Journal of the Geological Society, London, Vol. **161**, 2004, pp. 1053–1056. Printed in Great Britain.

Discussion on sea-level change and facies development across potential Triassic–Jurassic boundary horizons, SW Britain

Journal. Vol. 161, 2004, pp. 365–379

Tony Hallam & Paul Wignall write: we welcome the paper by Hesselbo *et al*. (2004) for its thorough description and careful evaluation of the environmental significance of an important group of system boundary strata, but beg to take issue on a number of key points.

(1) Most importantly, while we agree with the evidence of a regressive phase within the Cotham Member, we maintain in marked contrast to their opinion that there was also a sharp regression followed by a rapid transgression at the junction of the Langport Member (of the Lilstock Formation) and Blue Lias. It is pertinent before discussing the evidence in SW England further to take into account the broader European context (Hallam & Wignall 1999, with many relevant references cited therein). The evidence is especially clear in Germany where there was extensive shallowing in the latest Rhaetian marked by progradation of sandstone over shales. In northern Frankonia (Bavaria) fluvial sandstones infill channels incised into late Rhaetian marine strata and are overlain by marine Hettangian. The sea-level rise in the earliest Hettangian (*planorbis* Zone) was evidently rapid, with the limit of marginal marine sandstones in the eastern part of southern Germany being pushed back at the expense of fully marine shales to the maximum extent achieved during the whole Hettangian Stage.

Both in the north and south of Europe a similar pattern of successive sea-level fall and rise can be inferred. A clear end-Triassic regressive pulse can be recognized in the Danish Basin, while both in southern Sweden and NW Poland the upper Rhaetian is missing and there is an unconformity at the base of the Jurassic. In the Northern Calcaerous Alps of Austria, widespread emergence at the end of the Triassic is recognized, with the creation of karst surfaces on emergent reef complexes, while in the few areas of more continuous sedimentation in basinal settings the base of the Jurassic is marked by eroded limestone clasts from emergent areas, or by an exceptional red mudstone horizon interpreted as marginal marine, in the midst of blue-grey fully marine deposits.

In England north of the southern Midlands the marine Hettangian Blue Lias Formation rests with a hiatus on an eroded surface on the Rhaetian Penarth Group, with the upper Langport Member of the Lilstock Formation missing. Collectively the evidence is about as convincing as can be expected from the stratigraphic record of a notable regressive–transgressive couplet at the system boundary.

Naturally, since the margins of the Bristol Channel were an area of relative subsidence at the critical time, with the most continuous marine sections known in northern Europe, any effect of sea-level fall is likely to have been minimized. Nevertheless such evidence is present. As regards the foreshore of the south Devon coast, Hallam (1988) argued for emergence at the end of deposition of the Langport Member (White Lias) on the basis of knife-sharp truncation of *Diplocraterion* burrows, many of which are eroded down to the base of their U-shaped burrow (Wignall 2001, fig. 8). An obvious prediction from this is that eroded limestone clasts might be expected at the base of the black shale directly overlying the Langport in adjacent areas of more continuous sedimentation, with no local indication of erosion, such as at St Audrie's Bay on the Somerset coasts. Such evidence was sought successfully by Hallam (1990) who, working in conjunction with Alastair Ruffell, found angular clasts of pale micrite, ranging in length up to 5 cm, within the bottom 2 cm of the basal Blue Lias black shale. Such clasts were not common, but about 20 were found in half-an-hour's intensive collecting. That Hesselbo *et al*. have failed to confirm this discovery probably relates to the much deteriorated quality of the coastal outcrop after the time the discovery was made in 1988. It is most decidedly not due to a 'mistaken identification of some other feature of this horizon such as the development of carbonate nodules' (Hesselbo *et al*. 2004, p. 374). It is worth adding here that the clasts occur in association with a layer of phosphatic nodules, suggesting a significant stratigraphic horizon.

Subsequent support for Hallam's claim was achieved by Wignall (2001) following examination of superb, newly revealed exposures on the Devon coast. He confirmed an episode of end-Langport Member erosion, which initially exposed a semi-lithified substrate on the seafloor, with truncated *Diplocraterion* burrows cross-cut by sharper-margined burrows. Seafloor lithification was completed later and both *Diplocraterion* and firm-ground burrows are cross-cut by borings. Oysters are occasionally found cemented to the top surface. Subsequently local erosion, brecciation and redeposition of the topmost beds of the White Lias took place. In our view this represents very shallow-water deposition of the Langport Member. A near-identical development of the Langport Member at Long Itchington on the East Midlands Shelf was similarly interpreted (Radley & Swift 2002). In total these features indicate a dramatic increase of a broad range of erosion and erosion-and-redeposition features in the topmost beds of the Langport Member suggestive of shallowing. In contrast Hesselbo *et al.* (p. 377) favour deposition 'in areas where the ramp was steep and the facies characterized by gravity-flow deposits' during base-level rise. Apart from the unlikelihood of such flows coming to rest on steep ramps it overlooks the observation that the larger clasts at the base of the breccia are essentially *in situ* having only been moved a few centimetres, with the result that bedding features can be traced laterally from undisturbed strata (Wignall 2001, fig. 4). In their review of modern carbonate depositional environments, Inden & Moore (1983) stress the frequency of early cementation and break up into clasts by storm action. Episodic emergence above sea level facilitates this early cementation, and can help to explain the frequency of limestone clasts of various sizes throughout the White Lias.

The subsequent fissuring of the top surface of the breccia provides the best evidence of subaerial emergence in the South

Dorset exposures, with both vertical and subhorizontal features present. These were illustrated in Wignall (2001, fig. 4) and also in Hesselbo *et al.* (fig. 10, eastern end of sketch). At Long Itchington the emergence may have been somewhat longer because Radley & Swift (2002, p. 271) observed that the top of the breccia bed is 'truncated by an iron-impregnated erosion surface' suggesting significant subaerial oxidation. Hesselbo *et al.* do not comment on this evidence for emergence and choose instead to interpret the Langport/Blue Lias contact as a flooding surface rather than an amalgamated sequence boundary and flooding surface. The sub-horizontal fissures in the breccia are infilled with laminated shale of identical lithology to the overlying basal bed of the Blue Lias Formation. Wignall (2001) interpreted them as a passive fill of dissolution cracks *after* Langport deposition. Hesselbo *et al.* (2004, p. 372) curiously argued that fine-grained settling from suspension of such material was 'incompatible with passive infilling of dissolutional features'. They alternatively suggested that the shales were deposited contemporaneously with the breccia beds and therefore implied that the depositional regime alternated between quiet, euxinic deposition and phases of brecciation and debris flow emplacement. However, evidence for powerful erosive events, which caused in situ seafloor brecciation, and channel erosion, is at odds with a euxinic depositional setting for the upper part of the Langport Member.

(2) Hesselbo *et al.* make the rather strange claim that the micrite of the Langport Member is secondary to an original coarser-grained rock such as a grainstone, on the basis of what they claim to be small-scale swaley and hummocky cross-stratification. We find this claim unconvincing; there is no supporting stratigraphic evidence in any of the substantial number of thin sections we have seen. We also note that the supposed swaley and hummocky features have been alternatively interpreted as scour-and-fill features produced by a distinct phase of erosion followed by a subsequent phase of settling of fine material that draped a scoured topography (Wignall 2001). There is therefore no need to evoke the presence of calcarenitic material moulded by storm waves required in Hesselbo *et al*.'s alternative.

Hesselbo *et al*.'s claim for a debris flow origin for much of the Langport Member to account for the widely dispersed clasts which do not support each other, is less implausible, but claims of movement down a carbonate ramp into water of some depth are more dubious. Even if a modest ramp gradient of 0.3° is assumed, then a ramp developed over the distance shown in their figure 12 (60 km) implies water depths at the base of the ramp well in excess of 300 m. However, they suggest that the ramp was steep (Hesselbo *et al*. 2004, p. 377). The gradient could of course be highly localized but similar facies are widespread (Radley & Swift 2002) and requires there to be numerous short, steep ramps. There is no supporting evidence in the form of frequent, rapid lateral thickness and facies variations.

(3) In the absence of Triassic ammonites the best location of the Tr–J boundary remains in dispute. Hesselbo *et al*. give a good, comprehensive review of the different positions adopted, confirming the view held by many of us that St Audrie's Bay is not a suitable candidate, as has been claimed, for a global stratotype. Their carbon isotope data certainly gives a valuable new tool in the form of negative anomalies, but their proposal of a new boundary at the top of the Cotham Member is open to dispute for at least two reasons. They cite palynological work by Orbell (1973) in support, but in fact analyses by several palynological researchers in SW England and Wales has led to equivocal and indecisive results (J.B. Riding, pers. comm.). This confirms the global picture, because a sharp biotic turnover of palynomorphs at the purported system boundary has been recognized only in a few places (Hallam 2002).

The use of the first negative carbon isotope anomaly close to the Tr–J boundary in the classic New York Canyon section in Nevada, surely the least unpromising candidate for a global stratotype, fails to resolve the best location for the boundary (Guex *et al.* 2003, 2004; cf. Ward *et al*. 2004). It occurs there at the same level as an *Arcestes* species, normally taken as an Upper Triassic ammonite, and 4 m below the oldest undoubted Jurassic ammonite, *Psiloceras*. Thus, the lowest negative $\delta^{13}C$ excursion is clearly well within the Triassic.

Let us, however, end our comments on a point of agreement, namely the likely importance of Central Atlantic Magmatic Province (CAMP) volcanism as a major factor controlling environmental change, with potential bearing on the end-Triassic mass extinction. Manifest slump phenomena have been recorded in the Langport Member by Hallam (1960) and the underlying Cotham Member by Mayall (1983) and Simms (2003). Such phenomena are unusual, if not unique, in the extra-Alpine Mesozoic epicontinental seas, and calls for a rather special explanation. Because, as is widely known, volcanic eruptions are commonly associated with seismic activity, we consider the slumps are more likely to have been produced by CAMP-related earthquakes than by a bolide impact, as suggested by Simms (2003). The only good evidence for bolide impact in SW England is an ejecta layer dating from pre-Rhaetian times, several million years before the main mass extinction phase (Walkden *et al*. 2002).

27 May 2004

Stephen Hesselbo, Stuart Robinson and Finn Surlyk reply: *(1) Sea-level change represented by the top of the Langport Member (Lilstock Formation).* In their discussion, Hallam & Wignall review the evidence for sea-level change at about the Tr–J boundary in a pan-European context. The issue is not whether the sections that they list show evidence of sea-level change; that is to be expected of paralic or shallow-platform successions. The key uncertainties are how the strata that they describe correlate with the strata under discussion in southern Britain, and how great are the hiatuses represented by the lowstand surfaces. We are unaware of any convincing evidence to show that surfaces formed in the continental European sections during times of relative sea-level lowstand correlate precisely with the top surface of the Langport Member in the southern UK.

The sharp truncation of the trace fossil *Diplocraterion* near the top of the Langport Member is a feature that Hallam (1988) has placed special emphasis upon as an indicator of sea-level fall. However, it is in the nature of *Diplocraterion* to show truncation of the higher ends of tubes and spreite as the animal was able to migrate up and down through the sediment column in response to sediment deposition and erosion (Goldring 1964; Bromley 1996, p. 289). Evidence for sediment erosion is not the same as evidence for sea-level fall.

There is indisputable evidence for early lithification occurring during deposition of the Langport Member, but early lithification does not require subaerial exposure (e.g. review in Tucker & Wright 1990), and the break-up of early-lithified carbonate can be simply explained by slope-related gravity transport processes operating on partially cemented marine substrates (e.g. Ineson & Surlyk 1995). The preponderance of early cementation and brecciation (in thin beds) towards the top of the Langport Member is compatible with reduced sedimentation rate (resulting from deepening and sediment starvation).

With regard to the supposed incompatibility between euxinic deposition and powerful erosive events, we note that it is not uncommon for gravity-flow or storm processes to temporarily introduce oxygenated waters into otherwise dysaerobic environments, resulting in laminated shales interbedded with debrites or burrowed, cross-stratified calcarenites. Recent geochemical work on the Oxford Clay, a similar lithofacies to that of the Blue Lias, has additionally shown that euxinia is in many cases an intermittent phenomenon on a timescale of years (Kenig *et al*. 2004).

In the absence of a detailed description, it is difficult to comment on the origins of the 'iron-impregnated' surface reported by Radley & Swift (2002) from the top of the Langport Member in the English Midlands. Two possibilities unrelated to Triassic–Jurassic subaerial exposure are: (1) oxidation of a pyritized surface by modern oxidizing ground waters; or (2) the occurrence of a ferromanganese oxide crust. Notable examples of the latter are in Jurassic marine limestone successions in Southern Italy (Sicily), as recently re-described by Martire & Pavia (2002) and Di Stefano *et al*. (2002).

Hallam & Wignall claim that 'fissuring' of the top surface of the Langport Member provides the best evidence of subaerial emergence. It would be inappropriate here to go over again the published arguments against a karst origin for these features, which are based upon carefully illustrated field observations. Nevertheless, we stress that all the observations can be accounted for on the basis of straightforward process sedimentology, and some of the observations are incompatible with subaerial dissolution of a hardened limestone, or previously unknown in that context (Hesselbo *et al*. 2004, pp. 372–374).

(2) Sedimentary processes on the Langport Member ramp. We have illustrated in our paper the evidence showing that swaley and hummocky cross-stratification in the Langport Member does not simply drape an underlying scoured topography but comprises both concave-up and convex-up cross sets and subsets. The occurrence of this style of cross stratification is incompatible with 'passive settling' of 'fine' (clay to fine silt) material on a scoured topography as these smaller grain sizes require still water to settle and such a process would not produce cross-stratification. Apropos discussion of widths of the hypothesized carbonate ramp and depths of water, it should be noted that figure 12 is schematic and is intended to illustrate only the positions of localities within a generalized ramp facies framework. Nevertheless water depths in the order of hundreds of metres would not have been impossible during deposition of the basal Blue Lias if the ramp or ramps resulted from modest fault block tilting. Also, slope facies may easily be widespread on shorter steeper ramps if those ramps had phases of progradation, as is implied in our figure 12, and the resultant facies would be diachronous.

(3) Position of the Tr–J boundary. Hallam and Wignall suggest that the negative carbon-isotope anomaly reported by Guex *et al.* (2003) co-occurs with the Late Triassic ammonite *Arcestes* and therefore cannot be regarded as a potential marker for the Tr–J boundary. However, their identification of the horizon of the negative excursion is strongly influenced by a single outlying data point that may well represent analytical or diagenetic 'noise'. In that case, the negative excursion occupies principally the 3 m of section immediately above the highest occurrence of *Arcestes* and 5 m below the lowest occurrence of *Psiloceras*, perfectly compatible with the level of the excursion in the UK successions (Hesselbo *et al*. 2002, 2004) and elsewhere (Ward *et al*. 2001; Pálfy *et al*. 2001).

7 July 2004

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Tony Hallam, School of Geography, Earth and Environmental Sciences, University of Birmingham, Birmingham B15 2TT, UK (e-mail: a.hallam@bham.ac.uk)

Paul Wignall, School of Earth Sciences, University of Leeds, Leeds LS2 9JT, UK

Stephen P. Hesselbo, Department of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK (e-mail: stephen.hesselbo@earth.ox.ac.uk)

Stuart A. Robinson, Lamont-Doherty Earth Observatory of Columbia University, 61 Route 9W, Palisades, NY 10964, USA

Finn Surlyk, Geological Institute, University of Copenhagen, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

Scientific editing by John Howell