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Palaeosol stratigraphy across the Permian–Triassic boundary, Bogda Mountains, NW China: Implications for palaeoenvironmental transition through earth's largest mass extinction

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ABSTRACT

Upper Permian and Lower Triassic palaeosols from northeastern Tethyan localities exposed within the Bogda Mountains, NW China, provide a wealth of information regarding long-term palaeoclimatic and palaeoenvironmental variations. Wuchiapingian palaeosols are characterized by intense redoximorphy, accumulation of vascular plant matter, accumulation of clay minerals and Fe-oxides, slickensides, and clastic dikes, suggesting a soil moisture regime that ranged from perennially wet to distinctly seasonal in soil moisture budget. Changsinghian to early Induan palaeosols include subsurface accumulations of clay and carbonate as well as surficial accumulations of organic matter, indicative of sub-humid to sub-arid soil moisture and variable soil moisture regimes. Induan to Olenekian palaeosols contain pedogenic $CaCO_3$ accumulations and gypsum pseudomorphs, indicating a drier environment characterized by net soil moisture deficiency. Elemental composition of palaeosol matrix was used to estimate palaeoprecipitation through the chemical index of alteration minus Potassium (CIA-K) proxy. Estimates from various Wuchiapingian strata indicate relatively stable palaeoprecipitation. During the late Changsinghian and early Induan, palaeoprecipitation appears to have decreased from 1100 to 230 (±180) mm/year over less than 100 m of vertical stratigraphic section. In the Induan and Olenekian, palaeoprecipitation appears much less stable than in Wuchiapingian, with values vacillating from 290 to 1014 mm/year. The transition to a relatively unstable precipitation state coincides generally with the Permian–Triassic boundary, and may reflect climatic disturbances associated with the end-Permian extinction event in addition to altered atmospheric circulation patterns resulting from regional tectonics, moisture availability, and expansion of the subtropical high pressure belt.

1. Introduction

Palaeoenvironmental reconstructions based on palaeosol morphologies are developed from Upper Permian to Lower Triassic strata that crop out in the Bogda Mountains, Xinjiang Uyghur Autonomous Region, northwest China (Fig. 1a). The strata were deposited in an entirely terrestrial, primarily fluvio-lacustrine setting (Carroll, 1998; Lou et al., 2000; Shao et al., 2001; Tang et al., 1994; Wartes et al., 2000; Yang et al., 2007, 2010). Within this work, we report field descriptions, mineralogy and geochemistry from >200 palaeosol profiles.

The end-Permian extinction was the largest extinction of the Phanerozoic when upwards of 84% of marine genera underwent extinction (Sepkoski, 1989). Furthermore, climate-sensitive proxies from Gondwana and Euramerica indicate that the Permian period was a time characterized by drastic climate change, including changes associated with glacial–interglacial and glacial–non-glacial variability during the Early and Middle Permian (Fielding et al., 2008; Montañez et al., 2007), which culminated in a transition from an icehouse world similar to modern conditions to a greenhouse world with no significant high-latitude ice during the Late Permian (Peyser and Poulsen, 2008; Tabor and Poulsen, 2008). Not only did ecosystems likely experience substantial climate variability during the Late Permian as a result of changing atmospheric and oceanic circulation patterns associated with the demise of high-latitude ice sheets, but they also responded to whatever processes triggered the mass extinction (e.g. Arche and López-Gómez, 2005; Lozovsky, 1998; Michaelson, 2002; Retallack, 1999; Retallack and Krull, 1999).

The Permian–Triassic boundary has been recognized in the Bogda Mountains and elsewhere in the Junggar and Turpan basins of the Kazakhstan plate in northwest China (Lou et al., 2000; Ouyang and

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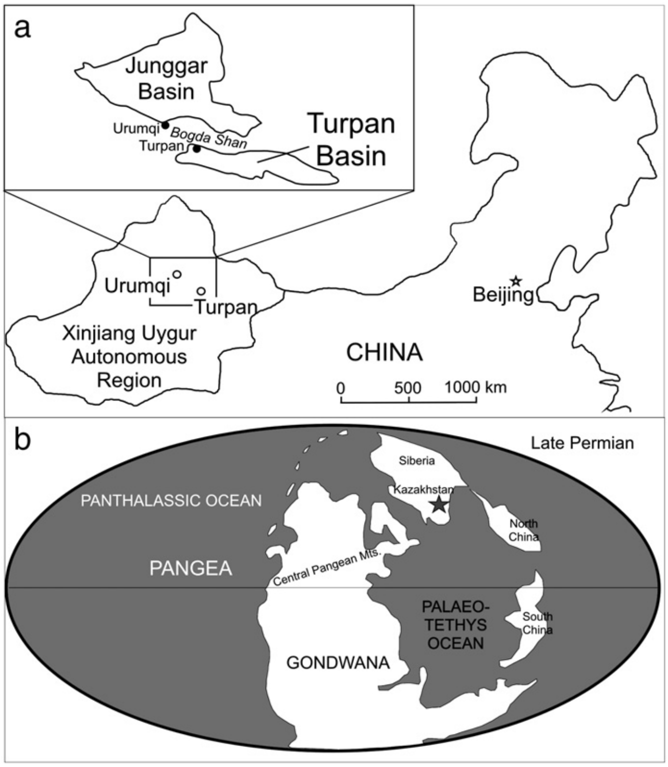

Fig. 1. a. The Bogda Shan is located along the northern edge of the Turpan Basin in Xinjiang–Uygur Autonomous region of northwestern China. The field locations are approximately 50 km north of Turpan. b. Global palaeogeographic reconstruction for the Late Permian (255 Ma) taken from Scotese (2002). The general location of the study area is designated by the star. This region is part of the Kazakhstan plate.

Norris, 1999; Yang et al., 2007, 2010; Zhu et al., 2005). Because there are no apparent unconformities of long duration in the Upper Permian and Early Triassic strata, the Bogda Mountains sections provide an ideal setting to examine Late Permian and Early Triassic climatic change, as well as to evaluate environmental changes and their possible links to ecological devastation caused by, or related to, the mass extinction event.

Although Upper Permian to Lower Triassic terrestrial strata are exposed in Euramerica, Russia, and China, the best-studied sections crop out in India, South Africa, Australia, and Antarctica, which occupied high-latitude positions in Gondwana during the Permian–Triassic (Collinson, 1997; Michaelson, 2002; Retallack, 1999; Retallack et al., 2005, 2006; Tiwari, 2001). The Turpan–Junggar basin of northwest China was located adjacent to the Tarim and Angaran blocks in the vicinity the Palaeo-Tethys Ocean (Zharkov and Chumakov, 2001), which likely occupied a mid-latitude position (Fig. 1b) during the Permian and Early Triassic. Therefore, this study provides a new Permian-Triassic data set from a region about which very little is known (Metcalfe et al., 2009).

2. Geologic setting and depositional framework

The studied sections, Tarlong and Taodonggou, crop out in the southern foothills of the Bogda Mountains, bordering the northwest margin of the Turpan Basin (Fig. 1a). The Bogda Mountains contain a sedimentary succession that ranges in age from Devonian to Neogene (Zhang, 1981). The strata at Tarlong and Taodonggou are Lower Permian to Lower Triassic in age.

Northeast-trending palaeocurrent data indicate that the Jueluotage Mountains were the dominant source of detrital sediments during the Late Permian and Early Triassic (Shao et al., 1999). Deposition occurred in intermontane graben basins, related to collision and shearing of the Turpan–Junggar and Tian Shan plates south of the study area (Allen et al., 1995; Carroll et al., 1995; Liu, 2000). Uplifted blocks of the intermontane basins provided an additional source of locally derived sediments (Şengör and Natal'in, 1996; Yang et al., 2010). Local sediments include igneous fragments from andesite, dacite, rhyolite, and basalt (Shao et al., 2001). The tectonic setting during the Late Palaeozoic is not fully understood, but regionally, the geology reflects some combination of arc volcanism, collision, and extension (Carroll et al., 1995; Wartes et al., 2002).

Two stratigraphic sections were measured in the Tarlong area, Tarlong North and Tarlong South, which are exposures on northern and southern limbs of a WNW-plunging syncline, respectively (Fig. 2). Two sections were also measured in the Taodonggou area: Taodonggou East and Taodonggou West. These Taodonggou sections fall approximately along strike and are separated by ~1 km. Correlation among the sections is largely based on lithostratigraphy, biostratigraphy, and cyclostratigraphy (Cheng et al., 1996; Liao et al., 1987; Wartes et al., 2002; Yang et al., 2007, 2010; Zhang, 1981; Zhu et al., 2005).

Here, we use the stratigraphic terminology of Yang et al. (2007, 2010. The measured sections have been divided into 3 cyclostratigraphic units (Figs. 2b and 3), or low order cycles (LOC): the Wutonggou LOC, the Jiucaiyuan LOC, and the Shaofanggou LOC. Correspondence between lithostratigraphic formations and the cyclostratigraphic units is shown in Fig. 2b. The Wutonggou LOC corresponds to the Wutonggou Formation and most of the Guodikeng Formation, whereas the Jiucaiyuan LOC includes the uppermost Guodikeng and all of the Jiucaiyuan Formation. The Shaofanggou LOC is equivalent to the Shaofanggou Formation. Correlations based on physical tracing of beds, isotope stratigraphy (Fig. 2), and cyclostratigraphy (Yang et al., 2010) are shown in Fig. 3.

The overall depositional system is interpreted to have been fluvial to lacustrine throughout the Wutonggou, Jiucaiyuan, and Shaofanggou LOCs. The Wutonggou LOC at Taodonggou West and within the lower 600 m at Tarlong North is interpreted to represent fluvial and deltaic deposition, with a thin (~25 to 50 m) interval of lake margin to littoral non-deltaic siliciclastic deposits. The upper ~325 m of the Wutonggou LOC at Tarlong North is interpreted to be primarily lacustrine: lake margin to littoral and deltaic in origin. This interval, however, is much thinner stratigraphically at the Taodonggou West section, which is likely the result of Taodonggou being distal to the basin margin whereas Tarlong North was proximal to a major depocentre. The relative differences in basinal position are evidenced by facies: proximal deltaic facies likely record delta-switching events (Yang et al., 2010) that influenced local subsidence patterns and subsequent stratal thickness variations between the sections. Strata within the Triassic Jiucaiyuan and Shaofanggou LOCs are interpreted to as mudflats (lake plain), fluvial channels, and floodplain environments (Yang et al., 2010).

The Permian–Triassic boundary has been placed within a ~30 m thick zone in the upper part of the Wutonggou LC in the Taodonggou West stratigraphic section based on biostratigraphy (Liao et al., 1987; Liu, 2000). The Permian–Triassic boundary is not well constrained in the Tarlong study area. Here, on the basis of cyclostratigraphic correlation to the Taodonggou section a Permian–Triassic transition zone is defined which approximately corresponds to the ~120 m thick zone defined here between 910 m and 1030 m at Tarlong North.

3. Field and laboratory methods

Fieldwork focused on the identification, description, and sampling of palaeosols. Sections were dug back ~20 to 60 cm so that fresh surfaces were obtained for description and sampling. Palaeosols from the sections were recognized and described using the methods of Retallack (1988), Kraus and Aslan (1993) and Kraus (1999). Palaeosol and sediment colours were identified from dry samples using Munsell colour charts (Munsell Color, 1975). Palaeosol matrix was sampled