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The Bude Formation (Lower Westphalian), SW England: siliciclastic shelf sedimentation in a large equatorial lake

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ABSTRACT

The 1300-m-thick turbiditic Bude Formation was deposited in a lake, Lake Bude, but disagreement persists over whether the environment was a deltaic or deep-water fan. The tectonic setting of the lake was the northern flank of a northerly advancing Variscan foreland basin, close to the Westphalian palaeo-equator. Palaeocurrents indicate sediment sourcing from all quadrants except the south.

There is a dm–m scale cyclicity, whereby sandstone bodies comprising amalgamated event beds alternate with mudstone intervals containing non-amalgamated event beds. The 'ideal' cycle is a symmetrical coarsening-up/fining-up cycle, consisting of three facies (1, 2 and 3) arranged in 12321 order. Facies 3, in the middle of the cycle, is an amalgamated sandstone body up to 10 m thick which interfingers laterally with thin (cm) mudstone layers. The sandstone body comprises amalgamated beds of very fine sandstone which are largely massive and up to 0·4 m thick. Channels are absent except for scours up to 0·2 m deep which truncate the interfingering mudstone layers. Sandstone bodies are inferred to be tongue-shaped in three dimensions. Facies 1 and 2, completing the 12321 cycle, are respectively dark-grey fine and light-grey coarse, varved(?) mudstone containing thin (< 0·4 m) sandstone event beds. Fossils and burrows indicate that facies 1 and 2 were deposited, respectively, in brackish (rarely marine) and fresh water. Hence, the ideal cycle (12321) reflects an upward decrease then increase in salinity (brackish–fresh–brackish); this is attributed to the lake sill being periodically overtopped by the sea, due to glacio-eustatic sea-level oscillations. The resulting oscillations in lake depth produced the coarsening-up/fining-up (regressive–transgressive) cyclicity, the central sandstone body representing the regressive maximum.

Event beds are interpreted as river-fed turbidites deposited during catastrophic storm-floods. Combined-flow ripples and other wave-influenced structures occur in event beds throughout the ideal cycle, suggesting deposition of the entire Bude Formation above storm wave base. The proposed environment is a shelf, of continental-shelf dimensions, but lacustrine instead of marine. Sandstone bodies are interpreted to be river-connected tongues or lobes. The absence of cycles containing nearshore or emergent facies is attributed to: (i) the lake sill preventing the water level from falling below sill level, thereby insulating the lake floor from eustatically forced emergence; and (ii) relatively distal deposition, beyond the reach of shoreline progradations. The lack of palaeoflow from the south is attributed to a (now eroded?) deep-water trough lying to the south, in front of the northerly advancing orogen.

Some facies 2 laminated mudstone beds grade laterally into massive and/or contorted beds, interpreted as *in-situ* seismites (Facies 4), consistent with an active foreland basin setting. Development of seismites was possibly facilitated by gas bubbles and/or weak cohesion in the (fresh water) bottom mud.

The late Quaternary Black Sea, with its broad northwestern shelf, is probably a good physiographical analogue of Lake Bude, and was likewise fresh at times.

INTRODUCTION

Although the literature on ancient lake sediments is growing rapidly, descriptions of Palaeozoic and older lake deposits are few (but see, for example, Donovan *et al.*, 1976; Van Dijk *et al.*, 1978; White & Youngs, 1980). The facies analysis described here deals with a Palaeozoic lake and provides insights into the depositional environments, hydrodynamics, biology, palaeoclimate, and bottom-mud properties of this ancient lake.

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GEOLOGICAL SETTING

The Bude Formation is of early Westphalian age (see below), and was deformed in the Variscan orogeny in Late Carboniferous time (Hobson & Sanderson, 1983). Post-orogenic Stephanian–Triassic rocks lie unconformably above (the New Red Sandstone; Laming, 1982). The Bude Formation was deposited in a foreland basin ahead of the northward-verging and northward-advancing Variscan deformation front (Selwood & Thomas, 1988). Deposition took place on the northern, non-orogenic flank of the basin, as shown by a lack of palaeocurrents from the south (see below) and by the compositionally mature (Burne, 1973), fine-grained nature of the Bude Formation.

Lying conformably beneath the Bude Formation is the Crackington Formation, consisting of Namurian–Westphalian shales with turbidite-like sandstone beds (Edmonds *et al.*, 1975).

STRUCTURE AND STRATIGRAPHY OF THE BUDE FORMATION

Exposure of the Bude Formation is poor inland. By contrast, the N–S coastal stretch from Hartland Point to Widemouth Sand (Fig. 1) offers superb strike-perpendicular exposure of the entire formation in near-continuous cliffs and wave-cut platforms (see cliff sections by Lovell, 1965; King, in Freshney *et al.*, 1972; Freshney *et al.*, 1979). The strata are folded in a series of mainly upright chevron folds trending E–W. Freshney & Taylor (1972) measured a composite section along the coast and calculated a total thickness of about 1300 m. Correlation along the folded and faulted cliffs is aided by a number of fossiliferous 'key shales' which punctuate the otherwise unfossiliferous succession (Fig. 2; Freshney & Taylor, 1972).

The base of the Bude Formation is defined as the top of the Hartland Quay Shale (Fig. 2; Freshney *et*

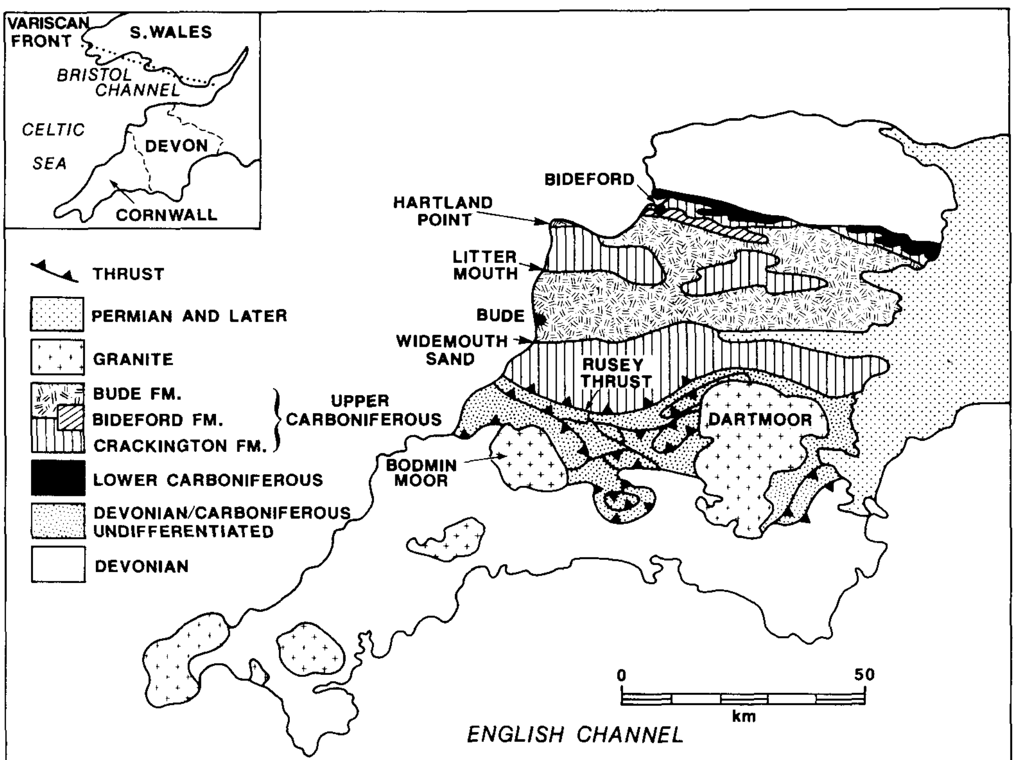

Fig. 1. Geological sketch map of SW England. After Edmonds *et al.* (1975, Fig. 1), Selwood & Thomas (1988, Fig. 3.1), and British Geological Survey 1:50 000 Sheets 292 and 293.

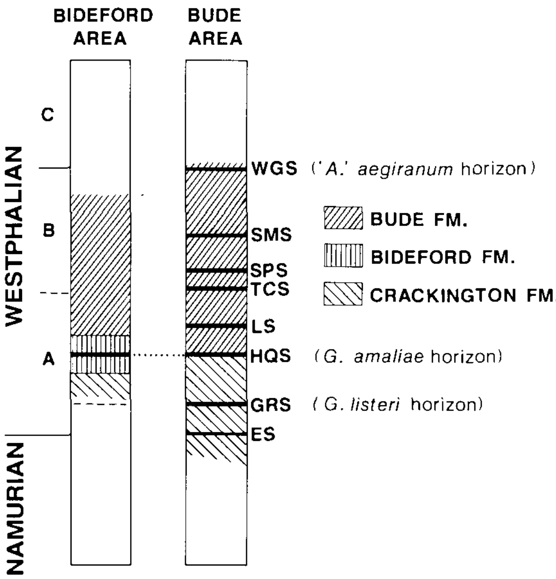

Fig. 2. Westphalian stratigraphy of the Bideford and Bude areas. Key shales (from Freshney & Taylor, 1972): ES = Embury Shale; GRS = Gull Rock Shale; HQS = Hartland Quay Shale; LS = Longpeak Shale; TCS = Tom's Cove Shale; SPS = Saturday's Pit Shale; SMS = Sandy Mouth Shale; WGS = Warren Gutter Shale. Goniatite horizons (from Freshney *et al.*, 1979): *G.* = *Gastrioceras*; '*A*' = '*Anthracoceras*'. The *G. amaliae* horizon is known in both the Bideford and Bude areas.

al., 1979), above which the 'massive sandstones of Bude Formation type become progressively more common' (Freshney & Taylor, 1972, p. 467). The age of the Bude Formation is late Westphalian A to earliest Westphalian C, based on goniatites in the Hartland Quay Shale and in the Warren Gutter Shale (Fig. 2; Ramsbottom *et al.*, 1978; Freshney *et al.*, 1979). Thus, deposition spanned about 5 Ma (Harland *et al.*, 1982).

PALAEOGEOGRAPHIC SETTING

There is general agreement that: (i) the Bude Formation was deposited entirely offshore, based on the presence of turbidite-like beds and the lack of evidence for emergence (Reading, 1963; Goldring & Seilacher, 1971; Freshney *et al.*, 1979; Melvin, 1986); and that (ii) the water body was fresh or brackish for most of the time, because body fossils and trace fossils are scarce and because the only definite marine fossils are confined to the key shales (Goldring & Seilacher, 1971; Freshney & Taylor, 1972; Burne, 1973). Hence, an offshore lacustrine environment is indicated (Goldring & Seilacher, 1971; Higgs, 1986a). The lacustrine interpretation is supported by the presence of a fossil fish unknown elsewhere (White, 1939), since endemic fishes are common in lakes (Beadle, 1981). The marine key shales imply intermittent marine incursions into the lake, suggesting that the lake was near sea-level (Goldring, 1978). The lake, 'Lake Bude' (Higgs, 1986a), lay within 5° of the palaeomagnetically-determined equator (e.g. Scotese *et al.*, 1979). A humid climate is indicated by the prevalence of Coal Measures among coeval deposits elsewhere in Britain.

Although the extent of Lake Bude is unknown, the fact that marker beds (e.g. Tom's Cove Shale, Fig. 2) can be traced over the entire N–S extent of the coast section suggests that the lake was larger than the present N–S outcrop length. This figure can probably be increased by 50% to allow for Variscan tectonic shortening (Coward & Smallwood, 1984). The east and west shores of Lake Bude are unknown because the Bude Formation passes beneath the sea to the west and beneath younger strata to the east. The position of the south shore, which was possibly an advancing mountain front (above), is uncertain because coeval rocks are unknown south of the Bude Formation. The north shore may have coincided with the southern margin of the 'Bristol Channel landmass', invoked by previous authors (e.g. Anderton *et al.*, 1979) to explain why early Westphalian palaeocurrents simultaneously flowed south in SW England (Bideford and Bude formations) and north in southernmost Wales.

Coeval deltaic deposits of the Bideford Formation crop out immediately north of the Bude Formation (Figs 1 & 2; Edmonds *et al.*, 1979), and are possibly shoreline equivalents of the Bude Formation (Melvin, 1986; Higgs *et al.*, 1990). However, the nature of the lateral transition from the Bideford Formation into the Bude Formation is unknown due to poor exposure. A major thrust fault may divide the two successions (Reading, 1965), and two separate nappes are suggested by vitrinite reflectance evidence (Cornford *et al.*, 1987). Hence, the palaeogeographic relationship between the two successions is uncertain.

PREVIOUS INTERPRETATIONS

Reading (1963) drew attention to a 'fundamental difference of opinion' regarding the depositional environment of the Bude Formation. He was referring to the conflict between Owen (1934, 1950), who favoured a deltaic origin, and Ashwin (1957), who proposed deposition from turbidity currents at the foot of a continental slope. Remarkably, this shallow-vs.-deep conflict persists: Freshney *et al.* (1979) suggested deposition on a delta slope in water depths of only a few metres at times, while Burne (1969) and Melvin (1976) both postulated a 'relatively deep-water' fan. Melvin (1986, 1987) attempted to reconcile the shallow-vs.-deep controversy by proposing a fan which developed in 'relatively shallow (shelf as opposed to abyssal) water depths'. This interpretation was challenged by Higgs (1987), who argued for non-fan shelf deposition (see also Higgs, 1986a; Higgs *et al.*, 1990).

The palaeobathymetry of the underlying Crackington Formation, and the Dinantian shales below that, is also uncertain: a pro-shelf, 'basin floor' was invoked by Thomas (1982, Fig. 3.5; 1988), but shelf deposition is an alternative possibility, especially in view of the neritic origin of Upper Devonian and basal Dinantian sediments that lie conformably below (Goldring, 1971).

FIELD-WORK

Field-work was confined to the folded coastal exposures. In the cliff face, individual beds can be traced for up to 100 m, to the cliff top, but can seldom be correlated into the next fold limb because marker beds are scarce. Beds can be traced across the wave-cut platform for up to 100 m.

Ten representative, stratigraphically dispersed cliff sections up to 112 m thick were measured at a scale of 1:20 (see Higgs, 1986b, for details). Three of the sections represent a single stratigraphic interval, between the Tom's Cove Shale (TCS) and the Saturday's Pit Shale (SPS; Fig. 2), enabling an assessment of N–S lateral variability to be made (Higgs, 1986b).

FACIES

Four facies are recognized: (1) dark and (2) light mudstone with intercalated sandstone beds; (3) amalgamated sandstone beds; and (4) disturbed beds.

Facies 1: dark mudstone with graded sandstone beds

Description

Facies 1 occurs as units of dark-grey mudstone (silty claystone in the classification of Tucker (1982, Fig. 3.2), 0·02–2·0 m thick, containing sharp-based, graded sandstone beds up to 0·2 m thick, spaced up to 0·3 m apart (described separately below). The mudstone is laminated, consisting of graded couplets up to 2 mm thick (average *c.* 0·5 mm). Each couplet is sharp-based, and grades upward from very-fine silt into dark clay (Fig. 3a). The mudstone consists mostly of illite and quartz (Freshney *et al.*, 1979) and contains 0·5–1·0% organic carbon (four samples; Higgs, 1986b). Comminuted plant debris is common on parting surfaces and in thin sections (Fig. 3a).

A few facies 1 units are unusually thick, up to 10 m, with sparse sandstone interbeds (Fig. 3b). These units include the key shales, which contain calcareous concretions and possible fish coprolites (Freshney *et al.*, 1979). The key shales contain the only body fossils known in the Bude Formation. Goniatites and pelagic bivalves occur in the Longpeak, Sandy Mouth and Warren Gutter Shales (Fig. 2; Freshney *et al.*, 1979); in each case, the fossils occur in an interval less than 0·2 m thick (R. T. Taylor, pers. comm., 1985), associated with finely disseminated pyrite. Other key shales have yielded fish (White, 1939; King, 1966), and specimens of an eocarid crustacean (White, 1939). Microfossils appear to be absent (E. C. Freshney, pers. comm., 1984; Palmer, 1984).

Various burrows occur in facies 1. Retrusive *Diplocraterion parallelum* is common (Fig. 3c; see Fürsich, 1974), whereas *Arenicolites* sp., *Phycodes* sp., *Skolithos* sp., and *Teichichnus* sp. are rare (Higgs, 1986b). These five ichnogenera are invariably confined to the top few centimetres of sandstone interbeds, with their apertures at the upper surface. Besides these 'bed-top' burrows, *Planolites* sp. is common, between sandstone interbeds.

Interpretation

The three goniatite-bearing horizons are clearly marine. Outside these horizons, the lack of pyrite and unequivocally marine body fossils and trace fossils suggests brackish or fresh water. A brackish environment seems the more likely because the carbon-to-sulphur (C/S) ratio in four out of five Facies 1 mudstone samples was below 10 (Higgs, 1986b), whereas C/S ratios in fresh-water deposits generally exceed 10 (Berner & Raiswell, 1984). Furthermore, the diversity of burrows may indicate brackish rather than fresh water, because there is evidence that fresh-water burrowers were few prior to Permian time (Miller, 1984). All six ichnogenera in facies 1 are consistent with a brackish environment, having been reported in (marine-influenced) brackish Palaeozoic

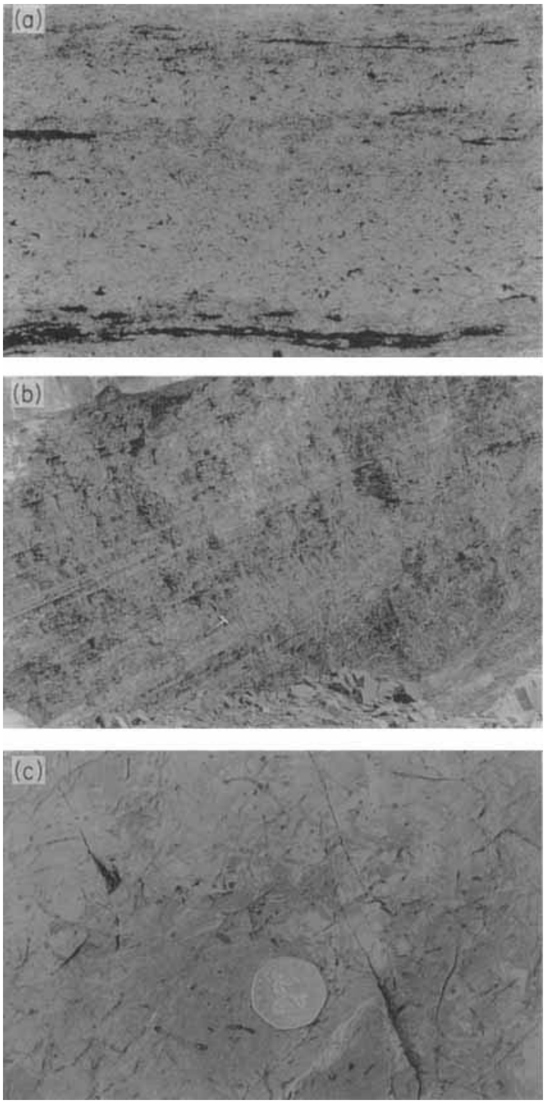

Fig. 3. (a) Facies 1 mudstone in thin section, showing two sharp-based, light–dark complets, each consisting of very-fine silt grading up into clay. The large, dark, elongated particles are carbonized plant fragments. Field of view = 0·5 mm high. Plane-polarized light. (b) Thick Facies 1 mudstone unit. Saturday's Pit Shale, at Saturdays Pit, UK National Grid reference SS 2026 0681. (c) *Diplocraterion parallelum* in transverse section on the upper surface of an event bed in Facies 1. Note characteristic dumb-bell shape (e.g. Goldring, 1962); some dumb-bells are curved, due to lateral shifting of successive U-tubes (Fürsich, 1974).

deposits elsewhere (Chisholm, 1970; Goldring & Langenstrassen, 1979; Tevesz & McCall, 1982; Eagar *et al.*, 1985). Similarly, fish and eocarid crustacea, both of which occur in facies 1, are known in brackish Palaeozoic deposits elsewhere (Brooks, 1969; Miles, 1971).

The bed-top burrows reflect colonization by opportunistic organisms during the late stages of each sand-depositing event (Frey & Pemberton, 1984). Rapid sedimentation is reflected in the retrusive behaviour of the *Diplocraterion parallelum* animal (Fürsich, 1974). The fact that burrow apertures do not extend above the sandstone bed suggests that the animal died when normal mud deposition resumed; death may have been due to: (i) clay particles clogging the respiratory apparatus (S. G. Pemberton, pers. comm., 1986); or (ii) development of oxygen deficiency in the bottom water between sand-depositing events. If oxygen was indeed restricted, the presence of *Planolites* and eocarids implies dysaerobic rather than anaerobic conditions (cf. Rhoads & Morse, 1971); both of these taxa are known to have been tolerant to low levels of oxygen in the late Palaeozoic (Brooks, 1969; Jordan, 1985). Bottom water was oxygen-deficient during deposition of the key shales, as shown by preservation of fish remains and by C/S ratios of less than 1 in each of two key-shale samples analysed (Berner & Raiswell, 1984; Higgs, 1986b).

The fine grain size of the mudstone indicates suspension deposition from a low-velocity current. The current was probably a river-fed surface plume, reflecting the low density of inflowing river water relative to the brackish lake water (cf. Bell, 1942). Interlamination of silt and clay suggests a fluctuating current, while the regularity of the lamination suggests a seasonal control. Hence, the couplets may be varves, as suggested by Goldring & Seilacher (1971), each couplet representing an annual wet-season–dry-season cycle. A similar monsoonal climate has been inferred from Westphalian deposits elsewhere in England (Broadhurst *et al.*, 1980).

With regard to water depth, sandstone beds with wave-influenced structures (see below) are common in facies 1. Beds are generally spaced no more than 1 m apart, suggesting that facies 1 was deposited above storm wave base. In the key shales, however, wave-worked beds are up to 5 m apart, so parts of the key shales could have accumulated below storm wave base.

Facies 2: light mudstone with graded sandstone beds

Description

Facies 2 is coarser than facies 1. It consists of light-grey mudstone (clayey mudstone to clayey siltstone) with sandstone interbeds, and occurs as units 0·02–10·0 m thick (Fig. 4a). Many sandstone beds show load casts. Sandstone beds are seldom more than 0·05 m apart, and are commonly amalgamated (Fig. 4b), as shown by internal planar sutures and by thin mudstone partings. Amalgamated sandstone bodies thicker than an arbitrary 1 m are assigned to Facies 3.

The mudstone consists of light–dark couplets like those of facies 1, except coarser grained. Within each couplet, a sharp-based layer of fine to medium silt grades up into a darker layer of silty clay. Couplets are up to 3 mm thick (average *c.* 0·5 mm). The mineral composition is quartz, illite, kaolinite, chlorite, siderite and minor feldspar (R. J. Merriman, pers. comm., 1987). Organic carbon in five samples was 1·8–2·9% (average 2·3%; Higgs, 1986b). Plant fragments are conspicuous in thin section.

Facies 2 includes a rare, coarser subfacies, comprising millimetre-scale couplets in which medium or coarse silt grades up into clayey silt. The rock splits readily on carbonaceous and micaceous partings, revealing tracks and trails (see below). Units of this flaggy subfacies are up to 2 m thick, including intercalated sandstone beds.

Facies 2 lacks body fossils and burrows. Xiphosurid trackways (*Kouphichnium*) and fish trails (*Undichna*) occur in the flaggy subfacies (King, 1966; Goldring & Seilacher, 1971; Higgs, 1988; Tyler, 1988).

Interpretation

The lack of marine fossils in Facies 2 suggests either fresh or brackish water. A fresh-water interpretation is preferred, because: (i) bed-top burrows are absent, in contrast to their presence in (brackish) Facies 1; and (ii) C/S ratios, measured on five mudstone samples, exceed 10 (22·8–39·0; Higgs, 1986b). Xiphosurids and fishes are known to have tolerated fresh

Fig. 4. Facies 2. Hammer is 31 cm long. (a) Interbedded mudstone and sandstone. (b) Three amalgamated sandstone beds. Top to the left. The 5-cm bed immediately left of the hammer head is mudstone. Planar partings mark the sutures between adjoining beds.

water in Carboniferous time (Størmer, 1955; Miles, 1971). The bottom water was aerobic, at least for the flaggy subfacies, as shown by the presence of tracks and trails.

The sediment-supplying current was weak, as indicated by the overall fine grain size. The regular coarse–fine lamination suggests a wet–dry seasonal control. The mud was probably deposited from a continuous river-fed turbidity current (e.g. Gould, 1960; Houbolt & Jonker, 1968), because the lake, being fresh, would have been prone to underflows from denser turbid river inflow. The current may have flowed year-round, but would have slackened during the dry season. Continuous underflow would have ensured aerobic bottom conditions.

The presence of wave-influenced structures in sandstone interbeds (see below) suggests deposition of facies 2 above storm wave base. The xiphosurid trackways also suggest relatively shallow water, because fossil xiphosurids and/or xiphosurid traces are known only in lacustrine and marine rocks which appear, on other evidence, to have been deposited in water no deeper than a few tens of metres (Bandel, 1967; Goldring & Seilacher, 1971; Anderson, 1975; Chisholm, 1983).

The graded sandstone beds of Facies 1 and 2

Description

Individual (non-amalgamated) sandstone beds are up to about 0·2 m thick in facies 1 and 0·4 m thick in facies 2. The sandstone beds in facies 1 are identical to those in facies 2, apart from being thinner and lacking load casts. The grain size never exceeds very fine sand. Beds are generally tabular for at least tens of metres laterally, but in some cases, a 0·1–0·2-m bed pinches out laterally within 5 m.

An 'ideal' Bude Formation sandstone bed consists of four divisions (Figs 5 & 6a). A massive basal division (B) consists of ungraded very fine sand. This grades up abruptly, over about 0·01 m, into a 0·02–0·05-m interval of medium to coarse silt showing parallel lamination (P), followed by ripple cross-lamination (R). The silt is capped by 0·02–0·05 m of light-grey mudstone (M). The ideal sandstone bed is less common than incomplete variants: of the lower three divisions, any one or two may be missing (Fig. 6b), but the order in which divisions occur is invariable.

The ripples of division R are in sharp contact with the M division, and commonly weather out. Ripples

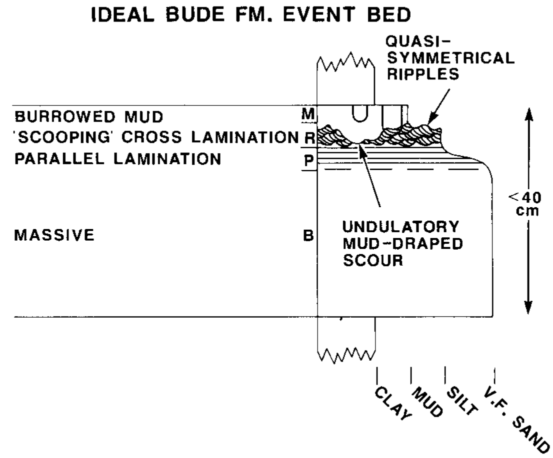

Fig. 5. Schematic ideal Bude Formation sandstone bed, showing the four successive divisions, B, P, R and M (see text). The background mud is shown as being of clay grade (i.e. Facies 1). When the background mud is Facies 2 instead, the M division lacks burrows.

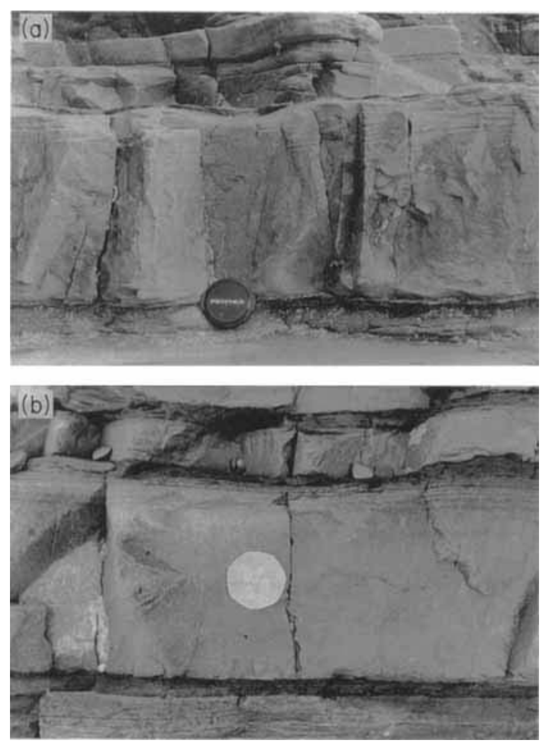

Fig. 6. (a) An ideal Bude Formation sandstone bed, encased in Facies 2 mudstone. All four divisions are visible, starting at the base: massive, B; parallel laminated, P; ripple cross-laminated, R; and mud, M (slightly recessive). Note quasi-symmetrical ripple profiles. Lens cap is 5·3 cm across. (b) Sandstone bed showing only the B, P and M divisions. Coin is 3·0 cm across.

are typically straight to slightly sinuous, and quasi-symmetrical, with a ripple symmetry index (RSI) ranging from 2 to 3. Ripple spacing is 0·05–0·15 m, and height is 0·005–0·02 m. Cross-lamination, viewed perpendicular to ripple crestlines, consists of gently climbing lamina-sets of irregular size and form. Set boundaries are erosional, and are gently curved to scooping (cf. de Raaf *et al.*, 1977). Laminae are tangential to the lower set boundary, and show preferential one-way dips. In some beds, lamina-sets contain sparse burrows.

A few sandstone beds show much larger ripples, with straight to slightly sinuous, bifurcating crestlines, and rounded, symmetrical crest profiles. Spacing is 0·3–1·0 m, and height is 0·02–0·1 m. All examples observed by the author are internally structureless.

Hummocky cross-stratification (HCS) was seen in about 20 beds (Higgs, 1983, 1984). HCS is omitted from the ideal sequence because of its scarcity. Most beds with HCS have a basal massive division; some beds are capped by cross-lamination. A few thin (cm) sandstone beds show 'type 3' ripple-drift cross-lamination (Walker, 1963). The associated ripples fade upward into mudstone and are strongly asymmetrical (RSI 4–6).

Sandstone beds show various sole marks, including flutes, grooves, longitudinal scours, prod marks, bounce marks and rare gutter casts. In some beds, tool marks are multidirectional.

In some cases, a sandstone bed is incised by an undulatory mud-draped scour. Such scours are usually lined by a 1–2-mm silt veneer (cf. Fig. 5 in Walker *et al.*, 1983). Scour surfaces are similar in form and dimensions to HCS set boundaries (e.g. Brenchley, 1985), suggesting a possible genetic link. Mud-draped scours can cause sand beds to pinch out laterally.

Interpretation

Each sand bed was produced by a waning-energy depositional event, as evidenced by its sharp base and grading; hence, the sandstones are 'event' beds (cf. Seilacher, 1982). The quasi-symmetrical ripples are intermediate in morphology between current ripples and wave ripples (Higgs, 1984); they are interpreted as combined-flow ripples formed under the joint action of a unidirectional, sediment-supplying current and a wave-induced oscillatory current (Higgs, 1984; cf. Harms, 1969). In the cross-lamination, wave action is suggested by the strongly curved set boundaries (cf. de Raaf *et al.*, 1977), while a superimposed unidirectional current is indicated by the one-way lamina dips (Higgs, 1984).

There is additional evidence for wave action, namely the HCS (Brenchley, 1985) and the multidirectional tool marks (Gray & Benton, 1982). Mud-draped scours are also commonly attributed to wave action (Walker *et al.*, 1983; Brenchley, 1985; Craft & Bridge, 1987), and gutter casts are perhaps also indicative of waves (Brenchley, 1985).

The large ripples are interpreted as wave ripples, based on their symmetrical form and on the presence of bifurcations (Reineck & Singh, 1980). These ripples are unsuitable for calculating ancient sea conditions, because the ripple index exceeds 7·5 (Allen, 1981). Large wave ripples are well known in modern and ancient sediments, but only in coarse sand or gravel (Leckie, 1988), unlike the very fine sand of the Bude Formation.

The evidence for bottom-feeling waves suggests that the depositional events were storms. The sand-supplying current may have been: (i) a slump-generated turbidity current; (ii) a wind-driven current; or (iii) a river-fed turbidity current (i.e. underflow), induced by catastrophic storm-flooding. Slump-generated turbidity currents are the least likely alternative, because they tend to wane continously once deposition has begun (Harms *et al.*, 1982), resulting in a graded bed (Middleton & Hampton, 1976), whereas the lack of grading in the massive (B) division of the Bude beds (Fig. 5) suggests deposition from a continuously replenished current. Wind-driven currents are known to move sand on modern shelves during storms (Swift *et al.*, 1986), but their ability to deliver and deposit a graded sand bed is unproven. The third alternative, river-fed turbidity currents, are likely to have occurred in Lake Bude, given the evidence for fresh to brackish water (above), which would have favoured underflow. Similar ephemeral underflows in modern lakes can transport coarse sand (Lambert *et al.*, 1976), and are probably responsible for sand interbeds up to 0·3 m thick in the Recent bottom muds of Swiss lakes (Houbolt & Jonker, 1968; Lambert *et al.*, 1976; Sturm & Matter, 1978). River-fed turbidity currents also occur in the sea (Wright *et al.*, 1988; Higgs, 1990). Hence, river-fed turbidity currents, induced by storm-flooding, are considered the most likely sediment-supply process for the Bude Formation event beds.

Deposition of the ideal event bed can be divided into four stages, corresponding to the B, P, R and M divisions. In Stage 1, the B division is inferred to have been deposited by fall-out from a two-component flow comprising a river-fed turbidity current and a wave-induced oscillatory current. The river-fed current was initially dominant, as shown by the prevalence of unidirectional sole marks. Deposition was sufficiently rapid to prevent traction lamination from forming (Middleton & Hampton, 1976), while the lack of grading implies that the current velocity decreased spatially but not temporally (Harms *et al.*, 1982). Stage 2 reflects a weakening of the river-fed current, and therefore of the overall combined flow, allowing finer sediment (silt) to settle. The P division was deposited during this stage, either by a pulsing unidirectional flow or by an asymmetrical reversing flow (cf. Harms *et al.*, 1982). In Stage 3, the overall current weakened, and the R division was deposited under oscillatory flow; the river-fed current was still active, imparting an asymmetry to the flow, as shown by the one-way lamina dips. Finally, in Stage 4, currents returned to their normal fair-weather velocities, and storm-suspended clay and silt settled to form the M division, perhaps over several days. Deposition of 'background' mud then resumed.

Fig. 7. General view of the Bude Formation, showing three amalgamated-sandstone units (Facies 3), of which the lowest is in the foreground, at beach level. Pack in left foreground and hammer at centre for scale. All three sandstone units show planar partings (see text). In the central unit, the undulatory surface at the level of the hammer head is a mud-draped scour. This and other mud-draped scours are responsible for the lenticularity evident in this sandstone unit. Interval 2–13 m in Fig. 14(a).

Facies 3: amalgamated sandstones

Description

Facies 3 is defined as any sandstone unit thicker than 1 m, and includes the 'thick massive sandstones with few visible structures' (Reading, 1963) which characterize the Bude Formation (Fig. 7). Units are up to 10 m thick. The sandstone is predominantly massive, and the grain size never exceeds very fine sand. Dewatering structures are common and fossils are absent. Lensoidal ferruginous concretions up to 2 m long, orientated parallel to bedding, occur in some thicker units, as does pressure-solution cleavage (Beach, 1977).

The bases of facies 3 units are planar (i.e. non-channelled), and show flutes, longitudinal scours, or tool marks. Tops are flat, or show quasi-symmetrical ripples. There are also hummocky upper surfaces; some of these are mud-draped scours (see below), while others are possibily formed by compaction-drape over large concretions. Many facies 3 units show bedding-parallel planar partings, 0·05–0·4 m apart vertically (Fig. 7). Other units have partings which are undulatory instead of planar, and less continuous laterally. Planar or undulatory partings are the only visible 'structure' in some facies 3 units.

Facies 3 commonly contains angular, bedding-parallel mudstone slabs, 0·01–0·1 m thick and 0·05–0·5 m long. The mudstone is of facies 1 or 2. Slabs can occur closely spaced along a single horizon, with massive sandstone occupying the gaps between slabs; lamination can be correlated from slab to slab in some cases.

Two kinds of mud-draped scour occur in and on top of facies 3 units: (i) undulatory scours (Fig. 7), identical to those described above, except that the mud drape may be just a string of disconnected mudstone slabs; and (ii) concave-up scours, 1–30 m across and up to 1 m deep (Owen, 1934, plate 42A; McKeown *et al.*, 1973, plate 5B). The latter have gently curved walls which merge smoothly into a flat or gently curved floor. The three-dimensional geometry of concave-up scours is unknown; they could be pan-shaped (cf. Goldring & Aigner, 1982), or channel-shaped. Some scour surfaces, of both the undulatory and concave-up types, show symmetrical ripples.

Some facies 3 units, traced laterally, split into two subunits, separated by a mud layer (of Facies 2) up to 0·2 m thick (Fig. 8). The mud layer is cut out laterally by the overlying sand bed due to scouring (Fig. 8). Such scours, up to 0·2 m deep, are the only type of channel observed in the Bude Formation. At least one facies 3 unit splits apart laterally into many subunits (Fig. 9); remarkably, the mud layers are all truncated at about the same place, defining a 'pinch-out belt' only 3–5 m wide (Fig. 9). Facies 3 units seldom extend more than 5 m laterally without splitting or merging. On a regional scale, Freshney & Taylor (1972, p. 464)

Fig. 8. An amalgamated-sandstone unit (Facies 3) splitting laterally into two subunits separated by a truncated mudstone layer. Top to the right. Note that the underlying sandstone bed has a parallel-laminated upper (P) division, and is warped due to differential compaction. Coin is 3·0 cm across.

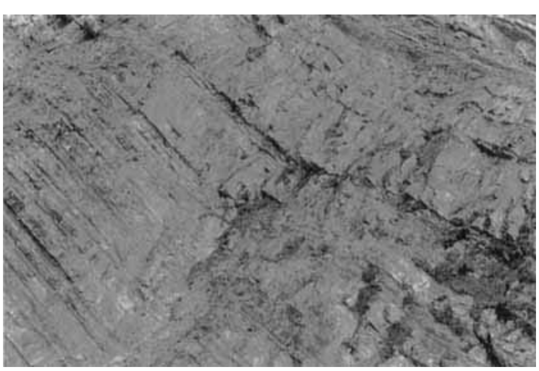

Fig. 9. Cliff face, viewed looking up from beach, showing a thick amalgamated-sandstone unit (centre) splitting laterally (downward) into numerous subunits separated by mudstone layers. The sandstone unit is about 3 m thick. Field of view is the interval 44–74 m in Figs 10 (Upton column) & 13.

found that the 'thick sandstones were normally too impersistent for correlative purposes.' Thick sandstones in the TCS–SPS interval extend no more than a few kilometres in a N–S direction, and terminate laterally by splitting (Fig. 10).

Interpretation

Facies 3 consists of amalgamated event beds, as shown by: (i) the internal planar partings and discontinuous mud layers, whose vertical spacing suggests that individual event beds are up to 0·4 m thick; and (ii) the similarity between sole- and top markings in Facies 3 and those in event beds in Facies 1 and 2. The overall massive character reflects the dominance of the B division in the constituent event beds (cf. Fig. 5).

The mudstone slabs are interpreted as *in-situ* remnants of formerly continuous mud layers, based on their lateral correlatability. Their angular, tabular morphology suggests that they formed by brecciation of the parent mud layer. Brecciation presumably took place at some depth below the sediment–water interface, where the mud was suitably brittle. The triggering disturbance was possibly an earthquake, or rapid deposition of an event bed.

Each truncated mudstone layer, or 'split', is thought to be a formerly continuous mud layer that was locally eroded during the next sand-depositing event. The sand, in effect, occupies and overspills a shallow (cm) channel. Channel width is unknown: in some cases, the channel 'fill' can be traced 20–30 m laterally, as far as the cliff top. The superimposition of successive splits (Fig. 9) suggests that successive sand-depositing flows followed the same track.

Planar partings are interpreted as sutures between event beds. Any mud which formerly intervened was eroded. Erosion also removed the structured cap of the lower event bed in some cases, but not in others (Fig. 8). Undulatory partings possibly reflect *in-situ* disturbance of planar partings by an earthquake (see facies 4 below for further evidence of earthquakes).

The undulatory mud-draped scours are probably due to wave action, as mentioned above. The concave-up scours may also be formed by waves: a possible modern analogue occurs on the Bering shelf, where elliptical, flat-bottomed scours, 10–30 m in diameter and 0·6–0·8 m deep, are incised into very fine sand and coarse silt, and are attributed to storms by Larsen *et al.* (1979). The presence of symmetrical ripples on (some) mud-draped scours supports a wave origin.

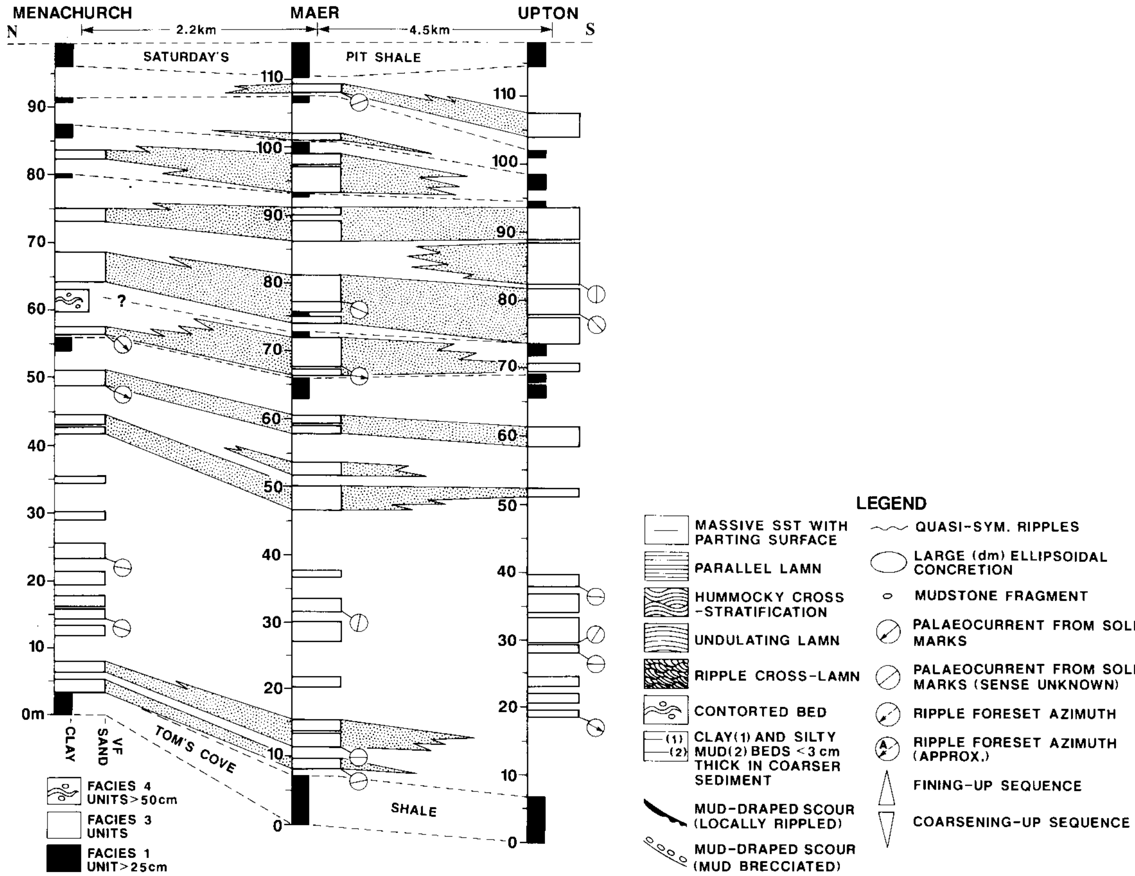

Fig. 10. Three correlated cliff sections of the Tom's Cove Shale – Saturday's Pit Shale interval. The Maer section is 0·5 km north of Crooklets Beach, Bude. For simplicity, the log omits: Facies 1 units thinner than 0·25 m; all facies 2 units; and facies 4 units thinner than 0·5 m. The Upton section is given at larger scale in Fig. 13. Dashed lines represent correlations between facies 1 units. Distances between sections are uncorrected for tectonic shortening, estimated as 40% (Goldring & Seilacher, 1971).

Discussion: sand-body geometry

Individual facies 3 sand bodies in the TCS–SPS interval 'shale-out' to the north and south (Fig. 10). This suggests that sediment transport was from the east or west, which is consistent with palaeocurrent data (below). The sand bodies are thought to be E–W-orientated tongues, comparable to the shelf 'sand lobes' of Brenchley & Newall (1982, Fig. 10). The tongues have planar upper and lower surfaces, and 'frayed' lateral margins (cf. Fig. 9). They are up to a few kilometres wide (N–S), and of unknown length. Internally, tongues consist of amalgamated sand beds; each sand bed is inferred to occupy (and overspill) a shallow channel, and to pinch out laterally due to depositional thinning. Tongues may have connected landward, to the east or west, with deltas from which the sand-depositing flows originated.

Facies 4: disturbed beds

Description

Two kinds of disturbed bed are recognized: (i) *mixed beds*, each comprising a non-laminated, homogeneous or quasi-homogeneous bed of mudstone, muddy siltstone or very fine muddy sandstone, with or without dark mudstone fragments (Fig. 11a, b, c); and (ii) *contorted beds*, differing only in that they are thicker and contain one or more discontinuous, folded sand layers, 0·01–0·1 m thick (Fig. 11d). Mixed beds and contorted beds are synonymous with the 'slurried beds' and 'slump beds', respectively, of Burne (1970); the latter terms are avoided here because of their genetic connotations.

Mixed beds are tabular and 0·01–0·4 m thick. They commonly contain one or more sand laminae, up to 0·02 m thick, with diffuse contacts, and commonly with millimetre-size load balls (Fig. 11a, c). Many mixed beds are bipartite, consisting of a layer of silt or sand, grading up into a darker, muddier layer (Fig. 11b, c). Contorted beds are thicker (up to 20 m). Internal folds have recumbent to inclined axial planes (Fig. 11d), with preferential E–W orientation and southward vergence (Enfield *et al.*, 1985).

Some disturbed beds have a sharp base, commonly with tool or scour marks, while others grade downward into undisturbed sand or mud of facies 2. Bed tops are sharp and generally planar, although some show symmetrical sand volcanoes (Burne, 1970).

Most disturbed beds contain angular to subrounded mudstone fragments, 0·01–0·05 m thick and up to 0·5 m long (Fig. 11a). Mudstone fragments are tabular, orientated subparallel to bedding, and commonly exhibit diffuse margins (Fig. 11a). Disturbed beds generally show dewatering sheets and pipes (Fig. 11a, b).

Disturbed beds extend at least 20–30 m laterally, to the cliff top. One contorted bed was correlated for 3·5 km along the coast by King (1966, and in Freshney *et al.*, 1972). At some localities, a disturbed bed grades laterally, over a distance of 1–5 m, into an undisturbed

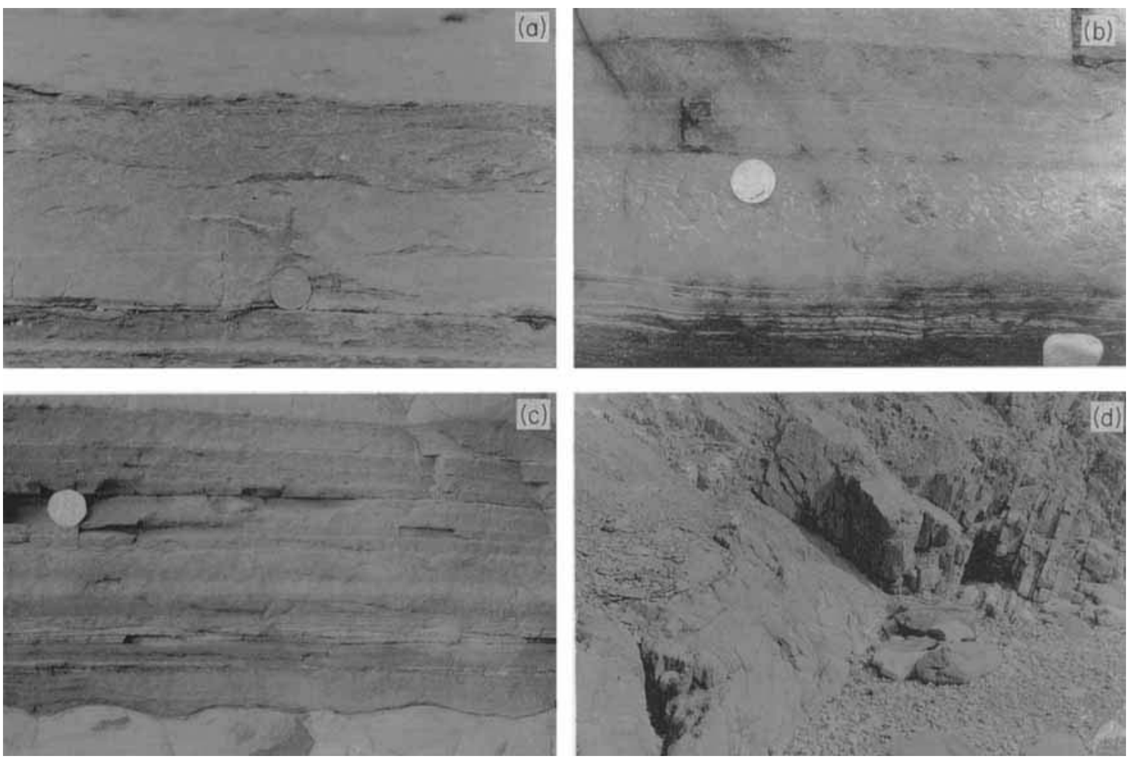

Fig. 11. (a) A mixed bed (Facies 4), extending from base of the coin to the abrupt colour change near (receding) top of view. The bed consists of clayey siltstone with dark mudstone fragments. Coin is 2·6 cm across. Note the laterally persistent sand lamina about 1 cm above coin. (b) Mixed bed consisting of massive very fine sandstone, grading up into a few millimetres of massive mudstone (immediately above coin). Coin is 2·8 cm across. The bed is sharply overlain by 5 mm of Facies 2 laminated mudstone, and sharply underlain by facies 2 mudstone. Note abundant dewatering structures. (c) Bipartite mixed bed encased in facies 2 mudstone. Coin is 3·0 cm across. The bed is 4 cm thick, and its base is 6 cm below base of coin. The bed consists of massive pale siltstone, grading up abruptly into darker mudstone, capped by a siltstone layer with irregular load balls. (d) A contorted bed comprising folded sand layers in a matrix of silty mudstone to clayey siltstone. Backpack at base of scree (upper left) marks top of bed. Note that stratification in the basal 1–2 m of the bed (immediately left of the small, dark cave) changes laterally (toward viewer) from uncontorted to contorted and diffuse, suggesting *in-situ* disturbance. Interval 34–42 m in Fig. 14(b).

interval of facies 2 mudstone with or without sandstone interbeds (Fig. 11d; see also Melvin, 1986, p. 25).

Interpretation

The fact that disturbed beds can grade laterally and downward into undisturbed facies 2 sediment suggests that they formed *in situ* from a facies 2 precursor. Disturbed beds were presumably formed at the sediment–water interface, because: (i) any overburden would have inevitably collapsed; and (ii) some beds have sand volcanoes. The substantial thickness (up to 20 m) of individual disturbed beds implies that the precursor mud was weakly cohesive to depths of many metres. The deformation was probably caused by a shock, induced by one of the following: storm waves; a tsunami; or an earthquake. Storm waves can probably be discounted because the tops of disturbed beds show no evidence for wave reworking. An earthquake is a more likely mechanism than a tsunami because it induces repeated (cyclical) shaking, which is more favourable for liquefaction than is a single impulse (Seed, 1968). Hence, facies 4 beds are interpreted as earthquake beds, or 'seismites' (Seilacher, 1969, 1984).

The inferred seismic shock caused: (i) homogenization of lamination in the precursor mud, possibly due to liquefaction of the coarser (silt) laminae; (ii) collapse and folding of thick (cm) sand interlayers, if any; (iii) development of load balls beneath thin sand layers; and (iv) brecciation of cohesive mud interlayers. Water expelled during and after deformation gave rise to dewatering sheets and sand volcanoes. Bipartite mixed beds are thought to reflect simultaneous disturbance of a sand (or silt) event bed and an overlying mud layer; disturbance did not necessarily extend to the bottom of the sand bed, whose lack of structure may instead by primary. In disturbed beds with sole marks, the latter are probably 'inherited' from a precursor event bed which now forms the base of the disturbed bed.

A possible prerequisite for the development of disturbed beds is weak cohesion in the parent mud, due to deposition in fresh water (van Olphen, 1977). Another likely contributor is the presence of biogenic gas (methane) bubbles, which form at much shallower depths (cm) in fresh-water sediments than in marine sediments (Curtis, 1977).

Similar contorted beds have been attributed to *in-situ* disturbance by earthquakes in marine and fresh-water deposits ranging from Precambrian to Quaternary (Selley *et al.*, 1963; Sims, 1975; Brenchley & Newall, 1977; Hempton & Dewey, 1983; Mayall, 1983; Brodzikowski *et al.*, 1987; Davenport & Ringrose, 1987). Some of these examples show folded sand layers in mud, as in the Bude Formation, but none of the cited publications mentions homogenization of mud lamination. In contrast, Seilacher (1969) invokes such homogenization for some Miocene seismites: these beds have an upper 'soupy zone' in which 'a kind of liquifaction' has destroyed the primary parallel lamination (Seilacher, 1969, p. 157).

An earthquake origin for facies 4 is consistent with a foreland basin setting. Contorted beds increase in abundance toward the top of the Bude Formation (e.g. Freshney *et al.*, 1979, Fig. 3), perhaps reflecting the approach of the orogenic front.

PALAEOCURRENTS

Sole marks in the studied intervals of the Bude Formation show a polymodal palaeocurrent pattern, with flow from all quadrants except the south (Fig. 12b). The dominant flow direction was toward the ESE (Fig. 12b). For discrete stratigraphic intervals, between successive key shales, the palaeocurrent pattern varies from unimodal to polymodal, and the dominant flow direction ranges from SE to SW (Freshney *et al.*, 1979, Fig. 6; Melvin, 1986, Fig. 10).

Ripple crestlines show less directional dispersion than sole marks (compare Fig. 12c and 12b); most crestlines are orientated within 30° of ENE–WSW. This clustering supports the conclusion that the ripples are wave-influenced types (above), and suggests that the direction of storm-wave approach varied little throughout deposition of the Bude Formation. The steeper flanks of the (combined-flow) ripples face south, indicating that the unidirectional component of the combined flow invariably approached the ripples from their northern side.

FACIES SEQUENCES AND CYCLICITY

The Bude Formation shows a pronounced cyclicity, in the form of sandstone units 1–10 m thick (Facies 3) alternating with mudstone-rich intervals of similar thickness (Figs 7, 13 & 14). To define cycles objectively, a facies relationship diagram was constructed (Fig. 15a; de Raaf *et al.*, 1965). The diagram reveals that Facies 1 and 3 are seldom in contact (i.e. Facies 2 usually intervenes), and that Facies 2 is overlain as

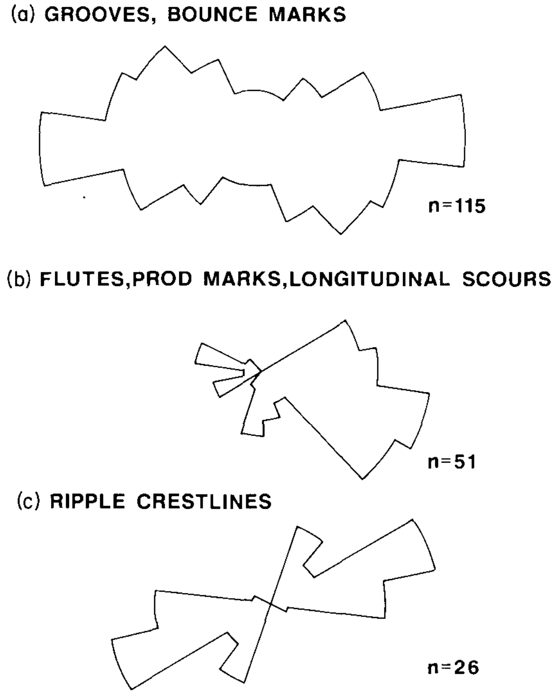

Fig. 12. Rose diagrams of various palaeocurrent indicators measured in this study.

often by facies 1 as by facies 3. These results could arise from either of two types of facies succession (Fig. 15b): (i) a regular succession of symmetrical, three-facies cycles (12321); or (ii) an intermixing of 121- and 232-type cycles. Inspection of the logs (Figs 13 & 14) shows the second model to be correct, except that some 12321 cycles are present locally (see below).

Ideal cycle

The 'ideal' (most complete) Bude Formation cycle is a symmetrical 12321 cycle (Figs 16 & 17, and 68–73 m in Fig. 13). The cycle has a coarsening-up component (123), and a fining-up component (321), not necessarily of the same thickness. The facies 1 interval at the top and bottom of the cycle is up to 1 m thick and contains thin (mm–cm) event beds. Stratigraphically adjacent to these facies 1 units is a facies 2 interval, up to 2 m thick, with thicker (up to 0·3 m) event beds. At the centre of the cycle is a facies 3 sandstone unit 1–10 m thick. Event beds with wave-influenced sedimentary structures (quasi-symmetrical ripples, HCS, mud-filled scours or multi-directional tool marks) can occur at any level in the cycle (Fig. 16).

A notable feature of the ideal cycle is its non-gradational character: a thick central (amalgamated) sandstone unit is flanked by much thinner sandstone layers (Fig. 16). This may reflect 'advanced amalgamation', whereby mud interlayers are missing (eroded), possibly due to the weak cohesion (high erodibility) of fresh-water muds. The result is that cycles are vertically 'compressed' (Fig. 18b) and, as documented by Melvin (1986), lack well-developed upward-thickening or -thinning sequences.

Departures from the ideal cycle

The cycles are rarely as complete or symmetrical as the ideal cycle. Most cycles are of 121- or 232-type. In some cases, facies 1 is directly overlain by facies 3, possibly due to complete erosion of a (weakly cohesive) facies 2 mud layer.

Interpretation of the cyclicity

The average cycle period is 7700 years, assuming that the Bude Formation spans 5 Ma (see above), is 1300 m thick (Freshney & Taylor, 1972), and that the average cycle thickness is 2 m (e.g. Fig. 14). However, this figure is a gross approximation: the true cycle period could be substantially longer if, for example, the thickness of the Bude Formation has been overestimated due to repetition by thrusting, such as that described by Enfield *et al.* (1985).

Any explanation for the cyclicity must include a mechanism for producing salinity fluctuations, because facies 1 was deposited in brackish (sometimes marine) water, while facies 2 (and 3?) was deposited in fresh water. Increases in salinity were probably caused by marine incursions, rather than by arid evaporative episodes, because (i) some facies 1 intervals are marine; and (ii) Lake Bude was in an equatorial (i.e. humid) setting. The salinity fluctuations cannot be due to deposition in front of a switching delta complex (e.g. Elliott, 1986), because this causes salinity changes in the *surface* water only, not the bottom water (e.g. Bates, 1953). The most likely explanation is that the lake sill, or outlet level, was intermittently overtopped by the sea. Each of the resulting marine incursions caused the lake to become brackish or (rarely) marine. Repeated drowning of the sill could have been caused by eustatic sea-level oscillations, or by intermittent bursts of accelerated tectonic subsidence. The eustatic mechanism is preferred, for three reasons: (i) the Late Carboniferous was a time of repeated glaciations (Frakes, 1979), so

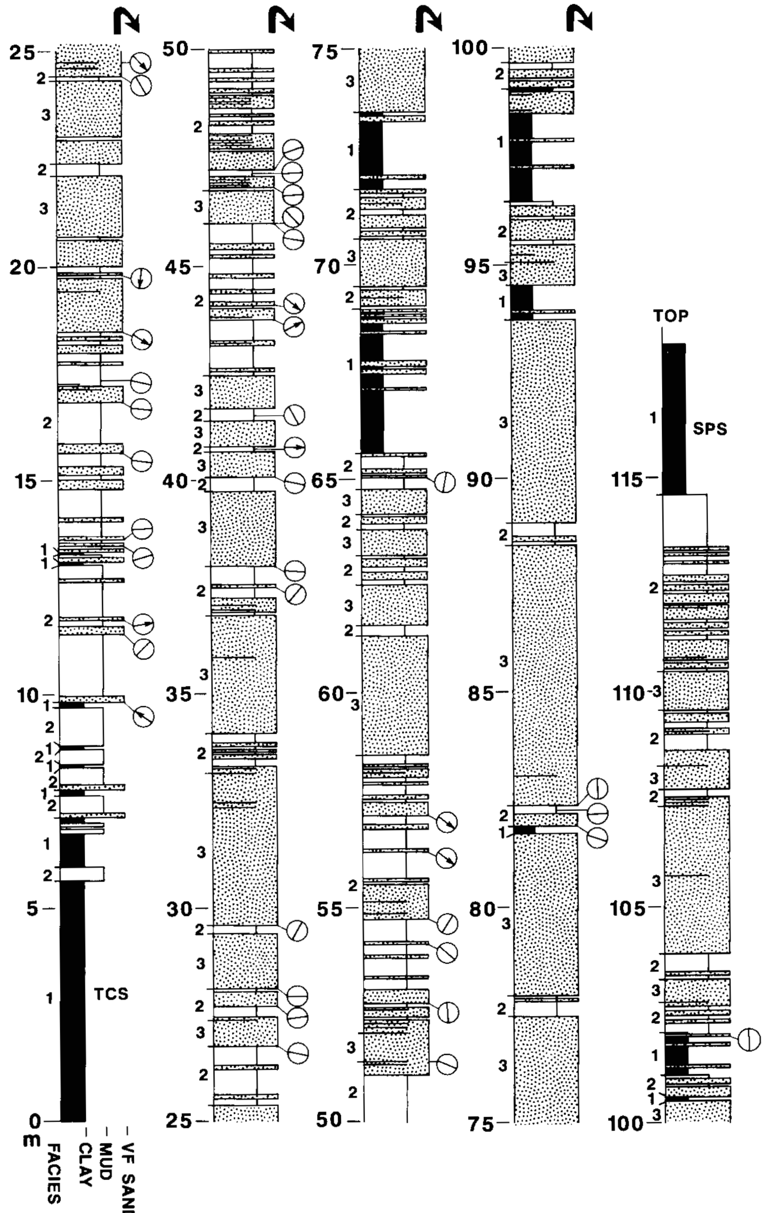

Fig. 13. Measured section between the Tom's Cove Shale (TCS) and Saturday's Pit Shale (SPS) at Upton, 1·5 km south of Summerleaze Beach, Bude. See Fig. 10 for legend. Base of section is at SS 2000 0501. For simplicity, sand/silt beds thinner than 3 cm are omitted, as are ripples. Also, facies 4 is included in facies 2; this is justifiable because the former is formed *in situ* from a facies 2 precursor (see text).

glacio-eustatic sea-level oscillations are to be expected; (ii) glacio-eustatic oscillations of similar periodicity (1000–10 000 years) to the possible 7700-year Bude cycle (see above) are known from the Pleistocene (e.g. Holmes, 1965, Fig. 522; Matthews, 1984, Fig. 1E; Lorius *et al.*, 1985, Fig. 4b); and (iii) some intervals consist of several consecutive, thicker-than-average cycles (e.g. Figs 10 & 13), and this clustering would be

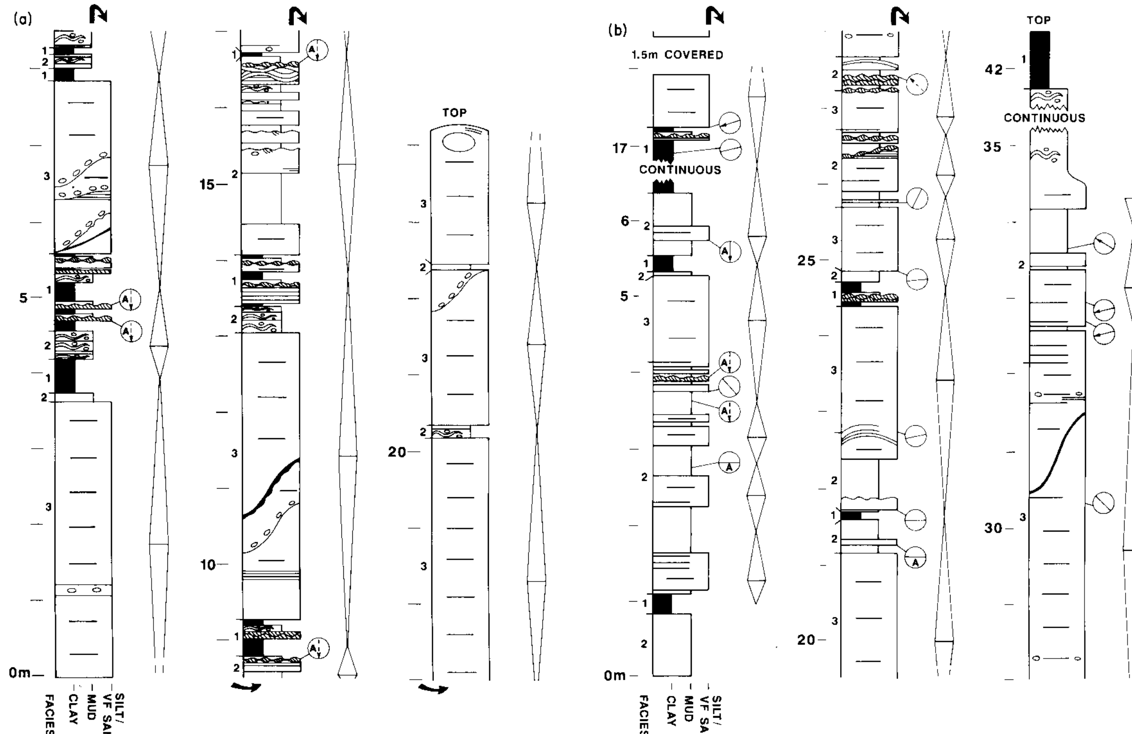

Fig. 14. Large-scale measured sections. See Fig. 10 for legend. For simplicity, sand/silt beds thinner than 3 cm are omitted, and facies 4 is included in facies 2. (a) Cliff section at northern end of Summerleaze Beach, Bude. Top of section is 20 m stratigraphically below Saturday's Pit Shale. (b) Cliff section at Warren Little Beach, 4 km north of Bude. Base of section is at SS 2009 1085. Section is between Sandy Mouth Shale and Warren Gutter Shale (see Fig. 2).

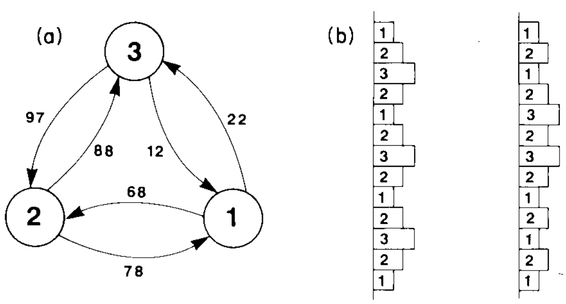

Fig. 15. (a) Facies relationship diagram, summarizing all (upward) facies transitions measured in this study. Numbers inside circles are facies numbers. Facies 4 is included in facies 2, since the former is formed *in situ* from a facies 2 precursor. (b) Two hypothetical facies successions consistent with the facies relationship diagram (see text).

difficult to explain tectonically. The cause of such short-period (7700 years) oscillations in global ice volume is unclear; they are possibly related to astronomical forcing of palaeoclimate, although the shortest period predicted by the Milankovich astronomical theory is about 20 000 years (e.g. Berger, 1980). Alternatively, if the thickness of the Bude Formation has been overestimated (above), then the Bude cycles could indeed represent the Milankovitch 20 000-year cycle.

Each marine incursion would have caused the lake to deepen, as the rising eustatic sea-level overtopped the sill. This deepening could explain why Facies 1, representing the incursions, has the finest grain size. Hence, the cycles are thought to reflect fluctuations in water depth as well as salinity (Figs 19 & 20), each cycle recording a regression (coarsening up) followed by a transgression. During each *transgression*, the rising sea eventually spilled over the sill, converting the lake outlet into a marine strait (Fig. 20b). If the sea-level rose high enough, a salt-water wedge in the strait introduced sea water into the lake, compensated by surface outflow of lake water. Normally, at such times, the lake became brackish (not marine), indicating that the sill was not drowned very deeply (compare

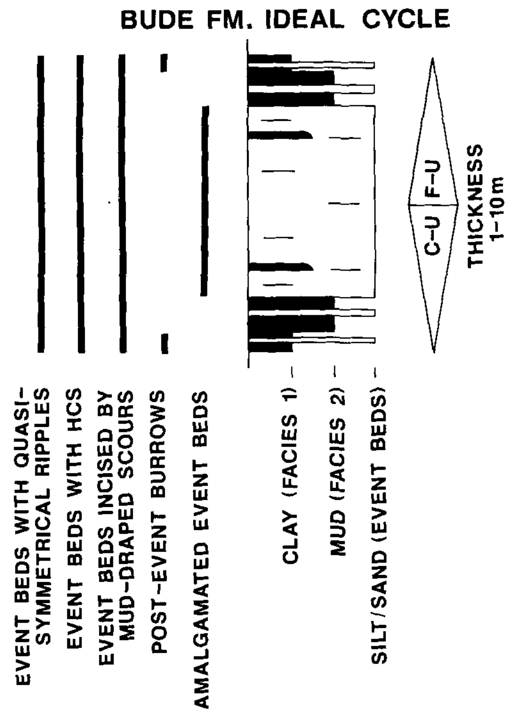

Fig. 16. Ideal Bude Formation cycle. HCS = hummocky cross-stratification; C-U = coarsening-up; F-U = fining-up; — = planar partings (bed sutures). Compare with Fig. 18.

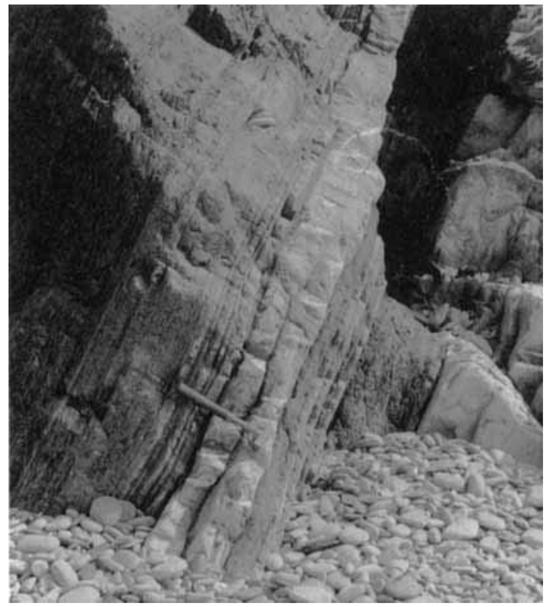

Fig. 17. Symmetrical cycle. Top to left. A central amalgamated sandstone unit passes up and down into pale mudstone with thin sandstone beds (Facies 2), followed by dark mudstone (Facies 1). Hammer is 31 cm long. Interval 5·3–6·9 m in Fig. 14(b).

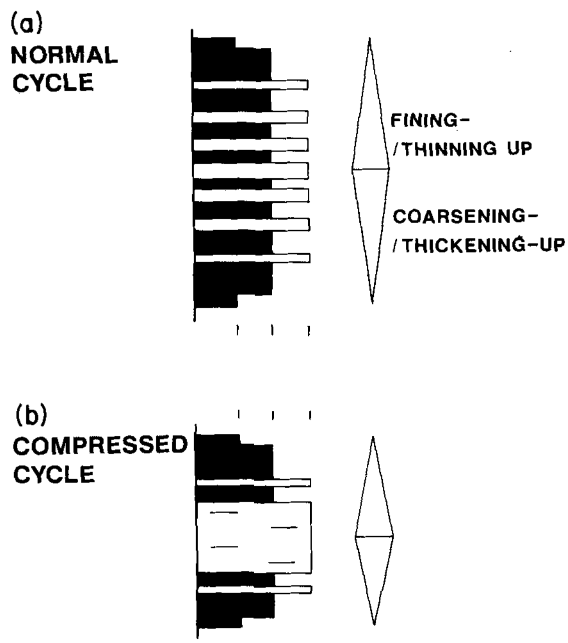

Fig. 18. (a) Hypothetical 'normal' cycle. (b) Typical Bude Formation 'compressed' cycle (see text). Horizontal dashes represent bed sutures. Compare with Fig. 16.

b and c in Fig. 20). A modern analogue is the Black Sea, whose sill depth is such that annual additions of sea water (flowing in through the Bosporus) and fresh water (runoff and precipitation) are about equal (Gunnerson & Özturgut, 1974; Scholten, 1974); mixing of these waters gives an overall salinity of about 20‰.

As eustatic sea-level continued to rise, the lake level necessarily rose in harmony, and the lake deepened. Occasionally, sea-level rose to such a height that the lake became fully marine (Fig. 20c), forming a land-locked sea.

Regressions were caused by a drop in eustatic sea-level, whereby the lake (and sill) shallowed. Eventually, sea-level fell below the crest of the sill, or at least to such a level that the salt wedge pinched out before entering the lake (Fig. 20a), as in modern Lake Maracaibo (Hyne *et al.*, 1979). With the marine connection severed, the lake freshened, probably due to displacement and mixing of brackish lake water with underflow fresh water introduced by turbid rivers (Higgs, 1986b), a process termed 'desalination' (Holdsworth & Collinson, 1988). Analogous freshening

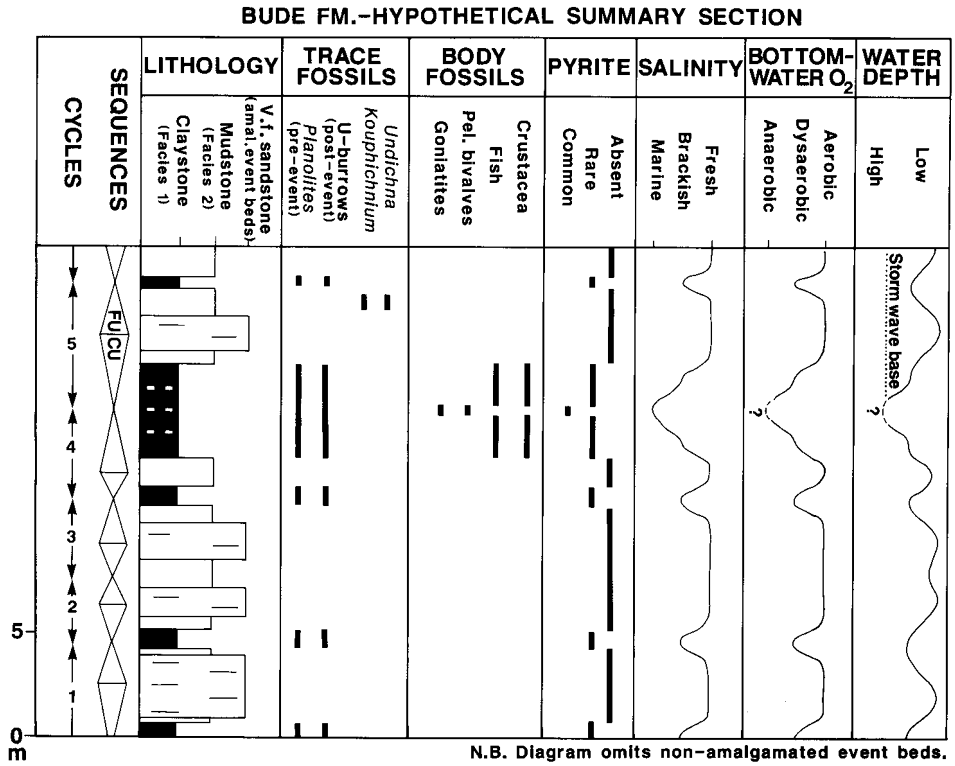

Fig. 19. Hypothetical section of the Bude Formation, showing inferred fluctuations in water depth, salinity and bottom-water oxygenation. The section included a goniatite-bearing 'key' shale, indicated as being deposited partly below storm wave base (see text; white squares = concretions). Omitted for clarity are non-amalgamated sandstone beds in Facies 1 and 2. CU = coarsening-up; FU = fining-up.

occurred in the Black Sea, which was almost fresh during the last glaciation, having been virtually marine during the preceding interglacial (Ross & Degens, 1974; Scholten, 1974).

In the case of 232-type cycles, fresh water may have persisted throughout the cycle, implying that sea-level failed to rise high enough for the salt wedge to reach the lake (Fig. 20a, inset). On the other hand, 121-type cycles lack a central facies 3 sand body (i.e. sand tongue), suggesting that they were deposited *between* sand tongues, or distally beyond them.

Larger cycles?

A higher order of cyclicity may be present, in addition to the small cycles described above. This is suggested by: (i) the occurrence of several thicker-than-normal (key) shale units; and (ii) the clustering of numerous thick sandstone units, each belonging to a small cycle, near the middle of the TCS–SPS interval, suggesting that this interval is a large symmetrical cycle (Figs 10 & 13), as suggested by King (1967). The TCS–SPS interval may represent a 'first-order' eustatic sea-level oscillation, upon which are superimposed many 'second-order' oscillations, responsible for the small cycles. The central cluster of thick sand units would correspond to the first-order lowstand, when the lake level would have remained low and constant (at sill level) for an extended period, punctuated only when exceptionally high *second*-order highstands overtopped the sill. Given that the TCS–SPS interval is about 100 m thick (Fig. 10), it represents about 380 000 years, assuming that the thickness and time-span of the Bude Formation are 1300 m and 5 Ma, respectively. Hence, it is possible that the TCS–SPS interval reflects the 400 000-year Milankovitch cycle (e.g. Berger, 1980).

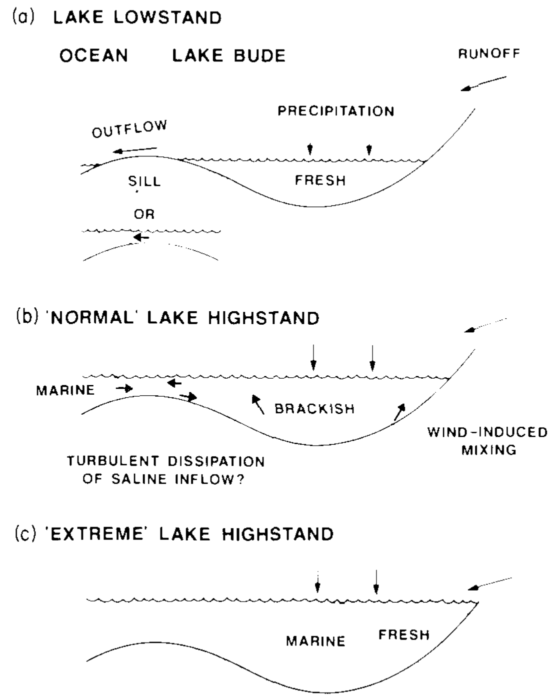

Fig. 20. Model showing how eustatic sea-level oscillations may have caused fluctuations in salinity and water depth in Lake Bude. Shown in (b) are inferred mixing processes, whereby marine water and fresh water entering the lake are mixed, forming brackish water. The diagrams depict fair-weather conditions and neglect the inferred shelf-and-trough physiography of Lake Bude. During marine episodes (i.e. c), Lake Bude was actually a land-locked sea rather than a lake. Similarly, during brackish episodes (b), the name 'lake' is not entirely appropriate (cf. modern Black Sea).

Lack of nearshore and emergent facies

The Bude Formation was deposited largely, if not entirely, above storm wave base, as indicated by the fact that wave-influenced structures occur throughout the succession, spaced no more than a few metres apart. It is curious that such a thick (1300 m) shallow-water succession is totally devoid of nearshore and emergent facies, implying that the shoreline never prograded into the Bude area. This is attributed to: (i) the lake sill preventing the lake water level from falling below sill level (cf. Fig. 20a); in this way the Bude depositional environment (assuming it was below sill level) was protected from eustatically induced exposure, unlike shallow-*marine* environments; and (ii) deposition of the Bude Formation in a relatively distal setting, beyond the reach of shoreline progradations.

DEPOSITIONAL PROCESSES AND ENVIRONMENT

Depositional environment

As explained earlier, the Bude Formation is lacustrine, based on evidence for subaqueous deposition combined with evidence that the basin water was fresh at times. Fresh-water conditions imply that the lake was hydrologically open (i.e. had an outlet; Allen & Collinson, 1986). Deposition took place: (i) below fair-weather wave base, as shown by the absence of nearshore features; and (ii) mostly or entirely above storm wave base, as shown by the presence of wave-influenced structures throughout the succession. The interpreted environment was a shelf, of continental-shelf dimensions, but lacustrine rather than marine (Fig. 21). The shelf may have been long-lived,

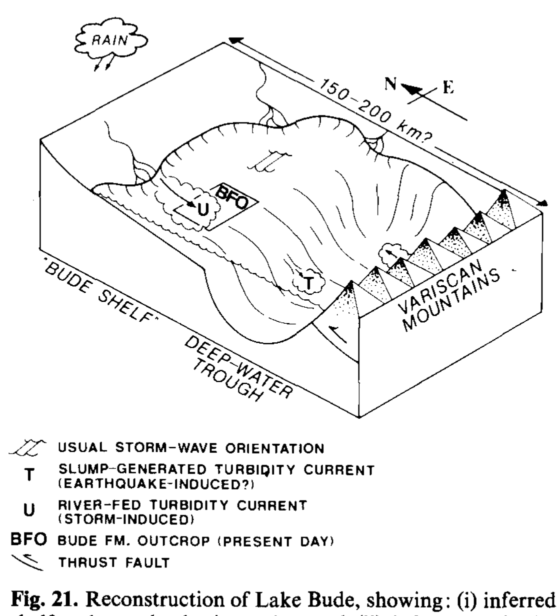

Fig. 21. Reconstruction of Lake Bude, showing: (i) inferred shelf-and-trough physiography; and (ii) inferred hydrodynamic processes on the 'Bude shelf' during storms. A river-fed underflow is shown issuing from one of the rivers discharging onto the shelf. Episodic turbidity currents, possibly generated by earthquake-induced slumping, are inferred in the deep-water trough.

inherited from Devonian time, because Devonian strata conformably below are neritic (see above). A shelf setting is consistent with *in-situ* seismites, whose failure to slide laterally suggests a very low depositional gradient, since subaqueous sliding and slumping can occur on slopes of only 0·5° (Coleman, 1981); the average gradient of modern shelves is 0·1° (Shepard, 1963). A shelf setting is also consistent with the polymodal palaeocurrents of the Bude Formation (Fig. 12), because underflows could have been deflected to flow *across* the (gentle) shelf gradient by wind-driven currents.

South of the shelf lay an inferred deep-water trough, formed by tectonic loading at the orogenic front (Fig. 21; see below). The trough explains the lack of northward-flowing palaeocurrents in the Bude Formation (Fig. 12). The trough also explains the 'problem' (Selwood & Thomas, 1988, p. 22) that despite an orogenic front nearby, the Bude Formation lacks wildflysch: in other words, any wildflysch would have been 'trapped' by the trough.

Velocity of river-fed turbidity currents

The event beds in the Bude Formation are interpreted as storm beds, deposited by river-fed turbidity currents induced by catastrophic storm floods (above). An insight into the velocity of river-fed turbidity currents on the 'Bude shelf' can be gained from modern analogues. River-fed turbidity currents, or underflows, are well known in Swiss lakes (Houbolt & Jonker, 1968; Lambert *et al.*, 1976; Sturm & Matter, 1978) and in Lake Meade, Nevada (Gould, 1960). However, none of these lakes is a good physiographical analogue of the Bude shelf; most are deep (100–350 m), steep-sided, flat-bottomed troughs.

In Lake Meade, underflows persist for months, and sometimes travel the entire 200 km length of the lake; the flows maintain a velocity of about 7 cm s^{-1}, even though the lake-floor gradient is less than 0·1° (Bell, 1942; Gould, 1960). An underflow peaking at more than 30 cm s^{-1}, lasting at least 2 days, was recorded on a 2° delta slope in the Walensee (Switzerland) during spring thaw, following a period of heavy rainfall (Lambert *et al.*, 1976). Simultaneous current measurements 1·5 km further offshore, on the flat central plain, yielded maximum velocities of 2–3 cm s^{-1}. Further measurements on the same delta slope 4 years later, following a summer rainstorm, yielded a maximum velocity of 50 cm s^{-1} (Lambert & Hsü, 1979). In the sea, river-fed underflows of up to 30 cm s^{-1} have been observed on the 0·6° delta slope of the Yellow River, China, at 4–11 m depth (Wright *et al.*, 1988).

Extrapolation of the foregoing velocity data to Lake Bude is tenuous because the Bude shelf was probably less steep than most of the underflow survey sites. However, to compensate for this factor, none of the modern surveys was made during a *catastrophic* storm. Hence, it seems reasonable to infer velocities of 30–40 cm s^{-1} for Bude shelf underflows. A 30-cm s^{-1} current would have been quite capable of transporting very fine sand in suspension (Sundborg, 1967, Fig. 1). The estimate of 30–40 cm s^{-1} is consistent with the presence of flutes in the Bude Formation: experimental flutes were made using a mean current velocity of 45 cm s^{-1} by Allen (1971).

DISCUSSION

Tectonic implications of the deep-water trough

The presence of a deep-water trough would imply that the foreland basin hosting Lake Bude was at an early, underfilled stage of development (cf. Allen *et al.*, 1986); i.e. the orogen had not yet fully emerged, so that sediment production lagged behind tectonic loading and resultant basin subsidence, giving rise to deep water (Covey, 1986). The trough fill may have comprised turbidites and olistostromes; such deposits crop out to the south between Dartmoor and Bodmin Moor (Fig. 1; Isaac *et al.*, 1982), but they are older (Dinantian and Namurian), and were probably deposited in a (marine) precursor of the 'Bude trough'. The missing Bude-age trough deposits may have been deposited on top of their Namurian counterparts and later eroded.

Ancestry of Lake Bude

The Bude Formation rests conformably on the turbiditic Crackington Formation, which is marine at the base, but contains brackish or fresh intervals higher up (Freshney & Taylor, 1972). This suggests that the Crackington and Bude formations were deposited in a water body which became progressively isolated from the open sea. Isolation may have been caused by the Variscan mountain front advancing (northward) across the areas east and west of SW England sooner than across SW England itself (Higgs, 1986b, c); this would have converted a formerly continuous E–W seaway into a progressively enclosed water body, flanked on each side by northward-propagating mountains. Sole marks in the Crackington Formation show mainly E–W and W–E transport directions (Thomas, 1988), consistent with mountains to the east and west. Ultimately, the water body became enclosed, forming a lake. The lake sill, which functioned as an outlet during lowstands and as a sea-water inlet during highstands (Fig. 20), presumably lay north of the flanking mountains. The sill probably lay somewhere along the NW margin of the lake, because the only connection between Britain and the open ocean in early Westphalian time lay to the west (Ramsbottom, 1970).

Modern analogue of Lake Bude

No modern lakes show the same combination of physiography, climate and tectonic setting inferred for Lake Bude. A good *physiographical* analogue is the Black Sea, which, like Lake Bude, has a broad (northwestern) shelf flanked by a deep trough. Another similarity is that the water level in the Black Sea has varied with Quaternary oscillations in eustatic sea-level, causing salinity to fluctuate from fresh to brackish (above). River-fed turbidites are possibly deposited on the Black Sea shelf during catastrophic storms; unfortunately, the author is unaware of any cores or relevant hydrographical data that confirm this supposition (cf. Ross *et al.*, 1978). The Black Sea further resembles Lake Bude in being hydrologically open and dominated by siliciclastic sediments (Ross *et al.*, 1978).

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REFERENCES

ALLEN, J.R.L. (1971) Transverse erosional marks of mud and rock: their physical basis and geologic significance. *Sediment. Geol.*, **5**, 167–385.

ALLEN, P.A. (1981) Some guidelines for reconstructing ancient sea conditions from wave ripplemarks. *Mar. Geol.*, **43**, M59–M67.

ALLEN, P.A. & COLLINSON, J.D. (1986) Lakes. In: *Sedimentary Environments and Facies* (Ed. by H. G. Reading), pp. 63–94. Blackwell Scientific Publications, Oxford.

ALLEN, P.A., HOMEWOOD, P. & WILLIAMS, G.D. (1986) Foreland basins: an introduction. In: *Foreland Basins* (Ed. by P. A. Allen & P. Homewood). *Spec. publs int. Ass. Sediment.*, **8**, 3–12.

ANDERSON, A. (1975) Limulid trackways in the Late Palaeozoic Ecca sediments and their palaeoenvironmental significance. *South Afr. J. Sci.*, **71**, 249–251 & 312.

ANDERTON, R., BRIDGES, P.H., LEEDER, M.R. & SELWOOD, B.W. (1979) *A Dynamic Stratigraphy of the British Isles.* George Allen & Unwin, London, 301 pp.

ASHWIN, D.P. (1957) *The structure and sedimentation of the Culm sediments between Boscastle and Bideford, North Devon.* PhD thesis, Imperial College, University of London.

BANDEL, K. (1967) Isopod and limulid marks and trails in Tonganoxie Sandstone (Upper Pennsylvanian) of Kansas. *Univ. Kansas paleont. Contrib.*, **19**, 1–10.

BATES, C.C. (1953) Rational theory of delta formation. *Bull. Am. Ass. petrol. Geol.*, **37**, 2119–2162.

BEACH, A. (1977) Vein arrays, hydraulic fractures and pressure-solution structures in a deformed flysch sequence, S.W. England. *Tectonophysics*, **40**, 201–225.

BEADLE, L.C. (1981) *The Inland Waters of Tropical Africa*, 2nd edn. Longman, London, 475 pp.

BELL, H.S. (1942) Density currents as agents for transporting sediment. *J. Geol.*, **50**, 512–547.

BERGER, A. (1980) The Milankovitch astronomical theory of paleoclimate: a modern review. *Vistas Astr.*, **24**, 103–122.

BERNER, R.A. & RAISWELL, R. (1984) C/S method for distinguishing freshwater from marine sedimentary rocks. *Geology*, **12**, 365–368.

BRENCHLEY, P.J. (1985) Storm influenced sandstone beds. *Mod. Geol.*, **9**, 369–396.

BRENCHLEY, P.J. & NEWALL, G. (1977) The significance of contorted bedding in Upper Ordovician sediments of the Oslo region, Norway. *J. sedim. Petrol.*, **47**, 819–833.

BRENCHLEY, P.J. & NEWALL, G. (1982) Storm-influenced inner-shelf sand lobes in the Caradoc (Ordovician) of Shropshire, England. *J. sedim. Petrol.*, **52**, 1257–1269.

BROADHURST, F.M., SIMPSON, I.M. & HARDY, P.G. (1980) Seasonal sedimentation in the Upper Carboniferous of England. *J. Geol.*, **88**, 639–651.

BRODZIKOWSKI, K., GOTOWALA, R., HALUSZCAK, A., KRZYSZKOWSKI, D. & VAN LOON, A.J. (1987) Soft-sediment deformations from glaciodeltaic, glaciolacustrine and fluviolacustrine sediments in the Kleszczów Graben (central Poland). In: *Deformation of Sediments and Sedimentary Rocks* (Ed. by M. E. Jones & R. M. F. Preston), *Spec. publs geol. Soc. London*, **29**, 255–267.

BROOKS, H.K. (1969) Eocarida. In: *Treatise on Invertebrate Paleontology, Part R, 1* (Ed. by R. C. Moore), pp. 332–345. Geological Society of America & University of Kansas.

BURNE, R.V. (1969) *Sedimentological studies of the Bude Formation.* PhD thesis, University of Oxford.

BURNE, R.V. (1970) The origin and significance of sand volcanoes in the Bude Formation (Cornwall). *Sedimentology*, **15**, 211–228.

BURNE, R.V. (1973) Palaeogeography of South West England and Hercynian continental collision. *Nature phys. Sci.*, **241**, 129–131.

CHISHOLM, J.I. (1970) *Teichichnus* and related trace fossils in the Lower Carboniferous of St Monace, Scotland. *Bull. geol. Surv. Gt Br.*, **32**, 21–51.

CHISHOLM, J.I. (1983) Xiphosurid traces, *Kouphichnium* aff. *variabilis* (Linck), from the Namurian Upper Haslingden Flags of Whitworth, Lancashire. *Rep. Inst. geol. Sci. London*, **83/10**, 37–44.

COLEMAN, J.M. (1981) *Deltas—Processes of Deposition and Models for Exploration*, 2nd edn. Burgess, Minneapolis, 124 pp.

CORNFORD, C., YARNELL, L. & MURCHISON, D.G. (1987) Initial vitrinite reflectance results from the Carboniferous of north Devon and north Cornwall. *Proc. Ussher Soc.*, **6**, 461–467.

COVEY, M. (1986) The evolution of foreland basins to steady state: evidence from the western Taiwan foreland basin. In: *Foreland Basins* (Ed. by P. A. Allen & P. Homewood), *Spec. publs int. Ass. Sediment.*, **8**, 77–90.

COWARD, M.P. & SMALLWOOD, S. (1984) An interpretation of the Variscan tectonics of SW Britain. In: *Variscan Tectonics of the North Atlantic Region* (Ed. by D. H. W. Hutton & D. J. Sanderson), *Spec. publs geol. Soc. London*, **14**, 89–102.

CRAFT, J.H. & BRIDGE, J.S. (1987) Shallow-marine sedimentary processes in the Late Devonian Catskill Sea, New York State. *Bull. geol. Soc. Am.*, **98**, 338–355.

CURTIS, C.D. (1977) Sedimentary geochemistry: environments and processes dominated by involvement of an aqueous phase. *Trans. R. Soc. London*, **A286**, 353–372.

DAVENPORT, C.A. & RINGROSE, P.S. (1987) Deformation of Scottish Quaternary sediment sequences by strong earthquake motions. In: *Deformation of Sediments and Sedimentary Rocks* (Ed. by M. E. Jones & R. M. F. Preston), *Spec. publs geol. Soc. London*, **29**, 299–314.

DONOVAN, R.N., ARCHER, R., TURNER, P. & TARLING, D.H. (1976) Devonian palaeogeography of the Orcadian Basin and the Great Glen Fault. *Nature*, **259**, 550–551.

EAGAR, R.M.C., BAINES, J.G., COLLINSON, J.D., HARDY, P.G., OKOLO, S.A. & POLLARD, J.E. (1985) Trace fossil assemblages and their occurrence in Silesian (mid-Carboniferous) deltaic sediments of the Central Pennine Basin, England. *Spec. publs Soc. econ. Paleont. Miner.*, **35**, 99–149.

EDMONDS, E.A., MCKEOWN, M.C. & WILLIAMS, M. (1975) *South-West England. British Regional Geology.* British Geological Survey, 138 pp.

EDMONDS, E.A., WILLIAMS, B.J. & TAYLOR, R.T. (1979) Geology of Bideford & Lundy Island, Sheets 292, with 275, 276, 291 and part of 308. *Mem. geol. Surv. Gt. Br.*, 143 pp.

ELLIOTT, T. (1986) Deltas. In: *Sedimentary Environments and Facies*, 2nd edn (Ed. by H. G. Reading), pp. 113–154. Blackwell Scientific Publications, Oxford.

ENFIELD, M.A., GILLCRIST, J.R., PALMER, S.N. & WHALLEY, J.S. (1985) Structural and sedimentary evidence for the early tectonic history of the Bude and Crackington Formations, north Cornwall and Devon. *Proc. Ussher Soc.*, **6**, 165–172.

FRAKES, L.A. (1978) *Climates Throughout Geologic Time.* Elsevier, Amsterdam, 310 pp.

FRESHNEY, E.C., EDMONDS, E.A., TAYLOR, R.T. & WILLIAMS, B.J. (1979) Geology of the country around Bude and Bradworthy, Sheets 307 & 308. *Mem. geol. Surv. Gt. Br.*, 62 pp.

FRESHNEY, E.C., MCKEOWN, M.C. & WILLIAMS, M. (1972) Geology of the coast between Tintagel and Bude, Sheet 322 (part). *Mem. geol. Surv. Gt. Br.*, 92 pp.

FRESHNEY, E.C. & TAYLOR, R.T. (1972) The Upper Carboniferous stratigraphy of north Cornwall and west Devon. *Proc. Ussher Soc.*, **2**, 464–471.

FREY, R.W. & PEMBERTON, S.G. (1984) Trace fossil facies models. In: *Facies Models*, 2nd edn (Ed. by R. G. Walker), pp. 189–207. Geoscience Canada Reprint Series 1, Geological Association of Canada, Toronto.

FÜRSICH, F.T. (1974) On *Diplocraterion* Torell 1870 and the significance of morphological features in vertical, spreiten-bearing U-shaped trace fossils. *J. Paleontol.*, **48**, 952–962.

GOLDRING, R. (1962) The trace fossils of the Baggy Beds (Upper Devonian) of North Devon, England. *Paläont. Z.*, **36**, 232–251.

GOLDRING, R. (1971) Shallow-water sedimentation as illustrated in the Upper Devonian Baggy Beds. *Mem. geol. Soc. London*, **5**, 80 pp.

GOLDRING, R. (1978) Sea level lake community. In: *The Ecology of Fossils* (Ed. by W. S. McKerrow), pp. 178–181. Duckworth, London.

GOLDRING, R. & AIGNER, T. (1982) Scour and fill: the significance of event separation. In: *Cyclic and Event Stratification* (Ed. by G. Einsele & A. Seilacher), pp. 354–362. Springer-Verlag, Berlin.

GOLDRING, R. & LANGENSTRASSEN, F. (1979) Open shelf and nearshore clastic facies in the Devonian. *Spec. pap. Palaeont. London*, **23**, 81–97.

GOLDRING, R. & SEILACHER, A. (1971) Limulid undertracks and their sedimentological implications. *Neues Jb. Geol. Paläont. Abh.*, **137**, 422–442.

GOULD, H.R. (1960) Turbidity currents. In: *Comprehensive Survey of Sedimentation in Lake Mead, 1948–49* (Ed. by W. O. Smith *et al.*), *Prof. pap. US geol. Surv.*, **295**, 201–207.

GRAY, D.I. & BENTON, M.J. (1982) Multidirectional palaeocurrents as indicators of shelf storm beds. In: *Cyclic and Event Stratification* (Ed. by G. Einsele & A. Seilacher), pp. 350–353. Springer-Verlag, Berlin.

GUNNERSON, C.G. & ÖZTURGUT, E. (1974) The Bosporus. In: *The Black Sea—Geology, Chemistry and Biology* (Ed. by E. T. Degens & D. A. Ross), *Mem. Am. Ass. Petrol, Geol.*, **20**, 99–114.

HARLAND, W.B., COX, A.V., LLEWELLYN, P.G., PICKTON, C.A.G., SMITH, A.G. & WALTERS, R. (1982) *A Geologic Time Scale.* Cambridge University Press, 131 pp.

HARMS, J.C. (1969) Hydraulic significance of some sand ripples. *Bull. geol. Soc. Am.*, **80**, 363–396.

HARMS, J.C., SOUTHARD, J.B. & WALKER, R.G. (1982) Structures and sequences in clastic rocks. *Soc. econ. Paleont. Miner. Short Course*, **9**.

HEMPTON, M.R. & DEWEY, J.F. (1983) Earthquake-induced deformational structures in young lacustrine sediments, East Anatolian Fault, southeast Turkey. *Tectonophysics*, **98**, T7–T14.

HIGGS, R. (1983) The possible influence of storms in the deposition of the Bude Formation (Westphalian), north Cornwall and north Devon. *Proc. Ussher Soc.*, **5**, 477–478.

HIGGS, R. (1984) Possible wave-influenced sedimentary structures in the Bude Formation (Lower Westphalian, south-west England), and their environmental implications. *Proc. Ussher Soc.*, **6**, 88–94.

HIGGS, R. (1986a) 'Lake Bude' (early Westphalian, SW England): storm-dominated siliciclastic shelf sedimentation in an equatorial lake. *Proc. Ussher Soc.*, **6**, 417–418.

HIGGS, R. (1986b) *A Facies Analysis of the Bude Formation (Lower Westphalian), SW England.* PhD thesis, University of Oxford.

HIGGS, R. (1986c) The 'South British Fault': 200 km of Stephanian–early Permian dextral wrenching on an E–W fault through the Bristol Channel. In: *Controls of Upper Carboniferous Sedimentation in North West Europe* (Ed. by B. M. Besly *et al.*), pp. 19–20.

HIGGS, R. (1987) The fan that never was?—Discussion of 'Upper Carboniferous fine-grained turbiditic sandstones from southwest England: a model for growth in an ancient, delta-fed subsea fan', by J. Melvin. *J. sedim. Petrol.*, **57**, 378–379.

HIGGS, R. (1988) Fish trails in the Upper Carboniferous of south-west England. *Palaeontology*, **31**, 255–272.

HIGGS, R. (1990) Is there evidence for geostrophic currents preserved in the sedimentary record of inner to middle shelf deposits?—Discussion. *J. sedim. Petrol.*, **60**, 630–632.

HIGGS, R., READING, H. & XU, L. (1990) *Upper Carboniferous Lacustrine and Deltaic Sedimentology, SW England.* Guidebook, Field Trip A-10, 13th International Sedimentological Congress, University of Nottingham, UK, August 1990.

HOBSON, D.M. & SANDERSON, D.J. (1983) Variscan deformation in southwest England. In: *The Variscan Fold Belt in the British Isles* (Ed. by P. L. Hancock), pp. 108–129. Hilger, Bristol.

HOLDSWORTH, B.K. & COLLINSON, J.D. (1988) Millstone Grit cyclicity revisited. In: *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe* (Ed. by B. M. Besly & G. Kelling), pp. 132–152. Blackie, Glasgow.

HOLMES, A. (1965) *Principles of Physical Geology*, 2nd edn. Thomas Nelson, London, 1288 pp.

HOUBOLT, J.J.H.C. & JONKER, J.B.M. (1968) Recent sediments in the eastern part of the Lake of Geneva (Lac Léman). *Geologie Mijnb.*, **47**, 131–148.

HYNE, N.J., COOPER, W.A. & DICKEY, P.A. (1979) Stratigraphy of intermontane, lacustrine delta, Catatumbo River, Lake Maracaibo, Venezuela. *Bull. Am. Ass. petrol. Geol.*, **63**, 2042–2057.

ISAAC, K.P., TURNER, P.J. & STEWART, I.J. (1982) The evolution of the Hercynides of central SW England. *J. geol. Soc. London*, **139**, 521–531.

JORDAN, D.W. (1985) Trace fossils and depositional environments of Upper Devonian black shales, east-central Kentucky, U.S.A. *Spec. publs Soc. econ. Paleont. Miner.*, **35**, 279–298.

KING, A.F. (1966) Structure and stratigraphy of the Upper Carboniferous Bude Sandstones, north Cornwall. *Proc. Ussher Soc.*, **1**, 229–232.

KING, A.F. (1967) *Stratigraphy and structure of the Upper Carboniferous Bude Formation, north Cornwall.* PhD thesis, University of Reading.

LAMBERT, A.M. & HSÜ, K.J. (1979) Non-annual cycles of varve-like sedimentation in Walensee, Switzerland. *Sedimentology*, **26**, 453–461.

LAMBERT, A.M., KELTS, K.R. & MARSHALL, N.F. (1976) Measurement of density underflows from Walensee, Switzerland. *Sedimentology*, **23**, 87–105.

LAMING, D.J.C. (1982) The New Red Sandstone. In: *The Geology of Devon* (Ed. by E. M. Durrance & D. J. C. Laming), pp. 148–178. University of Exeter.

LARSEN, M.C., NELSON, H. & THOR, D.R. (1979) Geologic implications and potential hazards of scour depressions on Bering Shelf, Alaska. *Environ. Geol.*, **3**, 39–47.

LECKIE, D. (1988) Wave-formed, coarse-grained ripples and their relationship to hummocky cross-stratification. *J. sedim. Petrol.*, **58**, 607–622.

LORIUS, C., JOUZEL, J., RITZ, C., MERLIVAT, L., BARKOV, N.I., KOROTKEVICH, Y.S. & KOTLYAKOV, V.M. (1985) A 150,000-year climatic record from Antarctic ice. *Nature*, **316**, 591–596.

LOVELL, J.P.B. (1965) The Bude Sandstones from Bude to Widemouth. *Proc. Ussher Soc.*, **1**, 172–174.

MATTHEWS, R.K. (1984) Oxygen isotope record of ice-volume history: 100 million years of glacio-eustatic sea-level fluctuation. In: *Interregional Unconformities and Hydrocarbon Accumulation* (Ed. by J. S. Schlee), *Mem. Am. Ass. petrol. Geol.*, **36**, 97–107.

MAYALL, M.J. (1983) An earthquake origin for synsedimentary deformation in a late Triassic (Rhaetian) lagoonal sequence, southwest Britain. *Geol. Mag.*, **120**, 613–622.

MCKEOWN, M.C., EDMONDS, E.A., WILLIAMS, M., FRESHNEY, E.C. & MASSON SMITH, D.J. (1973) Geology of the country around Boscastle and Holsworthy, Sheets 322 & 323. *Mem. geol. Surv. Gt. Br.*, 148 pp.

MELVIN, J. (1976) *Sedimentological studies in Upper Palaeozoic Sandstones near Bude, Cornwall and Walls, Shetland.* PhD thesis, University of Edinburgh.

MELVIN, J. (1986) Upper Carboniferous fine-grained turbiditic sandstones from southwest England: a model for growth in an ancient, delta-fed subsea fan. *J. sedim. Petrol.*, **56**, 19–34.

MELVIN, J. (1987) Upper Carboniferous fine-grained turbiditic sandstones from southwest England: a model for growth in an ancient, delta-fed subsea fan—Reply to discussion by R. Higgs. *J. sedim. Petrol.*, **57**, 380–382.

MIDDLETON, G.V. & HAMPTON, M.A. (1976) Subaqueous sediment transport and deposition by sediment gravity flows. In: *Marine Sediment Transport and Environmental Management* (Ed. by D. J. Stanley & D. J. P. Swift), pp. 197–218. Wiley, New York.

MILES, R.S. (1971) *Palaeozoic Fishes*, 2nd edn (after J. A. Moy-Thomas). Chapman & Hall, London, 259 pp.

MILLER, M.F. (1984) Distribution of biogenic structures in Paleozoic nonmarine and marine-margin sequences: an actualistic model. *J. Paleontol.*, **58**, 550–570.

VAN OLPHEN, H. (1977) *An Introduction to Clay Colloid Chemistry*. Wiley, New York, 318 pp.

OWEN, D.E. (1934) The Carboniferous rocks of the north Cornish coast and their structures. *Proc. geol. Ass. London*, **45**, 451–471.

OWEN, D.E. (1950) Carboniferous deposits in Cornubia. *Trans. R. geol. Soc. Cornwall*, **18**, 65–104.

PALMER, S.N. (1984) *The stratigraphy and structure of the area north of Bude, Cornwall*. BSc thesis, Portsmouth Polytechnic.

DE RAAF, J.F.M., BOERSMA, J.R. & VAN GELDER, A. (1977) Wave-generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. *Sedimentology*, **24**, 451–483.

DE RAAF, J.F.M., READING, H.G. & WALKER, R.G. (1965) Cyclic sedimentation in the Lower Westphalian of north Devon, England. *Sedimentology*, **4**, 1–52.

RAMSBOTTOM, W.H.C. (1970) Carboniferous faunas and palaeogeography of the south west England region. *Proc. Ussher Soc.*, **2**, 144–157.

RAMSBOTTOM, W.H.C., CALVER, M.A., EAGAR, R.M.C., HODSON, F., HOLLIDAY, D.W., STUBBLEFIELD, C.J. & WILSON, R.B. (1978) A correlation of Silesian rocks in the British Isles. *Spec. Rep. geol. Soc. London*, **10**, 81 pp.

READING, H.G. (1963) A sedimentological comparison of the Bude Sandstones with the Northam and Abbotsham Beds of Westward Ho! *Proc. Ussher Soc.*, **1**, 67–69.

READING, H.G. (1965) Recent finds in the Upper Carboniferous of south-west England and their significance. *Nature*, **208**, 745–747.

REINECK, H.E. & SINGH, I.B. (1980) *Depositional Sedimentary Environments*, 2nd edn. Springer-Verlag, Berlin, 549 pp.

RHOADS, D.C. & MORSE, J.W. (1971) Evolutionary and ecologic significance of oxygen-deficient marine basins. *Lethaia*, **4**, 413–428.

ROSS, D.A. & DEGENS, E.T. (1974) Recent sediments of the Black Sea. In: *The Black Sea—Geology, Chemistry and Biology* (Ed. by E. T. Degens & D. A. Ross), *Mem. Am. Ass. petrol. Geol.*, **20**, 183–199.

ROSS, D.A., STOFFERS, P. & TRIMONIS, E.S. (1978) Black Sea sedimentary framework. In: *Initial Reports of the Deep Sea Drilling Project*, 42(2) (Ed. by D. A. Ross, Y. P. Neprochnov *et al.*), pp. 359–373. US Government Printing Office, Washington.

SCHOLTEN, R. (1974) Role of the Bosporus in Black Sea chemistry and sedimentation. In: *The Black Sea—Geology, Chemistry and Biology* (Ed. by E. T. Degens & D. A. Ross), *Mem. Am. Ass. petrol. Geol.*, **20**, 115–126.

SCOTESE, C.R., BAMBACH, R.K., BARTON, C., VAN DER VOO, R. & ZIEGLER, A.M. (1979) Paleozoic base maps. *J. Geol.*, **87**, 217–277.

SEED, H.B. (1968) Landslides during earthquakes due to soil liquifaction. *J. Soil Mech. Fdns. Div. Am. Soc. civ. Engrs*, **94(SM5)**, 1055–1122.

SEILACHER, A. (1969) Fault-graded beds interpreted as seismites. *Sedimentology*, **13**, 155–159.

SEILACHER, A. (1982) General remarks about event deposits. In: *Cyclic and Event Stratification* (Ed. by G. Einsele & A. Seilacher), pp. 161–174. Springer-Verlag, Berlin.

SEILACHER, A. (1984) Sedimentary structures tentatively attributed to seismic events. *Mar. Geol.*, **55**, 1–12.

SELLEY, R.C., SHEARMAN, D.J., SUTTON, J. & WATSON, J. (1963) Some underwater disturbances in the Torridonian of Skye and Raasay. *Geol. Mag.*, **100**, 224–243.

SELWOOD, E.B. & THOMAS, J.M. (1988) The tectonic framework of Upper Carboniferous sedimentation in central southwest England. In: *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe* (Ed. by B. M. Besly & G. Kelling), pp. 18–23. Blackie, Glasgow.

SHEPARD, F.P. (1963) *Submarine Geology*. Harper & Row, New York, 557 pp.

SIMS, J.D. (1975) Determining earthquake recurrence intervals from deformational structures in young lacustrine sediments. *Tectonophysics*, **29**, 141–152.

STØRMER, L. (1955) Merostomata. In: *Treatise on Invertebrate Paleontology, Part P, 2* (Ed. by R. C. Moore), pp. 4–41. Geological Society of America & University of Kansas.

STURM, M. & MATTER, A. (1978) Turbidites and varves in lake Brienz (Switzerland): deposition of clastic detritus by density currents. In: *Modern and Ancient Lake Sediments* (Ed. by A. Matter & M. E. Tucker), *Spec. publs int. Ass. Sediment.*, **2**, 145–166.

SUNDBORG, Å. (1967) Some aspects on fluvial sediments and fluvial morphology. *Geogr. Annlr*, **49A**, 333–343.

SWIFT, D.J.P., THORNE, J.A. & OERTEL, G. (1986) Fluid process and sea-floor response on a modern storm-dominated shelf: middle Atlantic shelf of North America. II. Response of the shelf floor. In: *Shelf Sands and Sandstones* (Ed. by R. J. Knight & J. R. McLean), *Mem. Can. Soc. petrol. Geol.*, **11**, 191–212.

TEVESZ, M.J.S. & MCCALL, P.L. (1982) Geological significance of aquatic nonmarine trace fossils. In: *Animal–Sediment Relations* (Ed. by P. L. McCall & M. J. S. Tevesz), pp. 257–285. Plenum Press, New York.

THOMAS, J.M. (1982) The Carboniferous rocks. In: *The Geology of Devon* (Ed. by E. M. Durrance & D. J. C. Laming), pp. 42–65. University of Exeter.

THOMAS, J.M. (1988) Basin history of the Culm Trough of southwest England. In: *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe* (Ed. by B. M. Besly & G. Kelling), pp. 24–37. Blackie, Glasgow.

TUCKER, M.E. (1982) *The Field Description of Sedimentary Rocks*. Open University Press, Milton Keynes, 112 pp.

TYLER, D.J. (1988) Evidence and significance of limulid instars from trackways in the Bude Formation (Westphalian), south-west England. *Proc. Ussher Soc.*, **7**, 77–80.

VAN DIJK, D.E., HOBDAY, D.K. & TANKARD, A.J. (1978) Permo-Triassic lacustrine deposits in the Eastern Karoo Basin, Natal, South Africa. In: *Modern and Ancient Lake Sediments* (Ed. by A. Matter & M. E. Tucker), *Spec. publs. int. Ass. Sediment.*, **2**, 225–239.

WALKER, R.G. (1963) Distinctive types of ripple-drift cross-lamination. *Sedimentology*, **2**, 173–188.

WALKER, R.G., DUKE, W.L. & LECKIE, D.A. (1983)

Hummocky stratification: significance of its variable bedding sequences. Discussion. *Bull. geol. Soc. Am.*, **94**, 1245–1249.
WHITE, A.H. & YOUNGS, B.C. (1980) Cambrian alkali playa–lacustrine sequence in the northeastern Officer Basin, South Australia. *J. sedim. Petrol.*, **50**, 1279–1286.
WHITE, E.I. (1939) A new type of palaeoniscoid fish, with remarks on the evolution of the actinopterygian pectoral fins. *Proc. zool. Soc. London*, **B109**, 41–61.
WRIGHT, L.D., WISEMAN, W.J., BORNHOLD, B.D., PRIOR, D.B., SUHAYDA, J.N., KELLER, G.H., YANG, Z.-S. & FAN, Y.B. (1988) Marine dispersal and deposition of Yellow River silts by gravity-driven underflows. *Nature*, **332**, 629–632.

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