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Carbon and conodont apatite oxygen isotope records of Guadalupian–Lopingian boundary sections: Climatic or sea-level signal?

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ABSTRACT

The Guadalupian–Lopingian (G–L) boundary (260.4 Ma) is one of the major extinction events in Earth history that coincides with the Emeishan Large Igneous Province and palaeoclimatic changes. Carbon isotopes of whole rock samples were studied in order to document changes in the global carbon cycle. In contrast to earlier studies, we observe no major negative excursion in $\delta^{13}C$ in the middle Capitanian. A positive $\delta^{13}C$ excursion is observed in the latest Capitanian with a 1.5‰ increase registered in the *J. xuanhanensis* Zone to *C. postbitteri hongshuiensis* Subzone, followed by a decrease of 1‰ within the *C. postbitteri postbitteri* Subzone and a 2‰ decrease in the *C. dukouensis* to *C. asymmetrica* Zone. Oxygen isotopes of conodonts from two G–L boundary sections were measured in order to reconstruct conodont habitat and potential changes in water temperature. Oxygen isotope ratios of gondolellid conodonts are higher in comparison to oxygen isotope ratios measured on hindeodid conodonts suggesting that gondollelids lived in cooler and thus deeper waters compared with hindeodids. The oxygen isotope record reconstructed from gondollelid conodonts suggests warming of water temperatures of about 4 °C in the late Capitanian (*J. postserrata* to *J. granti* Zone), cooling of about 6 to 8 °C across the G–L boundary and in the earliest Wuchiapingian, and again significant warming in the Wuchiapingian (*C. dukouensis* to *C. liangshanensis* Zone). The temperature increase can be correlated with the main phase of Emeishan volcanism suggesting that climatic warming may have resulted in an intensified hydrological cycle, fertilisation of the oceans and enhanced primary productivity, the latter documented in the positive late Capitanian carbon isotope excursion. However, changes in sea-level seem to parallel reconstructed water temperatures suggesting that changes in water depth in combination with superimposed climatic changes may be responsible for the observed temperature changes. This study documents that oxygen isotope studies on Permian conodonts should be performed on mono-generic conodont samples and that oxygen isotopes not only provide valuable palaeoclimatic information but also may help to constrain the life habitat of conodonts.

1. Introduction

The end-Guadalupian or late Capitanian mass extinction (Shen and Shi, 2009; Wignall et al., 2009a) is considered one of the most severe extinction events in Earth history (Jin et al., 1994; Stanley and Yang, 1994). The cause of this mass extinction event is still debated with various triggering mechanisms being proposed to have initiated the major decrease in marine and terrestrial faunal diversity, especially, the significant loss of fusulinacean foraminifera (Stanley and Yang, 1994; Yang et al., 2004; Bond et al., 2010), corals (Wang and Sugiyama, 2000), and brachiopods (Shen and Shi, 2002, 2009). Radiometric and biostratigraphic dating indicated that the Emeishan volcanic eruption coincided with this mass extinction (Zhou et al., 2002; Wignall et al., 2009b; Sun et al., 2010). This temporal link has led some researchers to suggest that flood basalt volcanism was a potential trigger of the extinction event (Courtillot et al., 1999; Wang and Sugiyama, 2000; Zhou et al., 2002; Wignall et al., 2009a). In addition to volcanism, a catastrophic methane outburst and oceanic anoxia were suggested as other potential causes of the mass extinction (Isozaki, 1997; Retallack et al., 2006). The most prominent sea-level lowstand during the Palaeozoic occurred during the Capitanian–Wuchiapingian interval (Haq and Schutter, 2008; Wignall et al., 2009a) and resulted in erosional unconformities developed in many shelfal or platform successions around the World (Jin et al., 2006). The regression and reduction of shallow water habitats may have severely affected marine life and contributed to the biotic crisis. Capitanian cooling (Kamura event) was suggested to have influenced marine

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faunal assemblages during the late Guadalupian (Isozaki et al., 2007a, b). High $\delta^{13}C$ values measured on shallow marine Capitanian carbonates from mid-Panthalassan sections were interpreted as indicative of high primary productivity, enhanced organic carbon burial and thus low atmospheric CO_2 levels culminating in comparatively cool climatic conditions (Isozaki et al., 2007a, b). In addition, Fielding et al. (2008) reported sedimentary evidence for Capitanian ice fields in eastern Australia. However, the precise causal links between these potential causes and the biotic crisis have not been resolved. For example, high-resolution palaeotemperature records which are crucial for the understanding of the relationship between volcanism, climatic changes, glaciations, sea-level changes and the end-Guadalupian mass extinction are not available for the Guadalupian–Lopingian transitional interval.

This contribution focuses on the reconstruction of palaeoenvironmental changes during the late Guadalupian and early Lopingian. Carbon isotopes of whole rock carbonates were studied to confirm changes in the global carbon cycle during this major crisis. Oxygen isotopes of conodont apatite which have been shown to faithfully record palaeotemperatures of ancient oceans (Joachimski and Buggisch, 2002; Joachimski et al., 2004, 2006; Trotter et al., 2008) were used to reconstruct the palaeotemperature history in the Guadalupian and early Lopingian. Palaeotemperatures reconstructed from oxygen isotopes of gondollelid and hindeodid conodonts from the Penglaitan and Tieqiao G–L boundary sections (Guangxi Province, Southwest China) are discussed in the light of conodont habitat, potential climatic changes as consequence of Emeishan flood basalt volcanism and changes in water depth as consequence of sea-level changes.

2. Geological setting

The study area is situated in Laibin County (Guangxi Province, Southwest China) where Permian sediments are extensively exposed along the emerged riverbanks of the Hongshui River (Fig. 1). During Permian times, this area was situated within the Yunnan–Guizhou–Guangxi Basin of the South China block (Sha et al., 1990; Shen et al., 2007), which was situated near the equator in the Eastern Tethys (Metcalfe, 2002). The Yunnan–Guizhou–Guangxi Basin represents the southern extension of the Jiangnan Basin which was continuously subsiding between the Yangtze and Cathaysian blocks during Permian times (Wang and Jin, 2000).

The Tieqiao section is located on the northern bank of the Hongshui River, about 5 km southeast of Laibin. The Penglaitan section represents the Global Stratotype Section and Point (GSSP) of the Guadalupian–Lopingian series boundary (Jin et al., 2006) and is exposed on the southern bank of the Hongshui River, some 20 km east of Laibin. Structurally, the Penglaitan and Tieqiao sections are located on the eastern and western flanks of the Laibin syncline, respectively (Fig. 1). Both sections are characterised by continuous deposition in the late Guadalupian and early Lopingian, are well-dated by means of conodont biostratigraphy (Mei et al., 1998; Jin et al., 2006), and were not affected by significant structural deformation or metamorphism (Jin et al., 2006). Whilst in most other sections in the world the G–L boundary is characterised by a depositional hiatus as a consequence of the end-Guadalupian global regression, the Penglaitan section is considered as one of the most complete Guadalupian–Lopingian (G–L) boundary sections worldwide (Mei et al., 1998; Jin et al., 2006).

The depositional environment of the Laibin sections during most of the Guadalupian to early Wuchiapingian was a basin to slope setting with water depth estimated around 200 m (Wang and Jin, 2000). The lithostratigraphic units are the Maokou and Wuchiaping/Heshan Formations which are dominated by mostly basinal carbonates and radiolarian cherts. The uppermost 10 m of the Maokou Formation is the Laibin limestone member, which is mainly composed of pale grey medium to thick bedded massive limestones in the lower part and thin to medium bedded crinoidal tuffaceous limestones in the upper part (Shen et al., 2007). The Laibin limestone is

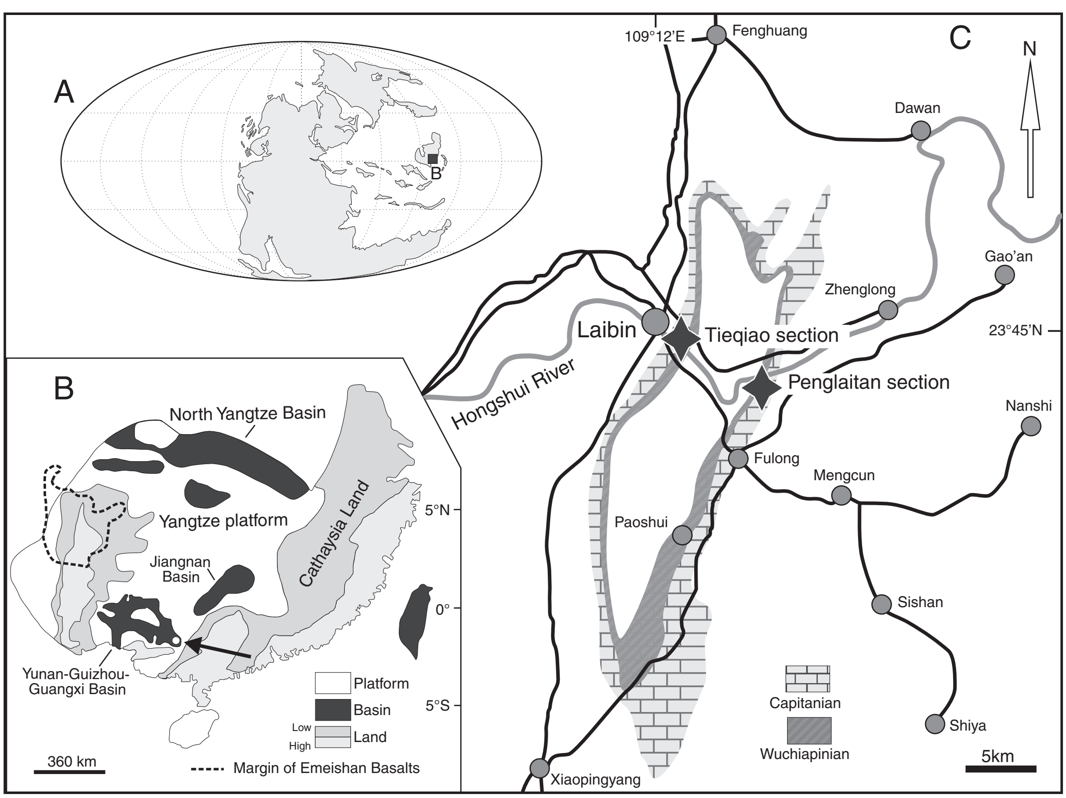

Fig. 1. (A) Late Permian palaeogeographic reconstruction with location of the study area on South China plate (http://www2.nau.edu/rcb7/260moll.jpg), (B) palaeogeography of South China during the Wuchiapingian (Wang and Jin, 2000; Shen et al., 2007; palaeolatitude after Zhang, 1997; Fang et al., 1992) and modern outcrop area of Emeishan volcanic province (Ali et al., 2002), and, (C) location of Penglaitan and Tieqiao sections (Jin et al., 2006).