is characterized by an increase in the the size and density of Planolites and Chondrites burrows, but no additional ichnotaxa, or by the appearance of deposit-feeding infaunal bivalves, such as Nucula or Lucina. At level 5 more complex feeding traces, such as Teichichnus and Zoophycos may be found, as well as the body fossils of infaunal grazers such as the anhurid gastropod Drepanoichius. Level 6 is indicated by the appearance of sediment-dwelling ichnotaxa ("domichnia": Seilacher 1964), such as Thalassinoides, which indicate the establishment of domiciles within the substrate (see Figure 9). Above level 6 all groups may be included to produce high-diversity, high-equitability assemblages, indicating normal aerobic biofacies.

Although the transition from infaunal biofacies levels 1 to 7 (Figure 7b) represents a progressive increase in community diversity and equitability, the extensive reworking of sediment that occurs at higher biofacies levels may reduce the number of taxa ultimately preserved, thus making evaluation of true diversity trends quite difficult. For example, in cases where later episodes of burrowing completely overprint earlier records, and only a single ichnotaxon may be preserved, diversity and equitability estimates...
cannot be made. In cases where burrowing is extensive, careful evaluation of gradations of ichnofabric (e.g. Droser & Bottjer 1986) may be the best proxy for trends in the infaunal community.

**EPIFAUNAL BIOFACIES** The predominantly epifaunal biofacies series (Figure 7b) is described from strata with greater than 90% of the total resident fauna composed of epifaunal taxa. As with the infaunal biofacies series, epifaunal biofacies level 1 reflects anoxic conditions above and below the SWI, as indicated by laminated sediments, the highest levels of preserved TOC, and no evidence of in situ benthic habitation. Level 2 is characterized by laminated sediments containing only ammonites. These taxa are believed to have been nektobenthonic scavengers and predators that were mobile within the water column, and thus could feed on dead or dying organisms at or near the toxic SWI. Level 3 is characterized by pioneer suspension feeders, a group of organisms referred to broadly as “flat clams,” which are ubiquitous inhabitants of Phanerozoic organic-rich facies (Kauffman 1988a). These taxa, which include mainly epibyssate suspension-feeding bivalves (e.g. *Inoceramus*) or free-living to pediculate brachiopods (e.g. *Leiothyris*) characterized by extremely thin shells, broad/flat shapes, and reclining to slightly erect life habits, were highly adapted to life at the boundary between anoxic and dysoxic to oxic environments. As indicated in Figure 7b, the lower levels of dysoxic biofacies in the epifaunal sequence are characterized by a redox boundary coincident with the SWI. The flat clams which characterize these “benthic boundary biofacies” had special adaptations to low oxygen levels and H₂S, which may have included chemosymbiotic relationships with sulfur-oxidizing bacteria (Kauffman 1988a, Kauffman & Sageman 1989, Sageman et al 1991). As a result, they were the most common epifauna to colonize low-oxygen, OC-rich substrates. At biofacies level 4, pioneer flat clams are joined by successor suspension feeders, taxa that lived as epibionts on the larger flat clam “shell islands” (Kauffman 1981). Level 5 is indicated by the first appearance of opportunistic epifaunal colonizers, such as small bivalves and ostracodes, indicating that the substrate surface was habitable to taxa other than the highly tolerant flat clams. Level 6 is marked by a significant increase in diversity with the addition of various surficial grazers, scavengers, and carnivores, and indicates the initial development of an ecologically complex community. At level 7, all epifaunal groups are included, resulting in the high diversity and equitability levels characteristic of normal aerobic biofacies.

**BIOFACIES SUMMARY** Application of the biofacies model described above to sections of the Hartland Shale Member that have been measured and collected in great detail (see example, Figure 9) highlights several points concerning black shale biofacies. First, taxa occur either as resident components of benthic communities (they comprise >90% of an assemblage, are distributed throughout a sample interval, and/or show evidence of long-term in situ habitation), or as event components (they occur in low number, comprise <10% of an assemblage, occur on single bedding
planes, and/or show evidence of benthic colonization followed by abrupt mass mortality). The high frequency of event biofacies in many black shales indicates the dynamic nature of low-oxygen benthic environments. Secondly, at higher biofacies levels of the dysaerobic range (levels 5 and 6), fossil assemblages are typically dominated by mixed infaunal-epifaunal biofacies. “Pure” biofacies only occur in the lower part of the dysaerobic range. Finally, whereas changes in biofacies levels have been assumed to reflect mainly changes in benthic oxygen and the position of the redox boundary, major shifts in biofacies type are commonly associated with changes in lithofacies and indicate that the nature of the substrate may play a major role in the colonization of low-oxygen tolerant taxa. Ultimately, biofacies models can be seen as a tool not only to reconstruct paleo-oxygenation trends, but also to determine paleoecological factors that control community development under conditions of oxygen deficiency.

The biofacies model described herein is only one of several including that of Savrda & Bottjer (1991)—who developed a similar scheme based mainly on analyses of ichnotaxa in Miocene and Cretaceous organic-rich facies, and that of Wignall & Hallam (1991)—who describe a biofacies model for British Jurassic black shales, focused mainly on megafoils. Although there are numerous differences between these three models, reflecting both different paleoecological approaches, as well as data bases widely separated in space and time, there is also a remarkable convergence among the three upon a common framework (see Figure 7c). All three models address the complexities of infaunal and epifaunal life habit groups and the importance of short-term event communities. 

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Figure 7  Major features of black shale biofacies models. Characteristics of the Rhoads & Morse (1971) biofacies are shown relative to current terminology for correlative states of benthic oxygenation (based on Tyson & Pearson 1991). Biofacies model based on studies of the Cretaceous Greenhorn Formation, Western Interior basin (modified from Sageman 1991) illustrates revision of the Rhoads & Morse (1971) scheme—including recognition of variation in the position of the redox boundary and the development of (a) predominantly infaunal and (b) predominantly epifaunal biofacies, finer subdivision of the anoxic to aerobic transition (dysaerobic biofacies), and detailed paleoecological characterization of each biofacies based on trends in trophic specialization, species diversity, and sediment fabric. Trophic pie charts and faunal examples represent averages of all samples analyzed in study of the Hartland Shale Member (Sageman 1991). [For further details see Sageman (1989), Kaufman & Sageman (1990), and Sageman et al. (1991). (c) Comparison of the Greenhorn-based model to biofacies models developed by Savrda & Bottjer (1986, 1987), Savrda et al (1991), and Wignall & Hallam (1991; see also Wignall & Myers 1988, Wignall 1990) illustrating consistent recognition of similar oxygen-related biofacies in strata of Jurassic, Cretaceous, and Miocene age, from both North America and Europe. 

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Application of Biofacies and Geochemical Techniques

To illustrate how the biofacies models can be combined with geochemical data to interpret depositional environments of ancient black shales we show two examples: 1. the Holocene Black Sea (Figure 8); and 2. the Upper Cenomanian-Lower Turonian basinal marine sequence from the U.S. Western Interior Seaway (Figure 9). The data in Figure 8 illustrate the use of geochemical indicators to detect possible changes in oxygenation for a case in which the sediments are entirely finely laminated and devoid of benthic fauna. These so-called sapropels were deposited during an interval from about 8–2 ka in the deepest part of the Black Sea. Although the deep-water mass has been presumed to have been euxinic throughout this period, Calvert (1990; see also Pedersen & Calvert 1990) suggested that much of the Black Sea Unit II sapropels were deposited under oxic conditions (with high productivity). The data shown in Figure 8 may argue for a brief episode of oxygenation that would have caused the decrease in V/(V+Ni), increase in Mn/Al, and decrease in TOC and pyrolysis HI at about 40 cm in the core. Note the transition from oxic Unit III to inferred euxinic Unit II at the base of the core. At present, we do not understand the origin of the mid-Unit II apparent oxygenation event. However, it does appear that geochemical data are sensitive indicators of changes in oxygen concentrations at very low levels.

Figure 8  Selected geochemical indices—V/(V+Ni), (Mn/Al) wt. ratio, % organic carbon (TOC), and pyrolysis hydrogen index (HI)—for laminated, deep-basin Black Sea Unit II sapropel (in giant gravity core 66; Dean & Arthur, in preparation). Note that TOC, HI, and V/(V+Ni) all decrease centered at about 40 cm depth whereas Mn/Al increases to a peak. Such changes in these indices might record a brief interval of oxygenation in the deep Black Sea at that time (perhaps 4000 ka).

Shown in Figure 9 are our data for a 23-meter section of the Hartland Shale and the lower part of the Bridge Creek Limestone Member of the Greenhorn Formation. The Hartland Shale is dominantly laminated to sparsely bioturbated with relatively uniform OC concentrations ranging upwards to 4.5% and HI values of about 400 mgHC/gOC. Bioturbation and biofacies analyses suggest deposition primarily under dysoxic conditions that allowed an epifaunal community to survive throughout deposition of most of the unit. Note that species richness and abundance do not necessarily indicate the severity of oxygen depletion suggested by other indicators. The redox boundary is interpreted to have been mainly at or just below the SWI. Sedimentation rates are relatively slow (pelagic) throughout and the V/(V+Ni) values are uniformly high (>0.75). Such high values might suggest deposition under euxinic conditions, but DOP values (not shown) of 0.4 to 0.7 suggest “restricted” but not inhospitable bottoms, in agreement with the biofacies and textural evidence. Mn/Fe values are extremely low, with the exception of one limestone bed, supporting the interpretation of low redox conditions at the SWI. The Hartland Shale is interpreted as a fully marine black (marly) shale unit deposited in an overall transgressive epicontinental sea that was oxygen-depleted but not euxinic. Productivity and high OC fluxes may have played a role in maintaining the redox boundary close to the SWI, but low dissolved oxygen concentrations in the bottom-water mass, perhaps derived by transgression of an external OMZ into the basin from Tethys, may have been important.

A significant change in depositional environment occurs at the contact between the Hartland Shale and Bridge Creek Limestone. The lower part of the Bridge Creek Limestone (S. gracile zone) is characterized by more intense bioturbation, biofacies indicating amelioration of redox conditions, low TOC, low HI values, a decrease in V/(V+Ni) values, and fluctuating Mn/Fe. All of these parameters indicate fairly oxic depositional conditions. As these conditions occurred during a transgressive episode, some aspect of water mass stratification must have changed significantly. It is difficult to invoke decreasing productivity in surface waters, and, more likely, the transgression altered the water balance of the basin and led to production of oxygenated “deep” waters within the Western Interior Seaway.

Strongly cyclic alternations between well-oxygenated and dysoxic bottom conditions prevailed for the remainder of the Bridge Creek Limestone as depicted by large variations in bioturbation index, biofacies composition, carbonate and OC concentrations, and V/(V+Ni) and Mn/Fe values. The cyclicity may reflect modulation of climate and water mass conditions by orbital insolation variations (Milankovitch cycles; e.g. Pratt...
1984, Barron et al 1985, Kauffman 1988b). The Pratt model suggests that periodic increases in runoff from highlands to the west of the seaway on 20–100 ky timescales caused surface salinity decreases and increased stratification with concomitant oxygen depletion in bottom waters. The variations in carbonate content appear to reflect dilution cycles (Pratt 1984, Arthur & Dean 1991). However, even these cycles are debated, and Eicher & Diner (1989), for example, prefer productivity variations as a cause, with the carbonate-rich units representing higher productivity. Deposition of this part of the Bridge Creek Limestone corresponds to the early stages of the global Cenomanian/Turonian OAE discussed later.

PHANEROZOIC BLACK SHALE DEPOSITION: TRENDS AND MODELS

Studies of black shales are legion. Yet few deposits are known in detail and definitive paleoenvironmental models are lacking for most deposits. The debates that characterize interpretation of Cretaceous black shales apply to Paleozoic examples as well. However, for the Paleozoic, it is even more difficult to establish organic carbon accumulation rates and other parameters needed to quantify carbon fluxes and establish the relative role of biologic productivity. As discussed earlier, oxygen depletion, either in silled basins, OMZs, or globally, has been interpreted as playing a major role in the formation of black shale units, but there are a number of examples attributed to higher productivity associated with upwelling zones as well.

There apparently were times in the past during which black shales were more widespread. The occurrences of major units has been synthesized by North (1979, 1980), Bois et al (1982), and Ulmishek & Klemme (1990); most of the events recognized are referred to in terms of hydrocarbon generation. Leggett (1980) was one of the earliest to point out the apparent coincidence of black shales across parts of Europe in relatively narrow time intervals in the early Paleozoic. Many of the major Phanerozoic episodes of black shale deposition correspond to positive carbon isotope excursions (Figures 10 and 11) which may indicate increased burial of isotopically light organic matter during these black shale depositional intervals. Note also that the most widespread events appear to correspond with overall higher sea level stands. Because, in contrast to the late Mesozoic, there is no substantial preserved deep-sea record, we cannot adequately assess the possible global extent of oxygen-depleted waters or OC-rich facies. However, it does appear that most black shales were deposited in somewhat deeper-water shelf settings even though continental flooding meant that epicontinental seas were more widespread at those times.

Although black shale deposits typify the Cambrian through Lower Silurian in general, data compiled by Ulmishek & Klemme (1990) indicate that Silurian black shales were particularly effective hydrocarbon source beds. For example, they suggest that OC-rich, graptolitic strata of Silurian age covered approximately 42%—a huge proportion—of the total Silurian depositional area. The organic matter was entirely of marine origin (type II), most likely because terrestrial vascular plants were only just evolving (see Thickett 1984 and Leventhal 1991 for examples and models). Upper Devonian–Early Mississippian black shales are perhaps the most heralded of Paleozoic units. Ulmishek & Klemme (1990) claim that OC-rich strata extended over about 21% of the marine depositional area in latitudinal positions generally lower than 45°. Again, the organic matter was dominantly marine (Type II), despite the more widespread vascular plants (for detailed examples see Cluff 1980, Ettensohn 1985, 1992). The Late Pennsylvanian–Early Permian is another time of widespread OC burial (Ulmishek & Klemme 1990). The dominant type of organic matter is terrestrial plant material (type III). This is consistent with the known distribution of coal deposits which peaks in this interval. In fact, by comparing sulfur isotope curves with carbon isotope curves and other data, Berner & Raiswell (1983) argue that the major positive carbon isotope excursion (Figure 10) resulted from burial of terrestrial organic matter in coals and associated black shales [for details of deposits and contrasting models see Heckel (1977, 1986), Boardman et al (1984), Zangerl & Richardson (1963), Wenger & Baker (1986), Cecil (1990), and Coveney et al (1991)].

The next major phase of widespread OC burial appears to have occurred in the Upper Jurassic following the major regressive period of the Permian through early Jurassic (Figure 10). Ulmishek & Klemme (1990) suggest that, on average, 27% of the marine Jurassic depositional area was typified by OC-rich strata, that the organic matter was dominantly Type II, and that the latitudinal extent of OC-rich facies was much greater than in the Paleozoic. Two major widespread Jurassic episodes have been cited (Jenkyns 1983, 1988; Jenkyns & Clayton 1986; Aigner 1980; Tyson 1986; Wignall 1990; see also Duff 1975, Miller 1990, Hollander & al 1993, Herbin et al 1991 for examples and models). The mid-Cretaceous (Aptian-Turonian) is also considered a major source-bed depositional interval, with black shales deposited over about 27% of the marine area (Ulmishek & Klemme 1991). The record can be resolved into several discrete episodes of widespread OC deposition (Figure 11; see Schlanger & Jenkyns 1976,
As for the general causes of black shale episodes, simple and more complex models of climate and ocean circulation can be used to effectively test assumptions about regions of upwelling and higher productivity, higher precipitation/runoff and consequent salinity stratification, areas of deep-water formation, changes in rate or mode of deep-water circulation, and the effects of changing temperature and oxygen solubility or nutrient concentrations on carbon flux, deep-water oxygen concentration, and organic carbon burial rates. Each of these parameters is potentially important in promoting widespread organic matter burial. For example, several studies have emphasized changing wind patterns and coastal upwelling to explain the distribution of black shales (e.g. Parrish & Curtis 1982, Scotese & Summerhayes 1986); numerical general circulation models have been employed to examine patterns of coastal upwelling expected from mid-Cretaceous (e.g. Kruis & Barron 1990, Barron 1985) and Silurian (Wenlockian; Moore et al 1993) changes in paleogeography.

Another side of the issue is the role of deep-water circulation and anoxia. For example, Wilde & Berry (1982) developed conceptual oceanographic models involving deep circulation changes for the overall pattern of black shale development in the Paleozoic. They suggested that the early Paleozoic ocean was dominated by largely oxygen-deficient deep waters whereas the later Paleozoic deep ocean became more oxic because of more efficient deep circulation associated with glacial climates, particularly in the Pennsylvanian-Permian. Episodes of heightened productivity and black shale deposition have been attributed to a variety of possible causes, including increased deep-water mass formation forced by rising sea level, impact-induced or volcanic-induced upwelling (e.g. Vogt 1989), and/or opening of a deep-water gateway. Somewhat more sophisticated box models can also be manipulated in order to examine the influence of different deep-circulation styles and rates on average deep-water properties. If one employs present-day parameters for exchange rates, preformed nutrients, river input, etc., but allows warm, saline bottom water (WSBW; Brass et al 1982) to form in low latitudes or by other means (mixing at oceanic fronts, Hay et al 1993) instead of sinking of cooler high-latitude water masses, then the average oxygen content of deep-water masses falls to about 36 μmol O₂/kg instead of the present average of about 160 μmol/kg (Sarmiento et al 1988). An average value of dissolved O₂ of 36 μmol/kg suggests that much of the bottom-water mass is anoxic. Ocean deep-water masses can also be made to go anoxic, using a model modern thermohaline circulation, by increasing deep-water residence times by a factor of 2–10 (see also Wilde & Berry 1982; Bralower & Thierstein 1984, 1987), depending on the turnover rate of the intermediate water mass.

Thus, widespread deeper-water anoxia can result from average surface-ocean productivities lower than the present with a warmer deep ocean and lower initial oxygen solubility. Such changes in deep-water turnover rates in either low or high latitudes may be cyclic and related to climatic changes induced by periodic changes in solar insolation, leading to the Milankovitch-like cycles in many Cretaceous black shales (e.g. Barron et al 1985, Herbert et al 1986).

CONCLUSIONS—PROBLEMS AND PROSPECT

Ancient black shales provide a fertile ground for future research because of their importance as hydrocarbon source rocks and their usefulness as environmental indicators in furnishing clues to secular changes in atmospheric and ocean chemistry and ocean circulation. It is a struggle, however, to uniquely define the full conditions under which a given black shale was deposited. No single parameter can be used to delimit the extent to which black shales reflect depocenter and eutinic conditions, with or without high biotic productivity. However, there is good potential for interdisciplinary studies using a combination of benthic biofacies and geochemical techniques, at least for defining oxygen gradients during deposition.

We do not yet fully understand the relative roles of organic carbon flux, dissolved oxygen (anoxia), and rate of sediment accumulation in producing relatively OC-rich sediments in modern environments, nor can we satisfactorily explain why preservation should be enhanced under anaerobic conditions, if it indeed is. Further careful studies of suitable modern environments are required to resolve these questions before ancient black shales can be satisfactorily interpreted in terms of the importance of productivity vs preservational phenomena. Better stratigraphic resolution and constraints on accumulation rates are required for ancient black shales.

There are clearly a number of temporal intervals during which marine black shale facies are much more widespread. These episodes are of interest because they implicate one or more global factors controlling organic matter production and burial. Widespread black shale deposition may provide evidence that oceanic nutrient concentrations were significantly higher than at present and/or that dissolved oxygen was profoundly depleted in oceanic deep waters as the result of any of several factors, including sluggish circulation, carbon loading, or low solubility of dissolved O₂ because of warm deep-water temperatures or low atmospheric pO₂. Considerably more investigations of the stratigraphic significance of black shale occurrence are needed. Likewise, these episodes are of interest...
because of the geochemical cycle linkages and the relationship between black shales with ore deposits and biotic extinctions. In all of this, we should keep in mind the admonition of Degens & Stoffers (1976) that “The well-mixed oxygenated ocean of today seems not to be the model for the past 600 m.y.”

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