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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A B S T R A C T

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 redosmorhpy, 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 CaCO3 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.

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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 Uygur 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 Kazakhsthan plate in northwest China (Lou et al., 2000; Ouyang and
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 (Sengör and Natalin, 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 LCs. 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 translated the lacustrine deposits, with some local areal extent and laterally extensive deposits (Yang et al., 2010) that influenced local subsidence patterns and subsequent stratigraphic variations between the sections. Strata within the Triassic Jiucaiyuan and Shaofanggou LOCs are interpreted to be mudflat (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.

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

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 Eurasia, 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).

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 Aslán (1993) and Kraus (1999). Palaeosol and sediment colours were identified from dry samples using Munsell colour charts (Munsell Color, 1975). Palaeosol matrix was sampled...
typically on a decimetre scale. Where present in the profiles, rhizoliths and nodules were sampled.

In this study, palaeosols were identified on the basis of pedogenic features not readily destroyed during diagenesis: 1) organic features, such as rhizoliths, root halos, and root moulds; 2) soil structure, which is recognized by the presence of peds, which are aggregates of soil material that form discrete shapes similar to those found in modern soils, and 3) horizonation associated with pedogenesis (Retallack, 1988, 2001). Horizons were designated using modern soil horizon terminology (O, A, E, B, and C; Table 1). Pedogenic carbonates were
described using the classification of Gile et al. (1966) and Machette (1985). Petrographic features were described using the terminology of Brewer (1964).

Using the assumption that processes that act upon and within modern soils were similar in the past, palaeo-pedogenic features can then be compared to modern analogs to infer ancient soil-forming processes and to interpret changes in soil-forming factors such as climate. Modern soil classification (e.g. Soil Survey Staff, 2006) relies upon the recognition and measurement of physical, chemical, and climatic parameters which often cannot be recognized or accurately measured in ancient soils (e.g. Mack et al., 1993). In this study, palaeosols are classified based on the presence (or absence) of well-developed palaeosol horizons preserved within each profile (Table 1), which can be linked to soil process (e.g. Birkeland, 1999; Buol et al., 1997). In this way, palaeosols have been designated on the basis of horizon types that preserve readily and which reflect processes that elucidate environment. Paleosol profiles have been grouped on the basis of presence of major morphological features and are referred to by their primary diagnostic horizon(s). A taxonomic name from the palaeosol classification scheme of Mack et al. (1993) is attributed to each class of palaeosols in the interpretation (Table 2).

Matrix samples were disaggregated prior to X-ray diffraction analyses by sonicating in dilute Na2CO3 solution. Matrix samples were then centrifuged to separate the $\approx 2\,\mu m$ equivalent spherical diameter (e.s.d.) fraction. Samples of the $<2\,\mu m$ e.s.d. fraction were prepared as oriented aggregates using filter-membranes (Moore and Reynolds, 1997). Each $<2\,\mu m$ sample was prepared at room temperature and analyzed after 1 M KCl saturation, 0.5 M MgCl2 saturation, and solvation with 1:4 glycerol–water solution (Jackson, 1979). After initial analysis, K-saturated samples were heated to 500 °C for at least 2 h and reanalyzed. XRD analyses were performed at Southern Methodist University using a Rigaku Ultima III X-ray Diffractometer.

Table 1: Soil horizons and modifiers relevant to study.

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Characteristics</th>
<th>Climatic/environmental significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>O</td>
<td>Organic matter accumulation</td>
<td>Typically found in low-lying, humid environments that have been water-saturated.</td>
</tr>
<tr>
<td>A</td>
<td>Mineral horizon of eluviation with highest amount of organic matter (unless O horizon present). Usually the uppermost horizon or below O horizon.</td>
<td>Not always present in palaeosols due to erosion of profile tops.</td>
</tr>
<tr>
<td>E</td>
<td>Zone of intense eluviation that removes, e.g. phyllosilicates, iron, aluminum and leaves sand and silt</td>
<td>–</td>
</tr>
<tr>
<td>B</td>
<td>Zone of illuviation and development of soil structure</td>
<td>–</td>
</tr>
<tr>
<td>C</td>
<td>Weakly weathered zone with little soil structure</td>
<td>–</td>
</tr>
</tbody>
</table>

Horizon modifiers

<table>
<thead>
<tr>
<th>Modifier</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>g</td>
<td>Gleyed horizon</td>
</tr>
<tr>
<td>h</td>
<td>Illuvial accumulation of organic matter</td>
</tr>
<tr>
<td>k</td>
<td>Carbonate accumulation greater than that of the parent material</td>
</tr>
<tr>
<td>ss</td>
<td>Slickensides present</td>
</tr>
<tr>
<td>t</td>
<td>Accumulation of phyllosilicates</td>
</tr>
<tr>
<td>w</td>
<td>Soil structure or colour present, but little evidence of illuviation</td>
</tr>
<tr>
<td>y</td>
<td>Accumulation of gypsum</td>
</tr>
</tbody>
</table>

Modifiers specific to the O-horizon

<table>
<thead>
<tr>
<th>Modifier</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>Decomposed organic material</td>
</tr>
<tr>
<td>e</td>
<td>Intermediate decomposition of organic material</td>
</tr>
<tr>
<td>i</td>
<td>Only slightly decomposed organic material</td>
</tr>
</tbody>
</table>
Table 2
Examples of paleosol descriptions from all major morphologies. HIM = hydroxy interlayered minerals, K = kaolinite, I = illite, S = smectite, C = pedogenic chlorite.

<table>
<thead>
<tr>
<th>Pedotype</th>
<th>Horizon depth (m)</th>
<th>Macromorphology</th>
<th>Variation from type profile</th>
<th>Nodule and rhizolith mineralogy</th>
<th>≤ 2 μm clay mineralogy</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;Bss&quot; n=47</td>
<td>Bssc (0 to 0.61)</td>
<td>Sandy mudstone with wedge shaped aggregates and secondary fine to medium angular blocky structure. Contains propagating slick planes. Matrix is 10R 3/3 with vermicular fine-coarse 10R 5/1 and G1 6/10 GY mottles.</td>
<td>Paleosol development of pedogenic features ranges from weak to strong. Wutonggou Bss paleosols largely contain Fe-oxides. Jiucaiyuan Bss paleosols often contain carbonate nodules.</td>
<td>Common cm-sized Fe nodules</td>
<td>HIM, K, I</td>
</tr>
<tr>
<td></td>
<td>Bss (0.61 to 1.50)</td>
<td>Sandy mudstone with weakly defined wedge shaped aggregates. Contains propagating slick planes. Matrix is 5YR 3/3.</td>
<td>Paleosol development of pedogenic features ranges from weak to strong. Lithologies range from conglomerates and sandstones to mudstones. Not all paleosols with &quot;Bb&quot; horizons contain Fe nodules. Redox depletions are present in some &quot;Bb&quot; paleosols. E horizons are only present within a few &quot;Bb&quot; paleosols.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>&quot;Bt&quot; n=28</td>
<td>Bw (0 to 0.07)</td>
<td>Mudstone, massive to fine subangular blocky structure. Matrix is 2.5YR 2.5/1.</td>
<td>Paleosol development of pedogenic features ranges from weak to strong. Lithologies range from conglomerates and sandstones to mudstones. Not all paleosols with &quot;Bb&quot; horizons contain Fe nodules. Redox depletions are present in some &quot;Bb&quot; paleosols. E horizons are only present within a few &quot;Bb&quot; paleosols.</td>
<td></td>
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<tr>
<td></td>
<td>C2Bt (0.07 to 0.50)</td>
<td>Silty mudstone with common granule lithic inclusions. Contain medium-coarse angular blocky structure. Discontinuous argillans coat ped surfaces. Clay-filled, vertically-oriented tubes are present. Matrix is 5Y 4/1.</td>
<td></td>
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<tr>
<td></td>
<td>Btc (0.50 to 0.79)</td>
<td>Mudstone with pebble inclusions. Contains fine to medium angular blocky to fine prismatic structure. Abundant clay and oxide coats are present on peds and lithics. Matrix is 10YR 5/3 with common med.-coarse 10R 3/2 mottles around Fe nodules and contains common 2.5YR 4/6 mottles coarse inclusions.</td>
<td>Common cm sized Fe nodules; oxides increase in abundance upwards</td>
<td></td>
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<tr>
<td></td>
<td>Bt (0.79 to 0.99)</td>
<td>Mudstone contains coarse angular blocky structure. Discontinuous clay skins coat peds. Clay-filled vertically-oriented tubules are present. Matrix is 10R 3/1 with abundant 10YR 4/1 coarse vermicular mottles.</td>
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<tr>
<td></td>
<td>2Bt (0.99 to 1.45)</td>
<td>Muddy, clay-rich granular conglomerate contains lamellar clays and vertically-oriented clay-filled tubules. Matrix is 5YR 3/1 to 10R 3/1.</td>
<td></td>
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<tr>
<td></td>
<td>C (1.45 to 2.11)</td>
<td>Muddy conglomeratic sandstone, massive.</td>
<td></td>
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<tr>
<td>&quot;Bgc&quot; n=19</td>
<td>Bgcm (0 to 0.30)</td>
<td>Sandy siltstone contains weak fine prismatic with secondary fine angular blocky structure. Horizon is indurated. Matrix is G1 4.5/N with common coarse 10R 3/3 mottles.</td>
<td>Some Bgc paleosols contain Bss horizons.</td>
<td>Rare Iron nodules and concretions</td>
<td>HIM, K</td>
</tr>
<tr>
<td></td>
<td>Bgc1 (0.30 to 0.56)</td>
<td>Sandy siltstone contains weak fine prismatic with secondary fine angular blocky structure. Matrix is G1 4.5/N with common coarse 10R 3/3 mottles.</td>
<td></td>
<td>Common iron concretions</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bgc2 (0.56 to 0.96)</td>
<td>Silty mudstone with coarse prismatic structure with secondary fine prismatic and fine angular blocky structure. Matrix is G15/0N with common 10YR 6/8 vermicular mottles.</td>
<td></td>
<td>Common Fe concretions</td>
<td></td>
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<tr>
<td></td>
<td>CBgc (0.96 to 1.36)</td>
<td>Mudstone is G1 5/0N with coarse 10YR 6/8 mottles.</td>
<td>Rare small concretions at top of unit</td>
<td></td>
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</tr>
<tr>
<td></td>
<td>(0.61 to 1.50)</td>
<td>Fissile claystone with platy structure. Contains organic matter inclusions. Weathers to a popcorn texture. Matrix is G1 2.5/10Y with common-abundant fine-medium G1 4/10Y mottles.</td>
<td>Ash layers do not overlie most O horizons. O horizon thickness varies from 5 to 20 cm.</td>
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</tr>
<tr>
<td></td>
<td>Qa (0.86 to 0.91)</td>
<td>Organic Matter with identifiable plant matter. Color is G1 2.5/1N with common fine to coarse 10YR 4/4 mottles.</td>
<td></td>
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</tr>
<tr>
<td></td>
<td>AO (0.91 to 1.11)</td>
<td>Platy mudstone containing unidentifiable plant matter. Matrix is G1 3/1N with fine 5Y 4/3 spherical to vermicular mottles.</td>
<td></td>
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<tr>
<td></td>
<td>A (1.11 to 1.25)</td>
<td>Mudstone is G1 3/1N with 2.5Y 4/3 common fine mottles.</td>
<td></td>
<td>S, HIM, K</td>
<td></td>
</tr>
<tr>
<td></td>
<td>AC1 (1.25 to 1.83)</td>
<td>Massive fine-grained sandy wacke that contains vertically-oriented, organic matter-rich bifurcating tubules. Matrix is 5Y 4/3 matrix.</td>
<td></td>
<td>S, HIM, K</td>
<td></td>
</tr>
<tr>
<td></td>
<td>AC2 (1.83 to 1.91)</td>
<td>Massive fine-grained sandy mudstone that is G1 3/1N with abundant coarse G1 2.5/1N mottles. Contains vertically-oriented dark tubules that are ~3 mm wide.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AC3 (1.91 to 2.20)</td>
<td>Massive fine-grained wacke that contains organic-matter filled vertically-oriented tubules. Matrix is 2.5Y 4/2 with 3Y 3/1 tubular color variation. Fine-grained arenite with flaser bedding. Contains rare organic matter. Matrix is 5Y 5/3.</td>
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<td></td>
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</tr>
<tr>
<td></td>
<td>C (2.20 to 3.28)</td>
<td>Fine-grained sandy mudstone with weakly developed fine to medium angular blocky structure. Discontinuous clay films coat peds. Matrix is 2.5Y 4/1.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>&quot;BtK&quot; n=3</td>
<td>Bt1 (0 to 0.31)</td>
<td>Fine-grained sandy mudstone with weakly developed fine to medium angular blocky structure. Discontinuous clay films coat peds. Matrix is 2.5Y 4/1.</td>
<td></td>
<td>S HIM, K, I</td>
<td></td>
</tr>
</tbody>
</table>

(continued on next page)
with CuKα radiation. Step-scan analyses were run between 2 and 30° 2θ using a step size of 0.04° 2θ and a count time of 2 s. Clay mineral identification is based upon the methods outlined in Jackson (1979) and Moore and Reynolds (1997).

Ninety-six matrix samples were ground to a mean e.s.d. of 100 μm and sent to University of Georgia Center for Applied Isotope Studies and analyzed for major-element composition using a Phillips (Panalytical) PW2420 wavelength-dispersive XRF spectrometer. Results are reported in oxide weight percent in Table 3. Palaeosol carbonate and oxide accumulations were analyzed by reflected- and transmitted-light petrography. Carbonate thin sections were stained following the procedure of Dickson (1965) to determine mineralogy and concentrations of elemental substitutions such as Fe2+ for Ca in calcite.

### 4. Palaeosol morphological analysis

Over 200 palaeosol profiles were excavated, measured and described in the Upper Permian and Lower Triassic strata in this study. Based upon the presence of characteristic pedogenic features, these profiles are divided into nine different groups of palaeosols which encompass the majority of palaeosol morphological variation in these Permian–Triassic strata. These groups of palaeosols are outlined below and are summarized in Table 2. X-ray diffractogram spectra from the B horizons of representative profiles from each group of palaeosols are shown. Clay minerals present include hydroxy-interlayered minerals (HIM), smectite, kaolinite, illite, and chlorite. The number of palaeosol profiles (n = k) is also given in the heading of each palaeosol group.

#### Table 2 (continued)

<table>
<thead>
<tr>
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<th>≤ 2 μm clay mineralogy</th>
</tr>
</thead>
<tbody>
<tr>
<td>“Btk” n = 3</td>
<td>B2 (0.31 to 0.68)</td>
<td>Fine-grained sandy mudstone contains fine to medium angular to subangular blocky structure. Discontinuous clay skins coat ped. Matrix is 5YR 3/1 with abundant vermicular G1 4/10GY mottles.</td>
<td></td>
<td></td>
<td>S, HIM, K</td>
</tr>
<tr>
<td>Bk1 (0.68 to 1.03)</td>
<td>Massive sandy mudstone that is 2.5YR 3/1 with common fine 7.5YR 4/1 mottles.</td>
<td>Vertically stacked stage II carbonate nodules (diameter is 0.5-5 cm); thin calcified vertical tubes.</td>
<td>HIM, S, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bk2 (1.03 to 1.51)</td>
<td>Massive sandy mudstone that is 2.5YR 2.5/2 with common fine vermiculor to spherical G1 4/10GY mottles.</td>
<td>Vertically stacked stage II carbonate nodules (diameter is 0.5-5 cm); thin calcified vertical tubes.</td>
<td>HIM, S, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C (1.51 to 1.94)</td>
<td>Massive granular wackestone, 5YR 4/2.</td>
<td></td>
<td></td>
<td>HIM, S, K</td>
<td></td>
</tr>
<tr>
<td>“O, Bk” n = 5</td>
<td>B (0 to 0.03)</td>
<td>Mudstone with fine angular blocky structure and organic cutans coating ped surfaces.</td>
<td>Most but not all “O, Bk” paleosols contain O horizons that range from 3 to 20 cm thick.</td>
<td></td>
<td>S, HIM, I, K</td>
</tr>
<tr>
<td>B (0.03 to 0.10)</td>
<td>Massive mudstone. 5Y 2.5/1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B (0.10 to 0.23)</td>
<td>Massive mudstone. 5Y 3/2</td>
<td>Ferroan calcite</td>
<td>S, HIM, I, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bk1 (0.23 to 0.43)</td>
<td>Mudstone with fine angular blocky structure and stacked carbonate nodules; root traces 7.5YR 3/2 with secondary 5Y 2.5/1</td>
<td>Ferroan calcite</td>
<td>S, HIM, K, C</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bk2 (0.43 to 1.48)</td>
<td>Mudstone with poorly defined fine angular blocky structure and stacked carbonate nodules. 10YR 4/2.</td>
<td>Vertic features are developed in some “Bk” paleosols. One paleosol contains gley colors and hematite nodules.</td>
<td>Stage I carbonate</td>
<td>HIM, S, I, K</td>
<td></td>
</tr>
<tr>
<td>Bk1 (0 to 0.06)</td>
<td>Mudstone with fine angular blocky to subangular blocky structure, 2.5YR 4/4. Contains calcans, tiny rhizoliths, filamentous carbonate (Stage I).</td>
<td>Stage II carbonate</td>
<td>S, I, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bk2 (0.06 to 0.42)</td>
<td>Very silty mudstone containing massive structure. 2.5YR 4/3. Contains stacked nodules rhizoliths (Stage II)</td>
<td>Stage II carbonate</td>
<td>S, I, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bk3 (0.42 to 0.99)</td>
<td>Mudstone with massive to very weak subangular blocky structure, 2.5YR 4/3. Tubular to nodular carbonate, none are &gt; 1 cm in diameter, 7 cm maximum length</td>
<td>Stage III carbonate</td>
<td>S, I</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bkm (0.99 to 1.12)</td>
<td>Very calcareous lamellar carbonate that is massive to lamellar. 10R 3/3. Stage III carbonate.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ev (0 to 0.07)</td>
<td>Silty matrix that is rich in out sized coarse sand grains. Lamellar texture.</td>
<td>Vertically oriented stacked calcareous nodules, possibly after gypsum.</td>
<td>Common discrete calcareous nodules with internal radiating crystals, most likely after gypsum.</td>
<td></td>
<td>HIm, S, I, K</td>
</tr>
<tr>
<td>By (0.07 to 0.40)</td>
<td>Sandy mudstone with fine angular blocky to massive structure. 2.5YR 4.5/6. Contains vertically oriented stacked calcareous nodules.</td>
<td>Stage II Stacked nodules; areas of ~2 cm long, 1 cm high filled with houndstooth spar in upper 2 cm</td>
<td>HIM, S, I, K</td>
<td></td>
<td></td>
</tr>
<tr>
<td>By (0.40 to 0.53)</td>
<td>Sandy mudstone with mass to weak angular blocky structure. 10R 4.5/6. Contains carbonate nodules with internal radiating crystals, likely after gypsum.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>By2 (0.53 to 0.77)</td>
<td>Massive sandstone that is laminated to bedded, gray</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2C (0.77 to 0.82)</td>
<td>Silt-rich sandstone that is laminated to bedded, gray</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SAMPLE</td>
<td>Section</td>
<td>Stratigraphic height (m)</td>
<td>Paleosol type</td>
<td>Horizon</td>
<td>Na$_2$O (wt.%</td>
</tr>
<tr>
<td>--------</td>
<td>--------</td>
<td>-------------------------</td>
<td>---------------</td>
<td>---------</td>
<td>------------</td>
</tr>
<tr>
<td>TR-249</td>
<td>Tarlong North-SV</td>
<td>532 Calcic Protosol Bk</td>
<td>1.44 2.67</td>
<td>Argillic Protot Bt</td>
<td>0.80 2.28</td>
</tr>
<tr>
<td>TR-226*</td>
<td>Tarlong North-SV</td>
<td>521 Argillisol Bt</td>
<td>1.43 2.11</td>
<td>Argilllic Protot Bt</td>
<td>0.67 1.93</td>
</tr>
<tr>
<td>TR-220</td>
<td>Tarlong North-SV</td>
<td>517 Histosol ACt</td>
<td>1.71 1.78</td>
<td>Argillisol Bt</td>
<td>0.55 1.35</td>
</tr>
<tr>
<td>TR-212*</td>
<td>Tarlong North</td>
<td>515 Calcic Argillisol Bt</td>
<td>0.84 2.18</td>
<td>Calcic Protosol Bk</td>
<td>0.76 2.29</td>
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</tbody>
</table>

S.G. Thomas et al. / Palaeogeography, Palaeoclimatology, Palaeoecology 308 (2011) 41–64
4.1. Group 1 — Palaeosols with “Bss” horizons (n = 47)

4.1.1. Description

“Bss” palaeosols are typically hosted in mudstone or sandy mudstone and are characterized by wedge-shaped aggregates (peds), sub-horizontal propagating slickensides, and clastic dikes that locally extend to a maximum depth of 60 cm from the interpreted palaeosol surface (Figs. 4A and 5a). Such palaeosols range in thickness that locally extend to a maximum depth of 60 cm from the interpreted palaeosol surface (Figs. 4A and 5a). These palaeosols are typically dusky red (10 R 3/3) and contain wedge-shaped aggregate structures, slickensides, and clastic dikes that reach ~0.5 m in length, similar to those described in the Lower Wutonggou LOC. However, these palaeosols are distinct from other “Bss” palaeosols in that they are typically dusky red (10 R 3/3) with light greenish grey (Gley 1/7/10GY) calcareous mottles and Stage II carbonate nodules that reach 2.5 cm in diameter.

4.1.2. Interpretation

Features such as slickensides and wedge-shaped peds in “Bss” palaeosols are common in soils where shrink-swell processes dominate. Clastic dikes crosscut palaeosol horizons, and are interpreted to reflect the infill of deep, vertical cracks that formed during desiccation of the soil profile. The hummock-and-swale structures identified in “Bss” palaeosols are interpreted to be mukkara, the subsurface expression of gilgai, which results from shrink-swell processes in fine clay-rich soils caused by pronounced seasonality in rainfall (Paton, 1974). Redoximorphic features (Vepraskas, 1994), such as gleying of matrix and Fe-oxide nodules are common within “Bss” palaeosols and are characteristic in soils where seasonal fluctuations in rainfall and soil-moisture availability occur. On the basis of these features, “Bss” palaeosols are classified within the Vertisol palaeosol order (Mack et al., 1993).

The dominance of HIM in the clay fraction is unexpected because smectite and 2:1 expandable phyllosilicates usually dominate the clay fraction in modern Vertisols (Buol et al., 1997). However, HIM is known to be a common constituent in some Vertisols (Yerima and van Ranst, 2005), including examples which form upon ephemeral lake plains in Iran (Heidari et al., 2008).

Palaeosols with “Bss” horizons that contain features indicative of strong redoximorphism are classified as gleyed Vertisols (Mack et al., 1993), indicating that although precipitation was seasonal, soils were likely water-saturated for an extended amount of time relative to the

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<table>
<thead>
<tr>
<th>SAMPLE Section</th>
<th>Horizon</th>
<th>Concentration (wt.%)</th>
<th>Na2O</th>
<th>MgO</th>
<th>Al2O3</th>
<th>SiO2</th>
<th>P2O5</th>
<th>K2O</th>
<th>CaO</th>
<th>TiO2</th>
<th>MnO</th>
<th>Fe2O3</th>
</tr>
</thead>
<tbody>
<tr>
<td>07-TR1-14 Tarlong South</td>
<td>Calcic Histic Histosol Bh1</td>
<td>0.94</td>
<td>2.29</td>
<td>14.02</td>
<td>57.01</td>
<td>0.18</td>
<td>2.09</td>
<td>20.06</td>
<td>0.65</td>
<td>21.82</td>
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</tr>
<tr>
<td>07-TR1-15 Tarlong South</td>
<td>Calcic Histic Histosol Bh2</td>
<td>0.91</td>
<td>2.45</td>
<td>15.14</td>
<td>51.52</td>
<td>0.17</td>
<td>2.32</td>
<td>17.17</td>
<td>0.67</td>
<td>17.94</td>
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</tr>
<tr>
<td>07-TR1-16 Tarlong South</td>
<td>Calcic Histic Histosol Bhk</td>
<td>0.53</td>
<td>2.42</td>
<td>14.02</td>
<td>50.91</td>
<td>0.18</td>
<td>2.09</td>
<td>20.06</td>
<td>0.65</td>
<td>21.82</td>
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</tr>
<tr>
<td>07-TR1-17 Tarlong South</td>
<td>Calcic Histic Histosol Bk</td>
<td>1.35</td>
<td>2.48</td>
<td>17.62</td>
<td>64.44</td>
<td>0.27</td>
<td>2.30</td>
<td>23.77</td>
<td>0.76</td>
<td>6.35</td>
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</tr>
<tr>
<td>07-TR1-65 Tarlong South</td>
<td>Calcic Histic Histosol Bk</td>
<td>0.56</td>
<td>1.05</td>
<td>4.98</td>
<td>16.67</td>
<td>0.14</td>
<td>0.88</td>
<td>73.44</td>
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<td>1.27</td>
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</tr>
<tr>
<td>07-TR1-94* Tarlong South</td>
<td>Calcic Histic Histosol Bk</td>
<td>0.50</td>
<td>2.93</td>
<td>17.31</td>
<td>56.91</td>
<td>0.19</td>
<td>2.04</td>
<td>8.07</td>
<td>0.77</td>
<td>12.80</td>
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</tr>
<tr>
<td>07-TR1-14* Tarlong North</td>
<td>Calcic Histic Histosol Bh1</td>
<td>0.94</td>
<td>2.29</td>
<td>14.02</td>
<td>57.01</td>
<td>0.18</td>
<td>2.09</td>
<td>20.06</td>
<td>0.65</td>
<td>21.82</td>
<td></td>
<td></td>
</tr>
<tr>
<td>07-TR1-15* Tarlong North</td>
<td>Calcic Histic Histosol Bh2</td>
<td>0.91</td>
<td>2.45</td>
<td>15.14</td>
<td>51.52</td>
<td>0.17</td>
<td>2.32</td>
<td>17.17</td>
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<td>17.94</td>
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</tr>
<tr>
<td>07-TR1-16* Tarlong North</td>
<td>Calcic Histic Histosol Bhk</td>
<td>0.53</td>
<td>2.42</td>
<td>14.02</td>
<td>50.91</td>
<td>0.18</td>
<td>2.09</td>
<td>20.06</td>
<td>0.65</td>
<td>21.82</td>
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</tr>
<tr>
<td>07-TR1-17* Tarlong North</td>
<td>Calcic Histic Histosol Bk</td>
<td>1.35</td>
<td>2.48</td>
<td>17.62</td>
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<td>07-TR1-65* Tarlong North</td>
<td>Calcic Histic Histosol Bk</td>
<td>0.56</td>
<td>1.05</td>
<td>4.98</td>
<td>16.67</td>
<td>0.14</td>
<td>0.88</td>
<td>73.44</td>
<td>0.30</td>
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<td>07-TR1-94* Tarlong North</td>
<td>Calcic Histic Histosol Bk</td>
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<td>2.93</td>
<td>17.31</td>
<td>56.91</td>
<td>0.19</td>
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<td>8.07</td>
<td>0.77</td>
<td>12.80</td>
<td></td>
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</tr>
</tbody>
</table>

* Standard error for palaeoprecipitation estimates is ±182 mm/year, as calculated by Sheldon et al. (2002).

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Table 3 (continued)
type palaeosol (Table 3; Fig. 4A). “Bss” palaeosols with concentrations of ferric nodules are considered ferric Vertisols or ferric gleyed Vertisols (Mack et al., 1993). Formation of iron-manganese nodules and concretions is common in Vertisols because of their high clay content and imperfect drainage, which can lead to episodic, often seasonal, changes in moisture content and concomitant availability of oxygen within the soil profile (Buol et al., 1997; Stiles et al., 2001).

Palaeosols containing Bssk horizons (containing carbonates as well as slickensides) are found in the Upper Wutonggou, Jiucaiyuan, and Shaofanggou LOCs and are classified as calcic Vertisols (Mack et al., 1993). Modern Vertisols that accumulate pedogenic carbonate occur where annual evapotranspiration exceeds precipitation (Driese et al., 2005; Gouldie, 1973, 1983; Nordt et al., 2006). This, however, need not imply extremely arid conditions. A recent study of modern and ancient carbonate-bearing Vertisols demonstrated that annual precipitation in such soils could be as great as 1400 mm/year (Nordt et al., 2006). Modern Vertisols are commonly found on landscapes where precipitation ranges from 250 to 1000 mm/year and occurs seasonally (Buol et al., 1997).

4.2. Group 2—Palaeosols with “Bt” horizons (n = 28)

4.2.1. Description

“Bt” palaeosols are hosted in an array of lithologies including mudstone, coarse sandstone, and muddy granular conglomerates. Pedogenic structures in these palaeosols include massive, fine to coarse angular blocky structure, and fine to coarse prismatic structure (Figs. 4B and 5b). Thicknesses of palaeosols with Bt horizons range from 30 to 370 cm. All “Bt” palaeosols contain horizons that preserve evidence for clay translocation and accumulation, such as clay-filled tubules, discontinuous and thin to continuous and thick argillans, and lamellar clays (Fig. 6b). These palaeosols may also contain organic matter- and hematite-filled tubules. These tubules, which range in length from approximately 2 to 10 cm, are oriented subvertically within the palaeosol profiles and commonly bifurcate.

“Bt” palaeosols vary in colour, but are typically brown (10YR 5/3), dark grey (5YR 4/1), reddish black (2.5YR 2.5/1), and dark reddish grey (10R 3/1) with common red (2.5YR 4/6) mottles. Although rare, some “Bt” palaeosols are very dark grey (Gley1 3/N). Ferric nodules as large as ~3 cm in diameter, concretions, and iron-rich ped coatings are common in palaeosols with “Bt” horizons. Palaeosols with “Bt” horizons rarely include horizons that contain propagating slickensides. HIM dominate the <2 μm e.s.d. fraction. Petrographic analysis indicates that skelsepic and masepic soil fabrics dominate (Brewer, 1964) and highly oriented clay laminae exhibit birefringent textures along channels (Fig. 6b). Palaeosols with “Bt” horizons typically occur within stratigraphic intervals that include a significant sandstone or conglomeratic component.

Most palaeosols with “Bt” horizons occur within the Wutonggou LOC, although one is found above the perceived Permian–Triassic boundary at Tarlong North. One “Bt” palaeosol has been recognized in the Upper Wutonggou LOC at Tarlong South. No “Bt” palaeosols were described at Taodonggou East or Taodonggou West.

4.2.2. Interpretation

Clay-filled tubules, thin and discontinuous to thick and continuous argillans, and lamellar clays are the prominent feature of palaeosols with “Bt” horizons, and these features are most likely formed due to illuviation of layer-lattice clay material. Because the most obvious feature of “Bt” palaeosols are horizons which contain evidence for
accumulation of clay-size material, they are classified as Argillisols (Mack et al., 1993), and likely correspond to Alfisols or Ultisols in USDA soil taxonomy (Soil Survey Staff, 2006). In modern environments, Alfisols are found typically in moist to sub-humid regions with moderate to cool soil temperature regimes whereas Ultisols are found generally in areas with annual average soil temperature ≥8° and where seasonal leaching occurs, or where pedogenesis occurs within an acid parent material (Buol et al., 1997). A prerequisite to the formation of Bt horizons is free drainage (Soil Survey Staff, 2006), suggesting that palaeosols with “Bt” horizons in the Upper Permian

Fig. 5. Field photographs. a. “Bss” palaeosol from Tarlong North. Pencil (13.2 cm long) for scale. b. “Bt” palaeosol from Tarlong North. Note the well-developed prismatic soil structure and incipient nodules. Pick head is ~19 cm in length. c. “Bgc” palaeosol from Tarlong North. Note the Fe-pisoids that are ~1 cm in diameter. d. “O” palaeosol from Tarlong North. The overlying bed is composed of possible volcanic ash. Pick is 62.5 cm in length. e. “Bk” palaeosol from Taidonggou West. Carbonate rhizoconcretions are prominent. Rock hammer, 28 cm in length, for scale. f. “By” palaeosol. Carbonate nodules are spherical and consist of coarse carbonate spar and houndstooth spar. Head of rock hammer for scale, 12 cm in length.
strata developed upon well-drained portions of the landscape, above the position of the local groundwater table.

The clay mineral assemblage within this group of paleosols does not show great compositional variance with depth, indicating that clay minerals may be largely inherited, i.e., depositional. Pedo-petrogenetic clay accumulations such as these indicate that leaching is the dominant process by which clay materials accumulated within these “Bt” soils (Dijkerman et al., 1967), and because “Bt” paleosols were developed primarily in strata that contain a sandstone or conglomeratic component, the coarser-grained material likely acted as a sieve for translocated clays within the soil horizon.

Where gleyed colours and mottling are present, paleosols with “Bt” horizons are classified as gleyed Argillisols (Mack et al., 1993) and their presence indicates that seasonal groundwater saturation was an important factor in the morphological development of these profiles. “Bt” paleosols that contain iron nodules are classified as ferric Argillisol. The formation of iron nodules, pisoliths, and tubules in soils is related to wetting and drying, which is often seasonal in occurrence.

Fig. 6. Matrix and nodule photograph and photomicrograph. a. Matrix from a Bs horizon with asepic to weak insepic fabric. Scale bar is 200 μm. b. Matrix from a Bt horizon with skelepic and insepic soil fabrics. Scale bar is 2 mm. c. Matrix from a BtG horizon. Note zones where iron oxide has been removed. Scale bar is 500 μm. d. Nodule found within “BtG” horizon. Core is composed of hematite whereas outer bands are composed primarily of goethite. Calcite cement is present. Scale is 2 mm. e. Matrix from a Btk horizon with skelepic and insepic soil fabrics. Scale bar is 2 mm. f. Nodule from a “Btk” horizon containing radial calcite texture and circumgranular cracks. Scale bar is 2 mm. g. Matrix with asepic soil fabric from a Bk horizon associated with an “O”, “Bk” paleosol. Scale bar is 2 mm. h. Microspar and spar dominated nodule from a Bk horizon associated with an “O”, “Bk” paleosol. Scale is 2 mm. i. Discrete organic matter found within carbonate nodule from Bk horizon of “O”, “Bk” paleosol. Dark bar next to sample is 2 cm. j. Carbonate nodule from Bk horizon associated with “Bk” paleosol. Note organic-matter along void and radial fibrous calcite texture. Structures composed of radial fibrous calcite are interpreted as Microcodium. Scale bar is 1 mm. k. Matrix from an Ev horizon associated with a “Ev” paleosol. This horizon is composed primarily of fine to coarse sand-sized particles within a silty matrix. Scale bar is 2 mm.
and is typical of cool, humid environments (Bhattacharyya and Kakimoto, 1982; Cornell and Schwertmann, 2003).

“Bt” palaeosols that contain feldspars and other unstable grains are interpreted as eutric Argillisols, whereas those that contain few unstable grains are interpreted to be dystric Argillisols (sensu Mack et al., 1993). Some “Bt” palaeosols can be considered dystric ferrigleyed Argillisols. Although classification of an Argillisol as eutric or dystric can assist in interpreting base status within the soil profile and possibly the degree of soil weathering, parent material must be taken into consideration, too. Here, “Bt” palaeosols formed on lithic- and feldspar-rich sandstones, conglomerates, and wackes. The variations among palaeosols with respect to the stability of the grains seem, at least in this case, to be largely controlled by the composition of the soil parent material.

4.3. Group 3—Palaeosols with “Bgc” horizons (n = 19)

4.3.1. Description

Palaeosols with “Bgc” horizons contain generally weakly developed soil structure, but where present soil structure includes fine angular blocky peds, fine prismatic peds, wedge-shaped aggregate structures, and fine platy structure. Palaeosols with “Bgc” horizons occur within mudstone and sandy mudstone lithologies (Figs. 4C and 5c) and range in thickness from 59 to 529 cm, the thickest representing a cumulate palaeosol (sensu Marriott and Wright, 1993).

Common matrix colours in “Bgc” palaeosols are (1) dusky red (2.5YR 3/2) matrix with dark greenish grey (Gley1 4/1 10Y) common vermicular mottles and (2) dark grey (Gley1 4.5/0N) matrix with common dusky red (10R 3/3) to brownish yellow (10YR 6/8) vermicular mottles. Zones of preferential iron-oxide removal are common within these palaeosols (Fig. 6c). One “Bgc” palaeosol contains an horizon with evidence for accumulation of translocated clay-size material, or a Bt horizon. Fossil organic matter is also commonly distributed throughout the profiles of palaeosols with “Bgc” horizons, but the organic matter components are not identifiable. The <2 μm e.s.d. fraction is composed primarily of HIM, with less abundant kaolinite. Most “Bgc” palaeosols contain Fe-oxide pisoids and nodules with maximum diameters of 4 cm (Fig. 6c, d). Fe-oxide nodules and pisoids in “Bgc” palaeosols are composed primarily of hematite and goethite, less abundant kaolinite and quartz, and some trace calcite. Fe-oxide pisoids are internally arranged into irregular sub-millimetre concentric rings that terminate against each other (Fig. 6d). Slickensides and wedge-shaped peds are commonly present.

Palaeosols with “Bgc” horizons are concentrated in the lower 335 m of the Wutonggou LOC at Tarlong North (Fig. 7). At Tarlong South, however, a “Bgc” palaeosol occurs within the upper Wutonggou LOC (Fig. 8). Correlative strata at Tarlong North contain palaeosols interpreted as well-drained (see Figs. 3, 8, and 9). This difference in observed palaeosol morphologies may be related to variations in landscape position between the sections. In the Taodonggou West stratigraphic section, “Bgc” palaeosols have been identified only in the Wutonggou LOC below 110 m, and no “Bgc” palaeosols have been described from the Taodonggou East section.

4.3.2. Interpretation

The dominant features of “Bgc” palaeosols are low chroma matrix, colour motting, and Fe-oxide nodules and pisoids. Soil redoximorphic conditions likely generated these features (Richardson and Vepraskas, 2000; Vepraskas, 1994), and thus, these palaeosols are classified as Gleysols (Mack et al., 1993). The formation of iron oxide nodules and pisoids are likely the result of variability in moisture availability and water chemistry and is typical of cool, humid, redoximorphic environments (Bhattacharyya and Kakimoto, 1982; Cornell and Schwertmann, 2003). The concentric rings of Fe-cements within
pisoids might result from wet–dry periodicity, possibly at the annual scale (Cornell and Schwertmann, 2003). Vertic features occur in some “Bg” palaeosols, which further implicate the effect of seasonal variations in soil-moisture availability (see Palaeosols with “Bss” Horizons) in contributing toward the morphology of these palaeosols.

4.4. Group 4—Palaeosols with “O” horizons (n = 13)

4.4.1. Description

Palaeosols with “O” horizons are characterized by in situ accumulation of vascular plant organic matter in the uppermost horizon (Fig. 4D). These O horizons are coals, and they typically overlie gray to buff mudrock, and many are stratigraphically associated with intervals that include volcanic tuff. The amount of identifiable vascular plant matter varies among “O” palaeosols, but encompasses the spectrum from predominantly unidentifiable (Oa horizon) to ~50% identifiable (Oe horizon) to ~50% identifiable (Oi horizon). Thicknesses of these palaeosols range from 22 to 254 cm, including O horizons and underlying horizons. O horizons are typically 5 to 20 cm thick. Rarely, palaeosols with “O” horizons include an underlying Bss horizon, but slickensides do not cross into the overlying O horizon. Most palaeosols with “O” horizons contain underlying mineral horizons that exhibit platy to massive soil structure, and were probably A or AC horizons (Figs. 4D and 5d). The <2 μm e.s.d. fraction is composed primarily of HIM with less abundant illite and kaolinite.

Although palaeosols with “O” horizons occur predominantly within the Wu tonggou LOC, they appear above the Permian–Triassic Boundary at Tarlong North (Fig. 7). A single palaeosol with an O horizon occurs in the Tarlong South stratigraphic section (Fig. 8), and three occur in the Taodonggou West stratigraphic section (Fig. 9). No “O” palaeosols occur in the Taodonggou East stratigraphic section (Fig. 10).

4.4.2. Interpretation

Based on the presence of an O horizon, these palaeosols are interpreted as Histosols (Mack et al., 1993). Profiles that contain an underlying Bss horizon are classified as vertic Histosols. In general, “O” palaeosols are interpreted to reflect a soil profile that is inundated with water, i.e., an everwet environment (aquic soil moisture regime; Soil Survey Staff, 2006). In modern environments, similar soils are found in the lowest-lying, wettest areas (Buol et al., 1997). Histosol O horizons are generally thinner when forming in warm and wet climates and thicker (>1 m) in cool and less humid climates (Schaeetzl and Anderson, 2005). The present thickness of O horizons in “O” palaeosols never exceeds 1 meter, so “O” palaeosols could be interpreted to have formed within a warm and wet climate. However, when considering a compaction ratio of 7:1 for ancient coal seams (Ryer and Langer, 1980), the thicknesses could indicate formation within either setting: warm and wet climate or cool and slightly drier climate, at least in terms of relative humidity.

The rare presence of slickensides below the O horizon in some “O” palaeosols indicates that pedogenesis occurred in an environment of changing soil moisture and, in this instance, soil moisture likely increased as pedogenesis progressed. This “O” palaeosol variant, the vertic Histosol, most likely reflects shallowing of the groundwater table, flooding of a pre-existing “Bss” palaeosol, and subsequent Histosol (O horizon) development after inundation of the underlying mineral horizons.

4.5. Group 5—Palaeosols with “Btk” horizons (n = 3)

4.5.1. Description

Palaeosols with “Btk” horizons are typically hosted in fine-grained sandy mudstone or mudstone. Two major features characterise these palaeosols: layers with calcium carbonate nodules and layers with evidence for translocation and accumulation of clay materials. Pedogenic structure in “Btk” palaeosols includes fine to medium angular blocky peds, often with less abundant fine prismatic structure. One “Btk” palaeosol also contains sub-horizontal propagating slickensides and wedge shaped aggregates within its uppermost horizon, flooding of a pre-existing “Bss” palaeosol, and subsequent Histosol (O horizon) development after inundation of the underlying mineral horizons.

Fig. 8. Tarlong South stratigraphic section. See Fig. 7 for key.
Matrix colours in palaeosols with Btk horizons are olive (5Y 5/3), olive grey (5Y 4/2 to 5Y 5/2), very dark grey (5YR 3/1), weak red (2.5 YR 4/1), and dark reddish grey (2.5 YR 3/1) in layers with argillans and very dusky red (2.5 YR 2.5/2) in layers that contain carbonate without argillans. The <2 µm e.s.d. fraction is primarily smectite with less abundant HIM and kaolinite. Carbonate nodules from “Btk” palaeosols are composed of micrite, microspar, and radial fibrous calcite that occludes detrital grains and clay-size materials (Fig. 6f). These nodules also exhibit circumgranular cracks (Fig. 6f). “Btk” palaeosols are associated with sandstone, carbonates and mudrocks that are interpreted as lake-plain and lacustrine deposits (Yang et al., 2007, 2010). Palaeosols with “Btk” horizons have been described only at Tarlong North (Fig. 7) in the mid-upper Wutonggou LOC and the Jiucaiyu LOC.

4.5.2. Interpretation

Based on the prominence of clay accumulation surrounding peds, and the presence of carbonate nodules in the lower horizons, “Btk” palaeosols are classified as calcic Argillisols (Mack et al., 1993). The textural characteristics of “Btk” carbonate nodules include micrite and circumgranular cracks, which are common alpha textures within pedogenic carbonates, but the nodules also contain radial fibrous calcite that is not consistent with typical pedogenic textures (Wright and Tucker, 1991). This phase could reflect closed-system
crystallization of calcite in void spaces of lacustrine deposits, prior to or after pedogenesis. Alternatively, the radial fabrics appear similar to those produced biologically by organisms such as Microcodium (Kabanov et al., 2008; Kosir, 2004; Wright and Tucker, 1991), and could thus reflect biologically-mediated crystallization of calcite during pedogenesis.

A few scenarios for the formation of “Btk” palaeosols can be postulated: (1) carbonate crystallization occurred during earlier lacustrine deposition and accumulation of clay-size material occurred later in well-drained, but relatively moist soil conditions, after the lake system withdrew from the lake plain; (2) accumulation of clay-size materials occurred in well-drained, but relatively moist soil conditions, and carbonate formation occurred during wet periods where the groundwater table inundated the lower portion of the soil; (3) carbonate formation occurred at first within a relatively dry soil moisture regime that allowed for the precipitation of carbonate, and was later overprinted by accumulation of clay-size materials that occurred in a well-drained, but relatively moist, soil environment; and (4) carbonate accumulated within the lower B horizons of a well-drained soil at the same time that clay accumulated in the overlying B horizons. Scenario (4) implies that carbonate nodules resulted from incomplete leaching of Ca$^{2+}$ ion from the soil profile, but it also requires environmental conditions that are more permissive of both Microcodium-produced radial-fibrous calcite textures and circumgranular cracking within a single soil-forming environment. Scenario (4) is similar to modern soils with Btk horizons, which form in xeric, or seasonal, soil moisture regimes (Soil Survey Staff, 1976) and is our preferred interpretation.

### 4.6. Group 6—Palaeosols with Stacked “O and Bk” horizons (n = 5)

#### 4.6.1. Description

“O, Bk” palaeosols are distinct in that they are characterized by not only layers of in situ accumulation of fossil plant matter (O horizons), but also by 7 to 45 cm thick horizons of gray to buff mudstone, sandy mudstone and fine-grained sandstone with calcium carbonate nodules (Bk horizons; Fig. 4E). These palaeosols range in total thickness from 45 to 148 cm. Carbonate in the Bk horizons includes stacked nodules and downward-tapering and bifurcating elongate tubules. Organic matter is concentrated within these Bk horizons as bifurcated, vertically oriented tubules and as coatings along ped surfaces (organans; Brewer, 1964).

O horizons are black, whereas Bk horizons range in colour from dark brownish grey (10YR 4/2) to dark grey (5Y 4/1.5) where carbonate is present and where no carbonate is found, the B horizon matrix is very dark grey (5Y 3/1). Pedogenic structure includes poorly to well-developed fine angular blocky peds in most B horizons, and massive to platy structure within O horizons. The <2 μm e.s.d. size fraction is composed primarily of HIM with less abundant smectite and trace illite and kaolinite. Mineral horizons are characterized by aseptic soil fabrics (Fig. 6g), and carbonate nodules contain micrite and microspar with cross-cutting sparry calcite that does not appear to define circumgranular cracks (Fig. 6h). Iron substitution within the calcite structure has been identified within the micritic and microspar textures through staining techniques. Micrite and microspar sometimes occlude identifiable vascular plant matter or bifurcating carbonized rootlets (Fig. 6i).

“O, Bk” palaeosols occur only within the uppermost Wutonggou LOC. They occur at 958 and 991 m in the Tarlong North stratigraphic section (Fig. 7), and they occur at 0.5, 1.3 and 42 m in the Tarlong South stratigraphic section (Fig. 8). These palaeosols are likely stratigraphically equivalent. No “O, Bk” palaeosols have been described at either Taodonggou section.

#### 4.6.2. Interpretation

The stacked carbonate nodules and downward-tapering CaCO$_3$ tubules are interpreted as rhizoliths that correspond to Stage II carbonate development (Gile et al., 1966; Machette, 1985), and horizons where these features are found are considered to be Bk horizons. Generally, calcic horizons in modern environments form in sub-humid to arid conditions where rainfall is less than 700 to 800 mm/yr (Birkeland, 1999; McFadden, 1988; Retallack, 2005).
Organic matter that fills bifurcating, vertically oriented tubules is interpreted as organic-matter filled roots. Organans are present and result from illuvial transport of organic matter. The ~20-cm-thick horizon of identifiable organic matter is considered to be an O horizon, which in modern environments is usually found in low-lying areas where precipitation is >1000 mm/year (Buol et al., 1997).

The most important pedogenic processes in the formation of “O, Bk” palaeosols resulted in the creation of alternating layers of pedogenic carbonate accumulation and organic matter accumulation. Thus, these palaeosols are considered to be carbonate rich Calcisols where calcium carbonate accumulation is the dominant feature and organic matter accumulation is limited to organans or a thin O horizon. Where the O horizon is the prominent feature, these palaeosols are classified as calcic Histosols (Mack et al., 1993). These palaeosol profiles are peculiar, as (1) surficial accumulation of organic matter, illuvial transport of organic matter and (2) subsurface accumulation of carbonates are generally exclusive processes in soil-forming environments. In this respect, the processes responsible for formation of palaeosols with stacked “O, Bk” horizons are somewhat enigmatic.

Multiple mechanisms are plausible for the co-occurrence of organic matter and calcium carbonate within stacked “O, Bk” horizons: (1) these are two distinct palaeosols and first, a calcic soil developed that was later flooded, altering the hydromorphic character of the soil profile and permitting the deposition of organic matter; or (2) the carbonate nodules do not reflect precipitation within a well-drained soil but rather formed in situ in a poorly-drained, closed system environment.

The presence of carbonate nodules within these profiles should not be used as an indication of semi-arid or arid climate. Micritic nodules have been found even in regions where mean annual precipitation is ~1500 cm (Aslan and Autin, 1998; Farrell, 1987). Carbonate nodule formation apparently requires episodic drying of the soil (Kraus and Hasiotis, 2006), and therefore the presence of these nodules could represent seasonal or periodic precipitation patterns.

Substitution of ferrous iron for calcium within calcite indicates that the second mechanism may be more reasonable. Low Mg–calcite is the chief carbonate mineral that precipitates within palaeosol profiles in open communication with the atmosphere (i.e. well-drained conditions) (Langmuir, 1997). Incorporation of Fe\(^{2+}\) into the crystal lattice would require reducing conditions, which generally prevail within organic matter-rich deposits (McCabe and Parrish, 1992; Shotyk, 1992), which could be manifested through a seasonally high water table.

In many organic-rich soils, especially bogs, acid conditions prevail via chemical weathering and oxidation of organic compounds (Shotyk, 1992). Carbonate precipitation, however, occurs under alkaline conditions so in these profiles, soil pH would be ~8.3 (Langmuir, 1997). Although this value is high for many organic soils, the characteristics of palaeosols with “O, Bk” horizons are consistent with those found in saline fens (Shotyk, 1992). Fens are considered minerotrophic in that they are fed by groundwaters rather than directly by precipitation (Gore, 1983). Fens are found typically in sub-humid climates with a notable dry season (Retallack, 2001), are in environments that can attain alkaline pH values, and are a plausible analog for palaeosols with “O, Bk” horizons.

4.7. Group 7—Palaeosols with “Bk” horizons (n = 23)

4.7.1. Description

Palaeosols with “Bk” horizons are typically hosted in mudstone or sandy mudstone. They are 10 to 225 cm thick, and characterized by massive, fine subangular, and fine angular blocky soil structure (Fig. 4C). The most diagnostic feature of these palaeosols is accumulation of calcium carbonate (Fig. 5e). Uppermost horizons tend to contain calcans, which are calcium carbonate coatings on ped surfaces (Brewer, 1964), and fine filamentous carbonate corresponding to Stage I carbonate accumulation (Gile et al., 1966). Underlying horizons typically contain vertically stacked carbonate nodules and tubular carbonate structures that reach up to 40 cm in length and >1 cm in diameter (Stage II; Gile et al., 1966). Rarely, the lowermost horizon consists of very calcareous lamellar carbonate, corresponding to a Bkm horizon (Stage III; Gile et al., 1966).

Matrix colours are typically reddish brown (2.5YR 4/3) and weak red (10R 4/3). Some “Bk” palaeosols contain abundant light greenish grey (G1 8/10Y) mottles that are ~1 cm in diameter or occur as thin (<1 cm thick) stratiform mottles that are ~<40 cm in length. Rarely, these palaeosols also contain iron-oxide inclusions within some carbonate nodules and discrete hematite nodules. Several examples of palaeosols with “Bk” horizons contain layers with wedge shaped aggregates, slickensides (Bksh horizons) and clastic dikes. HIM and smectite dominate the <2 μm e.s.d. size fraction in the uppermost horizon of this group of palaeosols with a “Bk” horizon, whereas the <2 μm e.s.d. size fraction in the lower horizons are dominated by smectite with less abundant kaolinite and illite. Carbonate nodules contain both alpha- and beta-fabrics (Wright and Tucker, 1991), and include features such as circumgranular cracks, domains of micrite and microspar, skeleton grains, and rhombohedral calcite spar cement, as well as radial fibrous calcite cement (Fig. 6j) that appears similar to Microcodium (Klappa, 1978).

“Bk” palaeosols are scarce within the Wu tonggou LOC among all stratigraphic sections (Figs. 7–10), but have been noted, particularly at Tarlong North, where they have been found in the middle to upper Wu tonggou LOC. In these instances, they are commonly in close stratigraphic association with thin, lacustrine limestone beds. Palaeosols with “Bk” horizons are common (1) in the upper 30 m of the Tarlong South stratigraphic section, within the Jucai yuan LOC (Fig. 8), (2) throughout the Jucai yuan and Shao fanggou LOCs in the Taodonggou West stratigraphic section (Fig. 9), and (3) from 65 to 163 m in the Jucai yuan Formation in the Taodonggou East stratigraphic section (Fig. 10). Palaeosols with “Bk” horizons are abundant in strata near the assumed position of the Permian–Triassic Boundary.

4.7.2. Interpretation

Vertically stacked carbonate nodules are interpreted as rhizoliths, and based on the prominence of calcium carbonate nodules, “Bk” palaeosols are classified as Calcisols (Mack et al., 1993). In modern environments, these soils form in sub-humid to arid environments where evaporation exceeds precipitation, and in the modern environments, precipitation is generally <700 mm/year (Birkeland, 1999; McFadden, 1988).

The presence of grey colours, iron nodules and vertic features in a few of these palaeosols suggests that they experienced fluctuation in soil drainage, moisture availability, and redoximorphic conditions during soil formation. These “Bk” palaeosol profiles are classified as vertic, gleyed Calcisols (Mack et al., 1993). These palaeosols, however, were probably better drained than gleyed Argillisols, gleyed Vertisols, and Gleysols which contain nodules composed of both goethite and hematite. This interpretation reflects that hematite precipitation preferentially occurs in environments with lower moisture than those environments in which goethite precipitates (Kampf and Schwertmann, 1982). With the exception of the noted variants, “Bk” palaeosols are interpreted to have formed under well-drained soil conditions upon stable lakeplain and floodplain sites.

4.8. Group 8—Palaeosols with “By” horizons (n = 4)

4.8.1. Description

Palaeosols with “By” horizons range from 24 to 82 cm in thickness, and are developed in silty sandstone and sandy mudstone (Fig. 4H). “By” palaeosols are characterized by vertically stacked nodules up to 2 cm in diameter that correspond to Stage II carbonate development
(Fig. 5f). Palaeosol matrix colours are red (10R 4.5/6 to 2.5YR 4.5/6). B horizons contain fine angular blocky pedds. In the type palaeosol, a coarse sand-rich lamellar horizon overlies the B horizon (Fig. 6k). The ~2 μm e.s.d. size fraction is composed primarily of smectite and HIM, with less abundant illite and kaolinite.

Carbonates in “By” palaeosols differ from “Btk” and “Bk” palaeosols with respect to texture: rather than consisting of micrite and microsparite, these nodules contain coarse calcite spar (houdtsoodt spar) and calcite that contains an internal radiating fabric that appears to be a permineralized replacement of fibrous gypsum crystals (Fig. 6i). Palaeosols with “By” horizons are found only in the Jiucaiyuan and Shaofanggou LOCs in the Taodonggou West (Fig. 9) and Taodonggou East (Fig. 10) stratigraphic sections.

4.8.2. Interpretation

Based on the presence of carbonate within “By” palaeosols, they could be classified as Calcisols (Mack et al., 1993), but the internal fabric of the nodules differs from the typical fabric associated with “Btk” and “Bk” palaeosols and their modern equivalents. Rather, the fabrics seen here are more commonly associated with replacement (e.g. Retallack et al., 2003; Tabor et al., 2007). We interpret the calcium carbonate within nodules from “By” palaeosols to be mostly diagenetic, and based on the common internal radiating fabric, the calcite has most likely replaced gypsum. Therefore, we consider these horizons to have been layers that accumulated gypsum or gypsic horizons (By) during Permian–Triassic time. The sandy layers overlying the By horizons may be eluvial horizons (Ev horizons) from which fine particles were dispersed and translocated. These palaeosols are thus classified as Gypsisols (Mack et al., 1993). Soils that accumulate gypsum in By horizons form in variably drained landscapes with arid climate, normally with <200 mm of precipitation per year (Watson, 1992). Given that we recognize eluvial horizons above By horizons, we interpret these soils to have developed upon stable portions of the landscape, with free drainage in an arid climate. Alternative interpretations, however, will be addressed below.

4.9. Group 9—Palaeosols with “AC, Bw” horizons (n = 104)

4.9.1. Description

Palaeosols with “AC, Bw” horizons are the most common palaeosols in the Permian–Triassic strata in the study area. Although these palaeosols are widely distributed throughout the Wutonggou, Jiucaiyuan, and Shaofanggou LOCs, their discriminating characteristics are limited to those that form the basis of palaeosol identification: soil structure, evidence for rooting, and horizonation. These palaeosols do not contain additional accumulations or features that would require a modifier other than Bw (soil structure with the absence of other features). Palaeosols with “AC, Bw” horizons show a great degree of variability in terms of their matrix colours and lithologies. Thicknesses vary from 5 to 478 cm. Due to the variability found in palaeosols with “AC, Bw” horizons, no type palaeosol has been designated for this group.

4.9.2. Interpretation

The main features of “AC, Bw” palaeosols are morphologies that distinguish them from non-pedogenically altered rock. Because these palaeosols are typically poorly-developed or contain no features to indicate processes other than general soil formation, they are classified as Protosols (Mack et al., 1993), which correspond to Entisols and Inceptisols using the USDA soil taxonomy (Soil Survey Staff, 2006). These palaeosols indicate pedogenesis, and although they can be used to understand depositional rates and duration of pedogenesis, Protosol morphologies provide little information regarding paleoclimate. They provide, however, evidence of depositional diastems and subaerial exposure in depositional environmental and sequence stratigraphic interpretations (Yang et al., 2007). Furthermore, their elemental composition may still provide insight into palaeweathering conditions and parent material composition.

5. Results of elemental analysis

Many studies have focused on employing empirical relationships to understand connections between soil properties, such as soil elemental chemistry, and soil-forming processes. Sheldon et al. (2002) delineated relationships between mean annual precipitation (MAP) and the chemical index of alteration minus potassium (CIA-K) and between salinization (S) and mean annual temperature (MAT). Elemental data from palaeosol profiles in the Bogda Mountains is presented in Table 3. Elemental data are listed for all sampled palaeosols, although only those that are interpreted to have formed in well-drained environments were considered for palaeoprecipitation reconstruction. The uppermost B horizon was used to calculate MAP (Sheldon et al., 2002). In general, this approach requires that palaeosol samples with >10% CaO be eliminated from consideration for the proxy method of estimating precipitation because such samples yield estimates that are too low (Kahmann and DiRiese, 2008; Sheldon et al., 2002). Due to the paucity of palaeosols in the uppermost Wutonggou LOC and in the Jiucaiyuan and Shaofanggou LOCs, these palaeosols have been included for analysis, although some caution must be exercised in the interpretation of the results. This approach yields time-averaged results, and thus variations in temperature or precipitation that occurred at timescales less than the duration of pedogenesis cannot be discerned (Kahmann and DiRiese, 2008).

In addition to the suitability of palaeosols for elemental analysis, the significance of the data must be considered. Sheldon et al. (2002) calculated the R² of the relationship between palaeoprecipitation and CIA-K to be 0.72. The importance of other soil-forming factors cannot be neglected, including other climate parameters like temperature (Rasmussen and Tabor, 2007). Parent material could be a particularly important factor influencing CIA-K values. For example, volcanic clasts are observed within the Wutonggou LOC and are less abundant in the Jiucaiyuan and Shaofanggou LOCs as sedimentary clasts become more abundant (Guan, personal communication). If significant provenance changes occurred, it would be expected that palaeosol elemental compositions could reflect changes in parent material provenance rather than changes in weathering.

CIA-K values of all sampled horizons range from 2.3 to 84.7. MAP is calculated using the equation: P = 221e^{0.187(CIA-K)} with a standard error of ±182 mm/year (Sheldon et al., 2002). In general, most of the calculated values for MAP are consistent with values for modern MAP for each palaeosol order. Calculated MAP values for all palaeosols designated suitable for this method range from 230 to 1270 mm/year.

Salinization, which is given by the molar ratio of [Na₂O+K₂O]/[Al₂O₃] has been weakly correlated with mean annual temperature, with a standard error of ±4.4 °C (Sheldon et al., 2002). By applying the salinization equation in Sheldon et al. (2002; their Eqs. 5) to the sampled Vertisols, Argillisols, calcic Argillisols, Protosols, Calcisols, and Gypsisols, MAT estimates range from 7 to 15 °C. With the exception of a zone of instability in estimated palaeoprecipitation noted ~1000 m in the composite section (Fig. 11), no significant stratigraphic changes or trends in temperature are observed across the Permian–Triassic boundary.

6. Stratigraphic synthesis

Permian–Triassic palaeosols from Tarlong and Taodonggou sections clearly demonstrate stratigraphic trends in morphology and in their mineralogical and geochemical compositions. Introduction of new palaeosol morphologies and disappearance of other morphologies through the stratigraphy and across the landscape reveal significant changes in soil-forming state factors (climate, organisms, landscape
Fig. 11. Composite diagram featuring paleosol morphologies as indicated by colour (denoted in key in Fig. 7). Hypothesized soil moisture regime is indicated (as based on paleosol morphology). Calculated palaeoprecipitation and palaeotemperature are included, on the basis of estimates derived from the CIA-K proxy. Values have a standard error of ±182 mm/year and ±4.4 °C for precipitation estimates (Sheldon et al., 2002).
position, parent material, and/or time; Jenny, 1941) and larger-scale processes, such as basin- to regional-scale tectonics, that occurred throughout the Late Permian and earliest Triassic in this region.

The overall depositional system remains fluvial to lacustrine throughout sections. In a general sense, the Wutonggou LOC is dominated by deltaic, lakeplain, and littoral lacustrine deposits and fluvial deposits. The Jiucaiyuan LOC is interpreted as fluvial at Taodonggou, but as lakeplain to littoral at Tarlong. The Shaofanggou LOC is interpreted to have been deposited within lakeplain to littoral environments (Yang et al., 2007; 2010). Palaeosols can be analyzed in the context of the corresponding dominant depositional environment (Fig. 3): (1) predominantly fluvial and deltaic environments; (2) predominantly lake margin to littoral non-deltaic and deltaic environments (3) variable environments ranging from fluvial, to shallow and profundal lacustrine and mudflat environments and (4) predominantly mudflat (lakeplain) and fluvial environments.

(1) Predominantly fluvial and deltaic environments with minor lake margin to littoral non-deltaic environments (Figs. 3 and 13; Yang et al., 2007, 2010) are observed from the base of Tarlong North and Taodonggou West to ~608 and ~250 m, respectively. The palaeosols present within these environments include Vertisols, Argillisols, Gleysols, and Histosols. In general, these palaeosols are thought to represent soil formation within environments characterized by perennially wet or one with seasonally variable soil moisture regimes. One calcic Argillisol and three calcic Protosols are present within this part of the section as well. Palaeosols in this part of the section are characterized typically by intense redoximorphy, vascular plant matter accumulation, and accumulation of clay minerals and Fe-oxides. These palaeosols are indicated as a relatively humid to seasonal soil moisture regime (udic/xeric; Soil Survey Staff, 2006). Using modern soils as analogs, these Permian palaeosols likely formed in humid environments with > 1000 mm of precipitation/yr. CIA-K palaeoprecipitation estimates are in accordance with this assessment, indicating MAP that ranged from 900 to 1240 mm/year, with the exception of the values obtained from the calcic Protosols, which are 530 and 650 mm/year. The calcic Protosols are interpreted to represent slight fluctuations to sub-humid conditions during a period of generally stable, humid climate.

(2) Predominantly lacustrine (lake shore and littoral) and deltaic environments are observed from ~608 to ~990 m at Tarlong North, from the base of Tarlong South to ~42 m, and from 250 to 260 m at Taodonggou West (Figs. 3 and 13; Yang et al., 2007, 2010). Palaeosols are relatively limited and poorly developed within this stratigraphic interval. CIA-K palaeoprecipitation estimates from three Argillisols and a calcic Argillisol sampled near the top of this interval range from 1030 to 1140 mm/year. Sandstone composition appears to be more mature throughout this section, which likely indicates a shift in provenance (Guan personal communication). The frequency of thin limestone beds increases towards the top of the Wutonggou LOC, indicating a gradual change in lake chemistry that may correspond to climatic change towards drier, sub-humid to semi-arid conditions (Yang et al., 2010).

(3) Varied environments ranging from fluvial, to shallow and profundal lacustrine and mudflat environments are observed from ~990 to 1044 m at Tarlong North, from ~42 to 140 m at Tarlong South, and from ~260 to ~325 m at Taodonggou West (Figs. 3 and 13; Yang et al., 2007, 2010). In this interval, palaeosols show a wide range of morphologies and include calcic Histosols and carbonaceous Calcisols, calcic Argillisols, Histosols, Vertisols, and Gleysols. These palaeosols reflect a variety of soil moisture conditions ranging from well-drained (Calcisols and calcic Argillisols) to poorly-drained (Histosols and Gleysols). The morphology of the profiles interpreted to have formed under poorly drained conditions probably does not reflect regional climate, but rather a local-to-regional shallow groundwater table (e.g., Buck and Mack, 1995; Mack, 1992). However, the morphology of profiles interpreted to have formed under well-drained conditions probably reflects regional climate that oscillated between semi-arid to sub-humid conditions. Palaeoprecipitation estimates derived from well-drained palaeosols formed in this interval range from 230 to 930 mm/year. We interpret this stratigraphic interval to indicate greater environmental and climatic instability.

(4) Predominantly mudflat (lakeplain) and fluvial environments are observed within the Jiucaiyuan and Shaofanggou LOCs at Taodonggou East and from ~325 to 533 m at Taodonggou West (Figs. 3 and 13; Yang et al., 2007, 2010). Vertisols, Calcisols, and Gypsids are the dominant palaeosol orders and they indicate soil moisture regimes characterized by evapotranspiration in excess of precipitation (ustic/aridic; Soil Survey Staff, 2006) and thus indicate generally greater aridity than other palaeosol types found in the Permian–Triassic strata of the field area. Our observations here indicate a sub-humid to semi-arid climate. Palaeoprecipitation estimates derived from CIA-K range from 290 to 1014 mm/yr. Palaeoprecipitation appears to be much less stable than in the Wutonggou LOC. The 1014 mm/year estimate is derived from a Gypsisol, and this value is inconsistent with the observed palaeosol morphology, which will be discussed below.

Lateral variations in palaeosol morphologies have been observed between age-equivalent strata, and palaeocatenary relationships were observed (e.g. relationships among palaeosols upon a floodplain) along laterally continuous outcrops over ~1 km (e.g. along strike at Tarlong North; Taodonggou East and Taodonggou West; Fig. 3). These relationships indicate that geomorphological controls on aggrading floodplains, lakeplains, and deltaplains likely influenced soil development, especially with respect to drainage and sediment accumulation (Bowen and Kraus, 1987). These variations, however, are limited to the observed palaeosol morphological types seen in each of the four groupings of major depositional environments.

In the case of the single Gypsisol sampled, the CIA-K estimate of 1014 mm/yr MAP is high in comparison to modern analogs (<200 mm; Watson, 1992), even after considering standard error (± 180 mm/yr: Sheldon et al., 2002). Many plausible scenarios can be considered: (1) this palaeosol is not suitable for CIA-K analysis; Sheldon et al. (2002) specifically preclude or exclude the use desert soils from this methodology; (2) the composition is largely detrital and reflects previous weathering; and (3) the atmosphere behaved differently than it does in the present. Given the similarities between the clay mineral assemblages of the By palaeosols and other palaeosols lower in the stratigraphy (Fig. 5), it appears that there could be a detrital component influencing the composition of the Gypsisol. The effect of the Early Triassic atmospheric composition (3) will be considered below.

In summary, MAP values remain relatively consistent throughout the Wutonggou LOC, ranging from 530 to 1270 mm/year, but with most estimates >1000 mm/year. Above this, MAP values show much greater variability, but are generally indicative of drier conditions. MAP values range from 230 to 1014 mm/year. The minimum value is derived from a Calcisol whereas the maximum MAP value comes from the Gypsisol. Although there is a gap in the data due to the paucity of sampled palaeosols, it nonetheless appears that precipitation in the Late Permian remained relatively stable until the Permian–Triassic Boundary.

7. Implications for the End-Permian extinction event

The cause(s) of the end-Permian mass extinction have been speculated upon widely, but there is no consensus regarding the mechanism(s). Hypotheses include: bolide impact (Becker et al.,
2001; Kaiho, 2001); loss of ecological niches as Pangea consolidated (Valentine and Moores, 1970); gradually increasing aridity concurrent with the consolidation of Pangea (Parrish, 1993), overturn of a superanoxic ocean resulting in rapid and catastrophic release of CO$_2$, CH$_4$, and H$_2$S, or a combination of all three gases (Isozaki, 1994; Knoll et al., 1996; Kump et al., 2005); elevated atmospheric CO$_2$, CH$_4$, and sulphate concentrations associated with volcanic emissions from the Siberian and Emaiashan Traps (Berner, 2002; Kidder and Worsely, 2004; Lo et al., 2002; Maruoka et al., 2003; Renne et al., 1995; Retallack and Jahren, 2008) resulting in the consumption of atmospheric oxygen, and low partial pressures of atmospheric O$_2$ (Sheldon, 2006; Sheldon and Retallack, 2002); or a combination of multiple events (Berner, 2002; Erwin, 1993).

The purported effects of mass extinction on land range from catastrophic soil erosion (Retallack, 1999, 2005; Sephton et al., 2001, 2005), shifts in sedimentation styles, i.e. from meandering to braided streams, interpreted as a result of plant extinction (Arche and López-Gómez, 2005; Michaelson, 2002; Ward et al., 2000), and a cessation of the deposition of coal (Faure et al., 1995; Retallack et al., 1996). In general, earliest Triassic palaeosols documented from Antarctica, Australia, and South Africa are typically gleyed and are interpreted to have formed in water-logged conditions during formation (Retallack, 1999; Retallack et al., 2003; Sheldon and Retallack, 2002; Tabor et al., 2007). The presence of these features has been used to infer low concentrations of atmospheric O$_2$ (Sheldon, 2006; Sheldon and Retallack, 2002).

A thin, ~10 cm thick interval of claystone breccia was observed at 828 m at Tarlong North. This is, however, tens of meters below the placement of the Permian–Triassic boundary, and at this time, does not appear to share any significant link to the end-Permian event, and no claystone breccias were found in direct association with the Permian–Triassic boundary. Thick fluvial and deltaic deposits are common within the Wutonggou LOC, and are interpreted to reflect intense provenance weathering and large sediment production and transport that resulted from hinterland uplift and basin subsidence in intense provenance weathering and large sediment production and transport that resulted from hinterland uplift and basin subsidence in intense provenance weathering and large sediment production and transport that resulted from hinterland uplift and basin subsidence in intense provenance weathering and large sediment production and transport that resulted from hinterland uplift and basin subsidence in intense provenance weathering and large sediment production and transport that resulted from hinterland uplift and basin subsidence.

The timescales over which tectonic processes occur are relatively long, the purported effects of mass extinction on land range from catastrophic soil erosion (Retallack, 1999, 2005; Sephton et al., 2001, 2005), shifts in sedimentation styles, i.e. from meandering to braided streams, interpreted as a result of plant extinction (Arche and López-Gómez, 2005; Michaelson, 2002; Ward et al., 2000), and a cessation of the deposition of coal (Faure et al., 1995; Retallack et al., 1996). In general, earliest Triassic palaeosols documented from Antarctica, Australia, and South Africa are typically gleyed and are interpreted to have formed in water-logged conditions during formation (Retallack, 1999; Retallack et al., 2003; Sheldon and Retallack, 2002; Tabor et al., 2007). The presence of these features has been used to infer low concentrations of atmospheric O$_2$ (Sheldon, 2006; Sheldon and Retallack, 2002).

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However, a major change in paleosol morphologies is observed across the Permian–Triassic boundary (Figs. 8–11). Notably, certain palaeosol morphologies disappear in the vicinity of the assumed Permian–Triassic boundary, including: (1) coal-bearing palaeosols (“O” and “O, Bk”), (2) palaeosols containing significant concentrations of iron–manganese nodules and pisoids (“Bgc” and some “Bt” morphs), and (3) palaeosols that accumulated abundant layer-lattice phyllosilicates in subsurface horizons (“Bt” and “Btk”). It should be noted that both (1) and (3) are observed above our placement of the Permian–Triassic boundary at Tarlong North.

The disappearance of abundant organic matter (“O” and “O, Bk” palaeosols) is consistent with reports from other terrestrial Permian–Triassic Boundary sections from Gondwana, in particular sites in Antarctica (Retallack and Krull, 1999), Australia (McLoughlin et al., 1997; Retallack, 1995; and Roberts et al., 1996) and India (Tiwari, 2001).

However, organic matter deposition appears to have occurred until at least the very latest Permian or perhaps even within the earliest Triassic at Tarlong North (Fig. 7). This suggests that dysaerobia and acidity may not have been ubiquitous features of earliest Triassic terrestrial systems as suggested by Retallack et al. (1996).

Furthermore, the demise of the organic matter-bearing soil horizons appears to have been caused by a long-term trend towards semi-arid conditions rather than by a single catastrophic event at the Permian–Triassic boundary. The lack of “Bgc” palaeosols and the decrease in abundance of “Bt,” and “Btk” palaeosols above the perceived Permian–Triassic Boundary also was likely a result of local or regional aridification, which appears to occur gradually over the studied interval. Observed stratigraphic trends in palaeosol morphologies cannot be explained wholly by catenary relationships. Although variations in palaeosol morphologies existed over the landscape, given stratigraphic intervals are dominated by a limited number of soil horizons. Thus, climate was likely the dominant control on long-term change in palaeosol morphologies.

Several studies of global Permian–Triassic sections indicate a reduction in moisture availability (i.e. increased aridity) during the latest Permian and Early Triassic (McLoughlin et al., 1997; Chumakov and Zarkhov, 2001), which have been interpreted to represent a weakened monsoon in inland Pangea (Parrish, 1995). Arid/semi-arid belts likely expanded during the Late Permian and into the Early Triassic as major atmospheric circulation belts shifted poleward (Kidder and Worsely, 2004; Chumakov and Zarkhov, 2001). Greenhouse emissions from volcanic and/or oceanic sources were likely released into the atmosphere during the latest Permian, furthering warming and possibly aridification (Kidder and Worsely, 2004).

During the Late Permian and Early Triassic, the Bogda Mountains fell within the semi-arid belt (Chumakov and Zarkhov, 2001). The presence of the Junggar–Turpan lacustrine system may have permitted the relatively humid climate observed here in the Lopingian Wutonggou LOC (e.g., Yemane, 1993). Sedimentological indicators (Yang et al., 2010) seem to reflect a gradual transition from a humid climate during the Permian to a sub-humid or semi-arid climate in the Triassic, with CIA-K proxy data indicating greater precipitation instability during the Early Triassic. Chaotic atmospheric circulation associated with both global warming and the shifting configuration of the Pangaean supercontinent could have led to the variable MAP estimates derived from Early Triassic palaeosols.

Gypsum, the presence of which is inferred from calcite pseudo-morphs in palaeosols within the Lower Triassic strata, is a typical precipitate of soils formed in desert environments (Watson, 1992), although it is not limited to desert environments. Soil gypsum is also a characteristic mineral formed within acid sulphate soils (Doner and Lynn, 1989). In modern environments, acid sulphate soils are found predominantly within coastal settings, but also within freshwater wetlands and areas with saline, sulphate-rich groundwater (Land and Water Quality Branch, 2006). Furthermore, the low pH conditions of acid sulphate soils result in a high degree of hydrolytic weathering reactions and leaching of base cations from the soil profiles. Many proposed causes and/or effects of the end-Permian event revolve around changes in atmospheric chemistry, and atmospheric chemistry ought to be reflected in the chemistry of the palaeosols. Increased sulphate concentration in the Early Triassic atmosphere (Maruoka et al., 2003) may have increased SO$_4$$^2-$ concentrations within soil waters such that gypsum was a common soil mineral (Tabor et al., 2007). This hypothesis would provide an explanation for why palaeoprecipitation estimates from Gypsisols at Taodonggou West are inconsistent with gypsum-rich soils formed within arid environments.

Tectonics and orographic controls may have also influenced palaeosol development. Through thermal subsidence, the basin could have experienced an increase in potential accommodation, creating a balance-filled to under-filled lacustrine system (Carroll and Bohacs, 1999). If this were the case, then the perceived climate change across the Permian–Triassic boundary at Tarlong and Taodonggou could be related to local effects. The depositional site has been interpreted by some to have existed within an intermittent setting, which may have promoted rain shadow effects, especially during times of uplift (Yang et al., 2010). The timescales over which tectonic processes occur are relatively long, 10$^6$–10$^7$ years (Allen, 2008). A radiometric date from zircons in an ash from the Wutonggou LOC yield an age of 253.12 ± 0.08 Ma (Yang et al., 2010), whereas the Permian–Triassic boundary is dated at 251 Ma (Gradstein et al., 2004). The tectonic history of the region is not
fully understood, so here we propose the Cascade Range as an appropriate analog to the Late Palaeozoic/Early Mesozoic Juelotage Mountains. Assuming uplift rates were on the order of 0.05 to 0.3 mm/year similar to the modern Cascades (Hren et al., 2007; Reiners et al., 2003), then over the course of two million years, 100 to 600 m of the hinterland would be exhumed. Palaeogeographic reconstructions of Zharkov and Chumakov (2001) indicate a reduction in orogenic structures from the Early Permian to the Early Triassic within southern Angara (their Figs. 1–3). Their reconstructions show the emplacement of an orogenic structure on the southern edge of Angara during the Late Permian that may have influenced air circulation and the amount of moisture brought into Angara from Palaeo-Tethys. The reconstructions show that shallow water carbonate and evaporite platforms along the western margin of Angara are replaced by terrestrial depositional environments from the Late Permian into the Early Triassic (Zharkov and Chumakov, 2001). The nearby shallow sea may have been an additional moisture source for the region and its closing may have contributed to increased aridity in the Latest Permian and Early Triassic. Circulation models predict wind directions from the west and from the south (Gibbs et al., 2002; Kiehl and Shields, 2005), which is consistent with the hypothesis presented here (Fig. 12).

These Permian–Triassic palaeosols differ from many previously documented acid sulphate soils and palaeosols in that they lack a significant jarosite or pyrite component (Kraus, 1998; van Dam and Pons, 1972; however, see also Maruoka et al., 2003; Retallack et al., 2007). In the modern scenario, sulphur is provided typically by marine sources. There is no evidence for a marine incursion in the Turpan Basin during the Early Triassic, nor is there any indication that soil parent material was derived from the weathering products of marine shales (Kraus, 1998). An alternative source of sulphur would have been necessary for acid sulphate formation, perhaps from the Early Triassic atmosphere. In modern acid sulphate soils, pyrite precipitates within a reduced environment and upon oxidation forms ferric hydroxide and sulphuric acid. If this reaction occurs in an environment where calcium carbonate is present, the pH is buffered and gypsum will precipitate (van Dam and Pons, 1972). Given the presence of palaeosol carbonates within nearby stratigraphic intervals, it is likely that a source of carbonate was present during soil formation, and thus gypsum could have precipitated upon oxidation of pyrite.

Early Triassic Gypsisols in the Taodonggou sections appear to have formed upon stable landscapes in well-drained environments. Well-drained soils form in open communication with the atmosphere, and are usually aerobic. However, in the case of the possible acid sulphate palaeosols presented here, conditions were likely reducing within the soil profile for some amount of time. Either these conditions resulted from (1) inundation of the soil profile or from (2) a state of low atmospheric oxygen, as suggested by Sheldon and Retallack (2002). If these are indeed palaeosols that formed as acid sulphate soils under low atmospheric PO2, then increased atmospheric PO2 would be necessary to oxidize the soils. Because four of these soils have been described within the Bogda Mountains sections, interpreting these as acid sulphate soils that formed within a low PO2 atmosphere would indicate highly volatile atmospheric oxygen concentrations during the Early Triassic.

8. Conclusions

Late Permian to Early Triassic palaeosols from the Bogda Mountains demonstrate variations in morphological, mineralogical and geochemical characteristics. Especially noteworthy are the presence of Late Permian palaeosols characterized by reductive depletion, ferric-manganese nodules, and organic matter accumulations, indicative of humid climates, which are replaced by palaeosols characterized by a lack of reductive morphology and subsurface accumulations of calcium carbonate and possibly gypsum, which indicate that evapotranspiration exceeded

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Fig. 12. Palaeogeographic reconstructions after Zharkov and Chumakov (2001). Dashed lines indicate location of regional northeast–southwest trending profiles that are displayed to the right (not to scale). Approximate basin position is denoted with a star. Arrows indicate the prevalent wind directions. Blue colors indicate ocean (light blue is shallow platforms) whereas the buff color indicates land and the gray color indicates mountains. Map and profile for the (a) Late Permian and (b) Early Triassic. These reconstructions show a marked change regionally in land surface area. Increasing land area, in addition to the development of mountainous regions to the west and southwest of the field area could have reduced moisture availability during the Late Permian and Early Triassic.
precipitation. Similarly, palaeo-precipitation estimates generated via the CIA-K proxy indicate a decrease in palaeo-precipitation from the Lopingian to the Early Triassic and more specifically a transition from relatively stable climate to unstable climate across the Permian–Triassic Boundary. Environmental and climatic instability was likely related to shifting circulation patterns, as they responded to possible uplift and global warming associated with greenhouse gas influx at the Permian–Triassic boundary.

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