**Carbon Cycle Perturbation Expressed in Terrestrial Permian–Triassic Boundary
Sections in South China**

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Abstract

Stable isotopes of inorganic and organic carbon are commonly used in chemostratigraphy to correlate marine and terrestrial sedimentary sequences based on the assumption that the carbon isotopic signature of the exogenic carbon pool dominates other sources of variability. Here, sediment samples from four Permian–Triassic boundary (PTB) sections of western Guizhou and eastern Yunnan provinces in South China, representing a terrestrial to marine transitional setting, were analyzed for $\delta^{13}\text{C}$ of organic matter ($\delta^{13}\text{C}_{\text{org}}$). These values were subsequently compared to published $\delta^{13}\text{C}$ values of carbonates ($\delta^{13}\text{C}_{\text{carb}}$) from the Global Stratotype Section and Point at Meishan and many other marine and terrestrial sections. A similar isotopic trend evident through all four sections is characterized by a negative shift of 2–3 ‰ at the top of the Xuanwei Formation, where we tentatively place the PTB. This negative shift also corresponds to a turnover in the vegetation and the occurrence of fungal spores, which is generally interpreted as a proliferation of decomposers and collapse of complex ecosystems during the end-Permian mass extinction event. Moreover, the absolute values of $\delta^{13}\text{C}_{\text{org}}$ are more extreme in the more distal (marine) deposits. The $\delta^{13}\text{C}_{\text{org}}$ values for the studied sediments are more variable compared to coeval $\delta^{13}\text{C}_{\text{carb}}$ records from marine records especially in the interval below the extinction horizon. We contend that the depositional environment influenced the $\delta^{13}\text{C}_{\text{org}}$ values, but that the persisting geographic $\delta^{13}\text{C}_{\text{org}}$ pattern through the extinction event across the four independent sections is an indication that the atmospheric $\delta^{13}\text{C}$ signal left an indelible imprint on the geologic record related to the profound ecosystem change during the end-Permian extinction event.

Keywords: Carbon Isotopes, Organic Matter, Permian–Triassic Boundary, South China, Mass Extinction

1. Introduction

The close of the Permian (ca. 252 Ma) is accompanied by the largest biodiversity loss in the Phanerozoic (Erwin, 2006). About 79% of marine invertebrate genera suffered extinction (Payne and Clapham, 2012) and some 70% of terrestrial animal species disappeared (Benton et al., 2004; Sahney and Benton, 2008; Ward et al., 2005), including insects (Labandeira and Sepkoski, 1993).

Land plants also underwent extinction and turnover across the Permian–Triassic transition (Vajda and McLoughlin, 2007; Vajda and Bercovici, 2014), but the timing of the extinction is not well constrained (McLoughlin et al., 1997; McElwain and Punyasena, 2007; Hilton and Cleal, 2007; Peng and Shi, 2009; Shi et al., 2010). Plant communities responded to climatic changes during the Late Permian (Rees, 2000; Rees et al., 2002) resulting in a more complex and diachronous extinction pattern across latitudes (Wang, 1996; Coney et al., 2007; Lindström and McLoughlin, 2007; Retallack et al., 2007; Yu et al., 2007; Xiong and Wang, 2011). In the earliest Triassic, the typical Permian floras were replaced by succession of pioneering lycopsids and ferns (Retallack, 1997; Looy et al., 1999; 2001; Wang and Chen, 2001; Hochuli et al., 2010; Yu et al., 2010; Hermann et al., 2011; Bercovici et al., 2015) or conifers (Schneebeili-Hermann et al., 2012). This drastic turnover is generally associated with a major change in depositional environments (Gastaldo et al., 2005; Bourquin et al., 2011; Bercovici et al., 2015) and indication of altered ecological conditions by the “fungal-spike” event (Eshet

et al., 1995; Visscher et al., 1996; 2004; 2011; Vajda, 2012). Biomarkers (lipids) also imply destruction of land vegetation causing dramatic increase in soil erosion during the end-Permian extinction event (e.g. Sephton et al., 2005; Xie et al., 2007; Wang, 2007; Luo et al., 2013).

The environmental stress during the end-Permian extinction event has generally been attributed to greenhouse gas-induced global warming (e.g. Payne and Clapham, 2012). Evidence for global warming is reflected in oxygen isotopes of unaltered fossils (e.g. Sun et al., 2012), indicating a warming of up to 9 °C in the tropical regions across the Permian–Triassic boundary (PTB) (Joachimski et al., 2012). Evidence supporting the hypothesis that CO₂ (e.g. Cui et al., 2013; Cui and Kump, 2015; Hermann et al., 2010; Clarkson et al., 2015) was the driving factor of the end-Permian warming derives from carbon isotopes ($\delta^{13}\text{C}_{\text{carb}}$), which exhibit a 2-4‰ negative excursion indicating a profound perturbation of the global carbon cycle in both marine and terrestrial environments (e.g. Ward et al., 2005; Korte and Kozur, 2010; Hermann et al., 2011; Zhang et al., 2015) during the end-Permian extinction event (EPE).

Although the link between increased CO₂ levels and global warming is evident, the exact trigger of CO₂ release through the end-Permian is a topic of continuing debate (Svensen et al., 2009). Many suggest that the most probable trigger of this event was the eruption of the Siberian Traps volcanism (Burgess et al., 2014; Burgess and Bowring, 2015). This has led to many studies on the environmental consequences of the Siberian Traps eruption, including the release of halogens that destroy the ozone layer, and ¹³C–depleted CO₂ generated by interactions between magma and organic-rich sediments (Svensen et al., 2009; Black et al., 2012, 2014; Cui et al., 2015). However, study of the

tempo and synchronicity of the EPE requires a solid correlation framework between marine and terrestrial sections. In addition, the identification of the PTB in terrestrial section remains challenging due to the scarcity of index fossils (e.g. [Marshall, 2005](#); [Ward et al., 2005](#)).

Unlike inorganic carbon isotopes, which have been widely used in stratigraphic correlation in carbonate-rich marine sediments (e.g., [Saltzman and Thomas, 2012](#)), organic carbon isotope chemostratigraphy has often been used in sediments that lack carbonates ([Erbacher et al., 2005](#)). In addition to stratigraphic correlation between marine and terrestrial sequences, carbon isotopes of terrestrial organic matter can also provide information on the paleoenvironment ([Gröcke, 1998](#)). The terrestrial and paralic EPE sections in Guizhou and Yunnan provinces in South China provide a unique opportunity to correlate terrestrial records with marine, largely due to the well-preserved macrofloral assemblages and organic matter in strata preceding the PTB ([Peng et al., 2005, 2006](#); [Yu, 2008](#)). This study characterizes the isotopic signature of four terrestrial PTB sections in Guizhou and Yunnan provinces in South China within a biostratigraphic framework and investigates the causes of the carbon cycle perturbation during the end-Permian extinction event.

2. Geologic Setting

The study area is located in western Guizhou and eastern Yunnan in southwestern China (Fig. 1), on the southwest margin of the Yangtze Craton. Laterally continuous Permian–Triassic sections record gradual changes from terrestrial to marine facies ([Bercovici et al., 2015](#)). The strata exposed in this region include the Middle Permian

Maokou Formation, Emeishan basalt Group, the Upper Permian Xuanwei Formation, the Lower Triassic Kayitou and the Dongchuan formations. Terrestrial exposures preserving the EPE are accessible in rural areas in western Guizhou and eastern Yunnan, including the Chahe section (48R 0385932, UTM 2953327), Zhejue section (48R 0394488, UTM 2943109), Jiuchaichong section (48R 0398848, UTM 2953327), and Mide section (48R 0444118, UTM 2888507), from west to east (Fig. 1). During the past three decades, major studies have been carried out on the terrestrial sections of the eastern Yangtze Platform by several working groups. Initially by paleobotanists at the Nanjing Institute of Geology and Paleontology ([Ouyang and Li, 1980](#); [Ouyang, 1982, 1986](#)). Regained interest for these sections led to numerous stratigraphical and paleontological studies led by the group of the China University of Geosciences Wuhan ([Wang and Yin, 2001](#); [Zhao, 2003](#); [Zhang et al., 2006](#); [Yin et al., 2007](#); [Yu et al., 2007, 2008, 2010, 2015](#); [Peng and Shi, 2009](#) and references therein), and again by the researchers of the Nanjing Institute of Geology and Palaeontology ([Fang, 2004](#); [Shen et al., 2011](#); [Zhang et al., 2015](#)).

The four studied sections were also lithologically and sedimentologically described in a recent publication by some of the authors of this study ([Bercovici et al., 2015](#)) and the stratigraphic observations were correlated to match the previous bed numbering scheme following [Zhao \(2003\)](#) and [Yu \(2008\)](#) on the basis of fining upwards genetic sequence represented by a coarser rock unit (sandstone to siltstone) capped by a fine unit (siltstone to mudstone) (Fig. 2). The terrestrial EPE sections of western Guizhou and eastern Yunnan show sub-vertical bedding, and strong deformation including numerous faults and zones of crushed rocks resulting from the Yanshanian orogeny, meaning that the studied sections are affected by tectonism.

The successions treated in this study comprise three formations; the Xuanwei Formation of Late Permian age, is equivalent to Wuchiapingian and Changhsingian and comprises light-colored tan and gray siltstones alternating with sandstone units (Bercovici et al., 2015). The Kayitou Formation conformably overlies the Xuanwei Formation, and is exposed as yellow, silty sandstone units (Bercovici et al., 2015). Resting conformably on the Kayitou Formation follows the Dongchuan Formation, which is composed of cross-stratified, alternating purple-red siltstone and sandstone units (Bercovici et al., 2015). The age of the transitional facies between the Xuanwei and Kayitou formations and the exact position of the PTB is a subject of debate, mainly because of the lack of a biostratigraphical definition for the PTB in terrestrial settings and due to the disappearance of fossils associated with paleoenvironmental changes and the onset of redbeds across the Kayitou and Dongchuan formations (Bercovici et al., 2015). Although the PTB in terrestrial settings are difficult to constrain, it is clear that the onset of the terrestrial EPE occurs at the boundary of the Xuanwei Fm. and the Kayitou Fm. We therefore refer the strata below the top of the Xuanwei Fm. as pre-EPE and the strata above the bottom of the Kayitou Fm. as post-EPE.

3. Biostratigraphy

Abundant fossil plant assemblages have been recovered from the Xuanwei Formation, represented by the typical Cathaysian Late Permian *Gigantopteris* flora and described in detail within Yu (2008). Spore and pollen assemblages were also documented from the Xuanwei Formation (Ouyang and Li, 1980; Ouyang, 1982, 1986; Wang and Yin, 2001; Zhang et al., 2006; Peng et al., 2006; Yin et al., 2007; Yu et al.,

2007, 2008, 2010; Peng and Shi, 2009), but the recovery is very low and limited to a few stratigraphic occurrences (Bercovici et al., 2015). Nonetheless, the assemblages recovered are almost exclusively represented by spores, consistent with the macroflora, dominated by a *Gigantopteris* wetland flora that inhabited Late Permian swamps and shallow lacustrine environments (Bercovici et al., 2015). The successions at the top of the Xuanwei Formation host fungal spores, including *Reduviasporonites* (Bercovici et al., 2015). The occurrence and proliferation of the saprophytic fungus *Reduviasporonites* has been linked to terrestrial ecosystem collapse associated with the EPE in many localities worldwide (Sephton et al., 2009; Visscher et al., 2004; Visscher et al., 2011; Vajda, 2012).

The Kayitou Formation has been interpreted as a coastal brackish lagoon facies and marks a major change in facies and depositional environment. Plant fossils are found as scarce, scattered, and small comminuted debris. New plant taxa appear, including the Triassic lycopsid *Annalepis* (Yu et al., 2010) and foliage and fertile organs of peltasperms. These collectively constitute a distinctive *Annalepis*–*Peltaspermum* assemblage (Yu et al., 2010). The first occurrence of typical Triassic (Induan) bivalves, spinicaudatans, and ammonites is reported from the basal part of the Kayitou Formation (Yu et al., 2010; Chu et al., 2013). The Kayitou Formation is barren of pollen and spores (Bercovici et al., 2015), and all fossils gradually disappear within the lowermost few meters of the Kayitou Formation. The purple-red Dongchuan is barren, with no trace of organic matter due to oxidizing conditions.

4. Methods

4.1. Carbon isotopes

In order to perform chemostratigraphy and to gain insights into the carbon cycle perturbation during the EPE, we collected samples from exposures at Chahe, Zhejue, Jiucaichong and Mide within the Xuanwei and Kayitou formations. The samples for bulk organic carbon isotope and total organic carbon (TOC) content analysis were collected at the top of each bed within fine-grained rocks.

Samples between 10 and 30 grams were collected from every bed within the four terrestrial sections, a total of 133 samples. All samples were washed with deionized distilled water (DDW) to remove any surface contamination prior to extraction. Five to ten gram samples were crushed in a tungsten carbide ball mill via a SPEX 8000 Mixer resulting in a homogeneous powdered sample (size $\sim 75 \mu\text{m}$). The grinding jar was cleaned between samples with DDW. To remove carbonate material (siderite was present in some samples), powdered samples were reacted with 10% HCl (w/v) in centrifuge tubes for 24 hours and rinsed with DDW repeatedly until pH was ~ 6 . Samples were then freeze-dried for isotope analysis. Isotopic analyses and weight percent data for carbon were performed by continuous flow on an elemental analyzer (EA; Costech ECS4010) coupled to an isotope ratio mass spectrometer (IRMS; Thermo Finnegan Delta Plus XP), via an open-split interface. All analyses were performed in the Stable Isotope Biogeochemistry Lab at The Pennsylvania State University. Powdered, decarbonated dried samples were weighed and sealed in tin boats for isotopic analysis. Samples were combusted at 1020 °C with a ‘zero blank’ helium atmosphere autosampler that has been retrofitted to include a custom vacuum purging and He-bleed system. Data are reported using delta notation relative to the Vienna Pee Dee Belemnite International Standard (V-

PDB) for carbon. Reference gases were calibrated relative to standard ANU sucrose and USGS 24 (graphite) for carbon in combination with in-house glycine, cornstarch, caffeine and Peruvian mud isotopic standards. Carbon isotope values were normalized to the VPDB scale using laboratory standards (calibrated with IAEA standards) and a two-point calibration following the methods of [Coplen et al. \(2006\)](#). Precision (1σ) for C isotopes was monitored with standards across all analyses and was determined to be 0.1‰. Accuracy for C isotopes was determined by measuring the difference between expected and measured values of standards (in-house standard from Peru) treated as samples across all analyses and was 0.2‰.

All samples were run with the ‘macro’ oxygen loop at 1.2 bars pressure to ensure complete combustion of samples. The efficiency of combustion and fidelity of isotope values were confirmed through the use of a thermally mature Devonian black shale house standard. Carbon percentage on the decarbonated fraction was produced on a Costech EA in conjunction with isotopic measurements. Carbon isotopic peak heights were calibrated to a Peru Mud standard of known elemental composition with a standard error of ± 0.1 wt. % for carbon.

4.2. Statistical t-test

T-test with Welch correction is widely used to test for the difference in the mean values of two populations when the variances are unequal ([Welch, 1947](#)) and when the sample size from two populations is different. This approach is adopted to evaluate whether the difference between two means (pre-EPE and post-EPE) at the given probability level (95%) is significant. A probability of $p = 0.05$ (95% probability of

making a correct statement) is usually used in statistical studies. The mean, standard error, degree of freedom, and t values used to evaluate the differences are listed in Table 1. Also listed on the Table 1 is the Welch corrected degree of freedom and the tabulated t value for comparison purpose. If the calculated t value exceeds the tabulated value we say that the means for the pre-EPE and post-EPE at each location are significantly different at that level of probability (with 95% confidence). The calculation is performed in R, using function t.test for two independent groups of data.

5. Results

5.1. Carbon isotopes

The $\delta^{13}\text{C}_{\text{org}}$ values vary in both absolute values and the magnitude of carbon isotope excursions (CIEs) across the four studied transects. Overall, the $\delta^{13}\text{C}_{\text{org}}$ values range from -20.5‰ to -30‰, with the most ^{13}C -enriched mean value occurring at Zhejue and the most ^{13}C -depleted mean value occurring at Mide (Table 1). The isotopic gradients amongst the four sections persist through the terrestrial extinction interval across the four sections as shown in Fig. 3. There appears to be a systematically decreasing trend from west to east with the exception of Jiuchaichong. However, the decline of the $\delta^{13}\text{C}_{\text{org}}$ values occurs at the boundary of the Xuanwei and Kayitou Fm. across the four successions. We calculate the magnitude of the carbon isotope excursion as the difference between the stratigraphically averaged $\delta^{13}\text{C}_{\text{org}}$ values for the interval prior to the terrestrial EPE (pre-EPE) and the interval post the terrestrial EPE (post-EPE) (Table 1). The magnitude of CIE across the terrestrial EPE is about -2.6‰ at Jiuchaichong, Chahe and Mide, and is only about -0.7‰ at Zhejue. The spread of data from the four successions appears to be

the largest in the longest section at Chahe, and is the least in the Zhejue section where the smallest magnitude of CIE is recorded. The Welch corrected t test suggests that at all studied locations, the pre-EPE and post-EPE population means are significantly different at the 95% confidence level (Table 1).

5.2 Total organic carbon

The wt.% TOC from the four successions ranges from 0.02% to 52.15%, with the highest mean value at Zhejue (4.8%) and the lowest mean value at Mide (0.3%). We explore the dependence of the wt.% TOC on sampled lithology by regression analysis on $\delta^{13}\text{C}_{\text{org}}$ and 1/TOC and color-coding on lithology (Fig. 4). We find no significant correlation between TOC and lithology in any of the studied sections ($r^2 \leq 0.24$, Fig. 4). Thus, the observed $\delta^{13}\text{C}_{\text{org}}$ is independent of the wt.% TOC and lithology. This trend is consistent across the four sections from Jiuchaichong to Mide for the pre- and post-PTB samples, suggesting no obvious relationship between the depositional environment and the observed magnitude of CIE (Fig. 4).

6. Discussion

6.1. Relationship between “fungal spike” and $\delta^{13}\text{C}_{\text{org}}$

A “fungal spike” is defined as the stratigraphic level where abundant *Reduviasporonites* spores occur (i.e., $\geq 95\%$ of palynomorphs) (Eshet et al., 1995). Some have proffered an affinity with *Tympanicysta* (green algae) (Afonin et al., 2001; Foster et al., 2002). However, geochemical and isotopic signatures indicate that *Reduviasporonites* are indeed fungi instead of green algae (Sephton et al., 2009). *Reduviasporonites* fungal

275 spores have been identified in numerous PTB locations, both marine and terrestrial,
276 including western Spitsbergen (Mangerud and Konieczny, 1993; Wignall et al., 1998),
277 North America (Utting, 1994), Israel and Italy (Eshet et al., 1995; Visscher et al., 1996),
278 Madagascar (Wright and Askin, 1987), South Africa (Steiner et al., 2003), and South
279 China (Ouyang and Utting, 1990; Bercovici et al., 2015). Most of these are associated
280 with the ecosystem collapse and are located very close to the PTB. *Tympanicysta*
281 reported in end-Permian successions in Meishan occurs several meters below the
282 extinction horizon, and thus may not be associated with ecosystem collapse (Ouyang and
283 Utting, 1990). The fungal spore spike reported by Steiner et al. (2003) at the base of the
284 Katberg Formation in the Karoo Basin, South Africa has later been suggested to be well
285 above the biostratigraphically defined mass extinction (Ward et al., 2005). Visscher et al.
286 (1996) suggest that the worldwide fungal proliferation at the PTB can only have
287 originated from arboreal vegetation dieback.

288 The co-occurrence of a fungal spike and a negative CIE at the PTB has been
289 recognized in a number of localities. Report of *Tympanicysta* (c.f. *Reduviasporonites*) by
290 Mangerud and Konieczny (1993) in Spitsbergen has been used as an important PTB
291 marker, which is coeval with significant prominent negative shift in $\delta^{13}\text{C}_{\text{org}}$ in EPE
292 successions in Spitsbergen (Wignall et al., 1998). The abundance spike of fungal spores
293 in the European Southern Alps reaches close to 100% at the level where $\delta^{13}\text{C}_{\text{carb}}$ ($\delta^{13}\text{C}$ of
294 carbonate) declines significantly across the Bellerophon–Tesero transition, which is close
295 to the PTB (Visscher et al., 1996). Similarly, the fungal spike also co-occurs with the
296 PTB in southern Israel, albeit with a more prolonged occurrence (Eshet et al., 1995;
297 Visscher et al., 1996).

298 Evidence for a Late Permian plant dieback that induced *Reduviasporonites*
 299 proliferation is available from limited sources. Taxonomic analysis of leaf remains and
 300 reproductive organs suggests that Late Permian conifers became extinct at the family
 301 level (Clement-Westerhof, 1984; Clement-Westerhof, 1987, Looy et al., 1999). In South
 302 China, the wetland biota represented by the *Gigantopteris* flora largely disappeared at the
 303 end of the Permian (Bercovici et al. 2015). Questions on the validity of using
 304 *Reduviasporonites* as a reliable marker for the PTB or ecosystem collapse have arisen
 305 because of the uncertain biological and ecological identity of the fossil remains. Recently
 306 Visscher et al. (2011) argued that indeed the global occurrence of *Reduviasporonites* as a
 307 type of soil-borne pathogenic fungi must have been associated with the worldwide Late
 308 Permian forest decline, itself the expected consequence of the catastrophic eruption of the
 309 Siberian Traps. Massive CO₂ and/or CH₄ release caused heat stress-related tree mortality
 310 (e.g. Allen et al., 2010), and may have been coupled to putative halogen gas injection into
 311 the stratosphere disrupting the ozone layer and causing increased UV-B radiation, leading
 312 to mutant pollen that has been preserved in the fossil record (Black et al., 2012).
 313 Bercovici et al. (2015) reported a fungal *Reduviasporonites* spore occurrence at
 314 the top of the Xuanwei Formation at Jiuchaichong and Mide, while Zhejue and Chahe
 315 show a diversity of other fungal spore species but no *Reduviasporonites*. The occurrence
 316 of *Reduviasporonites* in Jiuchaichong and Mide also occur at the negative 2.6‰ shift in
 317 $\delta^{13}\text{C}_{\text{org}}$, consistent with a global occurrence of *Reduviasporonites* close to the PTB
 318 irrespective of depositional environment, floral provinciality and climatic zonation as
 319 proposed by Visscher et al. (1996).
 320

6.2. Placement of the PTB

As the placement of the PTB at the Meishan GSSP is defined based on First Appearance Datum (FAD) of the conodont *Hindeodus parvus*, a strictly marine fossil, it is not possible to directly define the PTB on land until a detailed correlation has been performed. Currently, there are two opposed opinions, the view of researchers at the China University of Geosciences Wuhan (Wang and Yin, 2001; Zhao, 2003; Zhang et al., 2006; Yin et al., 2007; Yu et al., 2007, 2008, 2010, 2015; Peng and Shi, 2009 and references therein), who estimated the PTB within the upper part of the Xuanwei Formation, and that of researchers at the Nanjing Institute of Geology and Paleontology (Shen et al., 2011; Zhang et al., 2015), who estimated the PTB much higher, within the lower part of the Dongchuan Formation. No clear criteria exist for the definition of the PTB in the terrestrial realm, and correlations have been based on different criteria by different groups of researchers. Therefore, we discuss two ways that are relevant to this study about the placement of the PTB.

6.2.1. Biostratigraphy

The End Permian is complex and is likely punctuated by two extinction events in the marine realm (Song et al., 2013, Fig. 5). Therefore, the PTB is bracketed by an interval where typical Permian taxa disappear while new Triassic taxa appear, defining a mixed "Permian–Triassic" assemblage (Fig. 5). In the terrestrial sections, numerous plant fossils (Wang and Yin, 2001; Zhao, 2003; Zhang et al., 2006; Yin et al., 2007; Yu et al., 2007, 2008, 2010, 2015; Peng and Shi, 2009; Bercovici et al., 2015) characterize the flora across the Permian–Triassic transition. Two distinct plant assemblages occur within the

Xuanwei and Kayitou formations in western Guizhou and eastern Yunnan: The *Gigantopteris* flora of the Upper Xuanwei Formation, and the *Annalepis* – *Peltaspermum* assemblage in the basal part of the Kayitou Formation (Yu et al., 2010; Bercovici et al., 2015). The *Gigantopteris* flora (*Gigantonoclea guizhouensis* – *Annularia pingloensis* assemblage), which include wetland cordaitales, sphenopsids and pteridosperms, suggesting a warm and humid climate, has previously been attributed to the Late Permian (Changhsingian) whilst the *Annalepis* – *Peltaspermum* assemblage has been attributed to the Early Triassic (Induan; Yu et al., 2010). In the lowermost part of the Kayitou Formation, the FADs of typical Triassic (Induan) bivalves, spinicaudatans (Chu et al., 2013), and ammonites (*Ophiceras* sp., Yu et al., 2010), associated with the Triassic plant genus *Annalepis* in the coastal Mide and Tucheng sections (Yu et al., 2010), support this interpretation. The EPE in the terrestrial section is marked by a turnover and by the demise of the *Gigantopteris* flora, albeit some relicts persist within the *Annalepis* – *Peltaspermum* assemblage. Moreover, the EPE and terrestrial ecosystem collapse is marked by the occurrence of the fungal spore *Reduviasporonites*, which was identified at the top of the Xuanwei Formation at Jiucaichong and Mide sections (Bercovici et al., 2015).

Zhang et al. (2015) further interpret the disappearance of plant fossils as a biological signal in response to the EPE stating that: "The uppermost part of the Kayitou Formation and the overlying Dongchuan Formation is characterized by the deposits with little spores and pollen, therefore, indicates a rapid deforestation during the end-Permian in South China. The most probable cause is a widespread die-off of the vegetal cover leading to the destabilization of the channel banks and soil system on land and a

progressive increase of temperatures". We argue that the primary cause affecting fossil preservation is facies and associated depositional environments, and that we showed a major regional facies change between the Xuanwei and Kayitou formations that likely account for the preservational differences and the disappearance of fossils altogether in the basal few meters of the Kayitou Formation (Bercovici et al., 2015). The Dongchuan Formation is known to be barren because organic matter is destroyed under oxidizing conditions, as indicated by the purple-red color of the sandstone facies (Bercovici et al., 2015).

6.2.2 Carbon isotopes

High-resolution radiometric dating on ash layer Bed 25 and Bed 28 at Meishan suggests the PTB (252.17 ± 0.06 Ma) follows the onset of the EPE (~ 252.3 Ma) by about 130,000 years (Shen et al., 2011). The rapid drop in $\delta^{13}\text{C}_{\text{carb}}$ is contemporaneous with the onset of the EPE, and predates the PTB (e.g. Luo et al., 2011; Shen et al., 2011). This carbon isotope excursion event has been modeled extensively to estimate the rate and amount of C release from several proposed sources (Berner, 2002; Rampino and Caldeira, 2005; Cui et al., 2013; Cui et al., 2015). The negative CIEs in terrestrial organic matter have been used as a chemostratigraphic marker to correlate the PTB (see Korte and Kozur 2010 for a review). However, many factors complicate the CIEs expressed in continental successions partly due to the discontinuities in terrestrial sedimentation, inconsistent rates and hiatuses along with heterogeneities in vegetation and organic matter source, transport and diagenesis. Due to these complexities, $\delta^{13}\text{C}_{\text{org}}$ itself is not a reliable indicator of global CIE events without independent stratigraphic markers (de Wit

et al., 2002) because spikes of $\delta^{13}\text{C}_{\text{org}}$ toward more ^{13}C -depleted values could easily be caused by local perturbations, such as soil respiration, mixing of organic matter of different age, methanotrophic activity, and/or source variations (Krull and Retallack, 2000; Krull et al., 2000; Cui et al., 2011).

In light of this, biostratigraphy and geochronology has often been applied to continental successions to define the PTB independently without the recognition of CIEs (e.g., de Wit et al., 2002), with palynology being utilized in locations such as India and Madagascar (de Wit et al., 2002), vertebrate biostratigraphy being applied in South Africa (Smith, 1995; MacLeod et al., 2000) and high-precision U-Pb dating being used in South China (Shen et al., 2011) and the Sydney and Bowen Basins in Australia (Metcalf et al., 2014).

Despite the mixed Permian–Triassic fauna across the PTB at the highly condensed Meishan GSSP, we argue that the floral turnover and appearance of many Triassic taxa in the basal Kayitou Formation is contemporaneous to the marine PTB as recorded at the Meishan section. These evidence collectively support the potential placement of the PTB at Jiuchaichong, Chahe and Mide near the base of the Kayitou Fm. The Zhejue section lacks any Early Triassic taxa, so the placement of the PTB at this location is unsubstantiated. More data, and especially new radiometric dates to bracket the Kayitou Formation, are required to further resolve events and place the PTB more precisely in these terrestrial sections.

6.3 Controls on the CIE of $\delta^{13}\text{C}_{\text{org}}$

As [Tieszen and Boutton \(1989\)](#) stated: “An interpretation of a $\delta^{13}\text{C}$ value often requires an understanding of the $\delta^{13}\text{C}$ values of all potential contributing sources, and an understanding of potential fractionation effects associated with environmental and metabolic factors”. The $\delta^{13}\text{C}$ values of terrestrial C_3 plants (Calvin cycle) range from -35‰ to -20‰ ([Boutton, 1991](#)), which cover the range of values reported in this study (-20.5‰ to -30‰). Organic carbon isotopes of sedimentary rocks represent a mixture of preserved OM, including plant organs (e.g. roots and woody stems were more ^{13}C -enriched than leaves, [Badeck et al., 2005](#); [Cernusak et al., 2009](#)) and their product through microbial processing ([Meyers and Ishiwatari, 1993](#)). Consequently, the relative contributions of organic matter in sedimentary rocks from non-vascular plants (e.g. algae that lack woody and cellulosic tissues) *versus* vascular plants (e.g. trees, grasses and shrubs that have these tissues) can influence the $\delta^{13}\text{C}$ values. At ecosystem level, evergreens are more ^{13}C -enriched than deciduous plants ([Brooks et al., 1997](#); [Diefendorf et al., 2010](#)). For that reason, a shift in plant type can also affect the $\delta^{13}\text{C}$ values. For C_3 plants, the $\delta^{13}\text{C}$ values tend to be higher under dry relative to wet environments ([Diefendorf et al., 2010](#); [Kohn, 2010](#)). As a result, a change in moisture patterns linked to climate change can also trigger the change in $\delta^{13}\text{C}$ values in a systematic way.

Other physiological and environmental factors that affect the $\delta^{13}\text{C}$ values of plants are their photosynthetic pathways ([Farquhar et al., 1989](#); [O'Leary, 1981](#)), the atmospheric $p\text{CO}_2$ ([Popp, 1989](#)), ambient $\delta^{13}\text{C}$ value of the atmosphere ([Arens et al., 2000](#); [Tippie et al., 2010](#)), water availability ([Diefendorf et al., 2010](#)) and canopy structure ([Graham et al., 2014](#)). However, the $\delta^{13}\text{C}_{\text{org}}$ values measured in organic matter preserved in geological samples are further complicated by post-burial processes, which tend to

mobilize compounds that are depleted in ^{13}C relative to the bulk organic matter (OM) (Hayes et al., 1999). The measured organic matter in these terrestrial P-T sections represent an original mixture of carbohydrates, lipids, proteins, and other biochemicals. On geologic timescales, the selective preservation of more refractory compounds can complicate the isotopic signal from the source atmospheric CO_2 . Biomarker studies on the OM from the Chahe section suggest the sediments are not thermally mature (Zhao, 2003); thus, thermal alteration likely plays a minimal role on the preserved $\delta^{13}\text{C}_{\text{org}}$ signal at our studied sites.

Spatial variations of $\delta^{13}\text{C}_{\text{org}}$ at our studied sites could be due to a number of factors, such as the different source inputs of OM to soils, differential preservation or decomposition of certain organic components (e.g. lignin is more ^{13}C -depleted than polysaccharide, which is preferentially removed during biogeochemical processing, Benner et al., 1987; Schmidt et al., 2011), different watershed topography (Meyers and Ishiwatari, 1993) and microbial fractionation during decomposition (Schweizer and Cadisch, 1999; Wynn, 2007). However, further studies on OM provenance would be required to understand which mechanism is responsible for the observed isotopic gradients among the four studied sites. Vertical variations of $\delta^{13}\text{C}_{\text{org}}$ at our studied sites are mostly attributed to the changes in atmospheric $\delta^{13}\text{C}$ (Arens et al., 2000) and perhaps the atmospheric $p\text{CO}_2$ level (Schubert and Jahren, 2012). Moreover, the frequent avulsion and rapid channel migration can also play a role in the variations of $\delta^{13}\text{C}_{\text{org}}$ (e.g. Schmitz and Pujalte, 2007).

Sedimentary facies analysis suggests that the sources of OM from our studied sites are dominated by terrestrial OM (Bercovici et al., 2015), but are influenced by

marine input due to the facies change at the beginning of Triassic because of regional marine transgression (Lehrmann et al., 2007). Traditionally, the marine $\delta^{13}\text{C}_{\text{org}}$ records are interpreted as changes in relative proportions of sources of marine and terrestrial organic matter (Hedges and Parker, 1976; Sackett, 1964), under the assumption that the $\delta^{13}\text{C}_{\text{org}}$ from marine OM is isotopically enriched in ^{13}C compared to terrestrial OM (Deines, 1980). However, it has been noted more than 30 years ago that unlike most of the Cenozoic, marine organic carbon is more depleted in ^{13}C relative to terrestrial materials from the Paleozoic (Maynard, 1981) and the Mesozoic (Arthur et al., 1985; Dean et al., 1986). This anomalous trend has been interpreted by Popp (1989) and Arthur et al. (1985) as higher isotopic fractionation between marine carbonates and aquatic algae-derived organic matter under higher atmospheric $p\text{CO}_2$ conditions.

We compiled the average value of $\delta^{13}\text{C}_{\text{org}}$ during the pre-EPE and post-EPE intervals for globally distributed marine and terrestrial successions. These values were calculated based on the placement of the extinction event, or, in the case where the extinction event was not obvious (mostly terrestrial sections), on biostratigraphic evidence to independently determine the PTB. We then summed up the pre-EPE $\delta^{13}\text{C}_{\text{org}}$ values from all the marine and terrestrial sections listed in Table 2 to determine the mean. We also summed up the during- and/or post-EPE $\delta^{13}\text{C}_{\text{org}}$ values of all the marine and terrestrial sections listed in Table 2 but exclude the most extreme values in Krull and Retallack (2000). We found that the pre-EPE marine $\delta^{13}\text{C}_{\text{org}}$ value (-26.4‰) is about 1.5‰ more depleted than that in the terrestrial realm (-24.9‰), and that the post-EPE marine $\delta^{13}\text{C}_{\text{org}}$ value (-29.5‰) is about 2‰ more depleted than that in the terrestrial region (27.6‰).

In an effort to interpret the -2.6‰ CIE across the EPE at Jiuchaichong, Chahe and Mide (Fig. 3), we present two hypotheses with the mixing of terrestrial and marine carbon and the mixing of autochthonous and allochthonous OM inputs. We have demonstrated that the pre-EPE $\delta^{13}\text{C}_{\text{org}}$ values at Jiuchaichong, Zhejue and Chahe are all more ^{13}C -enriched relative to that at Mide and Meishan and the $\delta^{13}\text{C}_{\text{org}}$ value at Mide is also higher than that at Meishan (Fig. 3 and 5), consistent with the pre-Cenozoic observations that terrestrial $\delta^{13}\text{C}_{\text{org}}$ is more isotopically enriched than marine $\delta^{13}\text{C}_{\text{org}}$ (Arthur et al., 1985; Maynard, 1981). It is important to note that the mean values measured in Meishan are -28.4‰ for the pre-EPE, -28.2‰ for the during-EPE-interval with minimum value reaching -32.8‰, and -26.7‰ for the post-EPE (Fig. 6). This long-term positive shift in $\delta^{13}\text{C}_{\text{org}}$ values at Meishan has been interpreted by Riccardi et al. (2007) as a change in the dominant photosynthesis pathway from an algae or cyanobacteria-dominated mode to a non-oxygenic phototrophic green sulfur bacteria-dominated mode. The reverse tricarboxylic acid (TCA) cycle that green sulfur bacteria use during the incorporation of carbon results in a smaller carbon isotope fractionation than the Calvin cycle that algal or cyanobacteria use (van Breugel et al., 2005). This hypothesis seems to be supported by the observed abundance of isorenieratane, a biomarker for green sulfur bacteria, at the event horizon in Meishan and Australia (Grice et al., 2005; Luo et al., 2010). Although, a rise in marine $\delta^{13}\text{C}_{\text{org}}$ values could also reflect greater productivity, lower $p\text{CO}_2$, or shifting algal groups associated with the EPE.

6.4 Influence of organic component on $\delta^{13}\text{C}_{\text{org}}$ values

The composition of the organic residue (based on the palynofacies analyses; Table 3) for

21 samples was compared to the $\delta^{13}\text{C}_{\text{org}}$ values in order to evaluate its influence on the latter. The palynofacies data based on 300 organic particles provide reliable insights into the bulk composition of the organic material sampled for $\delta^{13}\text{C}_{\text{org}}$ values. Most samples are dominated (in terms of abundance) by wood fragments that are either charred (charcoal) or non-charred (phytoclads). Note that the small proportion of spores, pollen, fungal remains and even marine components, such as dinoflagellates, that may occur outside the palynofacies counts do not notably influence the $\delta^{13}\text{C}_{\text{org}}$ values as their mass is insignificant.

The comparison clearly shows that the fluctuations in $\delta^{13}\text{C}_{\text{org}}$ values in this study are not linked to the composition of the organic residue. Carbon isotope values range between -26.1‰ and -21.8‰ in the Chahe section. The two oldest samples in this suite contrast considerably in $\delta^{13}\text{C}_{\text{org}}$ values but have a similar palynofacies composition dominated by charcoaled plant remains indicating that the difference in $\delta^{13}\text{C}_{\text{org}}$ values is not linked to the organic content. The remaining samples have $\delta^{13}\text{C}_{\text{org}}$ values that similarly vary independently of the organic content. For example, samples with different organic composition can have similar $\delta^{13}\text{C}_{\text{org}}$ values – notably those samples from Zhejue section. A contrary pattern is evident among other samples where the organic composition is very similar but $\delta^{13}\text{C}_{\text{org}}$ values vary. The only exception may be sample 11C023 from bed 35 at Chahe in which the high content of pollen and spores (14%) may have influenced the more negative $\delta^{13}\text{C}_{\text{org}}$ value (Table 3).

6.5 Mixing of terrestrial and marine carbon

The Meishan record (Fig. 6) clearly indicates that source variations have influenced the marine $\delta^{13}\text{C}_{\text{org}}$, and leads one to question correlations based on $\delta^{13}\text{C}_{\text{org}}$ in general. On the other hand, the terrestrial data presented here indicate that persistent local offsets accompany the global shift toward lighter $\delta^{13}\text{C}_{\text{org}}$ values at the PTB (Fig. 3; Table 1). Sedimentological analysis performed by [Bercovici et al. \(2015\)](#) at our studied locations reveals that there is a vertical transition from alluvial plain facies to fluvial lacustrine facies towards the top of the Xuanwei Formation at Jiucaichong, Zhejue and Chahe, with the exception of Mide, which displays marine influence at the very top of the Xuanwei Formation, and then to marine facies at all locations in the Kayitou Formation, suggesting a regional transgression ([Bercovici et al., 2015](#)). Considering the sources of OM are likely local, the Early Triassic flooding event might have carried the same types of OM to the studied locations during the deposition of the Kayitou Formation. Therefore, we adopted the mean values of Late Permian and Early Triassic $\delta^{13}\text{C}_{\text{org}}$ from the marine Meishan section to run a simple mixing model to test the idea whether increased marine OM proportions contributed to the CIE observed in our studied locations.

The binary mixing equation we used is based on the study of [Schultz and Calder \(1976\)](#) to estimate the relative proportions of terrigenous and marine organic carbon (or proportions of autochthonous and refractory allochthonous organic carbon) in sediment samples. The mixing model is:

$$\delta^{13}\text{C} = f_t \delta^{13}\text{C}_t + f_m \delta^{13}\text{C}_m \quad (1)$$

where f_t and f_m are the fractions of terrestrial and marine organic carbon, and $f_t + f_m$ should equal to 1 to ensure mass balance. Rearranging Eq. (1), we get:

$$f_t = (\delta^{13}\text{C} - \delta^{13}\text{C}_m) / (\delta^{13}\text{C}_t - \delta^{13}\text{C}_m) \quad (2)$$

We used the Meishan Late Permian mean marine $\delta^{13}\text{C}_{\text{org}}$ value (-28‰; Fig. 6) for $\delta^{13}\text{C}_m$ and mean terrestrial $\delta^{13}\text{C}_{\text{org}}$ value (-24.9‰; Table 2) for $\delta^{13}\text{C}_t$, and used the measured $\delta^{13}\text{C}_{\text{org}}$ value for the $\delta^{13}\text{C}$ in Eq. (2) (-25‰ for Jiuchaichong, -23.8‰ for Chahe, and -26.2‰ for Mide). Doing so reveals a Late Permian f_t of >97% for Jiuchaichong and Chahe, but only 58% for Mide. This is consistent with Mide being mostly influenced by the ocean because of its location at the very top of the Xuanwei Formation. For the Early Triassic, $\delta^{13}\text{C}_{\text{org}}$ values in Jiuchaichong and Chahe converge with the $\delta^{13}\text{C}_{\text{org}}$ value in Meishan (~ -27‰), suggesting they might share the same OM source, perhaps more isotopically heavier plant input or the effect of green sulfur bacteria as proposed by [Riccardi et al., \(2007\)](#). In fact, the biomarker study at Chahe by [Zhao \(2003\)](#) suggests a shift on the type of OM from algal/bacteria-dominated source at Beds 64-67 (top of the Xuanwei Formation) to a mixture of higher plants and bacteria/algal at Beds 69-74 (bottom of Kayitou Formation). This is consistent with, based on our analysis, that a greater contribution of terrestrial plants would have driven the $\delta^{13}\text{C}_{\text{org}}$ towards more ^{13}C -enriched values.

An additional source of variability in terrestrial and marine sedimentary $\delta^{13}\text{C}_{\text{org}}$ is the allochthonous inputs, either aged soil OM or refractory kerogen from the erosion of sediments exposed on land ([Hayes et al., 1989](#)). Research on modern fluvial systems has shown that there is a significant contribution of aged refractory OM to the particulate organic carbon pool ([Raymond and Bauer, 2001](#); [Clark et al., 2013](#)). Although we cannot exclude this possibility, the persistent offsets of our study sections and the

correspondence with the global CIE argues against variations in allochthonous inputs as a major source of isotope change.

Interestingly, the $\delta^{13}\text{C}_{\text{org}}$ values in the studied terrestrial sections show a consistent -2.6‰ shift (with the exception of Zhejue), likely reflecting the global CIE. This is supported by the global compilation of $\delta^{13}\text{C}_{\text{carb}}$ recorded in marine environments, all showing a 2-4‰ negative CIE across the EPE (Korte and Kozur, 2010), indicative of a global atmospheric $\delta^{13}\text{C}$ shift of similar magnitude (Svensen et al., 2009; Cui et al., 2013). The local expression of the CIE in sedimentary organic matter is expected to differ from the global atmospheric excursion, as it is affected by changes in isotope fractionation with carbon fixation and due to changes in temperature, productivity, and carbon source, as well as differences in preservation biases.

Several mechanisms have been proposed to explain the global CIE, with the top candidates being the Siberian Traps volcanism (Berner, 2002; Beerling et al., 2007; Svensen et al., 2009; Aarnes et al., 2010; Luo et al., 2011; Black et al., 2012, 2014), or a combination of various sources associated with the Siberian Traps volcanism (Cui et al., 2013; Cui et al., 2015). Strong evidence exists to support that the mass extinction event is contemporaneous with the Siberian Traps eruption (Bowring et al., 1998; Shen et al., 2011; Burgess et al., 2014). An Earth system modeling study suggested that carbon emission from Siberian Traps volcanism or other related sources came in two multi-millennial pulses, followed by enhanced silicate weathering and increased organic carbon burial (Cui et al., 2015). Other lines of independent evidence have not demonstrated whether the carbon release has come in two pulses. Despite the intriguing coincidence between the extinction event and the Siberian Traps volcanism, the exact mechanism that

triggered the ^{13}C -depleted carbon is still a question of debate, which begs for further studies.

7. Conclusion

We present $\delta^{13}\text{C}_{\text{org}}$ and TOC data from four terrestrial sections across the EPE in western Guizhou and eastern Yunnan in the South China Yangtze block. The $\delta^{13}\text{C}_{\text{org}}$ values do not linearly correlate to the amount of TOC, suggesting end-member mixing is not a primary control on the carbon isotope composition. The absolute values of $\delta^{13}\text{C}_{\text{org}}$ vary from the more proximal Jiucaichong section to the more distal Mide section showing the most depleted values. The magnitude of the CIE of the four sections is variable, ranging from -0.7‰ to -2.6‰, although three of the four studied sections have shown shifts of -2.6 ‰. This indicates that depositional environment played a role in determining the $\delta^{13}\text{C}_{\text{org}}$ values, and that these geographic patterns in isotope composition persisted through the extinction event. Occurrences of fungal spores, including *Reduviasporonites*, at the top of the Xuanwei Formation in all four studied sections are indicative of terrestrial ecosystem collapse and decomposition of organic matter. Both biostratigraphy and the fungal spike suggest that the PTB should be located near or above the top of the Xuanwei Formation at Jiucaichong, Chahe and Mide, in accordance with the -2.6‰ CIE. We have demonstrated that the carbon isotope in the study area is dominated by the atmospheric $\delta^{13}\text{C}$ signal and that the carbon cycle perturbation during the event left an indelible imprint on the geologic record.

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Figure Captions

Figure 1. Topographic map of western Guizhou and eastern Yunnan showing the position of Chahe, Zhejue, Jiucaichong and Mide sections in bold. Geographic coordinates are given as a 10 km UTM grid (Bercovici et al., 2015).

Figure 2. Correlation of the four studied sections arranged from left to right according to their depositional environments. The PTB is determined by the LAD of *Gigantopteris* flora and the FAD of *Annalepis-Peltaspermum* at Jiucaichong, Chahe and Mide and Triassic spinicaudatans at Jiucaichong and Chahe, the FAD of *Annalepis-Peltaspermum* and *Ophiceras* sp. at Mide, the U/Pb dating at Bed 68 at Chahe by Shen et al., (2011), and by the fungal spike of *Reduviasporonites* spore at Jiucaichong and Mide. Also shown are the measured $\delta^{13}\text{C}_{\text{org}}$ and TOC for each studied section and the published $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ from the GSSP Meishan section based on Shen et al. (2011).

Figure 3. Statistical analysis on the pre- and post-PTB $\delta^{13}\text{C}_{\text{org}}$ values from the four studied sections. The black line within the box indicates median values and red diamond indicates mean values.

Figure 4. Cross plot of $\delta^{13}\text{C}_{\text{org}}$ and 1/TOC for the four studied sections, the color-coding indicates different lithologies shown in the legend.

Figure 5. Compilation of the current radiometric ages, organic geochemistry and biostratigraphy of the end-Permian mass extinction and PTB interval at the Meishan

1096 GSSP.

1097

1098 Figure 6. Statistical analysis on $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ of the GSSP Meishan section. Upper
1099 left) $\delta^{13}\text{C}_{\text{carb}}$ prior to, during and after the end-Permian extinction event based on the
1100 definition of Shen et al. (2011). Lower left) $\delta^{13}\text{C}_{\text{org}}$ prior to, during and after the end-
1101 Permian extinction event (EPE) based on the definition of Shen et al. (2011). Upper right)
1102 Box and whisker plot for $\delta^{13}\text{C}_{\text{carb}}$ of the pre-, during- and post-EPE, the mean values are
1103 shown in red diamonds; Lower right) Box and whisker plot for $\delta^{13}\text{C}_{\text{org}}$ of the pre-, during-
1104 and post-EPE, the mean values are shown in red diamonds. The blue shaded region
1105 represents pre-EPE, the green shaded region represents during-EPE, and the yellow
1106 shaded region represents post-EPE.

Figure 1

Figure 2

Figure3

Figure4**Jiuaichong** **Zhejue** **Chahe** **Mide**

Figure5

Figure 6

Table 1. Welch-corrected t test for the $\delta^{13}\text{C}_{\text{org}}$ values from four studied location during pre- and post-PTB conditions.

	Jiuchaichong		Chahe		Zhejue		Mide	
	Pre-	Post-	Pre-	Post-	Pre-	Post-	Pre-	Post-
Sample size	n=25	n=9	n=47	n=6	n=20	n=9	n=11	n=6
Mean	-25.0	-27.6	-23.8	-26.4	-23.5	-24.2	-26.2	-28.6
Standard deviation	1.3	1.3	1.3	1.3	0.3	0.7	1.0	1.1
Variance	1.6	1.7	1.7	1.7	0.1	0.4	0.9	1.1
CIE magnitude	-2.6		-2.6		-0.7		-2.4	
Minimum $\delta^{13}\text{C}_{\text{org}}$	-30.0		-28.7		-25.8		-30.0	
Maximum $\delta^{13}\text{C}_{\text{org}}$	-23.0		-20.5		-23.0		-25.1	
T value	5.265		4.756		3.287		4.659	
Degree of freedom	13.907		6.363		9.933		9.531	
Tabulated t value at 95% confidence	2.16		2.45		2.26		2.26	
P<0.05	yes		yes		yes		yes	

Table 2. Summary of the $\delta^{13}\text{C}_{\text{org}}$ from several marine/continental successions in East Greenland, West Spitsbergen, New Zealand, South China, Antarctica, South Africa, Australia, India, and Madagascar.

Location	Setting	Pre-extinction $\delta^{13}\text{C}_{\text{org}}$ (‰)	During- or Post- extinction/PTB $\delta^{13}\text{C}_{\text{org}}$ (‰)	Reference
East Greenland	Marine	-23	-29	Twitchett, (2001)
West Spitsbergen (Festningen)	Marine	~ -25	~ -32	Wignall et al., (1998)
New Zealand (Maitai Group)	Marine	-24.7	-30.7	Krull et al. (2000)
South China (Meishan)	Marine	-28.5	-26.6	Shen et al., (2011)
South China (Meishan)	Marine	-27	-28.3	Riccardi et al., (2007)
South China (Sichuan)	Marine	-28.2	-27.3	Cui et al., (2008)
Austria (Gartnerkofel-1 Core)	Marine	-24.6	-26.6*	Magaritz et al., (1992)
Canada (Williston Lake)	Marine	~ -29	~ -32	Grasby and Beauchamp, (2008)
Canada (Buchanan Lake)	Marine	-26.6	-29.4	Grasby and Beauchamp, (2008)
Southern Alps (Seis)	Marine	-26	-29.8	Siegert et al., (2011)
Southern Alps (Misci)	Marine	-27.3	-29.8	Kraus et al., (2013)
Southern Alps (San Antonio)	Marine	-27.2	-29.4	Kraus et al., (2013)
Croatia (Velebit Mountain)	Marine	-25.7	-27	Fio et al., (2010)
Northeast Italy	Marine (n-alkanes)	~ -27	<-32	Sephton et al., (2005)

Western Solvenia (Idrijca Valley)	Marine (Kerogen)	-27	-25.6	Schwab and Spangenberg (2004)
Western Solvenia (Idrijca Valley)	Marine (n-alkanes)	-27.7	-30.2	Schwab and Spangenberg (2004)
Average	Marine	-26.5	-29.3	
Antarctica (Graphite Peak)	Terrestrial (Paleosol)	~ -25	< -40 (extreme)	Krull and Retallack (2000)
Antarctica (Allan Hills)	Terrestrial (Paleosol)	~ -24	< -40 and > -17 (extreme)	Krull and Retallack (2000)
Australia (Murrays Run bore)	Terrestrial (Paleosol)	~ -26	~ -30	Retallack and Jahren (2008)
Australia (Wybung Head)	Terrestrial (Paleosol)	~ -25	~ -26	Retallack and Jahren (2008)
South Africa	Terrestrial (coal)	-22.6 (Late Permian)	-24.3 (Middle Triassic)	Faure et al., (1995)
South Africa (Carlton Heights in Karoo Basin)	Terrestrial	-22.9	-24.2	Ward et al., (2005)
India	Terrestrial	-24 ± 2	< -30 and > -21 (extreme)	de Wit et al., (2002)
Madagascar (Morondava)	Terrestrial	-28 ± 2	< -32 (extreme)	de Wit et al., (2002)
South Africa	Terrestrial	~ -27	~ -30	de Wit et al., (2002)
South China (Jiuchaichong)	Terrestrial	-25.0	-27.6	This study
South China (Zhejue)	Terrestrial	-23.5	-24.2	This study
South China (Chahe)	Terrestrial	-23.8	-26.4	This study
South China (Mide)	Terrestrial	-26.2	-28.6	This study
Average	Terrestrial	-24.8	-27.6	

*The five extreme $\delta^{13}\text{C}_{\text{org}}$ outliers in the Gartnerkofel-1 core measurement were removed according to Magaritz et al. (1992), but note that it was considered valid by Retallack and Krull (2006).

Table 3. Comparison of the pollen counts from the four studied sections and their correspondent carbon isotope values.

Bed #	Strat pos (cm)	Pollen sample:	Phytoclasts	Charcoal	Fungi	Miospores	Isotope sample:	$\delta^{13}\text{C}$ (‰)
Chahe								
64	10330	11C047	100	0	0	0	CHISO-64	-23.86
59	9745	11C041	98	2	0	0	CHISO-59	-24.20
56	9195	11C039	100	0	0	0	CHISO-56	-24.30
55	8970	11C037	97	3	0	0	CHISO-55	-24.24
35	5845	11C023	86	0	0	14	CHISO-35	-25.02
27	4230	11C018	100	0	0	0	CHISO-27	-23.42
18	3520	11C016	100	0	0	0	CHISO-18	-24.24
11	2220	11C010	11	89	0	0	CHISO-11	-26.08
5	1230	11C006	15	85	0	0	CHISO-5b	-21.77
Jiuchaichong								
23	4615	11J013	91	9	0	0	JCISO-23	-27.17
21b	4600	11J010	93	7	0	0	JCISO-21	-26.55
11	2795	11J007	100	0	0	0	JCISO-11-1	-24.35
4	1330	11J004	100	0	0	0	JCISO-04-1	-23.40
Zhejue								
49	2505	11Z025	98	2	0	0	ZJISO-49-3	-23.79
39	1655	11Z023	91	9	0	0	ZJISO-39-1	-23.52
30	590	11Z013	99	0	0	0.7	ZJISO-30-1	-23.06
30	580	11Z012	100	0	0	0	ZJISO-30-2	-23.46
29	450	11Z011	97	0	0	2.7	ZJISO-29-1	-23.19
28	285	11Z010	74	26	0	0	ZJISO-28-1	-23.00
28	230	11Z008	100	0	0	0.3	ZJISO-28-3	-23.15
Mide								

19	505	11M010	100	0	0	0	MDISO-19-1	-27.40
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