

Comment—Contrasting Deep-water Records from the Upper Permian and Lower Triassic of South Tibet and British Columbia: Evidence for a Diachronous Mass Extinction (Wignall and Newton, 2003)

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“Happy families are all alike, but every unhappy family is unhappy in its own way” Tolstoy (1878). As in this famous first line of *Anna Karenina*, each Permian–Triassic boundary section is unhappy in its own way. Wignall and Newton (2003) are to be commended for showing that this great mass extinction had different effects at different localities, such as deep oceanic warming in Tibet, but they have not demonstrated their chief claims that (1) mass extinction was globally diachronous by a half million years, or (2) due to dysoxia from oceanic stagnation. Furthermore, their foraminiferal disaster taxa are better explained by a methane-outburst hypothesis (Krull and Retallack, 1999; Berner, 2002), which they fail to mention.

There is a difference between diachronous mass extinction, where most species become extinct at different times in different places, and selective survival, where some species survive the mass extinction. Wignall and Newton’s (2003) claim for diachroneity comes from range truncation of foraminifera assigned to 7 genera in the basal 35 cm of Triassic beds in the Selong section of Tibet. These are a statistically and taxonomically inadequate basis for delayed mass extinction, especially considering that none of these Triassic foraminifera were fusulines, which remain the primary foraminiferal casualties of the Permian–Triassic life crisis (Stanley and Yang, 1996). The best understood Late Permian taxa near Selong are 43 species of brachiopods, which dwindled to 8 dwarfed brachiopod species in the “*Waagenites*” bed, above which no brachiopods were found despite deliberate search (Shen et al., 2000, 2001). All the brachiopods were gone before the first appearance of the conodont *Hindeodus parvus* just above the “*Waagenites*” bed, which includes a marked carbon isotopic excursion (Jin et al., 1996). Brachiopods are evidence of mass extinction coincident with the isotopic excursion and before the boundary-marking *H. parvus*, as in numerous other Permian–Triassic boundary sections around the world (Erwin, 1993; Jin et al., 1996). Wignall and Newton’s (2003) Selong foraminifera were thus survivors of the brachiopod mass-extinction 35 cm below.

The selective survival of foraminifera is unsurprising within the context of a Permian–Triassic methane-outburst hypothesis (Krull and Retallack, 1999; Krull et al., 2000). Modern foraminifera thrive around sewage and

natural gas outlets, creating large and well-calcified tests (Yanko et al., 1994). Disaster blooms of foraminifera also were associated with the methane-outburst crisis of the Eocene–Paleocene boundary (Thomas, 2003). Many foraminifera also thrive under dysoxic conditions (10–40 $\mu\text{M.kg O}_2$: Patterson et al., 2000; Platon and Sen Gupta, 2001). Modelling of Permian–Triassic methane outburst by Berner (2002) indicates that atmospheric reduction initiated by methane oxidation, added to carbon dioxide from volcanism and ongoing extinctions, could have lowered atmospheric oxygen levels from a Late Permian high of 35% to an earliest Triassic low of 12% within 20,000 years of the boundary. The difficulties of diminished atmospheric and oceanic oxygen were compounded by hypercapnia from methane oxidation to carbon dioxide. Among marine invertebrates, poorly ventilated corals and brachiopods were preferentially lost compared with more muscular molluscs (Knoll et al., 1996). Among plants, marginally aerated swamp vegetation succumbed, rather than plants of well-drained soils (Retallack et al., 1996). Among vertebrates, the survivors had more effective lungs and nasal passages (Retallack et al., 2003).

Supposed diachronous extinctions are linked with interpreted diachronous oceanic dysoxia by Wignall and Newton (2003), whereas isotopic evidence for methane outburst is globally synchronous (Krull et al., 2000). The methane-outburst hypothesis presents a mechanism for dysoxia (Berner, 2002) that is very different from the oceanic stagnation model invoked by Wignall and Newton (2003). Both stagnation and methanogenic hypoxia can result in well-bedded shales with little evidence of life. The stagnation model is characterized by carbonaceous black shales fed by surface-water biological productivity (Wignall, 1994). Methane-induced hypoxia on the other hand would not necessarily produce shales with high organic carbon content. Although Isozaki (1997) and Wignall and Newton (2003) claim that early Triassic shales are carbonaceous, I have been unable to find any supporting analytical data. Hundreds of published organic carbon analyses of marine earliest Triassic shales do not exceed 3 wt %, whereas Late Permian and Middle Triassic shales in the same sequences are much more carbonaceous (Suzuki et al., 1993; Kajiwarra et al., 1994; Wang et al., 1994; Wolbach et al., 1994; Wignall et al., 1998; Morante, 1996; Krull et al., 2000). Well-studied oceanic anoxia events of the Cretaceous produce thick, carbonaceous shales (Wignall, 1994; Kunht et al., 2002) with carbon contents (>10 wt % TOC) well in excess of thin carbonaceous shales at the methane outburst crisis of the Paleocene–Eocene boundary (Gavrilov et al., 2003; Beniamovski et al., 2003). Earliest Triassic oceanic dysoxia was, contrary to Wignall and Newton (2003), neither diachronous nor carbonaceous, but more likely synchronous and oligotrophic (Morante, 1996; Krull et al., 2000).

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Reply

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We are pleased that our paper has generated interest from Retallack and relish the opportunity to comment further on the intriguing timing of the end-Permian extinction. Perhaps not surprisingly, we do not agree with any of Retallack's assertions and deal with them here in the order he presented them.

Firstly, Retallack misconstrues the nature of the Permian foraminifera that survive into the Triassic in Tibet. These taxa are not disaster taxa in the understood sense, although foraminiferal disaster taxa, such as *Earlandia* are well known from post-extinction strata in many sections (see Hallam and Wignall 1997, which includes a definition of disaster taxa). Neither do they have attributes of dysaerobic taxa, but rather they are forms commonly encountered in shallow-marine platform carbonates of the Late Permian equatorial Tethyan realm. They turn up in Tibet, in the earliest Triassic, after they had gone extinct in their normal equatorial haunts. It is the same story for the inozoan sponge taxon that Retallack does not mention. However, these foraminifera did not survive for long as they went extinct within the Griesbachian Stage; so it is perplexing that Retallack considers this a selective survival. It is not a survival because all the foraminifera (probably 10 species in total) and the sponge went extinct! The implication that these extinctions are unimportant because they were non-fusuline taxa is simply spurious. The end-Guadalupian mass extinction eliminated the majority of fusulines and they had not recovered greatly by the time of the end-Permian mass extinction. The fusulines were not the primary foraminiferal casualties of this extinction event and Stanley and Yang (1996) did not claim other-

wise. Rather foraminifera with calcareous granular tests, which included the surviving fusulines as well as the diverse and abundant endothyrids, are the primary end-Permian casualties (except in Tibet where they survive a little longer, as we showed).

Secondly, Retallack refers to the excellent, detailed analysis of the brachiopod record of Selong by Shen and colleagues (2000, 2001). We do not doubt that many brachiopod species went extinct before the Permian–Triassic boundary (P–Tr) at Selong, as they show, and never claimed otherwise in our paper. This is part of a diachronous mass extinction story within this section.

Thirdly, Retallack challenges the ocean stagnation cause for the marine anoxia and instead invokes dysoxia caused by methane oxidation. However, his alternative does not hold water. Firstly, Berner (2002) did *not* claim that atmospheric oxygen levels could have plummeted from 35% to 12% in the space of 20,000 years across the P–Tr boundary. This would involve the loss of around 5×10^{19} moles of oxygen from the atmosphere. Berner's modeling showed that the negative C-isotope shift across the boundary could have been caused by the release up to 3×10^{17} moles of methane. Clearly, oxidation of this methane would have little impact on atmospheric oxygen levels. On the other hand, rapid release of this methane into the water column could have caused transient methanogenic anoxia. The Paleocene–Eocene thermal maximum (PETM) provides a possible example of this phenomenon (Dickens, 2000), but this is most unlike the P–Tr superanoxic event in both duration and intensity. The PETM event was brief (a few tens of thousands of years), only associated with transient, deep-water dysoxia, and linked with the emission of around half the amount of methane released during the end-Permian C-isotope event. The P–Tr superanoxic event lasted around 20 million years (it began at the start of the Late Permian and finished at the start of the Middle Triassic; Isozaki, 1997) and, at its height, true anoxia was intensely developed even within shallow water locations (Wignall and Twitchett, 2002). Furthermore, the onset of this anoxic intensification predates the $\delta^{13}\text{C}$ excursion (Twitchett et al., 2001; Wignall and Newton, 2003). It is impossible to envisage how the required vast volumes of methane could have been released continuously throughout this time interval, caused such intense oxygen restriction, and yet only left an isotopic signal at the P–Tr boundary.

Retallack dismisses our alternative warming-induced stagnation model because of a lack of organic-rich strata. However, numerous factors control organic carbon levels, including sediment accumulation rates, dilution by (for example) carbonate content, surface water productivity, and only modest primary productivity is perfectly compatible with anoxic conditions (Wignall, 1994). For this reason most workers assess oxygen levels using other criteria, such as trace metal enrichment, fabric indicators such as the presence of undisturbed lamination, pyrite framboid size distributions, and others (cf. Wignall and Twitchett, 2002; Wignall and Newton, 2003). Retallack falsely claims that Lower Triassic marine shales generally have low total organic carbon (TOC) contents not exceeding 3 wt % and that they are less organic C-rich than Upper Permian and Middle Triassic shales. These assertions are not supported by the literature to which he refers. Thus, Suzuki et

al. (1993, p. 711) noted that the P–Tr boundary black shales in their Japanese sections have “2.2% to 5.3%” TOC and this is after they have been heated to over 100°C, implying an original TOC content of “4 to 8%” (Suzuki et al., 1993, p. 715). The sulfur/carbon plot of Kajiwara et al.'s (1994, fig. 1) study clearly demonstrates that the P–Tr boundary black shale is more organic-rich than the bounding sediments. Our data from Spitsbergen (Wignall et al., 1998, table 1, fig. 8) show that lowest Triassic shales have comparable or higher TOC values than the underlying uppermost Permian shales. The Morante (1996) study was of non-marine to paralic strata and so is irrelevant in this context. Krull et al.'s (2000, fig. 4) section was developed in a volcanogenic sandstone succession, which, not surprisingly, has low TOC values although the highest values are in the Lower Triassic.

In finishing, we note that our claim of a diachronous P–Tr mass extinction also was based on evidence from British Columbia, which Retallack does not challenge. However, the BC data also provide crucial evidence for the diachroneity with a diverse radiolarian fauna disappearing over a meter below the loss of the benthic fauna (composed of abundant hexactinellid sponge spicules of diverse types). In our original paper we noted that the diverse radiolaria were replaced for a time by apparently simple radiolaria, which we referred to as sphaeroids. Subsequently, Peter Cejchan (pers. comm., 2003) has succeeded in extracting sphaeroids from their matrix to reveal that they are simple, thin-walled siliceous spheres that lack diagnostic pores or ornament of radiolarians. This discovery serves to emphasize the abrupt demise of the radiolarian populations. Thus the evidence from BC and Tibet indicates a distinct, phased nature of the P–Tr extinctions in the marine realm:

Phase 1: Abrupt elimination of radiolarian communities, probably several tens of thousands of years prior to phase 2, assuming typical pelagic sedimentation rates for the deep-water biogenic cherts of the BC section.

Phase 2: Mass extinction of marine benthos, including deep-water sponges in BC and many shallow-water taxa including brachiopods in Tibet.

Phase 3: Onset of a negative C-isotope excursion that may record methane release from hydrates in a warming ocean. Foraminifera and sponge refuges from the tropics turn up in high latitude Tibet.

P–Tr Boundary

Phase 4: Extinction of refuges in Tibet due to the delayed onset of dysoxia in this region.

In detail, the negative $\delta^{13}\text{C}$ excursion, and by implication the release of gas hydrates, occurs after the extinction phases, indicating that it is a consequence of the end-Permian environmental crisis and not a factor in the extinction story. A similar sequence of timings is seen in East Greenland, which has an expanded record of the end-Permian mass extinction (Twitchett et al., 2001).

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