Chapter 2: Deposition of Mudstones and Shales

Deposition of Mudstones and Shales: Overview, Problems, and Challenges

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The Problem

As pointed out in the introduction to this book, shales contain the lion's share of earth history that is recorded in sedimentary rocks. Understanding their deposition is key to the reconstruction of past oceans, landscapes, climates, and climatic cycles. Yet, whereas students of sandstones and carbonates have been able to learn a great deal about the deposition of these rocks from studying bedforms and sedimentary structures in the context of experimental data and direct observations in modern environments (e.g. Harms et al., 1982; Friedman et al., 1992), the situation is not as fortunate when we examine experimental and observational data to assess the significance of sedimentary features in fine-grained terrigenous clastics.

Experiments and Observations

What do we know? Unlike sands, the erosion and deposition of muds is not primarily governed by particle size and flow velocity. Interparticle cohesion and the degree of consolidation (proxies are water content and plasticity) play a major role. In the classical studies by Hjulström (1955) and Sundborg (1956) it was concluded that muds require larger current velocities for erosion than fine sands (because of cohesive forces), that velocities required for mud erosion can be of the same order of magnitude as required for erosion and transport of gravel, and that the flow velocity at which erosion begins depends on the degree of consolidation. The latter point has been confirmed and elaborated upon by (among others) Parthenaides (1965), Southard et al. (1971), and Lonsdale and Southard (1974). Subsequent studies by Einsele et al. (1974) and Migniot (1968) showed that the subject of mud erosion is unfortunately much more complex than originally determined by Hjulström (1955) and Sundborg (1956). Einsele et al. (1974) explored the influence of clay type, void ratio, and shear strength on threshold erosion velocities, and found that the "geologic history" of a mud plays a significant role as well. Under "geologic history", Einsele et al. (1974) include whether a mud was deposited from dilute or dense suspensions (slurries), the fabric that developed during settling and consolidation, and the distribution of inhomogeneities.

Young and Southard (1978) studied erosion of marine muds in the laboratory and with in situ measurements, and demonstrated that threshold erosion velocity decreases with an increase in the degree of bioturbation, and that there is a systematic increase in threshold erosion velocity with organic content in nonbioturbated sediments. They also found that the bioturbated marine sediments in their study required lower threshold erosion velocities than one would expect from predictions made in the works of Hjulström (1955) and Sundborg (1956). The latter authors predicted that muds should be more difficult to erode than fine sand, whereas Young and Southard (1978) found that the opposite was the case for the muds in their study. They explained the discrepancy with the lower bulk density of biologically aggregated sediments. Young and Southard (1978) also noted significant lateral variability in threshold erosion velocity in their seafloor study. Because bulk physical properties of the sediments varied very little, they suggested that the differences in threshold erosion velocity are related to more subtle and as yet unmeasured physical and biochemical parameters.

More recent data on erosion experiments in natural muds are reported by Schünemann and Kühl (1993), who found that biological factors, such as microbial mats, density of sea grass, and diatom populations have a stabilizing effect (they can increase threshold of erosion velocities by as much as a factor of 4), whereas grazers and burrowers lower threshold erosion velocities. A related useful study is that by Neumann et al. (1970) where in situ measurements of the stabilizing effect of microbial mats on sandy sediments are reported.

Because in seawater muds will flocculate and behave as aggregates rather than individual particles, flocculation is a major factor in the rapid settling of muds in nearshore environments. Kranck (1991) presents experimental data that indicate that shale microfabric features can be explained by the mechanics of flocculation, and gives references to other studies of flocculation. The settling modes of suspended particles in the marine environment are discussed by Syvitski (1991), who distinguishes settling as single particles (river mouth, eolian), floccules (halocline regions surrounding river plumes), organo-mineralic agglomerates ("marine snow"), and planktonic fecal pellets. The latter are a seasonal phenomenon (Syvitski, 1991), and can cause the coupling of organic carbon flux and clay deposition on the seafloor (Fischer, 1991).

With regard to observations of mud deposition in modern environments, important studies of that subject were for example conducted by the Senckenberg Institute in Wilhelmshaven/Germany under the direction of H.-E. Reineck. Important publications include those of Reineck et al. (1967, 1968), Reineck and Wunderlich (1969), Reineck and Singh (1972), Reineck (1974), Wunderlich (1969, 1978), and Aigner and Reineck (1982). Reineck et al. (1967, 1968) used a holistic approach to the study of the muddy North Sea shelf, and based on descriptions of sediment cores, petrographic analysis, and faunal analysis, arrived at a sedimentologic model that explained facies and faunal zonations in the study area. These papers also contain some of the earliest descriptions and process interpretations of sandy and muddy storm deposits in shelf sequences (Reineck et al., 1967, 1968; Reineck and Wunderlich, 1969; Reineck and Singh, 1972; Reineck; 1974). The study by Reineck and Wunderlich (1969) on the development of bedding and bedforms in a tidal setting contains excellent descriptions and insightful interpretations of sedimentary features in muddy sediments. Wunderlich (1969) investigated settling velocities, compaction, resuspension and erosion in tidal environments, and found that floccules can be important for suspension transport of sand (floccules serve as "parachutes" for sand grains) and the later development of sandy bedforms in areas that are not within the reach of bedload transport processes. He also found (Wunderlich, 1978), that flocculation can cause deposition of very thick (2 cm) mud deposits within very short time periods (30 minutes) during slack water periods. The study by Aigner and Reineck (1982) on storm deposits and proximality trends in North Sea muds extends earlier work on storm deposits in the North Sea, and has definite applicability to ancient shale successions (e.g. Schieber, 1989, 1992, 1994a) Many of the sedimentary features described in above studies, such as the various styles of flaser bedding, wavy bedding, and lenticular bedding, are summarized and pictured in the sedimentology textbook by Reineck and Singh (1980).

McCave (1970, 1971) also investigated the influence of tidal currents and waves on mud deposition in shallow marine settings. From field data and theoretical considerations he concluded that there is no need to assume that muds that are associated with sand were necessarily deposited under conditions of lower current velocity, lower wave activity, or greater water depth. An increase of the suspended mud concentration can change an area of sand deposition to one of mud deposition. That substantial amounts of mud can indeed be deposited in energetic environments when suspensions are concentrated enough is for example indicated by the presence of muddy shoreface deposits along the Guiana coast (Allison and Nittrouer, this volume).

Surface layers of fluidized mud are an important element in areas of "high energy" mud deposition of modern shelves and coastal areas (e.g. Nair, 1976; Rine and Ginsburg, 1985; Mallik et al., 1988; Kirby, 1991; Ross and Mehta, 1991). These fluid mud layers can very significantly dampen wave action (Nair, 1976; Mallik et al., 1988) and may in some areas be responsible for rapid deposition of thick graded mud layers during periods of stagnation (Kirby, 1991). Sedimentary features in some of these muds record lateral movement in form of a slurry-like flow (Rine and Ginsburg, 1985) and also indicate intermittent erosion and bedload transport of coarser particles (Rine and Ginsburg, 1985; Allison and Nittrouer, this volume; Kirby, 1991). Descriptions of sedimentary features and stratification sequences by these authors may prove helpful to identify and interpret fossil analogs.

A number of papers on shallow marine mud deposition has been published in recent years by Charles Nittrouer and collaborators (e.g. Allison and Nittrouer, this volume; Nittrouer et al., 1986; Alexander et al., 1991a, 1991b; Kuehl et al., 1986a, 1986b, 1988, 1991; Segall and Kuehl, 1994). The methods that were employed in these studies include high resolution seismic, coring, X-radiography, thin section examination, and grain size analysis. The geographic focus were shelf areas in the vicinity of large rivers that contribute huge quantities of mud to the shelf (Amazon, Yellow River, Yangtze River, Ganges-Brahmaputra). An essential element of these studies was the examination of petrographic thin sections of epoxy-stabilized muds, revealing mm-scale sedimentary features that can not be observed with other techniques (such as X-radiography and SEM). Using thin sections, these authors also paid close attention to details of thin silt and sand layers (sharpness of basal and top contacts, types and size distribution of constituents, thickness variations of internal laminae, micro-cross-laminae, rip-up clasts, soft sediment deformation, etc.), and interpreted them in terms transport and depositional processes, such as erosion and reworking of the seabed, congregational sorting, bedform migration, etc. Because comparable sedimentary features can be observed in ancient deposits, findings from these studies are relevant for the study of ancient shelf mud deposits. One particular feature, plasmic fabric, consisting of laminae of aligned clays and micas that alternate with laminae of randomly aligned clays, has to my knowledge not yet been observed in ancient equivalents. It is apparently produced by shear sorting in the boundary layer (Kuehl et al., 1988), for example when waves pass over a muddy surface (Allison and Nittrouer, this volume).

Important contributions by Nittrouer and collaborators are the establishment of sediment budgets, the mapping of accumulation rate distribution for the studied shelf areas (e.g. Nittrouer et al., 1986; Alexander et al., 1991), the recognition of subaqueous deltas with gently dipping topsets, steeply dipping foresets (clinoforms), and gently dipping bottomset deposits, and the characterization of muddy shoreface and tidal deposits associated with major rivers (Allison and Nittrouer, this volume; Alexander et al., 1991). The work by Allison and Nittrouer (this volume) on muddy shoreface deposits builds on earlier work by Allersma (1971) and Rine and Ginsburg (1985). These deposits accumulate under conditions of relatively high current and wave activity, and debunk the myth that all muds are deposited in relatively quiet water. Possible ancient analogs of muddy shorelines and tidal deposits as described by Allison and Nittrouer (this volume) and Alexander et al. (1991) may be found in the Precambrian of India (Singh, 1980), the Devonian of Pennsylvania (Walker and Harms, 1971), and the Pennsylvanian of Indiana (Kvale et al., 1989). Recognition of preserved clinoforms in Upper Cretaceous shales of Utah/USA has allowed application of the subaqueous delta model to the stratigraphic record (Leithold, 1993, 1994).

The study of sedimentation in modern ocean basins in conjunction with the Deep Sea Drilling (DSDP) and Ocean Drilling Programs (ODP) also provides a wealth of data concerning the deposition and characteristics of deep sea muds. The interplay of a range of modern deep sea depositional processes (high- and low-concentration turbidity currents, turbid layer flow, detached turbid layer flow, settling of hemipelagic suspensions, etc.) and resultant fine-grained deposits are for example summarized by Stanley (1983) for the eastern Mediterranean. Criteria to distinguish turbidites and contourites, based on modern examples, are given by Stow (1979) and Stow and Lovell (1979). Several reviews of deep sea clastic sedimentation that include discussions and descriptions of fine-grained facies are given by Pickering et al. (1986), Stow (1985), and Stow et al. (1996). A model for the recognition and interpretation of fine-grained turbidites in shale sequences was presented by Stow and Shanmugam (1980). Much of the data published in the DSDP and ODP progress reports still remains to be "mined" for information that could be relevant for the interpretation of shale deposition in ancient successions.

Small scale sedimentary features as for example reported by Alexander et al. (1991b), Kuehl et al. (1988, 1991), and Segall and Kuehl (1994) are not the only aspect of muddy sediments that can be used to deduce sedimentary conditions and processes. Microfabrics are also an important avenue of inquiry. They are the cumulative record of the history of transport, deposition, and burial of fine-grained sediments, and are mostly studied with electron microscopes (see introduction to chapter 3). Experimental work has shown that microfabrics can for example be related to depositional processes, such as flocculation and single grain sedimentation (Mattiat, 1969; O'Brien, 1972, 1981). The book "Microstructure of Fine-Grained Sediments" (Bennett et al., 1991a) contains a series of papers that employ microfabric analysis in the study of modern muds (e.g. Wartel et al., 1991; Shepard and Rutledge, 1991; Bryant et al., 1991; Reynolds and Gorsline, 1991; Chiou et al., 1991). In an overview paper, Bennett et al. (1991b) review the physico-chemical, bio-organic, and burial-diagenetic processes that affect accumulating muds, and give modern and fossil examples of fabrics produced by flocculation, bioturbation, deposition of biosediment aggregates ("marine snow"), compaction, and fecal pellets.

The amount of oxygen available at the sediment/water interface also has a profound impact on the nature of accumulating muds, and may determine the organic carbon content, the degree of bioturbation, preservation of laminae, and fissility. A symposium volume with the title "Modern and Ancient Continental Shelf Anoxia" (Tyson and Pearson, 1991a), provides a very useful entry into this subject. The first part of the book contains 11 studies of modern shelf anoxia in which the biotic consequences of such conditions are examined with a view towards underlying causative factors, such as salinity stratification, development of thermoclines, plankton blooms, bottom water circulation, seafloor topography, etc. The second part of the book consists of 16 studies of ancient black shale successions, in which paleoecological approaches, trace fossils, paleontology, sedimentology, organic and inorganic geochemistry, and basin analysis are utilized to determine the underlying reasons for the accumulation of large quantities of organic carbon in these rocks. The introduction by Tyson and Pearson (1991b) succinctly summarizes what we currently understand (and not understand) about the possible causes of black shale formation in the geologic record. Access to more recent work on this important topic of shale research is provided by several of the contributions to this volume (e.g. Brett and Allison, Ettensohn, Genger and Sethi, Hoffman et al., Jaminski et al., Macquaker et al., O'Brien et al., Schieber, Wetzel and Uchman, Leventhal).

The Practice of Shale Sedimentology

Having said all this, the question then is, how can we best determine the processes by which ancient muds were deposited? As is true for any other sedimentary rock, the final appearance of a shale or mudstone depends upon the interplay between a whole range of physical, biological, and chemical processes. Thus, there is often a choice of investigative approaches that one can pursue in order to determine under what environmental conditions and by what processes a mudstone or shale unit has accumulated. At times this choice is dictated by available data, at times by the investigators preferences, and in many instances by the limitations of our understanding. The various contributions to this book clearly illustrate this state of affairs.

There are basically three main avenues of inquiry: 1) sedimentological, 2) paleontological, and 3) geochemical. Their application is summarized in the following paragraphs.

Sedimentology: From my vantage point there are basically four approaches to the sedimentological study of shale successions, depending on the scale of observations one wishes to make. These approaches can also be understood as stages of increasing detail in an expanding investigation and are listed below:

1) the examination of outcrops and drill cores
2) the examination of more resistant interbedded lithologies
3) the examination of shale facies
4) the examination of microfabrics

Outcrop/core: A commonly held misconception is that shales have not much to offer in outcrop because they are too fine-grained and weather too readily. Outcrop studies of shales, however, can be highly rewarding and should be the first step (where possible) in any study of a shale unit. Whenever the opportunity exists one should take the time to examine the available outcrops. Although shales weather indeed more readily than other lithologies, weathering tends to accentuate subtle differences in composition and color that may barely be discernible in core material, and thus makes bedding and rhythmic or cyclic compositional changes readily apparent in outcrop (Macquaker et al., this volume). In fact, some features, such as the evenness and continuity of beds (or the lack thereof), or the magnitude of erosion surfaces as seen for example in the Chattanooga Shale (Schieber, this volume), positively require outcrop examination. Large scale shale-on-shale erosion and truncation surfaces probably represent powerful erosive events and are good candidates for sequence boundaries (Schieber, this volume). Their systematic detection in shale successions will most likely be essential for extending sequence stratigraphic concepts to shale sequences.

In cores compositional differences are often not as apparent as in outcrop, but cyclic or rhythmic compositional changes can usually still be recognized (to know of their existence from outcrop studies can help quite a bit in that regard). One major advantage of core material is the fact that the contact relationships between beds and layers are easily observed, and that one gets a much better appreciation of the abundance and distribution of easily weathered components, such as pyrite.

Resistant Lithologies: In many older studies of shale sequences, the examination of more resistant interbeds (largely sandstones and carbonates) has provided vital clues to the depositional history. Good examples for studies of that type are those by Pepper et al. (1954), Seilacher and Meischner (1964), Folk (1962), and many more that are referenced in Potter et al. (1980). These interbeds are typically more weathering resistant than the enclosing shales, and their sedimentary features (cross-lamination, scours, grading, etc.) are more readily visible. They tend to record unusual and comparatively rare events in the history of a shale sequence or extremes of sediment input, productivity, and climate. The latter may be marked for example by limestone beds (reduced terrigenous input?, dry climate?, sea level rise?), chert horizons, an abundance of phosphate nodules, whereas the former may for example represent the deposits of storms (HCS beds), transgressions (bone beds), floods, or sediment gravity flows (turbidites). Although the conditions that are recorded by these interbeds are usually not the same that prevailed during deposition of the interbedded shales, they nonetheless help to establish the environmental boundary conditions within which shale deposition occurred. Because of this, and because they provide sedimentological information more readily than the enclosing shales, examination of interbedded sandstones and carbonates is an essential step in any sedimentological investigation of shale successions.

Shale Facies: The description and interpretation of shale facies is the most commonly employed approach in the sedimentological study of shales. Examples of this approach are numerous and include studies by Schieber (1989, 1990a, 1994b) Macquaker and Gawthorpe (1993), Wignall (1989), Stow and Piper (1984), Lash (1987), Cluff (1980), O'Brien and Slatt (1990); Einsele and Mosebach (1955), Ettensohn et al. (1988), Eugster and Hardie (1975), Lineback (1970), and Sutton et al. (1970). Because observable features vary between shale sequences, there may be an emphasis on compositional parameters in some studies, and on sedimentary features in others. How much one can accomplish with a composition-based approach to shale facies is for example demonstrated in the study by Macquaker and Gawthorpe (1993). They distinguished five lithofacies types based on the content of clay, siliciclastic silt, biogenic components, authigenic carbonate, and the absence or presence of lamination, and interpreted the results in terms of length of transport path, water column processes, water depth, clastic dilution, and early diagenetic processes. Under favorable circumstances the examination of organic facies may also be useful for the study of shale environments (e.g. Bustin, 1988).

In shales that contain few sedimentary features, detailed composition-based approaches to shale facies probably provide the best means for a study of depositional environments. Many shale successions, however, contain recognizable and varied sedimentary features. Their careful study usually allows more detailed and direct conclusions about sedimentary conditions. Studies by Schieber (1989, 1990a, 1990b) are examples of shale facies studies where the emphasis is on small scale sedimentary features. Although quite inconspicuous or even invisible in outcrop, small scale sedimentary features in shales (ranging in size from a few cm to less than a mm) can reveal a wealth of information about depositional conditions and history of a shale. When polished slabs and petrographic thin sections are examined, many seemingly drab shales show a large range and variability of sedimentary features. Shales that I had an opportunity to study with this approach range in age from Proterozoic to Eocene and come from the Proterozoic Belt Supergroup of Montana, the Cambrian Wheeler Shale of Utah, the Ordovician Athens Shale of Alabama, the Devonain Chattanooga Shale of Tennessee, the Triassic Moenkopi Formation of Utah, the Posidonia Shale of Germany, the Mancos Shale of Utah, and the Green River Formation of Wyoming. In all these stratigraphic units small scale sedimentary structures can be used to detail sedimentary conditions (e.g. Cole and Picard, 1975; Schieber, 1986, 1989, 1990a, 1990b, 1994b).

Laminae are the most typically observed sedimentary feature in shales, and exhibit a large range in thickness and lamination styles (even, discontinuous, lenticular, wrinkled etc.), which can represent quiet settling, sculpting of the sediment surface by bottom currents, and growth of microbial mats respectively (Schieber, 1986). Internal lamina features, e.g. grading (a), random clay orientation (b), preferred clay orientation (c), sharp basal contacts (d), and sharp top contacts (e) may be interpreted as indicative of (a) event-sedimentation (e.g. floods, storms, turbidites), (b) flocculation or sediment trapping by microbial mats, (c) settling from dilute suspension, (d) current flow and erosion prior to deposition, and (e) current flow and erosion/reworking after deposition (Schieber, 1990a, 1990b). Because of their somewhat larger grain size, silt laminae are usually the most easily observed lamina type in shales. They also imply somewhat more energetic conditions, and may for example indicate deposition by density currents (grading, fading ripples), storm reworking and transport (graded rhythmites), wave winnowing (fine even laminae with scoured bases), and bottom currents (silt layers with sharp bottom and top). Gradual compositional changes between e.g. clay and silt dominated laminae, are another commonly observed feature, and are suggestive of continuous (although slow) deposition, possibly from deltaic sediment plumes and shifting nepheloid flows.

Other sedimentary features that can be observed in shales are mudcracks, load casts, flame structures, bioturbation, graded rhythmites (Reineck and Singh, 1972), fossil concentrations and lags, cross-lamination, and loop structures (Cole and Picard, 1975), all of which carry information about conditions of sedimentation. Shales may also reveal clay-filled mud cracks, brecciation due to desiccation, and sands or conglomerates that consist entirely of shale particles (Schieber, 1985). Because in the latter features the fills in cracks and between shale particles consist of the same components as the cracked substrate or the shale particles, there is little compositional contrast to reveal them on broken surfaces in outcrop. Polished slabs or petrographic thin sections are typically required to detect shale filled cracks and shale sands and conglomerates. The latter can for example form as a result of soil erosion (pedogenic particles; Nanson et al., 1986; Rust and Nanson, 1989), erosion of cracked mud crusts, and submarine scouring of mud substrates by strong currents.

Biologic agents may produce microbial laminae and protection of mud surfaces from erosion (e.g. Schieber, 1986; O'Brien, 1990), or may manifest themselves as bioturbation and destruction of primary fabrics. However, in many instances sufficient proportions of primary features survive bioturbation, and observation of bioturbation features can provide information about substrate firmness, event deposits (escape traces), and rates of deposition. Fecal pellets and pelletal fabrics are another by-product of organic activity, and they are probably more abundant in mudstones and shales than commonly recognized (Pryor, 1975; Potter et al., 1980; Cuomo and Rhoads, 1987). They are indicators of organic activity, and are best seen in thin section, where they typically differ from the shale matrix with respect to texture, color, and organic content. They may range in size from 0.2-2mm and are generally of elliptical outline. Compaction tends to flatten the pellets to varying degrees, and if matrix and pellets are similar in composition and composed of particles of similar grain size, their identifaction can be challenging. The composition of fecal pellets can give clues on whether they were produced by benthic or planktonic organisms (Cuomo and Bartholomew, 1991; Schieber, 1994b).

As is true for other lithologies, a single feature can typically not be used to pinpoint specific depositional conditions and environments. Assemblages of features, however, will probably turn out to be environmental indicators once more sedimentological case studies of shale sequences have been undertaken.

As pointed out in the introduction to this book, because many of the described features require thin sections or polished slabs for proper study, shale facies studies with an emphasis on sedimentary features constitute in essence a microfacies approach to shale sedimentology. From an increasing number of case studies there may in time evolve a shale microfacies concept, along with the definition of standard shale microfacies types and generally applicable facies models (Figs. 1 and 2). Development of a "shale microfacies" scheme may eventually solve our problems with shale classification and should in the end allow sound interpretation of shale environments.

Microfabrics: Shale microfabrics are variable and record the combined impact of processes active during transport, deposition, and burial. The emphasis is on particle orientation and particle contacts, because these can potentially be related to processes active during transport and deposition (Bennett et al.; 1991b). The "Argillaceous Rock Atlas" by O'Brien and Slatt (1990) contains case examples from a variety of sedimentary environments (as well as many references) and illustrates mudstone and shale fabrics with SEM photos and photomicrographs. With some effort fabric features of shales or combinations thereof can be related to specific facies types and sedimentary environments, as well as to basic processes of sedimentation, such as flocculation, settling of dispersed clays, agglutination on microbial mats etc. The current problem with the application of microfabrics to the interpretation of depositional conditions is the fact that muds undergo serious compaction and diagenetic changes on the way to becoming rocks. To determine from an SEM study which fabrics are secondary and which features are reflective of original depositional fabrics is difficult and requires considerable experience. In many instances the successful interpretation of microfabrics hinges on the availability of other information, such as small scale sedimentary features, paleogeographic setting, and paleoecological information (e.g. O'Brien et al., 1994).

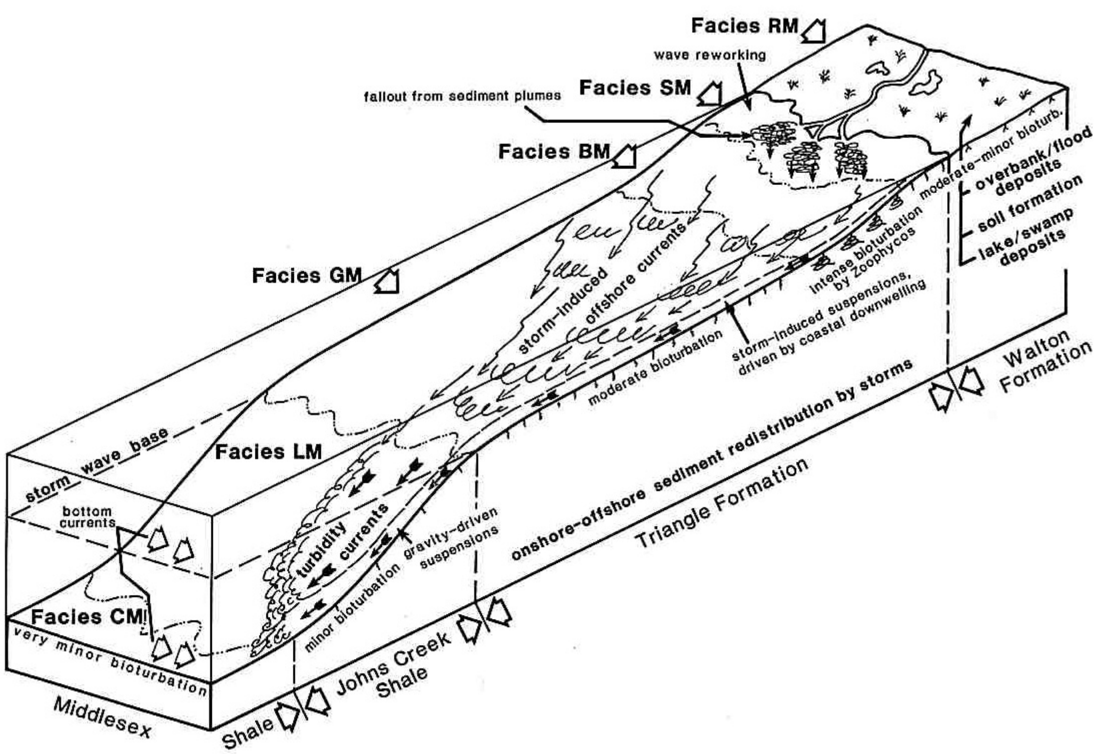

Figure 1: Example of a shale succession, the Upper Devonian Sonyea Group of New York, that was studied with a shale microfacies approach (Schieber, in preparation). The summary depositional model is for the lower Sonyea interval (Sutton et al., 1970). The model shows distribution of six shale facies with respect to laterally equivalent stratigraphic units (shown at base of block diagram), as well as inferred depositional processes. Facies RM consists of red to gray coastal plain mudstones (6 shale microfacies types); facies SM are sandy nearshore mudstones (2 shale microfacies types); facies BM are intensely bioturbated offshore mudstones (2 shale microfacies types); the graded mudstone facies GM contains abundant muddy tempestites (3 shale microfacies types); the laminated mudstone facies LM is characterized by fine-grained turbidites (2 shale microfacies types); and facies CM consists mainly of black shales (2 shale microfacies types). Boundaries of subaqueous facies marked with double-dotted dashed lines.

Paleontology: How paleontological approaches can be used to interpret sedimentary environments of shales is the subject of Chapter 5 of this book. Fossils can serve as indicators of salinity, water depth, water temperature, energy regime, water turbidity, sedimentation rates, and the nature of the substrate, and can thus provide many important clues to depositional conditions of a shale unit (see summary in Boggs, 1987). The paper by Brett and Allison (this volume) summarizes how the type, association, and preservation of body fossils can provide clues about depositional conditions and sedimentary environments. Trace fossils as well can provide a wealth of information about depositional sedimentary environments, and their application to investigations of shales and mudstones is summarized in the contribution by Wetzel and Uchman (this volume).

Geochemistry: There is a variety of ways in which geochemical data have been used to interpret sedimentary environments of shales. The bulk chemical composition of a shale is inherited from the source area, and thus many element concentrations are solidly controlled by the sediment source. Only elements that are added in significant quantities in the depositional environment are of interest, because in those instances there is a chance to differentiate the environmental component from the detrital one.

For example, for the purpose of paleosalinity studies, the distribution of boron has been extensively studied. Boron in the marine environment is incorporated into authigenic illites, and generally speaking marine shales will have larger boron contents than freshwater and brackish water shales (e.g. Harder, 1970). Because grain size, mineralogy, and diagenesis all have a bearing on the boron content of a given shale, it is essential to compare samples where these three variables are as similar as possible (Boggs, 1987). Fossils are typically better paleosalinity indicators than boron.

An early attempt at geochemical facies differentiation is the classical study by Adams and Weaver (1958), who suggested that Th/U is a sensitive indicator of depositional environments and geochemical facies in marine shales. Zelt (1985) conducted a more recent study of the utility of the Th/U ratio and concluded that Th/U is related to the organic carbon contents of marine shales, and that marine to brackish shales form a smooth continuous trend that can be used to infer the Eh of the depositional environment (which is controlled by oxygen content and early diagenetic conditions).

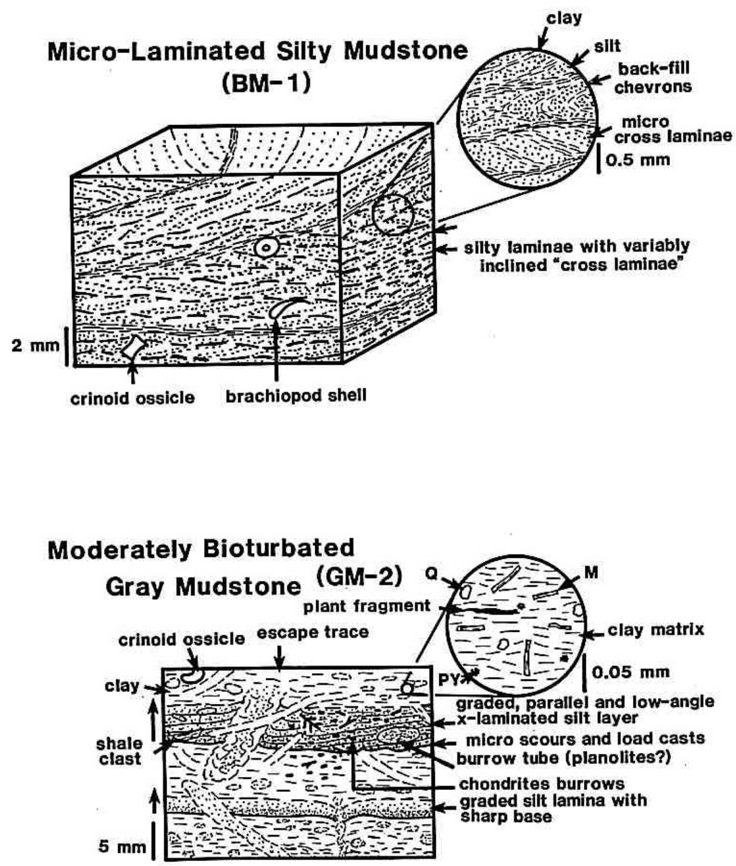

Figure 2: Examples of shale microfacies types from the Sonyea Group (see Fig. 1). Line drawings summarize features observed in one of the microfacies types from facies BM, and in one of the microfacies types from facies GM. Note that scale bars are variable. In the line drawing of BM-1, the micro-lamination represents bioturbation by Zoophycos, and micro cross-laminae are compacted Zoophycos back-fill meniscae. In outcrop this type of shale has a subtle wavy-laminated appearance and the intense bioturbation is not recognized. In the line drawing of GM-2 we see sharp-based graded silt/clay couplets that are interpreted as tempestites, and more intense bioturbation in layers between tempestites.

The majority of geochemical studies with an environmental emphasis has been focused on the degree of oxygenation of the depositional environment. The geochemical classification of sedimentary environments by Berner (1981) is a good example. Based on observations from modern sediments, Berner distinguished oxic, post-oxic, sulfidic, and methanic environments. The various environments were differentiated based on the presence or absence of authigenic iron minerals (e.g. pyrite, hematite, siderite), organic matter, and manganese minerals (e.g.

rhodochrosite, MnO2). The mechanisms of early diagenetic sulfide formation, especially the formation of sedimentary pyrite (Berner, 1984), have been a central theme of most recent geochemical investigations into ancient anoxia. Organic carbon, pyrite sulfur, degree of pyritization (DOP), and carbon to sulfur ratios (C/S) are used as variables in a variety of plots to discriminate oxic, anoxic, and euxinic environments. The whole subject is discussed and illustrated at length by Leventhal (Chapter 7, this volume).

Black shales typically are enriched in a wide variety of trace elements (e.g. Mo, As, Cd, Co, Ni, V, Pb, Zn, Mn, Cu) and various attempts have been made to utilize plots of trace elements and trace element ratios to discriminate black shales that were deposited under euxinic, anoxic, and oxic conditions. The general approach is exemplified by Hatch and Leventhal (1992). The latter publication and Leventhal (Chapter 7, this volume) provide further references.

Stable isotope studies of shale components are typically carried out on organic matter (carbon isotopes), carbonate minerals (carbon and oxygen isotopes), and on pyrite (sulfur isotopes). Carbon isotopes can be used to understand the sources or origins of the different types of organic matter that we may find in shales (Schidlowski et al. 1983), for example whether organic matter was derived from land plants or from marine organisms (see paper by Hoffman et al., this volume). Leventhal (Chapter 7, this volume) gives some examples and further references regarding the application of carbon isotope data. Carbon and oxygen isotopes in fossil shells and limestone interbeds within a shale can be used to differentiate between marine and non-marine sediments (e.g. Dodd and Stanton, 1975; Schopf, 1980), and paleotemperatures can be estimated from oxygen isotope data of unaltered biogenic marine carbonates (Shackleton, 1967; Emiliani and Shackleton, 1974; Schopf, 1980). Sulfur isotope data from pyrite primarily record bacterial sulfate reduction of seawater sulfate. Bacterial sulfate reduction is accompanied by an enrichment of pyrite sulfur with the light isotope (^{32}S), and the fractionation that occurs is expressed in $\delta^{34}S$. The fractionation relative to coeval seawater is fairly predictable at about 40 to 50 permil $\delta^{34}S$ (Faure, 1986). Sulfur isotope data of pyrite in shales either indicate this expected bacterial fractionation (knowledge of sulfur isotope composition of coeval seawater required), or they may show sulfur isotope values that are significantly heavier (more ^{34}S) than what one should expect. In the former instance one may assume that the depositional basin was connected to the world ocean (e.g. Strauss and Schieber, 1990), whereas in the latter case one can speculate on basin isolation and a limited sulfate pool (e.g. Smith and Croxford, 1973; Lambert, 1976; Goodfellow, 1987; Goodfellow et al., 1993). The timing of pyrite formation within the sediment, however, has a strong influence on its sulfur isotopic composition. In general, the later the pyrite formed, the "heavier" it is (increase of ^{34}S), and thus sulfur isotopic studies of pyrite should always be accompanied by petrographic examination of the samples (Strauss and Schieber, 1990).

Outlook

Which approach one takes depends on the questions one is interested in, the available equipment, available study material and data, and the amount of time that is available for the study. Ideally the investigation of a shale unit should be interdisciplinary. Once the conclusions from different avenues of inquiry converge, the likelihood that a depositional setting has been interpreted correctly greatly increases. Invariably, knowing about the larger context of a shale units deposition helps to focus in on more probable scenarios. To understand the “big picture”, basin analysis approaches including sequence stratigraphy, lateral tracing of extent and geometry, gamma ray profiles, wire-line logs, and other subsurface information are most helpful. The contributions by Bohacs, Schutter, and Macquaker et al. in chapter 1 are examples of this aspect of shale sedimentology.

Although there is quite a bit of information on the processes and mechanisms operating in modern environments where mud accumulates, applying this information to the sedimentary record is not as easy as it may initially appear. For example, although more and more small scale sedimentary structures are described from modern muds, we usually have no direct knowledge of the conditions (current velocity, density of suspension, etc.) that controlled the formation of these structures. We don’t see them in the making, and only infer about the controlling parameters from temporally and spatially very limited measurements in the water column above. The tacit assumption is that our measurements of today’s conditions can be representative for the uppermost centimeters to decimeters of modern sediments. In detail, however, the heterogeneity that is often shown by these “modern” sediments belies this assumption. Basically, our knowledge of modern mud deposition is two-dimensional and spotty, whereas the interpretation of the ancient record requires a tree-dimensional solution. Improved studies of modern environments may alleviate some of these problems, but a true improvement can only come from experimental studies of mud deposition. Unlike experimental studies of sand deposition where the main variables are grain size and flow velocity (incl. velocity profile), in muds we will have to consider in addition the water content, the viscosity, a host of compositional variables (which clay minerals, silt/clay ratios, amount of organic matter, type of organic matter, etc.), the influence of microbes (within the sediment or as mats covering the surface), fabric parameters, the compaction history, water expulsion, internal gas development, and probably a few more. Obviously, these experiments are going to be complex and most likely they will be a source of much frustration for those that attempt to conduct them. Because of the many variables, numerous permutations of depositional conditions will have to be explored before we can truly say that we have a firm grasp of mud deposition. Thus, it will probably take several decades before the task is completed.

There is also a comprehension gap between modern and ancient microfabrics. What we understand from modern microfabrics can not be uncritically transferred to the sedimentary record. Whereas in modern muds we can identify fabrics due to flocculation, dispersed settling, biosediment aggregates etc. (see above), comparable fabrics in the sedimentary record may not necessarily have the same origin. Compaction, as well as transformation and recrystallization of clays during diagenesis, makes survival of primary fabrics very unlikely. To identify relict fabrics, or to determine how a primary fabric will be transformed during compaction and diagenesis, is a difficult task at best. How microfabric is related to macrostructure, to sedimentary processes and environments, and how diagenetic processes influence primary fabrics, is still a wide open field of research with ample room for new discoveries and breakthroughs.

Basic questions that should always be kept in mind when conducting a study of a shale sequence are the following:

1) What was the source of mud?
2) How (in what form) was it transported to the depositional site?
3) What was the water depth, sedimentation rate?
4) What were the factors that sustained life (oxygenation, food supply, etc.)?
4) What happened to the mud since deposition?

In initial stages, these questions will provide the necessary focus for a shale study. Later on, as answers to these questions are arrived at, they will most likely provide a suitable framework for further investigations.

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