End-Permian terrestrial ecosystem collapse in North China: evidence from palynology and geochemistry

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Abstract

The Permian-Triassic Mass Extinction (*ca.* 252 Ma; PTME) is the most severe biocrisis of the Phanerozoic in both the oceans and on land. The crisis saw the collapse of terrestrial ecosystems in low, mid and high latitudes. Although terrestrial plant losses have been implicated as a driver of concurrent changes in terrestrial sedimentary environments and facies (e.g., fluvial style and/or grain size), the relationship between extinction and environmental change in the North China Plate (NCP) remains uncertain due to a paucity of plant macrofossils. We explore the relationship between terrestrial environments and changes in plant communities using a combination of sedimentology, palynology, geochemistry, mineralogy and charcoal data from a terrestrial succession in the Yiyang Coalfield located in the southern NCP. Our multiproxy approach places the end-Permian Terrestrial Collapse (EPTC) at the base of bed 20, below the level of the main (marine) PTME at the top of bed 21. The EPTC manifested as a rapid loss of vegetation accompanied by climatic warming and frequent wildfires. The main PTME was accompanied by warming, spikes in the Chemical Index of Alteration (corrected CIA, CIA*) and sedimentary Ni (Ni/Al) concentrations, and a transition from arid floodplain to fluvial facies in the sedimentary record. Our results suggest that wildfires induced by global warming during the early eruption phase of the Siberian Traps Large Igneous Province triggered terrestrial ecosystem collapse in the NCP prior to the PTME. Plant extinctions during the EPTC were accompanied by changes in sedimentology and environment, but there was no abrupt change in fluvial styles. Temporal coincidence suggests that shifts in end-Permian terrestrial ecosystems towards those tolerant of warmer and more environmentally stressed environments were driven by concurrent Siberian Traps volcanism.

Keywords: Permian-Triassic mass extinction; End-Permian terrestrial ecosystem collapse; Palynology; Charcoal; Extinction

1. Introduction

The Permian-Triassic Mass Extinction (PTME) is the greatest extinction event of the Phanerozoic (Erwin, 1994) and resulted in the loss of >81% of marine (Fan et al., 2020) and >89% of terrestrial (Viglietti et al., 2021) species. High-resolution isotope dilution thermal ionization mass spectrometry (ID-TIMS) zircon U-Pb dates show that the PTME was concurrent with the Siberian Traps Large Igneous Province (Burgess et al., 2014; Burgess and Bowring, 2015) and contemporaneous volcanism in South China (Zhang et al., 2021). Volcanism released large quantities of greenhouse gases and caused global environmental and climatic changes, including rapid global warming (Sun et al., 2012; Cui et al., 2021; Wu et al., 2021), large-scale perturbations to the carbon cycle (Shen et al., 2011; Bond and Grasby, 2017; Dal Corso et al., 2022), widespread wildfires (Shen et al., 2011; Chu et al., 2020; Lu et al., 2020, 2022) and soil erosion (Biswas et al., 2020; Lu et al., 2020, 2022; Aftabuzzaman et al., 2021; Kaiho et al., 2021). A combination of these factors likely led to the end-Permian terrestrial ecosystem collapse (EPTC), which resulted in significant changes in terrestrial biotas and sedimentary environments (Dal Corso et al., 2022).

Global climate changes led to significant shifts in terrestrial floras during the Late Permian. Studies of plant macro- and microfossils suggest that plant diversity experienced a rapid decline in Australia (Fielding et al., 2019; Mays et al., 2020), northwest China (Cao et al., 2008 and references therein), South China (Chu et al., 2020; Feng et al., 2020; Xu et al., 2022), and North China (Chu et al., 2015, 2019; Xiong et al., 2021). These losses included the disappearance of the *Glossopteris* flora and coals from Australia (Fielding et al., 2019; Mays et al., 2020), and the disappearance of the *Gigantopteris* flora and coals from South China (Chu et al., 2020; Feng et al., 2020; Xu et al., 2022). In the North China Plate (NCP), extinctions among terrestrial floras resulted in the loss of ~54% (14/26) of genera and ~88% (28/32) of plant species in the gymnosperm-dominated assemblages of the Sunjiagou Formation (Chu et al., 2015, 2019), which follows the stratigraphically lower losses of *Gigantoperis* floras in this region (e.g., Yang and Wang, 2012). However, due to the low resolution of existing plant macrofossil studies, the precise timing of the these extinctions in the NCP has not been determined. Further, detailed studies are needed to fully evaluate Late Permian floral changes in the NCP.

Changes in global climate and terrestrial floras during the Late Permian had significant effects on terrestrial sedimentary environments. Late Permian fluvial styles shifted from meandering to braided rivers in Australia (Sydney Basin) (Michaelsen, 2002), South Africa (Karoo basin) (Ward et al., 2000, 2005), India (Satpura and Pranhita-Godavari basins) (Tewari, 1999), and in the Ordos Basin in North China (Zhu et al., 2019, 2020) and this has been attributed to climate change and vegetation losses. However, recent studies have shown no sudden sedimentological change at the level of the plant extinction in Australia, where terrestrial environmental changes mostly occurred after the PTME (Fielding et al., 2019). It has since been reported that there were no significant sedimentological changes at the level of terrestrial plant extinctions in South Africa (Gastaldo et al., 2020) or South China (Wignall et al., 2020), where sedimentary facies changes also occur above the level of the PTME. A recent study from the NCP revealed significant sedimentological changes in Permian-Triassic terrestrial environments, though these were not accompanied by changes in fluvial style (Ji et al., 2022). However, due to the lack of detailed records of terrestrial plant communities, the relationship between these and their sedimentary environment in the NCP is unclear.

To evaluate the timing and nature of terrestrial ecosystem changes and their relationship with environmental change, we examined spore-pollen fossils, charcoal abundance, chemical index of alteration (CIA) values, kaolinite content, and Ni concentrations from the Dayulin section in the southern NCP.

2. Geological setting

During the Late Permian, the NCP was located at approximately 20° N on the northeastern margin of the Paleo-Tethys Ocean (Fig. 1a), with the Inner Mongolia uplift (IMU) to the north and the North Qinling Belt (NQB) or Funiu paleo-land to the south (Shang, 1997; Lu et al., 2020; Guo et al., 2022; Fig. 1b). Sediments in the study area were mainly derived from the NQB or the upper crust of the southern NCP (Shang, 1997; Wang et al., 2020; Fig. 1b).

In the Yiyang Coalfield (Henan Province), Permian-Triassic strata comprise the Shangshihezi, Sunjiagou and Liujiagou Formations (Fig. 1c). The Sunjiagou Formation comprises three members: the Pingdingshan Sandstone Member, the Tumen Member, and the Quanmen Member (Fig. 1c). The Pingdingshan Sandstone Member is primarily composed of conglomeratic sandstone, medium-coarse and feldspathic quartz sandstone, and thin layers of siltstone and mudstone (Fig. 1c) with a palynological assemblage dominated by rib-bisaccate gymnosperms (Ouyang and Wang, 1985). The Tumen Member is primarily composed of fine sandstone, siltstone, silty mudstone, and mudstone (Fig. 1c). This member is considered to be Late Permian in age based on the presence of the *Ullmannia-Yuania* fossil plant assemblage (Chu et al., 2015). The Quanmen Member comprises silty mudstone, thin layers of mudstone and siltstone, and interbedded limestone, lacking in age-diagnostic plants (Fig. 1c).

3. Materials and methods

We sampled 26 fresh mudstone samples from the Sunjiagou Formation at the Dayulin section (34.5017N, 112.1556E) in the Yiyang Coalfield of the southern NCP (Figs. 1a, 1b; sampling locations are shown in Figure 2a). Each sample was divided into two parts, one of which was crushed into particles approximately 1 mm in diameter for palynological analysis. The remaining part of the sample was crushed to pass through a 200 μm mesh and then divided into three subparts for analysis of (1) major elements, (2) trace elements, and (3) clay mineral compositions.

Major and trace elements were measured at the Beijing Research Institute of Uranium Geology. Major and trace elements analysis was undertaken with an X-ray fluorescence spectrometer (PW2404) and an inductively coupled plasma mass spectrometer (Finnigan MAT), respectively. The spectrometer was calibrated before use with standards of CRMs (GBW07427) and analytic precision was within 5%. Clay mineral composition was analysed using an X-ray diffractometer (D/max 2500 PC) at the State Key Laboratory of Coal Resources and Safe Mining (Beijing), and the data were interpreted using Clayquan 2016 software with a relative analysis error of ±5%. The analytic precision or error of all samples is based on reproducibility and repeats of the standard sample and standard samples were run after every five sample analyses.

Palynological isolation and identification was undertaken according to the China national standard (SY/T5915–2018) at the Chinese Academy of Geological Sciences (Beijing). A detailed description of the analytical methods used is available in the Supplementary Material. For each spore-pollen sample, more than 100 sporomorphs were identified by the point-counting method under transmitted light microscopy (Olympus BX 41). The organic residue was mounted in epoxy resin on microscopy slides, and organic matter analysis involved scanning two microscope slides (18 ×18 mm) of each sample according to the scheme developed by Batten (1996). All palynological slides are housed at the State Key Laboratory of Coal Resources and Safe Mining (Beijing). Percentages of spore and pollen taxa were calculated based on the sum of total sporomorphs. Palynological assemblages were identified by stratigraphically constrained cluster analysis (CONISS) using Tilia software (e.g., Zhang et al., 2022a).

We use Ni concentrations as a proxy for volcanism due to the relationship between sedimentary Ni content and volcanic eruptions and magmatic intrusions during the PTME interval (Rampino et al., 2017; Lu et al., 2020; Kaiho et al., 2021). Variations in spore-pollen composition through the studied strata were used to reflect the changes in paleo-vegetation and to reconstruct paleoclimatic conditions (e.g., Lu et al., 2021; Zhang et al., 2022a). The Chemical Index of Alteration (CIA; see Supplementary Material) values were used to restore the weathering trends and paleoclimate, with spore-pollen fossils and kaolinite content used for reference (e.g., Xu and Shao, 2018; Zhang et al., 2019, 2022b; Lu et al., 2020, 2021; Shen J. et al., 2022). Charcoal abundance was used as a proxy for paleo-wildfire (e.g., Glasspool et al., 2015; Chu et al., 2020; Lu et al., 2020; Zhang et al., 2022a).

4. Results and analysis

4.1 Lithologic shift and environmental records

In the study area, the Sunjiagou Formation records significant changes in lithology and depositional environment. At the base of our studied succession, above bed 1 (gray mudstone belonging to the Shangshihezi Formation), beds 2–14 comprise yellowish-brown conglomeratic sandstones and medium-coarse grained sandstones of the Pingdingshan Sandstone Member of the Sunjiagou Formation. Also present are purplish-red fine grained sandstone, siltstone, and mudstone (Fig. 1c). The thickness of sandstone beds decreases upwards in the succession, while the proportion of fine-grained sediments such as siltstone and mudstone increase (Fig. 1c), consistent with deposition in a delta plain/front setting (Wang, 2019). Beds 15–19 (belonging to the Tumen Member) are mostly greyish-green fine sandstones and mudstones that are rich in plant macro- (Chu et al., 2019) and micro-fossils (see section 4.2). Beginning in bed 20, and more prominently in bed 21 the sediments abruptly shift to purplish-red mudstones and sandstones of the Quanmen Member, which also includes paleosols contaning calcareous nodules (Figs. 1c, 2). The plant-rich greyish green sediments may record a mixture of humid waterlogged conditions (Figs. 2b, 2c) and have been interpreted as medial terminal fan deposits (Ji et al., 2022), while the purplish-red sediments represent floodplain deposits adjacent to meandering channel belts (Figs. 2b, 2e). Occasional scattered or layered calcareous nodules (Fig. 2d) within the mudstones probably represent calcretes formed within floodplain soil horizons subject to extended periods of subaerial exposure, and have been interpreted as being deposited on an arid floodplain (Ji et al., 2022). Beds 23–24 are mainly composed of purplish-red, fine-grained calcareous sandstone, siltstone, mudstone and grey thin-bedded limestone (Fig. 2b). Sandstone units contains parallel bedding, wave ripples (Fig. 2h) and load structures, which have been interpreted as indicative of deposition in a coastal mudplain (Ji et al., 2022). The Liujiagou Formation (bed 25 and upwards beyond the top of our studied succession) (Fig. 2b) is mainly composed of red sandstones that range from several decimeters up to 2 m thick, and have been interpreted to record low-sinuosity, braided fluvial channel systems (Ji et al., 2022).

4.2 Palynology fossils and paleofloral and paleoclimatological records

Eighteen spore and 25 pollen genera have been identified (Figs. 3, 4; Table S1) and these can be divided into three palynological assemblage zones (AZ) based on vertical variations in palynological content determined by CONISS (Figs. 3, S1). The compositions of AZ-I (samples YY-1 to YY-5; *Florinites–Lunatisporites–Alisporites*/*Chordasporites*) and AZ-II (samples YY-6 to YY-10; *Chordasporites*–Florinites/*Alisporites–Vesiccaspora*) are generally similar and include many co-occurring taxa, with the assemblages distinguished on abundance differences. AZ-I and AZ-II are dominated by gymnosperm pollen (72.2–80.7% and 71.2–83.6%, mean (x̄) = 77.5 and 78.3%, respectively), followed by fern spores (19.3–27.8% and 16.4–28.8%, x̄= 22.5 and 21.7%, respectively) (Figs. 3, S1; Table S1). In AZ-I, syncytia-bisaccate pollen dominate (20.0–27.5%, x̄= 23.4%) and include *Vesiccaspora*, *Klausipollenites*, *Florinites*, and *Cordaitina*; followed by ribbed-bisaccate pollen (12.3–24.1%, x̄= 15.1%), including *Chordasporites*, *Lueckisporites*, *Gardenasporites*, *Vestigisporites*, *Limitisporites*, *Taeniaesporites*, and *Protohaploxypinus*; bisaccate pollen are common (8.3–13.8%, x̄= 10.8%), including *Alisporites*, *Platysaccus* and *Pityosporites* (Figs. 3, S1; Table S1). In AZ-II, ribbed-bisaccate pollen dominate (14.4–24.1%, x̄= 19.9%), followed by syncytia-bisaccate pollen (13.6–21.1%, x̄= 18.0%), whilst bisaccate pollen are also common (10.1–11.7%, x̄= 11.2%) (Figs. 3, S1; Table S1). AZ-III (samples YY-11 to YY-26) records a rapid decline in spore-pollen abundance and diversity, and spores and pollen are largely absent (Fig. 3), except for a sporadic and low abundance ocurrences in sample YY-11 (Fig. 3; Table S1; see Supplementary Material).

The stratigraphic age for the lower and middle parts of the Sunjiagou Formation (AZ-I and AZ-II) has been determined biostratigraphically based on the characteristic elements and abundance of spore and pollen taxa. Paleovegetation and paleoclimate has been inferred based on the relationship between palynological fossils and their parent plants (Tables S1, S2). Palynological compositions from the lower and middle parts of the Sunjiagou Formation comprise characteristic elements from the Late Permian Changxingian Stage including *Lueckisporites*, *Alisporites*, and *Protohaploxypinus* (e.g., Hou, 1990; Ouyang et al., 1993, 2004; Balme, 1995; Liu, 2000; Gao et al., 2018). The spore-pollen composition, which is dominated by an abundance of gymnosperms with less common fern spores supports this age assessment (Fig. 3; see Supplementary Material). During this interval, paleovegetation in the study area was mainly dominated by xerophytic conifers and meso-xerophytic seed ferns, with abundant ferns, while lycophytes, cycads, and horsetails were rare and sporadically distributed, indicating that a semi-arid climate prevailed (samples YY-1 to YY-10; AZ-I and AZ-II) (see Supplementary materials).

4.3 Ni concentrations and volcanism

Results for Ni concentrations are shown in Figure 3 and Table S3. Ni concentrations vary from 20.6 to 49.6 ppb ($\bar{x} = 33.21$ ppb) with a clear peak at the top of bed 21 (Fig. 3; Table S3). Ni concentrations in terrestrial strata are usually influenced by total organic matter content (TOC) and lithological change (i.e. Al content) (e.g., Grasby and Beauchamp, 2009; Grasby et al., 2015, 2019; Fielding et al., 2019) and so Ni concentrations are normalized to TOC and Al. TOC contents in samples that are enriched in Ni are low (≤ 0.02 wt. %) (Wu et al., 2020; Fig. 3), and therefore TOC is not suitable for standardization in this case (e.g., Grasby et al., 2019). The Ni/Al ratios vary from 1.37 to 3.60 $\times 10^{-4}$ ($x = 2.27 \times 10^{-4}$), with a peak 1.58 times background (mean) values (Fig. 3; Table S3). The Ni (Ni/Al) peak is located within a negative $\delta^{13}C_{org}$ excursion (Wu et al., 2020; Fig. 3). The Ni (Ni/Al) peak is within the range of the Ni anomaly known from the PTME interval globally (Rampino et al., 2017 and references therein). Collectively these features suggest that the Ni and Ni/Al enrichment anomalies are related to volcanism in the Siberian Traps LIP (e.g., Rothman et al., 2014; Rampino et al., 2017; Fielding et al., 2019; Lu et al., 2020; Kaiho et al., 2021).

4.4 Chemical weathering indices and terrestrial weathering records

Clay mineral components are mainly illite-smectite mixed layers, followed by illite and kaolinite, with less frequent chlorite (Fig. 3; Table S4). The content of the illite-smectite mixed layers varies from 35 to 66% (x̄ = 50.2%), illite content varies from 28 to 63% (x̄ = 40.1%), and chlorite content varies from 0 to 17% (x̄ = 5.9%). Kaolinite content varies from 28 to 63% (x̄ = 40.1%) with peaks in beds 18 (10%) and 21 (15%) (Fig. 3).

Values of all samples deviate from the ideal weathering trend line on the A-CN-K (Al_2O_3-Cao*+NA_2O-K_2O) diagram (Fig. 5a), indicating that potassium metasomatism has affected the sediments (Fedo et al., 1995). This results in lower CIA values and higher Weathering Index of Parker (WIP) values, but does not affect the variation trend of the Chemical Index of Weathering (CIW) and sodium depletion index (τNa) values (Yang et al., 2018; Cao et al., 2019) (these weathering indices were calculated according to the formulae outlined in the Supplementary Material). Thus, potassium metasomatism was corrected according to previously published methods (Fedo et al., 1995; Huang et al., 2017).

In our study, the corrected CIA values (CIA*) vary from 73.6 to 83.8 (x̄= 78.2), the corrected WIP values (WIP*) vary from 25.5 to 54.4 (x̄ = 44.0), CIW values vary from 86.6 to 94.8 (x̄= 91.1), and τNa values vary from -0.86 to -0.54 (x̄= -0.72) (Table S3). Furthermore, CIA* values shows a significant positive correlation with CIA (r^2= 0.90) (Fig. 6a), suggesting that this correction eliminated the effects of K-metasomatism on CIA values without changing their temporal trends (Yang et al., 2018).

Chemical weathering is susceptible to the influence of changing provenance, sedimentary recycling, hydraulic or sedimentary sorting processes and diagenesis, thus potentially masking paleoclimatic information (e.g., Chen et al., 2003; Xu and Shao, 2018; Yang et al., 2018, 2020). In this study, all samples were distributed along the ideal weathering trend line of the upper crust in the southern NCP on the CIA (CIA*)-WIP(WIP*) plot (Fig. 5b), suggesting that the provenance of the southern NCP experienced a continuous and consistent supply and that samples are not affected by changes in provenance or sedimentary recycling (e.g., Garzanti et al., 2013; Yang et al., 2020). This conclusion is in accordance with previous results that the sediments of the Yiyang Coalfield were sourced mainly from the southern NCP during the Permian-Triassic transition (Shang, 1997; Wang et al., 2020). This is also consistent with studies on Th/U ratios that vary from 1.78 to 5.19 ($\bar{x}$ = 3.65), indicating that the samples are not affected by depositional recycling (e.g., Lu et al., 2020; Table S3). In addition, the poor correlation of Al_2O_3/SiO_2 ratios with CIA*($r^2 = 0.05$), WIP*($r^2 =$ 0.20), and τNa ($r^2 = 0.04$) (Figs. 6b–d) further indicates that samples have not been affected by hydraulic or sedimentary sorting processes or temporal variations in sedimentary provenance (Yang et al., 2018, 2020). These observations, together with high KI values ($\bar{x}$ = 0.36; Table S4) and its random distribution in ascending order, indicate that the samples in our study have not been affected by diagenesis (Cheng et al., 2019; Zhang et al., 2022b). We consider that variations in kaolinite content and CIA* values in our study are reliable indicators of terrestrial weathering and paleoclimate conditions. The kaolinite and CIA* values in this study (Fig. 3) suggest that the arid climates of the Late Permian were interrupted by two periods of relatively humid conditions. This is consistent with similar changes reported from the nearby Yuzhou Coalfield (Lu et al., 2020; Fig. 1b).

4.5 Charcoal abundance and wildfire records

The organic matter in our samples includes charcoal (Figs. 7a, 7b), fungi (Fig. 7c), plant spores and pollen (Fig. 7d), plant cuticles (Figs. 7e-g), bisaccate taeniate pollen, and bisaccate non-taeniate pollen according to the scheme developed by Batten (1996). Results of charcoal contents are shown in Figure 3 and Table S4, with charcoal content varying from 11 to 1023 grains ($\bar{x}$ = 203.2 grains) (Fig. 3; Table S4). Charcoal abundance was highest in samples YY-1 to YY-11, varying from 201 to 1023 grains ($\bar{x}$ = 395.3 grains), while charcoal abundance in samples YY-12 to YY-26 decreased significantly, ranging from 11 to 75 grains ($\bar{x}$ = 38.6 grains). Under transmitted light microscopy, the charcoal is opaque, pure black, and strongly fragmented with sharp edges, and does not fluoresce under fluorescence illumination (Figs. 7a, 7b). These features indicate that the charcoal has not undergone transport over long distances (e.g., Lu et al., 2020, 2022; Zhang et al., 2022a) and is a reliable indicator of wildfire in the study area. The arid and semi-arid climatic conditions prevailing in the study area (see Section 4.2) were favourable for wildfires during the Late Permian (e.g., Lu et al., 2020, 2022).

5. Discussion

5.1 Temporal patterns of terrestrial ecosystem change in North China

In the NCP, the lithostratigraphic units of the Sunjiagou Formation are diachronous (Shen B. et al., 2022). The position of the PTME is placed variably in the middle and upper parts of the Sunjiagou Formation in different sections (Chu et al., 2015, 2017, 2018; Tu et al., 2016; Cao et al., 2019; Guo et al., 2019; Lu et al., 2020, 2022; Wu et al., 2020; Zhu et al., 2019, 2020; Guo et al., 2022). In the Dayunlin section of the Yiyang Coalfield (Fig. 1b), the lower and middle parts of the Sunjiagou Formation are constrained to the Late Permian and the upper part of the Sunjiagou Formation is constrained to the Permian-Triassic transitional period based on the biostratigraphic (including conchostracans and plant mega-fossils) and chemostratigraphic data (Chu et al., 2015, 2017; Tu et al., 2016; Wu et al., 2020), but the precise placement of both the EPTC and the PTME are less well constrained.

Recent studies on the ID-TIMS zircon U-Pb age, magneto- and chemo- stratigraphy of the Sunjiagou Formation in the Shichuanhe section from the southern NCP suggest that the EPTC occurred ~270 kyr before the level of the marine PTME (Wu et al., 2020; Guo et al., 2022). The Shichuanhe section appears to contain a complete terrestrial sedimentary record of the Late Permian-Early Triassic transition, and related bio- (including ostracod and plant mega-fossils) and chemo- (including $\delta^{13}C_{org}$ and CIA patterns) stratigraphy have been extensively studied (Cao et al., 2019; Chu et al., 2019; Wu et al., 2020; Guo et al., 2022). In this context, the Shichuanhe section provides a key age and stratigraphic framework for Late Permian terrestrial ecosystem changes in the NCP, which can be used for global terrestrial and marine stratigraphic correlation (Guo et al., 2022; Fig. 1b). We use Shichuanhe as a reference section here to discuss the positions of the EPTC and the PTME in the Yiyang Coalfield.

Palynological and geochemical data provide effective methods to constrain the position of the PTME. In this study, palynological data constrain beds 2 to 20 of the Sunjiagou Formation to the Late Permian Changxingian Stage (Fig. 3; see section 4.2). The CIA* spike and accompanying Ni peak in the upper part of the Sunjiagou Formation (at the top of bed 21, Fig. 3) co-occur with a negative $\delta^{13}C_{org}$ excursion (Wu et al., 2020; Fig. 3). A similar pattern is known from the Yuzhou Coalfield in the southern NCP, where these features are diagnostic for the PTME (Lu et al., 2020). In addition, previous studies on the Shichuanhe and Yima sections (Fig. 1b) in the southern NCP showed that the CIA spikes in the upper part of the Sunjiagou Formation coincide with the marine PTME (Cao et al., 2019). Similar CIA peaks occur at the onset of the marine PTME in South China, including at the Xinmin, Chaohu, and Meishan sections (Shen et al., 2013; Chen et al., 2015; Zhao and Zheng, 2015; Cao et al., 2019; Fig. 8). Ni and Ni/Al enrichments are also known at the onset of the terrestrial and marine PTME levels around the world (e.g., Kaiho et al., 2001, 2006, 2021; Grasby et al., 2009, 2015; Rampino et al., 2017, 2020; Fielding et al., 2019; Lu et al., 2020). Such anomalies in the study area all predate the Permian-Triassic boundary (PTB) defined by previous studies (Tu et al., 2016; Fig. 1c). Therefore, we consider that the spikes of CIA* and Ni (Ni/Al) values in our studied section represent the level of the marine PTME.

Across the NCP, the EPTC occurred prior to the marine PTME (Guo et al., 2022; Lu et al., 2022). In this study, spore-pollen fossil abundance and richness decreased rapidly at the base of bed 20 in the middle part of the Sunjiagou Formation (Fig. 3). This dramatic palynological change was accompanied by disappearance of the *Ullmannia-Yuania* floral assemblage (Chu et al., 2015; Fig. 3), a rapid fall in TOC contents (from 0.3–0.01 wt. %) (Wu et al., 2020; Fig. 3), the onset of a negative $\delta^{13}C_{org}$ excursion (Wu et al., 2020; Fig. 3), a change in lithology (from greyish-green sandstone and mudstone to purplish-red mudstone at the base of bed 20) (Figs. 2, 3), the appearance of calcareous nodules (Figs. 2d, 3), frequent wildfires (Fig. 3), and a subsequent decrease in CIA* values (Fig. 3). These phenomena predate the peaks in CIA* and Ni values associated with the marine PTME and likely represent the onset of the EPTC in the study area (Figs. 2c, 3, 8). In the Shichuanhe section, the EPTC accompanied the onset of the negative $\delta^{13}C_{org}$ excursion, the disappearance of plant mega-fossils, the appearance of calcareous nodules, and a decrease in CIA values (Cao et al., 2019; Wu et al., 2020; Guo et al., 2022; Fig. 8). Similar patterns are seen from the Zishiya section (Wu et al., 2020), Yima (Cao et al., 2019; Fig. 8), and the ZK21-1 borehole (Yuzhou Coalfield) (Lu et al., 2020; Figs. 8, 9) of the southern NCP and the ZK-3809 borehole (Liujiang Basin) (Lu et al., 2022). Furthermore, similar phenomena have been recorded from equatorial Northern Hemisphere settings in southern China and Italy where they are interpreted as the onset of the EPTC approximately ~60 kyr before the marine PTME (constrained by detailed marine biostratigraphic constraints) (Biswas et al., 2020; Aftabuzzaman et al., 2021; Kaiho et al., 2021; Dal Corso et al., 2022; Fig. 9). We consider that the base of bed 20 represents the onset of the EPTC in the study area. Taking the onset of EPTC (252.21 ± 0.15 Ma) in the Shichuanhe (Guo et al., 2022) as a reference (Fig. 1b), we consider that the onset of EPTC in the NCP is roughly consistent with its onset in higher latitude South Africa (252.24 ± 0.11 Ma) (Gastaldo et al., 2020).

5.2 Potential links with volcanism and global warming

High-resolution ID-TIMs zircon U-Pb ages and detailed marine biostratigraphic evidence suggest that the EPTC and the PTME are closely related to the eruption of the Siberian Traps LIP (Burgess et al., 2014; Burgess and Bowring, 2015; Fielding et al., 2019; Gastaldo et al., 2020; Kaiho et al., 2021; Guo et al., 2022). Recent studies have shown that the EPTC was not synchronous across different latitudes/regions and climatic zones (Fielding et al., 2019; Gastaldo et al., 2020; Kaiho et al., 2021; Guo et al., 2022; Lu et al., 2022), and it may have been driven by different phases of the Siberian Traps LIP in different places (Burgess and Bowring, 2015).

During the first phase of Siberian Traps volcanism, flood-lava eruptions potentially drove the EPTC in high latitude areas such as Australia and South Africa as well as low- and mid- latitudes regions such as the NCP and northwest China (Burgess and Bowring, 2015; Burgess et al., 2017; Fielding et al., 2019; Gastaldo et al., 2020; Dal Corso et al., 2022; Guo et al., 2022; Fig. 9). In high latitude southern hemisphere settings, terrestrial environments in Australia (Bowen and Sydney Basins) and South Africa (Karoo Basin) record fluctuating carbon isotope records, increases in chemical weathering (CIA) rates, and floral and tetrapod changes (Fielding et al., 2019; Gastaldo et al., 2020; Frank et al., 2021; Viglietti et al., 2021; Fig. 9). Recent high-resolution ID-TIMS zircon U-Pb ages suggest that these high latitude changes (252.24 ± 0.11 Ma) (Gastaldo et al., 2020) roughly coincide with flood basalt volcanism in Siberia (252.27 ± 0.11 Ma) (Burgess and Bowring, 2015). Similarly, a series of phenomena associated with the EPTC in the study area (as described in Section 5.1 and below) are associated with the initial phase of Siberia Traps volcanism (Figs. 9, 10a, 10b) based on the new chrono- (252.21 ± 0.15 Ma) and chemo- (including $\delta^{13}C_{org}$ and CIA patterns) stratigraphic data from the reference Shichuanhe section (Cao et al., 2019; Chu et al., 2019; Wu et al., 2020; Guo et al., 2022). During this period, the lithologic shift from greyish-green to purplish red sediments (Figs. 2c, 3) and the appearance of calcareous nodules (Figs. 2d, 3) in the study area are consistent with an increase in surface temperatures (Tu et al., 2016; Wu et al., 2020; Zheng et al., 2020). Warmer climatic conditions can promote frequent wildfires through frequent lightning activity and may contribute to vegetation loss on land (e.g., Lu et al., 2020, 2022; Zhang et al., 2022a). In the context of this climatic warming, an increase in the terrestrial hydrological cycle will contribute to an increase in terrestrial weathering (e.g., Cao et al., 2019; Frank et al., 2021; Lu et al., 2021). However, the CIA* values in the study area shows a brief downward trend (Figs. 3, 8, 9). Similar CIA values have been recorded in terrestrial sections from the NCP and northwest China (mid-latitude) (as described in Section 5.1; Figs. 8, 9). We consider that the fall in CIA* values immediately above the EPTC may have been driven by an increase in soil erosion caused by terrestrial vegetation loss, which exposes fresher source materials characterised by lower

CIA values (e.g., Frank et al., 2021). Intriguingly, these early eruptions did not produce global changes in the $\delta^{13}C$ and Hg records (Dal Corso et al., 2022). Only northern latitude marine records downwind of the Siberian Traps have these shifts in $\delta^{13}C$ and Hg records, indicating limited atmospheric mixing of volatiles released during this early eruption phase (Grasby and Beauchamp, 2009; Grasby et al., 2011; Sanei et al., 2012).

In the second phase of LIP emplacement, the largest volcanic eruptions of the Siberian Traps (plus felsic volcanism in South China) released large amounts of greenhouse gases into the global atmospheric system and caused the eventual destruction of terrestrial ecosystems (e.g., Burgess et al., 2014; Burgess and Bowring, 2015; Kaiho et al., 2021; Zhang et al., 2021; Fig. 9). In this study, CIA* values increased rapidly, accompanied by the increase in kaolinite content, Ni concentration, and Ni/Al ratios (Figs. 3, 8, 9, 10c). Ni and Ni/Al enrichment anomalies provide potential evidence for a connection with the Siberian Traps (e.g., Kaiho et al., 2001, 2021; Grasby and Beauchamp, 2009; Grasby et al., 2015; Fielding et al., 2019; Rampino et al., 2017, 2020; Lu et al., 2020), and the occurrence of the CIA* spike indicates a significant increase in surface temperature during this time (Cao et al., 2019; Lu et al., 2020). Similar CIA spikes were observed in contemporaneous terrestrial and marine strata (Figs. 8, 9; as described in Section 5.1). Although warmer conditions may have been the main driver of the CIA* spike, the potentially corrosive effects of more intense precipitation and/or acid rain cannot be ruled out (Cao et al., 2019; Lu et al., 2020; Frank et al., 2021). During the PTME interval, the area around the Paleo-Tethys Ocean experienced a brief wetting phase (Winguth and Winguth, 2013), and sedimentological studies in the southern and northern NCP provide further evidence for this (Zhu et al., 2019; Ji et al., 2022). Similarly, acid rain caused by SO_2 release from the Siberian Traps has also been implicated in the PTME (Jurikova et al., 2020; Kaiho et al., 2021). The downward trend of CIA* values above the PTME may reflect a temperature rebound and/or the termination of acid rain effects, but persistently higher CIA* than the pre-PTME average (Figs. 8, 9, 10d) suggests a permanent transition to warmer climatic conditions in the extinction aftermath (e.g., Sun et al., 2012; Joachimski et al., 2022).

The emplacement of the Siberian Traps caused significant changes in the terrestrial environments via global warming. Our data supports a scenario in which global warming caused by the first eruptive phase of the Siberian Traps led to a rapid loss of terrestrial plants accompanied by (and partly resulting in) lithologic changes (a shift from greyish-green to purplish-red and the appearance of numerous calcareous nodules) (Figs. 2c, 2d, 3, 9, 10a, 10b). Often such phenomena are associated with more arid climatic conditions (Tu et al., 2016; Zheng et al., 2020). However, a recent sedimentological study suggests that this lithological change reflects an increase in oxidation and evaporation rather than rapid aridification (Ji et al., 2022). Instead we suggest that the colour change is due to the loss of the reducing power of organic matter following the loss of terrestrial biomass caused by plant extinctions (Figs 10a, 10b). Later, global warming caused by the most significant volcanic eruptions (in Siberia and South China) led to river rejuvenation in the study area (Fig. 10c). The concave-up erosional surfaces (Fig. 2e), conglomerates (Fig. 2f), and the trough cross-bedded sandstones (Fig. 2g) at the base of bed 22 indicate that sediments were deposited in fluvial channels during this interval. These individual channels have width-to-depth ratios characteristic of low-sinuosity channels, indicating an anastomosing rather than a meandering channel system (e.g., Ji et al., 2022). The conglomerates (Fig. 2f) and the dominance of massive bedding (Fig. 2e) within the channel fill also indicate rapid deposition and increased fluvial erosion during peak flow conditions, but the channel bases are not strongly erosional (Fig. 2e), which indicates only moderate variance of peak discharge (e.g., Ji et al., 2022). In sum, we consider that sedimentological changes occurred in the terrestrial Permo-Triassic environments in the study area, but there was no abrupt transition in fluvial styles around the PTME.

Terrestrial ecosystems are more likely to be adversely affected by global warming than their marine counterparts. As shown in this study, terrestrial environments underwent stepwise changes in response to intermittent shifts toward warmer and more stressful conditions, eventually culminating in the final collapse of terrestrial ecosystems (e.g., Dal Corso et al., 2022; Fielding et al., 2022; Figs. 9, 10). Volcanogenic global warming might have the greatest impact on terrestrial environments because these responded more quickly than marine settings to atmospheric changes (Figs. 2, 3, 8, 9, 10).

6. Conclusions

(1) 18 spore and 25 pollen genera have been identified in the study area through the Permian-Triassic transition interval. The age of the lower and middle parts of the Sunjiagou Formation are constrained to the Late Permian Changxingian Stage based on their spore-pollen fossil assemblages. Using biostratigraphic (spore-pollen fossils), chemostratigraphic (CIA* and Ni values), and sedimentological evidence (lithological change from greyish-green to purplish-red and accompanied by the appearance of calcareous nodules), we place the EPTC at the base of bed 20 and the PTME at the top of bed 21.

(2) The dominance of conifers, low kaolinite content and CIA* values, and large amounts of calcareous nodules in the studied strata indicate that arid and semi-arid climatic conditions prevailed through the Late Permian on the NCP. However, two increases in CIA* value (77.93 and 83.77) and kaolinite content (10% and 15%) point to two brief phases of more humid climate that interrupted the otherwise arid and semi-arid conditions.

(3) The EPTC predates the correlated level of the marine PTME in the southern NCP. The former was accompanied by rapid de-vegetation on land, frequent wildfires, and warming climatic conditions (inferred from a lithologic shift from greyish-green to purplish-red and the appearance of calcareous nodules). In contrast, the latter was accompanied by the spikes of CIA* and Ni (Ni/Al ratios) values and warming climatic conditions (inferred from the increased CIA* values).

Author contributions

PZ, MY, JL, SL, and JH designed the research. JL, PZ, MY, DB, LS, and JH analysed the data. PZ, JL, MY, LS, DB, and JH wrote the manuscript. All authors contributed to the interpretation of the data and the final manuscript.

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Competing interests

MY was employed by PetroChina. All authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

Data and materials availability

All data needed to evaluate the conclusions are present in the paper and/or the Supplementary Materials. All palynological slides are housed at the State Key Laboratory of Coal Resources and Safe Mining (Beijing).

Supplementary materials

Supplementary material for this article is available at [see Supplementary Information].

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Figure captions

Figure 1. Location and geological context for the study area. **a**, Paleogeographic reconstruction for the Changhsingian (Late Permian) showing the location of the NCP (modified from the ~252 Ma map of the webpage https://deeptimemaps.com/global-paleogeography-and-tectonics-in-deep-time) and approximate extent of the Siberian large igneous province (revised after Cao et al., 2008; Lu et al., 2020); **b**, Paleofacies map of the NCP during the Changhsingian (Sunjiagou Formation) showing the location of the study area (modified from Shang, 1997). **c**, Stratigraphic distributions from the uppermost Shangshihezi (SSHZ) to the lowermost Liujiagou formations of the Dayulin section in the Yiyang Coalfield. Position of the Permian-Triassic boundary (PTB) comes from Tu et al. (2016). Abbreviations: J = Junggar Basin; K = Karoo Basin; S = Sydney Basin; B = Bowen Basin; ZSY = Zishiya; SCH = Shichuanhe; DYL = Dayulin; F. = Formation; M. = Member; B. = Bed; Litho. = Lithology.

Figure 2. Field photos from the Dayulin section in the Yiyang Coalfield during the late Permian-early Triassic transition. **a**, Stratigraphic distributions from the uppermost Shangshihezi (SSHZ) to the lowermost Liujiagou formations of the Dayulin section in the Yiyang Coalfield showing the location of samples in this study; **b**, The Sunjiagou (beds 15 to 24) and Liujiagou (bed 25) formations with the upper dashed line marking the formational contact (the area highlighted in a box is enlarged in 2h); **c**, Lithological contact between beds 19 and 20 (marked by a dashed line) that records the onset of the end-Permian terrestrial ecosystem collapse (EPTC; the highlighted box is enlarged in 2d; **d**, Enlargement of area in box in 2c showing calcareous nodules from the Sunjiagou Formation; **e**, Lithological contact between beds 21 and 22 (marked by a dashed line) showing concave-up erosional surfaces with highlighted boxes enlarged in 2f and 2g; **f**, Enlargement from 2e showing details of the conglomeratic sandstone at the base of bed 22; **g**, Enlargement from 2e showing details of trough cross-bedded sandstone at the base of bed 22; **h**,

Enlargement from 2b showing details of wave ripples in bed 23. Note that the length of the pen in Figures 2d, 2f, and 2h is 15 cm. Abbreviations: F. = Formation; M. = Member; P. = Pingdingshan sandstone; Sa. = Sample.

Figure 3. Results of spore-pollen contents (including species and genera), generic richness, charcoal abundance, Ni concentrations, Ni/Al ratios, CIA* values, clay minerals, TOC, and $\delta^{13}C_{org}$ from the Dayulin section in the Yiyang Coalfield. Note the TOC and $\delta^{13}C_{org}$ data are from Wu et al. (2020). Abbreviations: F. = Formation; M. = Member; P. = Pingdingshan sandstone; Sa. = Sample; 1 = *Limatulasporites*; 2 = *Leiotriletes*; 3 = *Osmundacidites*; 4 = *Reticulatasporites clathratus*; 5 = *Laevigatosporites*; 6 = *Striotatospora multifasciatus*; 7 = *Crassispora*; 8 = *Densosporites*; 9 = *Spinozontriletes*; 10 = *Planisporites*; 11 = *Raistrickia*; 12 = *Cyclogranisporites*; 13 = *Lophotrilotes*; 14 = *Verrucosisporites*; 15 = *Anapiculatisporites*; 16 = *A. incundus*; 17 = *Calamospora*; 18 = *Punctatisporites*; 19 = *Inaperturopollenites*; 20 = *Cycadopites*; 21 = *Ephedripites*; 22 = *Perinopollenites*; 23 = *Potonieisporites*; 24 = *Crucisaccites*; 25 = *Pityosporites*; 26 = *Alisporites*; 27 = *Platysaccus*; 28 = *Klausipollenites*; 29 = *Vesiccaspora*; 30 = *Vesiccaspora giganteus*; 31 = *Vesiccaspora minor*; 32 = *Florinites*; 33 = *Cordaitina*; 34 = *Chordasporites*; 35 = *Lueckisporites*; 36 = *Gardenasporites*; 37 = *Vestigisporites* 38 = *Limitisporites*; 39 = *Taeniaesporites*; 40 = *Lunatisporites*; 41 = *Protohaploxypinus*; 42 = *Costapollenites globosus*; 43 = *Conifers Spp*; CIE = Organic carbon isotope excursion.

Figure 4. Selected photos of representative palynological genera from the Dayulin section in the Yiyang Coalfield (all scale bars = 20 μm). **a** = *Lophotrilotes* spp.; **b** = *Planisporites* sp.; **c** = *Spinozontriletes* sp.; **d** = *Cyclogranisporites* spp.; **e** = *Craassisporites* sp.; **f** = *Reticulatasporites clathratus*; **g** = *Calamospora* sp.; **h** = *Punctatisporites* spp.; **i** = *Vesiccaspora* sp.; **j** = *Klausipollenites* sp.; **k** = *Ephedripites* sp.; **l** = *Cycadopites* sp.; **m** = *Gardenasporites* spp.; **n** = *Chordasporites* spp.; **o** = *Crucisaccites* sp.; **p** = *Platysaccus* sp.; **q** = *Protohaploxpinus* spp.; **r** =

Lunatisporites spp.; **s** = *Verrucosisporites* sp.; **t** = *Costapollenites globosus*; **u** = *Folorintes* spp.; **v** = *Vestigisporites* spp.; **w** = *Klausipollenites* spp..

Figure 5. A-CN-K (Al_2O_3-Na_2O+CaO*-K_2O) diagram with CIA scale on the left and CIA (CIA*)-WIP (WIP*) diagram for the Dayulin section. **a**, A-CN-K diagram of mudstone samples from Changhsingian to early Induan with the chemical index of alteration (CIA) scale to the left, showing the possible influence of potassium metasomatism (e.g., Fedo et al., 1995); **b**, CIA (CIA*) -WIP (WIP*) diagram for discriminating sedimentary recycling and chemical weathering influences on the mudstone samples from Changhsingian to early Induan of the Dayulin section. Potassium metasomatic effects are corrected to obtain CIA* and WIP* values. After correcting the diagenetic K enrichment, the analysed mudstones and the average source rock display a linear relationship between CIA* and WIP*, which indicates a first-cycle and chemical weathering trend (Garzanti et al., 2013).

Figure 6. Plots of **a**, CIA vs CIA*; **b**, Al_2O_3/SiO_2 ratio vs CIA*; **c**, Al_2O_3/SiO_2 ratio vs WIP; and **d**, Al_2O_3/SiO_2 ratio vs K/Si from the Dayulin section in the Yiyang Coalfield.

Figure 7. Photomicrographs showing microstructure characteristics of palynofacies in the Dayulin section of the Yiyang Coalfield. **a**, overview showing characteristics of charcoal (transmitted light, sample YY-10); **b**, charcoal (transmitted light, sample #YY-9); **c**, fungi (transmitted light, sample YY-15); **d**, spore (transmitted light, sample #YY-10); **e**, **f**, and **g**, plant cuticles (transmitted light, samples YY-5, YY-8, and YY-11).

Figure 8. Records of the CIA contents across the PTME in terrestrial and marine sections including this study; the ZK21-1 terrestrial sequence from a borehole in the Yuzhou Coalfield of North China (NC, Lu et al., 2020); the Yima and Shichuanhe terrestrial successions from NC (CIA date from Cao

et al., 2019; CIE date from Wu et al., 2020; Zircon U-Pb age from Guo et al., 2022); the Dalongkou terrestrial succession from Xinjiang Province, northwest China (Cui and Cao, 2021); the Xinmin marine succession from Guizhou Province, South China (conodont zones from Zhang et al., 2014; CIA data from Shen et al., 2013 and Zhao and Zheng, 2015); and the Chaohu marine succession from the Lower Yangtze basin of South China (conodont zones from Zhao et al., 2007; CIA data from Chen et al., 2011 and Cao et al., 2019). Abbreviations: CIE = Organic carbon isotope excursion; NW China = northwestern China, PTME = Permian-Triassic mass extinction.

Figure 9. Correlation of volcanic and biotic events in marine and non-marine successions and magmatic phases of the Siberian Traps large igneous province (LIP) based on conodont zones (black lowercase letters), carbon isotope stratigraphy, Ni spikes, and U-Pb zircon dates (red "U-Pb"). Meishan data are from Jin et al. (2000), Shen et al. (2011), Burgess et al. (2014), Burgess and Bowring (2015); Grasby et al. (2017), Song et al. (2018), Wang et al. (2018), and Shen et al. (2019a). Global palaeogeographic map of the Late Permian (~252 Ma) modified from the webpage https://deeptimemaps.com/global-paleogeography-and-tectonics-in-deep-time. Note: a = *Clarkina changxingensis*; b = *C. yini*; c = *C. meishanensis*; d = *Hindeodus changxingensis*; e = *C. taylorae*; f = *H. parvus*; g = *Isarcicella staeschei*; h = *I. isarcica*. Chinahe data are from Chu et al. (2020). High-southern-latitude data are from Fielding et al. (2019), Gastaldo et al. (2020), and Frank et al. (2021). ID-TIMS zircon U-Pb age in North China is from Guo et al. (2022). Siberian Trap volcanic events are from Burgess and Bowring (2015) and Burgess et al (2017). Pale colors in the Siberian Traps LIP record show ranges including measurement errors. Abbreviations: LNL = Lower northern latitudes; Ni = Ni spike; EPTC = end-Permian Terrestrial Collapse; PTME = Permian-Triassic Mass Extinction.

Figure 10. Schematic reconstructions examining how the Siberian Traps large igneous province (SLIP) might have driven climatic and environmental changes in the Late Permian (a–c) and Early

Triassic (d) in North China. Abbreviations: WSW = west-southwest; SNCP = southern North China Plate; EPTC = end-Permian Terrestrial Collapse; CIE = Organic carbon isotope excursion; PTME = Permian-Triassic Mass Extinction.