More gaps than shale: erosion of mud and its effect on preserved geochemical and palaeobiological signals

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Abstract: Ths wht th fn-grnd mrine sdmtry rcrd rlly lks like.

It has been forty-one years since the publication of the first edition of *The Nature of the Stratigraphical Record* by Derek Ager (1973). In his provocative book, Ager suggested that the sedimentary record is riddled with gaps, and that short, recurring events may have had more effect on the rock record than long periods of cumulative slow change by processes operating at relatively constant rates. Ager's catchphrase 'more gaps than record' is not normally considered to be applicable to the marine fine-grained sedimentary record, which is commonly thought – or at least taught – to be virtually complete and therefore an excellent recorder of earth history and past environmental change.

Fine-grained sediment (Fig. 1), or mud, which includes siliciclastic clay and silt and the also mud-sized remains of once-living organisms, for example, coccoliths, cysts, frustules, and tests, is normally thought of as the sediment fraction that settles slowly from suspension to the bottom of channels and basins only when, and where, the water column is quiet (Shields 1936; Hjulström 1939; Sundborg 1956). Once deposited, mud is thought to be difficult to erode and re-entrain back into suspension due to its cohesive behaviour and/ or to the low energy of the environments in which it has been deposited. 'Aside from this it is hard to say anything definitive about the environment of a shale since most environments have periods and places of quiet water deposition' (SEPM Strata) (cf. Potter *et al.* 2005, p. 75). In marine environments normally unaffected by wave action, that is, in certain lagoons and in environments deeper than the lower shoreface, sedimentation – though little (<2.5 g/cm^2 ka) – is often thought to occur everywhere and all the time, thereby preserving a continuous – although often condensed (<1 cm ka^{-1}) – layer-cake record of earth history.

Although gravitational settling is an important depositional process in the ocean, its importance in the deposition of mud has been overstated.

Unlike in freshwater, mud in the marine environment is mostly present as aggregates with settling velocities and hydraulic behaviour that are very different from those predicted for individual particles by Stokes's Law and the Shields and Hjulström–Sundborg diagrams. Fresh, aggregated mud deposits with up to 90% water by volume have lower densities and shear strengths than non-aggregated mud deposits, and are therefore easier to erode than predicted by experimental curves.

This paper focusses on the origin and stratigraphic nature of stratification surfaces in shales. I show how some of these are erosional surfaces ranging from diastem to unconformity. Then, I discuss the processes in ancient and modern marine environments that generate an incomplete shale stratigraphic record. Finally, using a Toarcian example, I discuss how stratigraphic incompleteness in shale successions affects the interpretation of geochemical and palaeontological records of past environmental change.

Origin of stratification in shales

Stratification is a distinctive characteristic of many sedimentary rocks. Sedimentary successions are the product of a series of alternating episodes of sedimentation and nondeposition/erosion. Most sediments are deposited layer upon layer, and this layering, or stratification, relates to depositional process and reflects sediment accumulation over different timescales. Some strata are preserved intact, others are partly truncated, and yet others are fully eroded and excluded from the stratigraphic record. Preserved stratal thickness equates with time, but strata only record a fraction of the time contained in a succession; the remaining time is recorded as stratification planes, which represent intervals of nondeposition and intervals of deposition subsequently (partly) removed by erosion.
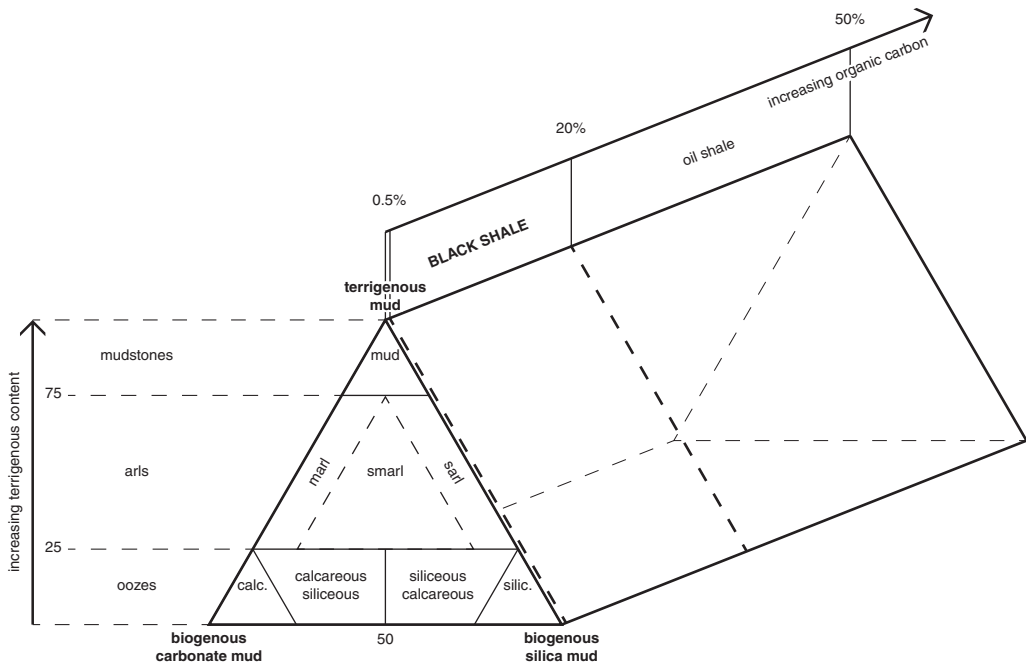
Fig. 1. Classification of shales based on composition (based on Hay *et al.* 1984). Shales are fine-grained sedimentary rocks with varying relative proportions of terrigenous mud and mud-sized biogenous components. Shales may contain up to 25% of terrigenous and biogenous grains coarser than mud (>62.5 μm). A shale with an organic matter content higher than the average marine rock, that is *c.* 0.5%, is referred to as a black shale. Black shale is thus the general term for any dark-coloured, fine-grained, organic matter-rich sediment. In the words of Stow, Reading, & Collinson (Reading 1996, p. 403): '[m]any black shales are hemipelagites; others, such as black cherts and organic matter-rich limestones, are pelagites; whereas still others are fine-grained turbidites.' A good review of the history of the terminology of fine-grained sedimentary rocks is given by Tourtelot (1960). Although the original meaning of the word 'shale' was 'laminated clayey rock' (not fissile), the historical usage of the word has been that of 'general class of fine-grained sedimentary rocks' (Tourtelot 1960). Many authors disapprove of this usage. They favour using the term mudrock as a collective noun for fine-grained sedimentary rocks and reserve the term 'shale' to mean laminated or fissile mudstone (e.g. Spears 1980). There is no reason why we should restrict our usage of the word 'shale' to laminated and/or fissile fine-grained sedimentary rocks, because 'lamination' has both a descriptive and a genetic definition with distinct sedimentological implications (cf. McKee & Weir 1953; Campbell 1967), and particularly because fissility is a secondary property largely related to weathering (Ingram 1953). Other authors prefer the term 'mudstone' for fine-grained sedimentary rocks in general (e.g. Macquaker & Adams 2003). I prefer to uphold the historic use of 'shale' with the meaning of a general class of fine-grained sedimentary rocks (including fine-grained, biogenous sedimentary rocks).

On the scale of an outcrop, a hand specimen, or in thin section, two main types of stratification can be recognized – lamination and bedding. Despite their importance, there is no generally accepted usage of these two terms. If thickness is taken as the criterion, and it often is, the limit is arbitrarily set at 1 cm (McKee & Weir 1953). Strata thicker than 1 cm are referred to as beds, and lamina is the term applied to strata thinner than 1 cm. The problem with this descriptive terminology is that it does not distinguish between strata that are 'separated from adjacent strata by surfaces of erosion, nondeposition, or abrupt change in character' (McKee & Weir 1953) and strata, whatever their thickness may be, that were formed by accretion of

sediment onto surfaces of net sedimentation without intervening episodes of erosion, nondeposition, or abrupt change in depositional conditions (Campbell 1967).

A bed is a layer of sediments bounded by surfaces produced during periods of erosion, nondeposition, or abrupt change in depositional conditions (Campbell 1967). Bedding surfaces thus represent breaks in sedimentation that may range from diastem (or nonsequence) to unconformity (viz. disconformity and paraconformity) and which pass laterally to correlative conformities. As pointed out by Macquaker & Howell (1999), a storm is erosive in one area and depositional elsewhere. Diastems are breaks in sedimentation that

are represented in other regions by a (few) bed(s), whereas disconformities and paraconformities mark time intervals that are represented elsewhere by deposits of formation value (cf. Barrell 1917). The presence of discontinuity surfaces in a succession results in a slow accumulation rate of a formation and more rapid accumulation of individual beds. Over longer time spans, stratigraphic successions contain more and longer hiatuses and, therefore, sedimentation rates decrease. Indeed, measured sedimentation rates for different depositional environments decrease as a power law function of the interval of time over which sedimentation rates are measured (Sadler 1981).

The term 'lamination' can be used descriptively to denote the thickness (<1 cm) of strata, or it can be used in a genetic sense to refer to strata generated by gradual sediment accretion (i.e. representing continuity of sedimentation) irrespective of their thickness. In principle, stratification within a bed should be termed lamination whatever the thickness of the individual layers (Kuenen 1966). No single process is responsible for the development of lamination. Although fluctuations in flow velocity can produce lamination (Otto 1938; Pettijohn 1957), they are not required. Kuenen (1966), for example, argued that different particles react differently to prevailing hydrodynamic conditions and accumulate alongside others of the same kind, and Jopling (1964) concluded that sorting processes related to flow separation are responsible for the development of cross-lamination. Certain kinds of horizontal 'lamination' are due to cessation of supply at intervals, but laminae 'are then merely thin beds, and no separate problem is involved' (Kuenen 1966).

The topic of stratigraphic incompleteness in fine-grained sedimentary rocks, or shales (Fig. 1), is intimately related to the origin and nature of stratification in these rocks. The problem normally comes in the semantics and interpretation rather than in the recognition of the presence of stratification in shales (Figs 2 & 3). Shale strata are often thin (<1 cm), because the sediment is fine and the effects of compaction on mud deposits are pronounced. This thin, or fine, stratification is often described as lamination. Although the term 'lamination' is in such cases used descriptively, genetic implications are subsequently attached to the term. Such 'lamination' is interpreted as having been produced by the settling of mud from suspension and preserved as a consequence of the lack of erosion by (strong) bottom currents after deposition and an absence of bioturbation due to anoxia. However, not all laminated muds were deposited from suspension: '[p]lanar strata (laminae) in muds may also be associated with the transport of sand-sized mud grains, as low-relief bed waves' (Bridge 2003). Petrographical studies (e.g. Macquaker et al. 2007,

2010; Ghadeer & Macquaker 2011, 2012; Trabucho-Alexandre et al. 2012; Plint et al. 2012; Plint 2013) have shown that many such laminae (in a descriptive sense) are actually thin beds (in a genetic sense), which may be internally (cross-) laminated. Therefore, they represent sedimentation episodes delimited by periods of erosion, non-deposition, or an abrupt change in depositional conditions, rather than continuous gravitational settling of mud. On the other hand, within shale successions, what is normally referred to as a 'bed' (Fig. 3 & cf. Fig. 8) actually records stratigraphic information at larger scale, such as bedset, parasequence, (sub)zone, or greater (e.g. Macquaker & Taylor 1996; Macquaker & Howell 1999). Thus, 'beds' used as correlation markers may not record short-lived processes, but instead reflect sedimentary processes acting over longer periods leading to correlations of much lower resolution than commonly thought (Macquaker & Howell 1999).

The presence of fissility in shales is commonly assumed to indicate that the rock is laminated. As discussed by Macquaker & Adams (2003), although this may sometimes be the case (cf. Oertel & Curtis 1972), many fissile shales are not laminated and their fissility is a secondary property largely related to weathering (Ingram 1953). Fissility is not a good rock classification criterion, because some mudstones are then merely freshly exposed shales. In nonlaminated shales, fissility is perhaps caused by separation along planes between (sometimes bioturbated) thin beds of slightly different lithological characteristics. Fissility should therefore not be taken to indicate the presence of lamination. Equating fissility with lamination and misinterpreting strata that are genetically thin bedded as being genetically laminated has caused researchers to overestimate the continuity of the shale sedimentary record and also the significance of bottom water anoxia in their deposition (Macquaker & Adams 2003).

Mud erosion

The physical behaviour of mud in sedimentary environments depends on many variables of which particle size is one of the most important. The terminal velocity of a spherical particle ($Re_p < {\sim}0.5$) settling in an unbounded Newtonian fluid is proportional to the density difference between particle and fluid and to the square of the diameter of the particle (Allen 1985). The upward component of turbulence generated by a current of only 0.0002 cm s^{-1} will keep a 2 μm particle in suspension (Pryor 1975). This means that clay (particles <4 μm) can only be deposited as individual particles in completely quiet, nonagitated water

Fig. 2. Stratification in Toarcian black shales from two intervals of a core through the Whitby Mudstone Formation, Whitby, North Yorkshire (UK). Squares with an area of 1 cm^2 each are included for scale. The arrows with an open arrowhead indicate a few prominent bedding planes, whereas the closed arrowheads point to examples of lamination planes. The most obvious bedding planes are those which are materialized by either an erosional surface with relief or a silt lag above it (left image, third arrow from above, for example). Other bedding planes are subtler (left image, fourth arrow from above, for example). All beds in the left image are thinner than 10 mm, and most are thinner than 5 mm, which would make them fall under the definition of laminae sensu McKee & Weir (1953). The interval shown in the right image is laminated, particularly the lower two-thirds shown. Bedding planes in the right image are subtler than in the left image due to reduced textural contrast between thin beds. Individual laminae are in general thinner than 1 mm. Unlike beds, laminae lack abrupt changes in texture and evidence for erosion at the stratification plane. Although the processes and environmental conditions behind their development are clearly different, McKee & Weir's descriptive terminology does not distinguish the true laminae sensu (Campbell 1967) in the right image from the 'laminae' in the left image.

masses, which is a situation that rarely occurs in most natural bodies of water.

In paralic and marine environments, mud tends to occur as micro- and macro-floccules (flocs), faecal pellets, pseudofaeces, and other organomineralic aggregates (e.g. marine snow). Four groups of processes bringing and keeping particles together have been shown to be particularly important in the formation of organomineralic aggregates in paralic and marine environments (Van Olphen 1963; Pryor 1975; Eisma 1986; van Leussen 1988): (i) compensation of the negative charge by adding cations to water (electrochemical coagulation), (ii) biogenic pelletization (faeces and pseudofaeces), (iii) particle collisions due to differential settling

velocities, Brownian motion, and turbulent motion of the water, and (iv) particle attachment due to adsorption to sticky organic matter. Biogenic aggregation may be the most important process controlling the transport and deposition of mud in paralic and shallow marine environments (Pryor 1975). Pelletized mud produced by filter and deposit feeders behaves as silt- and sand-sized sediment (Pryor 1975; Oost 1995) and is transported as bedload rather than suspended load (Fig. 4). Pelletized clays can survive rough handling and accumulate in relatively high-energy environments (Moore 1931; Pryor 1975; Oost 1995), which explains, for example, why mm- to cm-thick mud drapes found in subtidal deposits were deposited during one slack

Fig. 3. Thin bedding in the Toarcian Jet Rock, Port Mulgrave, North Yorkshire (UK). Traditionally, 'beds' in the lower Toarcian of NE England were defined based on the presence of calcareous concretions (Howarth 1955, 1962), of which there are two obvious examples in the upper image. The concretionary horizons coincide with horizons of faunal change, which were used to subdivide the succession into ammonite zones and subzones. These 'beds' are in the order of metres thick (cf. Fig. 8) and thus three orders of magnitude larger than bedding in the succession. Upper photo courtesy of Jonny Imber (University of Durham).

water period of the order of 20–30 minutes (Van Straaten 1954; McCave 1970; Terwindt & Breusers 1972; Visser 1980; de Boer 1998). Under normal flow conditions, turbulent shear will not break aggregates up because they are too small. Studies indicate that the smallest eddies in a fluid limit the maximum size of macroflocs (van Leussen 1988). In natural turbulent flow in estuaries and coastal seas, the size of the smallest turbulent whirls

on the Kolmogorov microscale is of the order of millimetres (Kranck 1975; Eisma 1986).

The erosional behaviour of mud is controlled by a large number of variables, including sediment composition and organic matter content, fabric, bioturbation type and intensity, and time available for compaction. The impact of these variables on mud erodibility is still poorly understood (e.g. Potter *et al.* 2005). It is generally agreed that water content
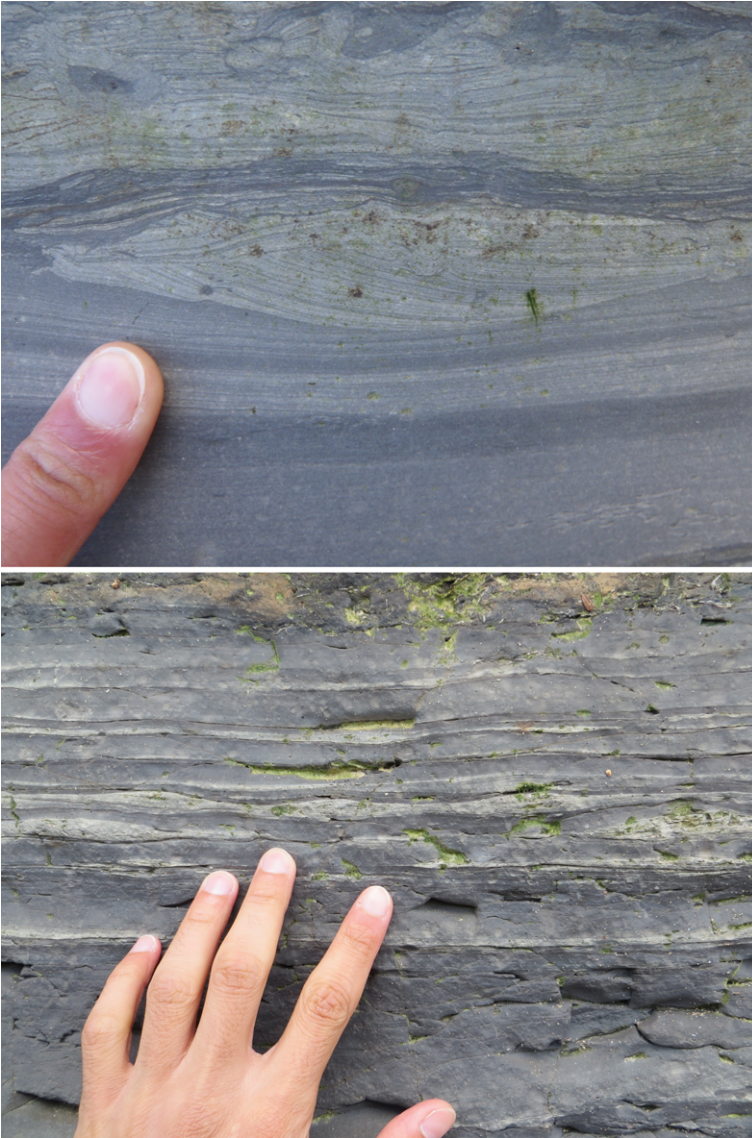
Fig. 4. Starved clay and fine silt ripples in Pliensbachian shallow marine shales, Staithes (upper image) and Jet Wyke (lower image), North Yorkshire (UK). In the upper image trough cross-lamination in fine silt and clay is seen. Note the presence of erosional surfaces and the alternation of light, coarser silt cross-laminae and dark, finer silt and clay cross-laminae. In the lower image, a train of combined flow fine silt ripples indicates that mud was deposited on the shallow seafloor above wave base by a combination of wave action and unidirectional flow.

and threshold erosion velocity show an inverse relationship (Southard *et al.* 1971; Lonsdale & Southard 1974), that is, soft, soupy muds require much less current strength to be eroded than firm, consolidated mud.

As mentioned above, in paralic and marine environments, freshly deposited muds are normally aggregated. Freshly deposited aggregated muds have open fabrics and water content in excess of 80% (Hedberg 1936; Migniot 1968; Terwindt & Breusers 1972), which makes them easier to erode than nonaggregated deposits. The latter have more stable subparallel fabrics with less water and thus higher densities and shear strengths. Mud transported in bedload as floccule ripples also generates denser fabrics than aggregated mud deposited by

settling from suspension, which has higher porosity (J. Schieber, pers. comm. 2013).

The transition from soupy mud to dense shale is not always rapid. Sorby (1908), for example, left a kaolinite suspension to settle for over a year and found it still to contain 75% water. Nevertheless, the stiffness of a mud deposit increases progressively and consequently its erodibility is reduced (Teisson *et al.* 1993). However, as pointed out by Partheniades (1965), for any shear stresses applied to a mud surface, there will always be mud particles whose bond to the bed will be small enough to be broken. Moreover, fluid stripping (Allen 1982), that is the direct action of the drag force exerted by a current on a bed, is not the only mechanism of erosion of mud; corrasion probably operates simultaneously in all natural environments. Corrasion is the blasting of the sediment interface by transported particles (including biological material) that serve as tools and result in the removal of fragments from the mud substrate at each contact. In turn, increasing boundary roughness results in a decrease in the effective resistance of mud beds. This may also be due to bioturbation or to the presence of larger stationary particles such as disturbing benthic organisms and objects, for example, manganese nodules (Lonsdale & Southard 1974). Recent experimental work using chalk showed that no significant changes in threshold erosion velocities (*c.* 19 cm s^{-1}) can be observed with changing consolidation time, but that the length of consolidation time affects erosion rates (T. Buls, pers. comm. 2013), that is, more consolidated deposits require longer intervals of erosion to remove the same amount of mud.

Recent flume experiments by Schieber *et al.* (2010) showed that erosion of soft, soupy muds (70 wt% water content) by currents at flow velocities between 15 and 35 cm s^{-1} results in the production of mud rip-up clasts. These rip-up clasts are transported by currents and, although their size diminishes during transport, they accumulate in areas of lower flow velocity and result in lenticular shale fabrics that are petrographically distinct from those that represent compacted faecal pellets or compacted burrow fills (Schieber *et al.* 2010). If this type of mass erosion, in which chunks of the bed are ripped up and carried down current, persists, the entire mud bed may be eroded. Indeed, Hawley (1981) suggested that the initial thickness of a mud bed is the main control on its preservation potential.

Young & Southard (1978) studied the erosion of marine muds in the field and in the laboratory and found that threshold erosion velocity decreases with an increase in the degree of bioturbation and increases with organic matter content in non-bioturbated beds. In their experiments, critical shear velocities for erosion of mud substrates ranged from 0.32 to 0.84 cm s^{-1}. The authors also found that erosion is often, but not always, initiated in biogenically disturbed parts of the bed. Bioturbation, in particular the burrowing activity of deposit-feeding organisms, leads to decreased compaction and increased water content of sediment (Rhoads 1970). Thus, bioturbation often results in mud deposits with lower densities and lower shear strengths that are relatively easy to erode. As pointed out by Meade (1972), equations based on physical relations between cohesive strengths of sediment and erosive velocities of water may have little relevance to resuspension mechanisms in the marine environment. Several complicating factors exist. They include variable water content, carbonate content – more carbonate resulting in greater shear strength – the effects of bacterial slimes, and clay mineralogy (which controls crystal morphology). Some processes can work both ways. Bioturbation, for example, makes mud more erodible when it makes it less dense, but decreases mud erodibility when it makes mud deposits more cohesive due to organic binding.

According to Young & Southard (1978), the most surprising result of their experimental work is that mud appears to be easier to erode than noncohesive fine sand. This result is contrary to the speculations of previous workers such as Shields (1936), Hjulström (1939), and Sundborg (1956), and may be explained, according to the authors, by the lower bulk density of biologically aggregated sediments. It is seldom noted that Shields (1936) had no data for Re* (boundary Reynolds number) smaller than about 2, and that Hjulström (1939) showed the fine-grained part of his curve representing cessation of transport as a dashed line presumably due to lack of data in that range (cf. Sundborg 1956). Einsele *et al.* (1974) concluded that the wide gap in the results of different authors and the great uncertainties in the field for fine-grained material of the Hjulström–Sundborg curve (Hjulström 1939; Sundborg 1956) are due to differences in the sample fabric.

Erosional features in shales remain largely underidentified because they are typically indistinct. Shales are highly compacted lithologies that exhibit little obvious macroscopic variability (Fig. 3). Burial of mud tends to obscure primary sedimentary fabrics (Allen 1985, p. 144, Fig. 8.5) unless there is early cementation. Moreover, owing to their drab colours, there is normally a low lithological contrast between adjacent units. Thus, while shale successions are considered to contain the most complete stratigraphic record, many unnoticed gaps exist in such successions. Schieber (1994, 1998) studied the Chattanooga Shale of central Tennessee and showed that during deposition it was prone to influence by storm waves and episodic erosive events. These events were of variable strength

Fig. 5. Small-scale ripples largely consisting of sand-sized biogenous aggregates (faecal pellets and pseudofaeces) Dollard Estuary, northern Netherlands. In the lower photo, the sand-sized mud aggregates can be seen. Photos courtesy of Poppe L. de Boer.

and/or duration and resulted in truncation surfaces beneath which a few centimetres to over a metre of section is missing. Importantly, these truncation surfaces cut underlying beds at very low angles or even become conformable making them essentially 'invisible'. Despite this difficulty, some authors have documented erosional features in shales, which are recognizable from outcrop scale down to microscopic scale (Figs 5, 6 & 7), and suggest 'a stratigraphic record littered by gaps' (Schieber 2003). Schieber shows, for example, how a 5 cm sample is all that is left of nearly 50 cm of potential rock record; that is a loss of 90% of recorded time. Baird (1976) studied the upper Devonian of western New York (Genesee Group) and found evidence

for erosive events scouring off soft surface mud of the seafloor. O'Brien (1996) shows abundant examples of Devonian shales of North America in which scour and discontinuity features indicate erosion by high energy events, and Schieber (1999) studied upper Devonian shales of New York (Sonyea Group) and concluded that mud was subject to rapid deposition and to frequent reworking.

Erosion and nondeposition in modern marine environments

Siliciclastic mud is, by volume, the most abundant sediment type in the ocean, and the dissolved and
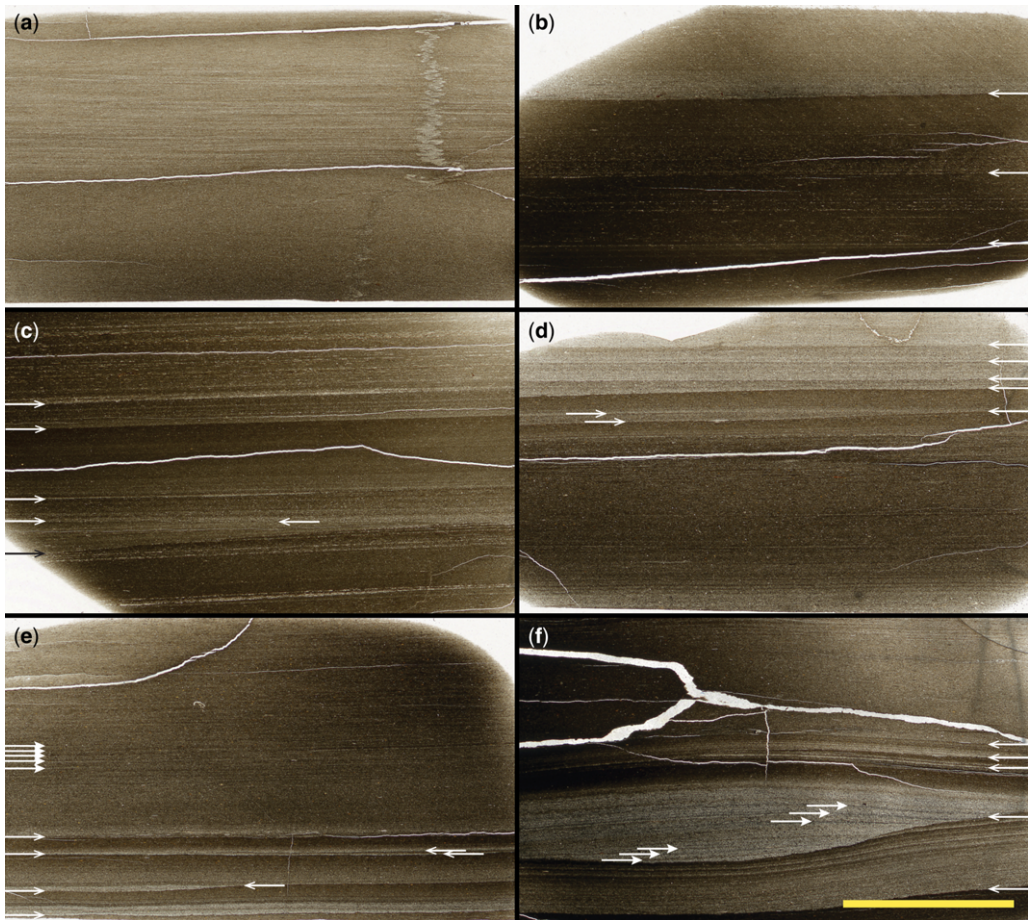
Fig. 6. (**a–f**) High-resolution scans of thin sections of Toarcian Jet Rock shales, Port Mulgrave, North Yorkshire (UK). Scale bar equals 1 cm. Sedimentary textures in shales, including sharp-based beds and thin silt lags, suggest that they were not deposited as a continuous rain of sediment and that the seafloor was episodically reworked. This observation is not really new: Sorby (1908) stated that 'the Lias near Whitby show[s] laminar structure due to currents' (p. 196).

particulate load of rivers constitutes the main source of sediment for the marine environment. Each year, about 15 billion (10^9) metric tons of terrigenous sediment are brought to the ocean by rivers (Milliman & Meade 1983). Although all terrigenous mud is ultimately derived from the continents, much of the suspended sediment in the ocean is not related to present-day river input of sediment, and the immediate source for most suspended sediment in waters of the continental shelf is the resuspension of older, possibly relict mud (Meade 1972).

Terrigenous sediment supply to the shelf is a local phenomenon; it occurs mainly at river mouths, with a very uneven geographical distribution. A few rivers in Asia supply about 70% of the total amount of terrigenous sediment brought to the ocean annually by rivers (Milliman & Meade 1983). Moreover, most terrigenous sediment – including mud – is (since the development of algae, filter feeders, and terrestrial vegetation) trapped more or less permanently in nearshore environments, such as subsiding deltas and large estuaries, and along the adjacent coastline, due to, among others, sediment aggregation (e.g. Gould & McFarlan 1959; Terwindt 1967; Wells & Coleman 1978; Walsh & Nittrouer 2003, 2009). Evidence suggests that most fine-grained river sediment that reaches the Atlantic continental shelf of the United States is transported back into estuaries and coastal wetlands and trapped there (e.g. Emery 1968a; Meade 1972; Walsh & Nittrouer 2009). The bottom

Fig. 7. Closeup of the 'Upper Striped Bed' of the Cleveland Ironstone Formation, Jet Wyke, North Yorkshire (UK) showing normally graded thin beds with erosional bases. Gutter casts with steep sides indicate that a significant amount of stratigraphy must be missing at these horizons.

sediments in many of these estuaries differ in their clay mineral composition from sediments carried by their tributary rivers, which suggests that they were derived from the erosion of older offshore deposits (Meade 1972). Similarly, mud in the mouth of the Western Scheldt, The Netherlands, is mostly the product of erosion of older marine mud deposits (Terwindt 1967). Concentrations of suspended matter in shelf waters often decrease rapidly seaward (e.g. McCave 1972), and sediment that escapes estuaries or is discharged by rivers onto the shelf tends to travel along the shore within a few kilometres of the coast rather than seaward (Manheim *et al.* 1970). Most mud transport along the Dutch coast, for example, takes place in a zone close to shore (Terwindt 1967).

Fluctuations in mud content in shelf waters are brought about by seasonal differences in biogenous mud production, by differences in the quantity of mud brought down by rivers, and by waves and currents stirring up previously deposited mud. The last process appears to be particularly significant, and properties of shelf seafloor sediments have a strong influence on the type of material suspended in overlying shelf waters. High concentrations of suspended matter in near-bottom waters of the continental shelf off Massachusetts can be related to bottom-dwelling organisms that either resuspend the sediment, or loosen the sediment so it can be resuspended by bottom currents (Rhoads 1963, 1970; Rhoads & Young 1971; Young & Rhoads

1971). In some cases, however, the suspended matter is organic and probably due to increased primary productivity due to nutrient input.

About 70% of the global area of the continental shelf is covered by (relatively coarse) material that was brought to the shelf during lowered sea levels associated with Pleistocene glacial episodes. Initially it was thought that these, relict, sediments are not in equilibrium with the prevailing present conditions (Emery 1968a). This is partly a legacy of early ideas about the continental shelf, which postulated the existence of an equilibrium shelf profile and its corollary, the graded shelf (Johnson 1919). This concept was challenged by Shepard (1932), who pointed out that chart notation indicated a non-graded condition of the shelf: '[t]he most outstanding feature of the sediments on the continental shelves is the general scarcity of outward decreasing gradation of texture.' Bottom-sediment charts compiled during World War II confirm that this pattern is rare and that grain size of shelf sediments is normally unrelated to distance from shore (Emery 1969). We now know that most shelf sediment shows a mixture of the influences of modern and preexisting conditions (Emery 1968a, Swift *et al.* 1971). Sediments on the Atlantic continental shelf of the United States, for example, are relict in composition but modern in texture (Meade 1972; Milliman *et al.* 1972). Away from the mouths of a few large rivers, the shelf is a sediment-starved environment in which sedimentation is dominated

by the reworking of older sediment by organisms and currents, rather than progradation of new detrital sediment contributed to the ocean by rivers and shore erosion or the accretion of biogenous material. Large swaths of the surface of modern continental margins consist of uncovered older sediments (as old as Eocene) and relict Pleistocene geomorphologic features (Shepard 1932; Northrop & Heezen 1951; Emery 1952, 1968a). For example, a 4–5-m thick mud deposit on the middle and outer shelf south of New England contains mollusc remains that indicate they were deposited when shelf water was substantially colder than at present (Meade 1972 and references therein). Radiocarbon dates of shell material in sands beneath this deposit indicate that the mud accumulated c. 10 ka ago when sea level was approximately 60 m below its present-day level.

Current systems on the continental shelf vary strongly in their nature (weather- or tide-dominated) and intensity. Measured current velocities on the modern Texas and Oregon shelves during fair-weather periods are in the $5-20$ cm s^{-1} range at depths between 70 and 140 m (Snedden et al. 1988; Hickey 1997). In other parts of the continental shelf, however, current velocities are so low that shells rest on the seafloor in hydrodynamically unstable positions (Emery 1968b). The importance of the episodic event on shelf sedimentation is well illustrated by looking at satellite images of turbid, stirred up shelf waters after the passage of hurricanes. Shanmugam (2008) documented shelf-wide sediment resuspension (his Fig. 3) and noted that part of the resuspended sediment is transported basinward over the shelf edge thus affecting deep water sedimentation too. Cyclonic waves can erode and transport sediment in deeper shelf environments at 200 m depth (Komar et al. 1972), that is to say, cyclonic waves can erode the seafloor of the continental shelf at all depths during major events. Cyclone-induced combined flows, a combination of unidirectional currents and oscillatory movement driven by waves, create near-bottom (20–45 m depth) velocities of up to $300-500$ cm s^{-1} (Morton 1988) and thus have significant erosive power. Storm conditions may affect the nature of shelf deposition far more than mild, 'normal' conditions (Swift 1970). So, most of the continental shelf is not a record of ongoing processes, and may have only a limited relationship to mean environmental conditions (Nittrouer & Sternberg 1981). Moore & Curray (1964) stressed the need to take a long term view of shelf processes, because episodic storm sedimentation becomes the dominant steady-state process on longer timescales (Swift 1970).

In deeper water, the common presence of hiatuses in Deep Sea Drilling Program (DSDP) cores belies the idea of a tranquil abyss where a continuous stratigraphic record is being formed (Moore et al. 1978). Before the DSDP, it was thought that the sedimentary record in the deep ocean should be very complete because the environment is quiet and the ultimate resting place of all sediment. According to Maury (1860, p. 312), at the deep seafloor 'the waters [. . .] are at rest [and there is] not motion enough there to abrade [biogenic remains], not current enough to sweep them about and mix up with them a grain of the finest sand.' This turned out not to be the case and, as pointed out by van Andel (1981), barely half of the history of the last 125 Ma is recorded in sediment in the South Atlantic.

Heezen & Hollister (1964) were among the first to infer strong bottom currents in the deep ocean based on sedimentary structures in bottom photographs and cores. Current velocities of $5-15$ cm s^{-1}, capable of eroding calcareous ooze, are common in the deep ocean. Erosion velocities range from 7 cm s^{-1} to 20 cm s^{-1} depending on whether the sediment has been more or less recently deposited (Southard et al. 1971). Critical velocities for surface erosion of pelagic red clay (Lonsdale & Southard 1974) are similar to those reported by Postma (1967) for an estuarine clay. They are more resistant to erosion than calcareous ooze ($30-100$ cm s^{-1}) though this is dependent on water content. However, the effective resistance of the red clay bed is decreased (12 cm s^{-1}) by the presence of manganese nodules by producing local areas of much higher instantaneous boundary shear stress by flow acceleration and separation.

Subsequent observations confirmed the presence of deep currents capable of eroding the seafloor and transporting sediment, and revealed the existence of eddies that can be about 300 times as energetic as the mean current (Hollister et al. 1984). Hollister & McCave (1984) showed that the deep sea stratigraphic record at sites of high eddy kinetic energy is much more fragmentary than generally supposed. It resembles weather-dominated shelf sedimentation more than the continuous accumulation presumed for the deep ocean. In the Nova Scotian Rise area, the mean deposit thickness of 1 ka net accumulation can be removed in a few weeks and redeposited in a few months. However, the signal of alternating erosion and deposition can be erased by bioturbation (Hollister & McCave 1984).

In pelagic regions, sediment supply is controlled by the productivity of surface waters. Regions underneath low productivity surface waters (e.g. gyres) are susceptible to nondeposition, particularly where the seafloor is deeper than the calcite compensation depth. The corrosiveness of seawater with respect to biogenous sediment is a major control on deep sea sediment removal and depends
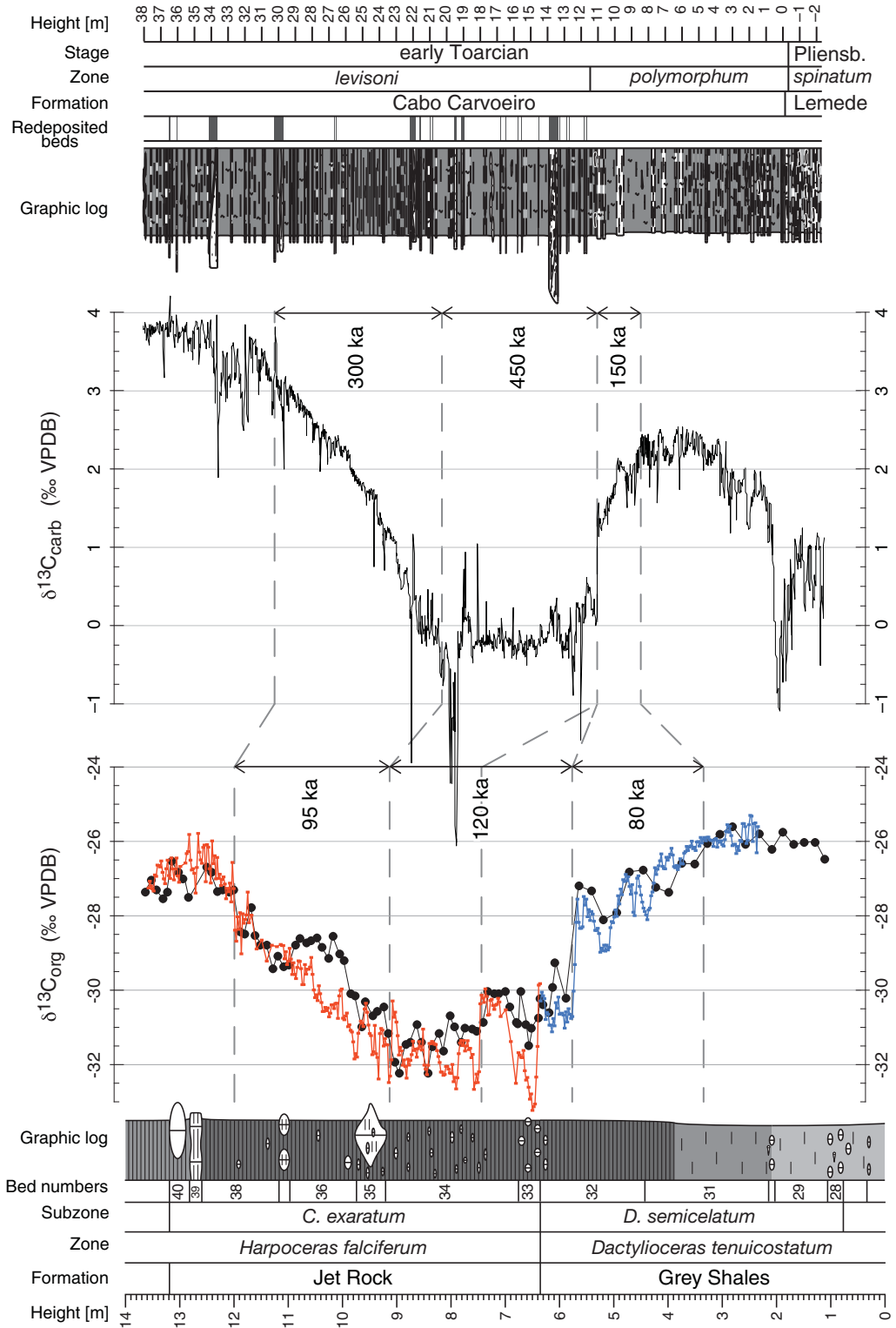
on the chemical properties of seawater, which change with depth. Moore *et al.* (1978) demonstrated that changes in the chemical nature and flow of bottom water are associated with the formation of hiatuses in the marine record, and it has been shown that the distribution of hiatuses in the South Atlantic is generally dependent on depth (Berger 1972). An extreme example of the significance of nondeposition in pelagic regions was recently documented in the South Pacific: a broad region, nearly the size of the Mediterranean Basin, experienced no sediment deposition over more than 80 Ma; that is, since the Late Cretaceous (Rea *et al.* 2006).

Effect on preserved signals: a Toarcian example

Carbon isotope excursions, recorded in lacustrine and marine shales, associated with Mesozoic oceanic anoxic events (OAEs) and Cenozoic hyperthermals (e.g. Palaeocene–Eocene Thermal Maximum) represent perturbations in the global carbon cycle and are synchronous worldwide. Despite an agreement on the stratigraphic definition of most of these events, there is currently no consensus concerning their durations (e.g. Sageman *et al.* 2006; McArthur *et al.* 2008; Suan *et al.* 2008; Dickens 2011; Zeebe *et al.* 2014). Estimates of the duration of the Toarcian OAE, for example, vary by a factor of four between 295 ka and 1100 ka (cf. McArthur *et al.* 2008). These differences obviously have important implications for the interpretation of geological processes and the calculation of their rates. If not due to differences in the stratigraphic definition of the intervals under consideration, what produces the differences in estimates for the duration of events?

Assuming that successive zonal and subzonal durations are equal (Westermann 1984), we can estimate the average duration of a Toarcian subzone at *c.* 540 ka (cf. Jenkyns 1988; Hallam 1997; Jenkyns & Clayton 1997). On the basis that the lower Toarcian Jet Rock of Yorkshire, UK, spans one ammonite subzone (*Cleviceras exaratum* Subzone; Fig. 8), we can say that it represents a

stratigraphic interval of approximately that duration. The shales of the Jet Rock contain regular microscopic laminae of about 50 μm thick. Hallam (1967) interpreted them as annual laminae (cf. Bradley 1931) and later used this interpretation to arrive at an independent calculation for the duration of the *C. exaratum* Subzone (Hallam 1997), which was subsequently used by Hesselbo *et al.* (2000). Although Hallam made an arithmetical error (Cope & Hallam 1998), his calculations result in a duration of 150.6 ka for the 7.53 m-thick *C. exaratum* Subzone, which is about a quarter of the time contained in the stratigraphic interval assuming equal duration of Toarcian subzones. If we use the longer duration of 1.1 Ma suggested by McArthur *et al.* (2000), then the sediments represent less than 14% of the duration of the ammonite subzone.

An alternative way to calculate the duration of the *C. exaratum* Subzone is to apply the rates of deposition of modern organic matter-rich epicontinental shelf sediments. The rate of sedimentation in shelf seas depends on several factors and can be quite variable. Indeed, the 'deposition of shelf muds is highly episodic' (Blatt *et al.* 1991, p. 270). A number of modern analogues may be used to estimate sedimentation rates in Mesozoic epicontinental shelf seas. The black muds of the Baltic Sea have been deposited at an average (uncompacted) sedimentation rate of 20 cm ka^{-1} (Kukal 1990). In Norwegian fjords, the average sedimentation rate of black mud is 0.5–3 mm a^{-1}, or 50–300 cm ka^{-1} (Kukal 1990). Sedimentation rates in turbidite intercalations in these basins may amount to as much as 1000 cm ka^{-1}. Applying the more conservative rates of deposition of the black muds of the Baltic Sea (but see Figs 2 & 6 where evidence for rapid deposition is shown), we find that a mere 188.25 ka would suffice to produce the entire 7.53 m-thick Jet Rock succession (*C. exaratum* Subzone) if its present-day thickness is 20% of the original thickness before compaction. This corresponds to 17–35% of the stratigraphic interval occupied by the deposits (i.e. 540–1100 ka), and it is not too far from Hallam's estimate based on sedimentary structures (Hallam 1997).

The shales of the *C. exaratum* Subzone in Yorkshire contain a record of the Toarcian OAE.

Fig. 8. Carbon isotope data through the lower Toarcian Whitby Mudstone (left) and Cabo Carvoeiro (right) formations from Yorkshire, UK, and Portugal, respectively. Lithological log for the Yorkshire succession and lower resolution carbon isotope dataset (in black) from Hawsker Bottoms adapted from Hesselbo *et al.* (2000). Higher resolution dataset from Kemp *et al.* (2005). The blue and red parts of the curve represent measurement of samples from Hawsker Bottoms and Port Mulgrave, respectively. Lithological log for the Portuguese succession from Hesselbo *et al.* (2007) and carbon isotope data from Suan *et al.* (2008). Estimates of the duration of the carbon isotope excursions from Cohen *et al.* (2007) and Suan *et al.* (2008). The stepwise pattern in the higher resolution Yorkshire curve ('beds' 32–34) is absent in the Portuguese curve. Despite the assumption that carbon isotope excursions associated with Mesozoic OAEs are synchronous, estimates for the duration of the event are shown to vary by a factor of three.

According to Kemp *et al.* (2005) and Cohen *et al.* (2007), the *c.* 9 m of lower Toarcian sediments that record the carbon isotope excursion, associated with the Toarcian OAE in Yorkshire, were deposited in 295 ka (Fig. 8). This implies an average sedimentation rate for those strata of *c.* 15 cm ka^{-1}. In view of the palaeogeographic setting of these sediments, these are rather low sedimentation rates, particularly considering that the same authors, based on high $^{187}Os/^{188}Os$ values at two concretionary levels, proposed an increase in continental weathering rates of 400–800% across this time interval (Cohen *et al.* 2004). At these sedimentation rates, it would take about 100 a to bury a 1.5 cm-thick ammonite lying horizontally on the seafloor, such as the exquisitely preserved specimens found in the Yorkshire succession. If we use the duration proposed by Suan *et al.* (2008) and by McArthur *et al.* (2000) for the same stratigraphic interval based on more robust cyclo- and chemostratigraphic studies (Fig. 8), that is 900–1100 ka, the problem is made 'worse'. The average sedimentation rate for the Yorkshire succession is then 4–5 cm ka^{-1}, and it would take 300–375 a to bury the same ammonite. It is very difficult to accept that an aragonitic shell could be exposed on the seafloor for hundreds of years without gaining any encrusting epifauna or being dissolved away completely (cf. Paul *et al.* 2008 for a similar discussion using a Blue Lias Formation example). Clearly, sedimentation rates were at times much higher than the measured sedimentation rate over the time span of the succession (Fig. 6). If we accept a sedimentation rate of 2 cm a^{-1} (cf. Walsh *et al.* 2004), which would bury an ammonite lying horizontally on the seafloor relatively quickly, we come to the conclusion that the entire Jet Rock succession could represent less than 2 ka of sedimentation, and that most of the time is recorded in bedding planes and other surfaces produced by erosion and nondeposition. This illustrates how measured sediment accumulation rates are inversely related to the time span for which they are determined (Sadler 1981).

In the *C. exaratum* Subzone ('beds' 33–40 Howarth 1962) of Yorkshire (Fig. 8), it is possible to recognize three ammonite faunal subdivisions (or horizons), each about 2.5 m thick (Cope & Hallam 1998). The first 'bed' belonging to each of the three horizons ('beds' 33, 35, and 37) coincides with a faunal change and contains calcareous concretions. In order to grow, these large calcareous concretions require Ca supply by diffusion from the sediment/water interface. The concretionary levels must therefore represent breaks in sedimentation. This supports the interpretation drawn from the $^{87}Sr/^{86}Sr$ record that the Jet Rock is condensed (at these concretion horizons) relative to other parts of the lower Jurassic sequence (McArthur

et al. 2000) and that during much of the time no net sedimentation occurred. A similar interpretation was proposed by Trabucho-Alexandre *et al.* (2012) to explain early diagenetic dolomitization of the lower Toarcian Posidonia Shale Formation in the Dutch Central Graben. Dolomite horizons were interpreted as representing relatively long breaks in sedimentation, perhaps linked to transgressive pulses, because if related to metabolic activity, they were kept relatively close to the sediment surface for prolonged periods of time. Surfaces of preferential cementation in the Blackhawk Formation of the Mancos Shale exposed in the Book Cliffs, Utah, have also been interpreted to occur at horizons where there were significant breaks in sediment accumulation due to sediment bypassing and winnowing of the seafloor (Macquaker *et al.* 2007).

The deposition of the lower Toarcian rock record of Yorkshire clearly only required a small fraction of the time available for sedimentation. The succession contains countless erosional surfaces that may range from diastem to unconformity (Figs 2, 3 & 6). The amount of time not represented by rock in the succession – 65% to >80% of the time – must therefore be contained in these erosional surfaces, and this explains why the accumulation rate obtained by dividing the thickness of the deposit by the length of the interval it occupies is much lower than the deposition rates measured directly in equivalent modern environments (Sadler 1981; Van Andel 1981).

Estimates of the duration of the negative carbon isotope excursion associated with the Toarcian OAE based on the study of lower Toarcian shales from Yorkshire are three times shorter than estimates based on the study of rocks from Portugal (Hesselbo *et al.* 2000; Kemp *et al.* 2005; Cohen *et al.* 2007; Suan *et al.* 2008). Negative carbon isotope excursions in Meso-Cenozoic successions are interpreted as involving the input of ^{13}C-depleted carbon from an unknown source to the exogenic carbon cycle. Thus, different estimates for their duration lead to different rates of carbon release or transfer between reservoirs and therefore to disparate interpretations of what the carbon source may be. The assumed short duration of the Toarcian carbon isotope excursion (295 ka), for example, has been considered by some authors as incompatible with slow rates of volcanogenic carbon degassing; these authors interpret the excursion as having been caused by rapid and massive dissociations of methane clathrates (Hesselbo *et al.* 2000; Kemp *et al.* 2005; Beerling & Brentnall 2007). Other authors argue for a longer duration of the excursion (900–1100 ka) and interpret the Toarcian OAE as part of a longer-term history of environmental change (Wignall *et al.* 2005; Suan
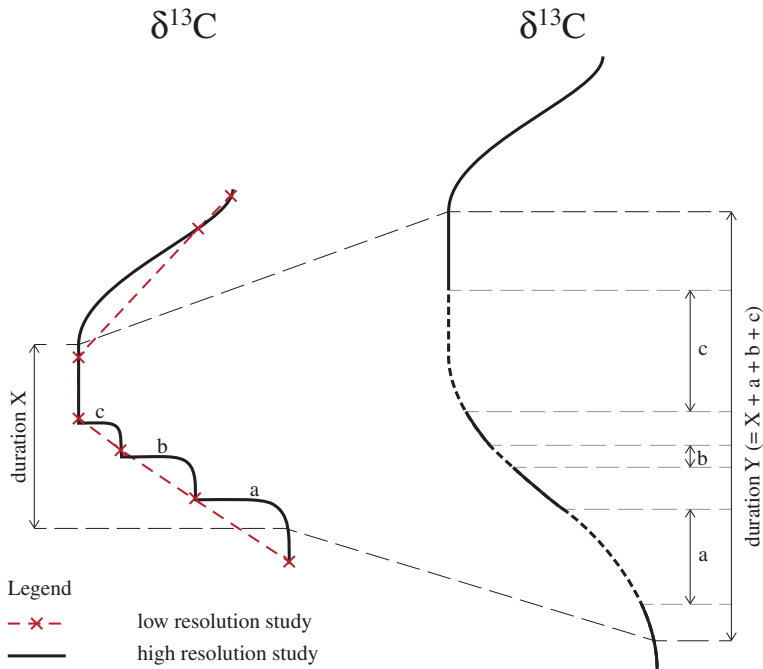
Fig. 9. Effect of stratigraphic incompleteness on geochemical records. On the left, the carbon isotope curve is shown at two resolutions: low-resolution studies do not resolve abrupt shifts, whereas high-resolution records reveal significantly more detail. On the right, a reconstruction of the curve if the succession were known to be incomplete. In this example, all abrupt shifts are artefacts of stratigraphic incompleteness at horizons a, b, and c. The missing time does not necessarily need to be in a single horizon. Several hiatuses of shorter duration below sampling resolution have the same effect as a single longer hiatus.

et al. 2008). I suggest that differences in estimates for the duration of the event (and other similar events) are due to different levels of stratigraphic completeness in successions studied by different authors, and that estimates are always likely to be underestimates of the actual duration of geological events.

In a few stratigraphic successions, the Toarcian carbon isotope excursion exhibits a stepwise pattern that has been interpreted as the record of three astronomically forced, rapid pulses of methane release (Kemp *et al.* 2005; Cohen *et al.* 2007; Hermoso *et al.* 2012). The recognition of the stepwise pattern is equivocal (e.g. Zeller 1964), and the process behind its development is open to more than one interpretation (Fig. 9). If the stepwise pattern were a record of several successive global perturbations in the carbon cycle, then we should expect to find the record of these perturbations worldwide. Although, as pointed out by Van Andel (1981), if we believe that we have located the deposits associated with a brief event in several stratigraphic successions, we ought to consider the extremely low probability that an event on short

timescales would be preserved at all, even in just a single section. The record of the Toarcian OAE in Peniche, Portugal, appears to be significantly more complete than the record of the same event in Yorkshire, UK. However, the stepwise pattern observed in the shallower, mud-rich Yorkshire succession is missing from the deeper, carbonate-rich Peniche succession (cf. Kemp *et al.* 2005; Suan *et al.* 2008). Given that 65% to >80% of the duration of the Toarcian OAE in Yorkshire is recorded as bedding planes, it appears that the steps observed in the carbon isotope record are artefacts of stratigraphic incompleteness, rather than a record of abrupt environmental change.

Conclusions

Away from the mouths of large rivers and associated submarine canyons, the seafloor is a sediment-starved environment in which reworking of older sediment by contemporary processes is, by volume, the main sedimentation process. In marine environments, nondeposition and erosion are the rule

rather than the exception. Mud, which includes terrigenous and biogenous fine-grained material, is mostly present in these environments as aggregates that behave as silt- and sand-sized particles. Aggregated mud deposits have high water contents (>80%) and are therefore relatively easy to erode. Despite being difficult to observe, erosional features in shales are common.

Marine shale successions are arguably the best archives of Earth history. Marine shales form conformable successions amenable to the multidisciplinary study required to reconstruct past environmental change. The degree of completeness of a shale succession is a critical factor in the interpretation of the geological record of climatic, oceanic, and biogeochemical processes, in the prediction of timescales of those processes, in the determination of the duration of events, and in the establishment of correlations between successions. Geological signals preserved in the shale record may be garbled due to stratigraphic incompleteness even when no hiatuses are equal to or longer than the scale of observation. However, as the resolution at which studies are carried out increases, so we approach the limits of the shale sedimentary record. In many cases, we may have already unknowingly exceeded them.

The stepwise pattern observed in carbon isotopes measured in the lower Toarcian of Yorkshire, for example, is likely to be an artefact of stratigraphic incompleteness. The Toarcian OAE is not as short as proposed by authors who studied the Yorkshire succession, and probably represents a much longer-term history of environmental change driven by processes acting on longer time scales.

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