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Major geological events in the Late Guadalupian and carbon-strontium isotope responses in the Yangtze platform, South China

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1 **Major geological events in the Late Guadalupian and carbon-strontium**
2 **isotope responses in the Yangtze platform, South China**

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Draft

Abstract

Two sections of marine carbonate strata, Podu and Lengshuixi, from the Yangtze platform, South China, are investigated in terms of major geological events and carbon-strontium isotope responses. The results show a large-scale regression and a Large Igneous Province (LIP) event occurred in succession in the short interval of the late Guadalupian, inducing a biodiversity crisis. The regression event, marked by an increase in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, is observed in the *Jinogondolella postserrata* Zone. It contributed significantly to the reduction in species habitat, leading to notable losses within the shallow water species assemblage. A subsequent LIP event occurred after the appearance of *Jinogondolella altudaensis*, releasing CO_2 and numerous other deleterious gases that further amplified the crisis in shallow water to deep sea environments. The corresponding decline in Sr and C isotopes could have been a response to the basaltic eruption (relative increase from mantle-derived Sr) and release of ^{12}C -enriched CO_2 . During the process of creation of the LIP, the decline in $\delta^{13}\text{C}_{\text{carb}}$ was mainly associated with ^{12}C -enriched CO_2 release from organic matter in sedimentary strata, and high temperature isotope fractionation during the volcanic eruption stage (exceeding 1200 °C). This study could further provide valuable clues regarding development process on the major geological events and the causes of the biodiversity crisis.

Keywords: regression, volcanism, biodiversity crisis, carbon-strontium, late Guadalupian

33 1. Introduction

34 The late Permian is an important period in Earth's history. Large-scale geotectonic
35 activity, Large Igneous Province (LIP) events, and mass extinctions were concentrated within
36 a short interval of less than 10 Myr. Although there has been a lot of research into these
37 events, the biodiversity crisis in the late Guadalupian, as a prelude to the end-Permian mass
38 extinction event, still attracts significant attention. The crisis was mainly restricted to tropical
39 invertebrate taxa with passive respiratory systems such as corals, bryozoans, brachiopods,
40 and ammonoids (Weidlich 2002; Wignall et al. 2009a).

41 In the Laibin sections of the Yangtze platform, South China (Global Stratotype Section
42 and Point (GSSP)), Mei et al. (1998) placed the biodiversity crisis approximately 2 m below
43 the top of the Laibin Limestone Member. Jin et al. (2006) placed the mass extinction around
44 the base of the *L. minima* Zone, because most keriotheca-walled fusulinaceans (major victims
45 of the extinction event) disappeared close to this level. These conclusions may be similar to
46 that of Texan, West Texas (Yang and Yancey 2000; Wignall et al. 2009a), and have been
47 accepted by most researchers.

48 Moreover, the negative C isotope excursion, a mark of the biodiversity crisis, occurred
49 in various patterns (Shen et al. 2013). A large negative carbon isotope excursion (ca. 6‰
50 from the Xiongjiachang section in Guizhou) was reported (Wignall et al. 2009b, 2012; Bond
51 et al. 2010), but not confirmed, in the Penglaitan section (Guangxi), the Xiaojiaba section
52 (South China: Wang et al. 2004; Kaiho et al. 2005; Chen et al. 2011; Wei et al. 2012), and
53 sections in Hungary and Greece (Wignall et al. 2012).

54 The causes of the biodiversity crisis and the large-scale perturbation of the carbon cycle

55 have been crudely attributed to: global climate change (Isozaki et al. 2007a, 2007b; Clapham
56 et al. 2009; Nishikane et al. 2014), the major regression (Shen and Shi 2002; Leven 2003;
57 Yang et al. 2004; Wignall et al. 2009a), a massive extension of deep-water anoxia (Weidlich
58 2002; Clapham et al. 2009), catastrophic methane release (Retallack et al. 2006), and the
59 eruption of the Emeishan LIP (Ali et al. 2002; Wignall et al. 2009b). Nevertheless, the link
60 between the major regression and the mass extinction is frequently highlighted (e.g., Wang
61 and Jin 2000; Zhou et al. 2002; Wignall et al. 2009b), based primarily on sedimentary facies
62 and carbon isotope evidence. Although some strontium isotopes have been mentioned in
63 previous studies, e.g., in the Laibin sections in South China (Wignall et al. 2009a), the
64 Abadeh section in central Iran (Liu et al. 2013), and the Boreal Realm (Spitsbergen) (Bond et
65 al. 2015), the number of samples was too small to distinguish available causative events.
66 Regarding the biodiversity crisis, it appears that further ascertaining the inherent causes of the
67 negative carbon isotope excursion remains key.

68 In this study, two sections of marine carbonate strata from the Yangtze platform, South
69 China, are investigated in terms of major geological events and carbon-strontium isotope
70 responses, which could further provide valuable clues regarding the causes of the biodiversity
71 crisis.

72

73 **2. Geologic setting**

74 The Yangtze carbonate platform in South China is bounded on the west by the Tibetan
75 Plateau, to the north by the Qinling-Dabie Orogen, and to the south by the Ailao Shan–Red
76 River Fault. Marine carbonate strata of the middle-late Permian, represented by the Maokou

77 Formation, are widely developed around the Emeishan LIP, and occur frequently in the
78 Yunnan, Guizhou, Sichuan, and Guangxi provinces of South China (Fig. 1).

79 The Emeishan LIP outcrops in the western Yangtze Craton, the Southeast
80 Songpan-Gangze Terrane, and the East Tibet Block. The LIP consists of massive volcanic
81 successions and numerous contemporaneous mafic and felsic intrusions, covering a
82 rhombic-shaped area of $\sim 2.5 \times 10^5 \text{ km}^2$ and surrounded by several deep large faults, which
83 include the Longmenshan-Anlinghe fault, the Ailao Shan-Red River slip fault, the Xiaojiang
84 fault, and the Xichang-Qiaojia fault. The thickness of the erupted basalts ranges from several
85 hundred meters to 5 km (Xu et al. 2001), and the original volume was substantially greater
86 than $3.8 \times 10^6 \text{ km}^3$ (Xu et al. 2010). Volcanic eruptions in the early to middle stages showed a
87 violent phreatomagmatic style and pyroclastic-phreatomagmatic volcanism. The depositional
88 thicknesses of individual mafic volcanoclastic beds approaches 200 m (Wignall et al. 2009b).
89 Remnant tuff layers of variable thickness (several to hundred meters) appear in the uppermost
90 basalt successions (Bowring et al. 1998; Zhu et al. 2011). Dating using the zircon U–Pb
91 isotopic method yields an approximate age of 262–259 Ma (He et al. 2007; Shellnutt et al.
92 2012; Zhong et al. 2014). However, it is difficult to accurately constrain the initial eruption
93 time of the Emeishan basalts, because of the complicated contact relationships between the
94 basalts and the Maokou carbonate strata (Emeishan flood basalts overlie Middle Permian
95 carbonates and are, in turn, overlain by the latest Permian terrestrial or marine clastic rocks in
96 the east and Triassic sedimentary rocks elsewhere, He et al. 2007). Wignall et al. (2009b),
97 placed the initial eruption and biodiversity crisis at the end of the *Jinogondolella*
98 *xuanhanensis* Zone in the late Guadalupian, based on conodont biostratigraphy in the

Xiongjiachang section. Subsequently, Sun et al. (2010) moved the time of initial eruption of the small-volume and very localized Emeishan basalts to ~263 Ma (around the *Jinogondolella altudaensis* Zone), but the later, kilometer-scale eruptions with widespread mafic and phreatomagmatic-style explosions were mainly assigned to the *J. granti* Zone (*J. Jinogondolella*).

According to He et al. (2003), the Emeishan LIP area is roughly divided into four zones: an intensely denuded zone (inner zone), a partially denuded zone (middle zone), a partial deposition zone (outer zone), and a continuous deposition zone. The sedimentary facies variations in the Yangtze platform are likely affected by the Emeishan LIP and Dongwu movement. The upper part of the Maokou Formation, which mainly consists of medium bedded to massive limestones, has a widespread unconformity with the Emeishan basalts in the west and center, and the Longtan Formation in the east (He et al. 2010).

Our study was limited to the top of the Maokou Formation. Two sections, Podu and Lengshuixi, are located in the outer zone of the LIP and in Qijiang and Shizhu counties, Chongqing city, South China, respectively.

The Podu section (upper Maokou Formation), which is 24 m thick, is divided into 15 beds and mainly consists of gray silt-fine crystal limestone with nodules and beds. The top of the Maokou Formation in this section is covered by ~1 m thick mudstone (Fig. S1a) which contains some carbonate breccia at the bottom, and shows an unconformity with the Longtan Formation.

The Lengshuixi section (upper Maokou Formation), which has a thickness of 29 m, is approximately 400 km from the Xiongjiachang section. The lithology contains gray thinly

bedded limestones and siliceous micrite divided into 38 beds. The uppermost part of the limestone is overlain by approximately 9 m thick shaly bed (Fig. S1b). The biostratigraphically-defined G-L (Guadalupian-Lopingian) boundary is placed at bottom of the Wujiaping Formation (Isozaki et al. 2004; Sun et al. 2008).

3. Sampling and methods

3.1. Sampling

Eighty-seven samples were collected from two sections in the Sichuan Basin. At least one sample was collected from each thin bed, and three samples were collected from each thick bed. Sampling near veins related to late cracks was avoided if possible. For the C and Sr isotopes and conodont stratigraphy analyses, the mass of each sample was approximately 10 kg.

3.2. Analytical methods

3.2.1 Carbon isotope analysis

The carbon isotope compositions were determined by the Isotopic Geochemistry Laboratory of the Chengdu University of Technology, Chengdu, China. Carbonate samples were extracted using a micro-drill, washed with deionized water, and dried. For each analysis, a 0.5-mg sample was placed in a tube of the Gas-Bench automatic sampling system, cleaned with high-purity helium, and dissolved with 100% H₃PO₄ at 70 °C for 2 h. The δ¹³C value was measured using a Thermo Fisher Scientific MAT253 isotope mass spectrometer. Vienna Pee Dee Belemnite (VPDB) was used as a standard sample, and the test accuracy was

±0.02‰.

3.2.2 Strontium isotope analysis

Strontium isotopes were analyzed at the State Key Laboratory of Oil and Gas Reservoir Geology and Exploitation of the Chengdu University of Technology, Chengdu, China. The samples (50 mg), which were washed with deionized water and crushed to less than 0.0750 mm, were dissolved using 1 M ultrapure acetic acid for approximately 12 h on an electric hot plate at 60 °C. A 1.5 mL pure solution was extracted after adding 1.5 mL 2.5 M HCl. The Sr purification and separation of Rb and other cations such as Ca^{2+} and Ba^{2+} was performed using AG 50W×12 cation exchange resin. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were then measured using a Thermo Fisher Scientific TRITON Plus Mass Spectrometer.

The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normalized to a $^{86}\text{Sr}/^{88}\text{Sr}$ ratio of 0.1194, and calculated from 150 measurements. An internal precision of $\sim 5 \times 10^{-6}$ (RSE) was maintained. The analytical precision was monitored using the NBS 987 standard. The mean measured value obtained for NBS 987 was 0.710247 ± 0.000008 (2σ , $n = 6$). All reported $^{87}\text{Sr}/^{86}\text{Sr}$ data were corrected to a nominal value of 0.710248 for NBS 987.

4. Results

4.1. Mn and Sr content

To determine the effect of diagenesis, the Mn and Sr contents in all samples were tested using an ELAN DRC-e inductively coupled plasma mass spectrometer (PerkinElmer SCIEX) in the Isotopic Geochemistry Laboratory of the Chengdu University of Technology, Chengdu,

China. According to previous research, samples with Mn (content) <250 ppm and Sr (content) >400 ppm, or Mn/Sr <0.625, were classified as well preserved (Bruckschen et al. 1999; Korte et al. 2003, 2005). The analysis results showed that Mn contents ranged from 11 to 223 ppm, Sr contents ranged from 490 to 2604 ppm, and Mn/Sr ratios were <0.295, indicating that effects due to diagenesis could be eliminated for all samples (Tables S1 and S2).

4.2. Carbon and strontium isotopes

All analytical results are shown in Tables S1 and S2 and Fig. 2. In the Podu section, the carbon and strontium isotopes varied from 0.88 to 4.40‰ ($\delta^{13}\text{C}_{\text{carb}}$ values) and 0.70689 to 0.70731 ($^{87}\text{Sr}/^{86}\text{Sr}$), respectively. During the early stage, negative C isotope excursions of -Δ1.71‰, from 4.06‰ to 2.35‰, and a corresponding increase in $^{87}\text{Sr}/^{86}\text{Sr}$ values from 0.70702 to 0.70731 were observed in Beds 8 to 7. The $\delta^{13}\text{C}_{\text{carb}}$ values then declined slightly. The $^{87}\text{Sr}/^{86}\text{Sr}$ values tended to small increase from Beds 6 to 3. The C and Sr isotope compositions simultaneously increased until the bottom of Bed 1. Finally, an abrupt negative isotope excursion in C (-Δ3.52‰) and a minimum $^{87}\text{Sr}/^{86}\text{Sr}$ (down to 0.70689) occurred below the Capitanian-Wuchiapingian boundary.

In the Lengshuixi section, the carbon and strontium isotopes varied from 0.01 to 3.05‰ ($\delta^{13}\text{C}_{\text{carb}}$ values) and 0.70689 to 0.70786 ($^{87}\text{Sr}/^{86}\text{Sr}$), respectively. The high Sr isotope values were very clear in the Lengshuixi section, but the variation in C isotopes was confined to a small range. The C isotope compositions in this section were probably disturbed and exhibited few fluctuations. Overall, the high $^{87}\text{Sr}/^{86}\text{Sr}$ values (0.70786) with the negative C

isotope excursions were still clear in Beds 26 to 23 (bottom), corresponding to Beds 8 to 7 in the Podu section. At the top of the section, a negative excursion trend of C and Sr isotopes was also seen (Fig. 2), but the variation was near that of the top of the Podu section. The apparent relationship between the C and Sr isotope variations in the two sections requires further analysis.

4.3. Conodont stratigraphy

Because the top of the Maokou Formation in the Yangtze Platform was mostly denuded, resulting in an uncertain sedimentary age, to obtain exact temporal constraints for the carbonate beds in the study sections for comparison with previous studies (regarding the timing of the biodiversity crisis, the volcanic eruption, and the negative carbon isotope excursions constrained by the conodont stratigraphy: Bond et al. 2010; Wignall et al. 2009a, 2012), this study employed high-resolution conodont stratigraphy. All samples were soaked in acetic acid solution for more than one month, and conodont samples were then picked using a binocular microscope after water clearing. The identified conodonts in the two sections are shown in Fig. 3. The timing of these sections was mainly constrained by *J. postserrata* and *J. altudaensis* (Fig. 2).

4.4 Continental materials in carbonate strata

Residual-bearing terrigenous material were found in some carbonate samples after dissolving by acetic acid. The residues contained typical magmatic or metamorphic mineral components, including mica, quartz, and other silicates (Fig. S2 and Table S3). These

minerals, which had high $^{87}\text{Sr}/^{86}\text{Sr}$ values, were found in Bed 7 of the Podu section and in Beds 23 and 21 of the Lengshuixi section, which may be associated with the regression event in that interval.

5. Discussion

5.1 Major geological events and carbon-strontium isotope responses

5.1.1 Early regression and isotope responses

A large-scale regression event occurred in the early stage of the late Guadalupian, confirmed by previous studies (Wang and Jin 2000; Shen et al. 2007; Wignall et al. 2009a; Bond et al. 2010), and might have severely impacted the shallowest marine species and sedimentary environment throughout the Tethys. Wignall et al. (2009a) stated that a major regression record existed within the basinal succession of the Jiangnan Basin, based on a facies analysis of the Penglaitan and Tieqiao sections (GSSP), South China. They provided regional evidence of the shallowest water facies, which had sharp-based and hummocky cross-stratification beds related to storm activity. The peak regression, which is diagnostic of the end of low-stand deposition, occurred within the relatively shallow-water Laibin Limestone Member up to the top of Bed 5b at the Penglaitan section, and Bed 5 at the Tieqiao section. The most likely timing of the mass extinction was probably associated with the regressive event at the beginning of the *J. granti* Zone (Wignall et al. 2009a). In the Podu section, the increasing Sr isotope ratios, from 0.70700 (top of Bed 10) to a maximum of 0.70731 (Bed 7), occurred in *J. postserrata* Zone. The regression likely began in Bed 10, and the peak occurred in Bed 7. Similarly, the increasing $^{87}\text{Sr}/^{86}\text{Sr}$ trend in the Lengshuixi section

simultaneously exists in Bed 27 to Bed 23. It is suggested that the Sr isotope ratio increase is a response to the regression event that results in a relative increase of continental Sr components in coeval sea water. The regression event, in terms of timing interval, was placed at the early stage of the late Guadalupian, slightly predating the shallow-water Laibin Limestone Member in the Penglaitan and Tieqiao sections. This regression probably spread over the Yangtze platform and the rest of the globe (Bond et al. 2015). The clear $^{87}\text{Sr}/^{86}\text{Sr}$ increase in sea water, as an indicator of the global regression, could be more dependable.

Our research also noted fluctuations in the Sr isotope values, and a few biological fossil relics (no intact biotic fossils) in the Lengshuixi section were found (Fig. 3B). This could be because there was clear dolomitization in the top of the Maokou Formation in the Lengshuixi section, or because late geological variations induced disturbances. Despite that, the evolutionary trend of Sr isotopes still indicated a clear correspondence with that of the Podu section (Fig. 2).

5.1.2 Late volcanic eruptions and isotope responses

The Emeishan LIP was an important geological event of the late Guadalupian in the study area. However, large-scale eruption of basalts mainly occurred after the regression event, as restricted by the conodont zone (*J. altudaensis* Zone), which strongly affected isotope records in the late Maokou carbonate strata. The Sr and C isotopes in the Podu section correspond with violent eruptions, i.e., $^{87}\text{Sr}/^{86}\text{Sr}$ decreased from 0.70714 to 0.70689 (the minimum in the study area), and $\delta^{13}\text{C}$ decreased from 4.40‰ to 0.88‰ at the upper of the carbonate, which likely corresponds to the negative carbon isotope excursion in the

Xiongjiachang and Gouchang sections, reported by Bond et al. (2010) (Fig. 2). The isotope evolution in the Lengshuixi section also shows similar characteristics. The temporal overlap between huge basalt eruptions and sudden isotope changes likely indicates an inherent relationship. The decline in $^{87}\text{Sr}/^{86}\text{Sr}$ values implies that the Sr provenance in the seawater carbonates is affected by Sr constituents from the mantle basalt reservoir. The $\delta^{13}\text{C}_{\text{carb}}$ fluctuations during the end-Guadalupian were probably not associated with bio-productivity variations because $\delta^{13}\text{C}_{\text{carb}}$ values after the early regression were consistently high and positive until the volcanic eruption (Fig. 2), although many of the shallowest marine species were lost (Wei et al. 2012). This situation was also apparent in the Xiongjiachang section (Wignall et al. 2009a; Bond et al. 2010).

5.1.3 CO_2 origin in volcanism and negative carbon isotope excursion

It is not debatable that the volcanism released massive amounts of CO_2 . Explosive eruptions displaying a violent phreatomagmatic style spread spectacular pyroclastic-phreatomagmatic lava from the Emeishan LIP event across the western Yangtze platform. Some mafic volcanoclastic beds approached 200 m (Wignall et al. 2009b; Bond et al. 2010), and tuff layers of different thicknesses (several to hundreds of meters) appeared in the uppermost basalt successions (Bowring et al. 1998; Zhu et al. 2011). Based on estimates from the Kilauea volcano, an eruption of $2 \times 10^6 \text{ km}^3$ of basalt could release $4 \times 10^{19} \text{ g}$ of CO_2 and more than 10^{19} g of SO_2 . These CO_2 emissions are approximately 15 times the current content of the atmosphere (Gerlach and Graeber 1985; Kamo et al. 2003). Xu et al. (2010) considered that the original volume of the Emeishan basalts was substantially greater than $3.8 \times 10^6 \text{ km}^3$.

If this estimate is true, the LIP could also have provided a huge amount of CO₂, equivalent to that of the Kilauea volcano. Furthermore, volcanic events triggering widespread metamorphism in the crust and assimilation (Marziano et al. 2008) can produce abundant CO₂ (Ganino et al. 2008, 2009; Korte and Kozur 2010), 3.6–8.6 times larger than the mass of magmatic CO₂, according to a metamorphic aureole within the Emeishan LIP (Fig. 4) (Ganino et al. 2008). Specifically, sediment-derived CO₂ could release a gigantic amount of ¹²C-enriched CO₂, due to significant organic matter splitting during volcanism, which would severely impact the carbon cycle and lead to negative carbon isotope excursions in coeval marine carbonates.

Furthermore, we noted a clear difference in the carbon isotope fractionation between metamorphism (below 800 °C) and volcanic eruption (above 1200 °C) in the CO₂-carbonate fluid system. The calculated isotope equilibrium fractionation results indicated that the $\Delta_{\text{CO}_2\text{-calcite}}$ value between metamorphism (below 800 °C) and the duration of the volcanic eruption (over 1200 °C) could have declined from ~4.0‰ to a minimum value (Fig. 4) (Bottinga 1969; Chacko et al. 1991). If true, such isotope fractionation be a new factor causing negative excursions of carbon isotopes.

5.2. Development processes of major geological events and biodiversity crisis

The large-scale regression and volcanism in the late Guadalupian are two typical geological events that could have induced the associated biodiversity crisis. The early regression in the late Guadalupian was probably associated with Emeishan mantle plume emplacement (He et al. 2003, 2010) and Dongwu Uplift. Generally, the abrupt doming of the

Earth's surface in the development stage of the mantle plume, marked by increasing $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in marine carbonates due to a relative increase of continent-derived Sr in coeval sea water, led to land area broadening and large-scale losses of shallow water species habitats, ultimately causing a decline in the number of shallow-water species containing tropical invertebrate taxa.

Following the regression event, the potential temperature in the upper crust increased because of the LIP event, inducing large-scale metamorphism in the sedimentary strata or wall rock around the magmatic system, and producing enormous amounts of CO_2 , SO_2 , Cl_2 , and other noxious gases. The tremendous amount of CO_2 and noxious gases from explosive volcanic eruptions played a crucial role in triggering the biodiversity crisis (Christie-Blick 2009; Wignall et al. 2009b). Because of their relatively fast diffusion velocities in the atmosphere, enormous amounts of CO_2 and noxious gases (CO_2 , SO_2 , Cl_2 , and fluorine-bearing volatile compounds) caused rapid environmental deterioration, ocean acidification (Bond et al. 2015; Clarkson et al. 2015), and the widespread (and possibly global) deaths of various species. Specifically, the salt-bearing strata in the Tethys domain contain abundant salt and gypsum or anhydrite layers, most of which cracked during the high-temperature volcanism and produced SO_2 , Cl_2 , or fluorine-bearing compounds that were released into the atmosphere and ocean and were highly lethal to certain species (Shi et al. 2017). Under the successive impacts of volcanism, the biodiversity crisis was steadily amplified from affecting shallow species to also impacting benthic species.

6. Conclusions

The major conclusions of this study are shown in Fig. 5 and are detailed below.

(1) Two events occurred in the late Guadalupian. The early regression (in the *J. postserrata* Zone), marked by high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, implies a predominant contribution from terrigenous materials. The late, large-scale volcanic eruptions (beginning in the *J. altudaensis* Zone) could have been an important cause of the abrupt and simultaneous decline in carbon and strontium isotopes, which are attributed to the basaltic origin and large-scale release of CO_2 from the upper crust.

(2) The regression could have caused a wide-ranging loss of species habitat and a consequent early biodiversity crisis in shallow water species assemblages. The volcanic eruption, which released abundant CO_2 and deleterious gases, was possibly a primary cause of the crisis, which was further amplified from shallow water to the deep sea.

(3) The LIP creation process, which included metamorphism and volcanism, could have produced abundant CO_2 of meta-sedimentary origin that may have been released into the atmosphere or ocean via abrupt eruption, impacting the carbon cycle, and leading to the negative carbon isotope excursions. High-temperature isotope fractionation during volcanic processes could also be an important cause of the negative excursion of carbon isotopes.

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Draft

Figure captions

Fig. 1. Distribution map of the Emeishan LIP, the location of geological sections, and paleogeography of South China (modified from Feng et al. 1996; Wang and Jin 2000). The distribution is divided into four zones: Inner Zone, Middle Zone, Outer Zone, and Continuous Deposition Zone (He et al. 2007; Shi et al. 2017). PD: Podu section (GPS: 28°45'9"N, 106°48'44"E), LSX: Lengshuixi section (GPS: 29°54'0"N, 108°16'29"E), XJC: Xiongjiachang section, NY Basin: North Yangtze basin, WY Basin: West Yangtze basin, JN Basin: Jiangnan basin, LY Basin: Lower Yangtze basin.

Fig. 2. Geological column on Podu and Lengshuixi sections and isotope records in South China. (A) Geological column of the late Guadalupian in the Podu and Lengshuixi sections, including carbon-strontium isotope secular changes, lithology, beds, and conodont stratigraphy. (B) Composite isotope records in South China (after Bond et al. 2010), containing the Laibin (GSSP; Wang et al. 2004; Wignall et al. 2009a), Xiongjiachang, Houchang, and Gouchang sections (Bond et al. 2010). Fm.: Formation, Lith.: Lithology, Wu.: Wuchiapingian, Lo. F.: Longtan Formation, Wu. F.: Wujiaping Formation, GLB: biostratigraphically-defined Guadalupian-Lopingian Boundary, *C.*: *Clarkina*, *J.*: *Jinogondolella*.

Fig. 3. Photos of biological fossils from the late Maokou Formation, South China. Scale bar: 300 µm. (A) Conodonts in the Podu section. 1-2, 4-5, 9: *Jinogondolella* sp. from beds 2, 3, 4, and 7, respectively (upper views). 3: *Jinogondolella altudaensis* from bed 3, upper oblique view (3a) and upper views (3b). 6: *?Hindeodus* sp. from bed 6, upper oblique view. 7-8: *Jinogondolella postserrata* (Behnken) from beds 6 and 7, upper (7a, 7b, and 8b), and upper oblique views (8a), respectively. 10: *Pseudohindeodus ramovsi* Gullo and Kozur from bed 7, lateral view. (B) Some biological fossil debris found in the Lengshuixi

541 section due to disturbance from clear dolomitization or late geological variations. Specimens 1-7 from bed
 542 1, 8-9 from bed 4, 10-11 from bed 8, 12-13 from bed 11, and 17-18 from bed 14, respectively. All
 543 specimens are displayed in lateral view.

544 **Fig. 4.** Modes for producing CO₂ and other gases and the carbon isotope fractionation curve in different
 545 stages of the LIP development process. The black bars (A to D) represent the relative amounts of CO₂, SO₂,
 546 and CH₄ released at various temperatures in the LIP development process, respectively. D (~300 °C): CO₂
 547 from metamorphic degassing of organic carbon; C and B (approximately 500 to 700 °C): decarbonation of
 548 dolomite producing CO₂; A (~900 °C): desulfation of anhydrite; over 1200 °C: CO₂ from wall-rock
 549 assimilation or the melting of carbonates produced during the near eruption stage (Ganino et al. 2008).
 550 Carbon isotope fractionation curve after Bottinga (1968, 1969) and Chacko et al. (1991).

551 **Fig. 5.** Skeleton plot of the development processes behind major geological events and the biodiversity
 552 crisis in the Late Guadalupian, South China. Intensity of the Emeishan LIP event is based on Sun et al.
 553 (2010). Lop.: Lopingian, Wu.: Wuchiapingian, Con.: Conodont Zones, *C. p. p.*: *Clarkina postbitteri*
 554 *postbitteri*, *J. alt.*: *Jinogondolella altudaensis*, *J. post-shan.*: *Jinogondolella postserrata-shannoni*.

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141x118mm (300 x 300 DPI)

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198x185mm (300 x 300 DPI)

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124x90mm (300 x 300 DPI)

Fig. 4. Modes for producing CO₂ and other gases and the carbon isotope fractionation curve in different stages of the LIP development process. The black bars (A to D) represent the relative amounts of CO₂, SO₂, and CH₄ released at various temperatures in the LIP development process, respectively. D (~300 °C): CO₂ from metamorphic degassing of organic carbon; C and B (approximately 500 to 700 °C): decarbonation of dolomite producing CO₂; A (~900 °C): desulfation of anhydrite; over 1200 °C: CO₂ from wall-rock assimilation or the melting of carbonates produced during the near eruption stage (Ganino et al. 2008). Carbon isotope fractionation curve after Bottinga (1968, 1969) and Chacko et al. (1991).

222x200mm (300 x 300 DPI)

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127x112mm (300 x 300 DPI)

Major geological events in the Late Guadalupian and carbon-strontium isotope responses in the Yangtze platform, South China

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Supplementary Figures

Figure S1. Photos of the stratigraphic boundary in the Podu section (a) and Lengshuixi section (b), South China.

Figure S2. Residue-bearing terrigenous materials found in carbonate samples after dissolving by acetic acid. The residues contained some typical magmatic or metamorphic mineral components, including mica, quartz, and other silicates. a, b, and c: silicate minerals in Bed 7, Podu section; d: silicate minerals in Bed 21, e and f: silicate minerals in Bed 23, Lengshuixi section, South China. 1, 5, 8, and 9: mica; 2 and 10: quartz; 3: andalusite; 4: titanite; 6: feldspar; 7: garnet (interpreted).

Supplementary Tables

Table S1. Isotope compositions and Mn and Sr contents of marine carbonate strata in the Podu section, South China.

Sample number	$\delta^{13}\text{C}$ (‰, PDB)	Sr $^{87}\text{Sr}/^{86}\text{Sr}$	2σ	Mn (ppm)	Sr (ppm)	Mn/Sr
NBS987		0.71022	5.99E-06			
NBS987		0.71025	7.79E-06			
PD-1-1	0.88	0.70689	2.35E-05	156	604	0.258
PD-1-2	2.08	0.70693	2.66E-05	192	651	0.295
PD-1-3	1.60	0.70699	4.16E-05	131	590	0.223
PD-1-4	3.05	0.70701	1.54E-05	93	620	0.149
PD-1-5	4.21	0.70702	3.43E-05	94	841	0.112
PD-1-6	2.69	0.70714	1.36E-05	128	963	0.132
PD-1-7	4.40	0.70704	2.29E-05	70	851	0.082
PD-2-1	4.08	0.70703	2.14E-05	51	799	0.064
PD-2-2	4.34	0.70707	1.72E-05	58	839	0.070
PD-2-3	4.08	0.70708	1.72E-05	63	831	0.076
PD-2-4	3.88	0.70709	1.64E-05	69	881	0.078
PD-2-5	4.08	0.70706	3.68E-05	81	843	0.097
PD-3-1	3.35	0.70702	1.07E-05	96	1097	0.088
PD-3-2	3.88	0.70706	2.31E-05	76	963	0.079
PD-3-3	3.57	0.70707	4.19E-05	63	886	0.071
PD-4-1	3.66	0.70704	3.21E-05	72	894	0.081
PD-4-2	3.82	0.70709	1.27E-05	54	878	0.062
PD-4-3	3.71	0.70707	8.98E-06	44	907	0.049
PD-5-1	4.01	0.70704	1.76E-05	53	901	0.058
PD-5-2	3.78	0.70703	1.22E-05	38	878	0.044
PD-6-1	3.73	0.70704	1.47E-05	44	914	0.048
PD-6-2	4.06	0.70704	1.50E-05	30	882	0.034
PD-7-1	3.93	0.70702	1.20E-05	35	872	0.041
PD-7-2	2.35	0.70720	2.32E-05	56	985	0.057
PD-7-3	3.83	0.70710	2.18E-05	23	814	0.029
PD-7-4	4.00	0.70731	1.34E-05	23	909	0.026
PD-8-1	4.06	0.70704	1.33E-05	19	835	0.022
PD-8-2	3.83	0.70702	2.50E-05	19	908	0.021
PD-9	3.70	0.70704	1.93E-05	17	876	0.019
PD-10-1	3.92	0.70700	2.98E-05	10	919	0.011
PD-10-2	3.62	0.70706	1.59E-05	12	831	0.015
PD-10-3	3.79	0.70706	1.21E-05	12	928	0.013
PD-10-4	3.79	0.70705	2.06E-05	11	888	0.012
PD-11-1	4.13	0.70709	1.34E-05	11	862	0.013
PD-11-2	4.00	0.70706	3.19E-05	18	773	0.023
PD-12-1	3.75	0.70712	8.69E-06	14	869	0.016
PD-12-2	3.76	0.70710	2.09E-05	14	847	0.016
PD-13-1	3.79	0.70703	1.30E-05	16	884	0.018
PD-13-2	3.68	0.70700	1.96E-05	14	1001	0.014
PD-14-1	3.71	0.70704	2.91E-05	13	890	0.014
PD-14-2	3.96	0.70697	2.95E-05	12	881	0.014

PD-14-3	4.13	0.70704	2.11E-05	18	874	0.021
PD-15	4.03	0.70709	1.54E-05	14	956	0.015

Table S2. Isotope compositions and Mn and Sr contents of marine carbonate strata in the Lengshuixi section, South China.

Sample number	$\delta^{13}\text{C}$ (‰, PDB)	Sr		Mn (ppm)	Sr (ppm)	Mn/Sr
		$^{87}\text{Sr}/^{86}\text{Sr}$	2σ			
NBS987		0.71026	6.20E-06			
NBS987		0.71023	6.64E-06			
NBS987		0.71027	1.80E-05			
NBS987		0.71025	9.05E-06			
LSX-1	0.01	0.70699	3.47E-05	86	864	0.100
LSX-2	2.43	0.70703	1.35E-05	65	1009	0.065
LSX-3	1.81	0.70704	3.76E-05	54	585	0.092
LSX-4	3.05	0.70711	9.54E-06	34	1031	0.033
LSX-5	2.82	0.70703	1.90E-05	84	604	0.139
LSX-6-1	2.80	0.70718	1.26E-05	31	1051	0.030
LSX-6-2	1.82	0.70725	4.46E-05	28	776	0.036
LSX-7	2.67	0.7071	1.08E-05	15	747	0.020
LSX-8	2.81	0.70692	2.66E-05	17	807	0.022
LSX-9-1	2.29	0.70689	9.65E-06	26	2347	0.011
LSX-9-2	2.22	0.70695	1.45E-05	36	939	0.038
LSX-10-1	2.12	0.70690	1.53E-05	42	1340	0.031
LSX-10-2	2.49	0.70700	1.76E-05	25	551	0.045
LSX-10-3	2.35	0.70712	4.26E-06	22	744	0.030
LSX-11-1	3.04	0.70706	1.17E-05	20	859	0.024
LSX-11-2	2.39	0.70699	2.47E-05	28	637	0.044
LSX-11-3	2.79	0.70695	1.21E-05	44	1254	0.035
LSX-12	2.77	0.70696	2.95E-05	23	977	0.024
LSX-13	2.84	0.70703	2.07E-05	21	800	0.026
LSX-14	2.60	0.70693	1.71E-05	23	675	0.035
LSX-15	3.01	0.70721	2.16E-05	13	771	0.017
LSX-16	2.22	0.70696	2.29E-05	24	555	0.043
LSX-17	2.32	0.70698	4.17E-05	24	612	0.039
LSX-18	2.69	0.7072	1.47E-05	35	490	0.071
LSX-19	2.34	0.70706	2.48E-05	33	773	0.043
LSX-20	2.72	0.70731	2.97E-05	26	1110	0.023
LSX-21	2.61	0.70720	6.69E-05	19	619	0.030
LSX-22	2.87	0.70727	2.45E-05	29	594	0.048
LSX-23-1	2.63	0.70719	1.67E-05	72	542	0.132
LSX-23-2	2.68	0.70725	9.24E-06	43	1222	0.035
LSX-23-3	2.62	0.70786	1.57E-05	33	1005	0.032
LSX-24	1.86	0.70765	1.29E-05	66	731	0.091
LSX-25	2.64	0.70762	1.38E-05	33	821	0.041
LSX-26	2.91	0.70772	2.98E-05	26	931	0.028
LSX-27	2.52	0.70700	1.11E-05	34	2604	0.013

LSX-28	2.19	0.70699	5.32E-06	223	996	0.224
LSX-29	2.82	0.70715	2.50E-05	43	2115	0.020
LSX-30	1.92	0.70710	1.41E-05	104	573	0.182
LSX-31	3.03	0.70733	7.31E-06	15	1170	0.013
LSX-32	2.91	0.70746	9.67E-06	35	1210	0.029
LSX-33	2.33	0.70721	4.73E-05	26	852	0.031
LSX-34	2.72	0.70739	1.32E-05	22	1247	0.017
LSX-36	2.63	0.70708	1.35E-05	35	2305	0.015
LSX-38	2.78	0.70730	1.43E-05	40	612	0.066

Table S3. Electron probe analyses of representative minerals numbered in Figure S2.

Sample number	Chemical content in wt%								Sampling location	Mineral
	SiO ₂	Al ₂ O ₃	FeO	K ₂ O	CaO	MnO	TiO ₂	MgO		
1	36.56	15.92	22.86	1.34	9.37	—	10.05	3.91	PD-7	Mica
2	98.70	0.24	1.04	—	—	—	—	—	PD-7	Quartz
3	49.20	50.80	—	—	—	—	—	—	PD-7	Andalusite
4	32.53	1.66	—	—	28.30	—	37.51	—	PD-7	Titanite
5	37.10	14.75	34.79	4.13	0.64	1.62	1.73	5.25	LSX-21	Mica
6	70.02	15.49	—	14.49	—	—	—	—	LSX-21	Feldspar
7	35.38	—	33.45	—	1.58	29.60	—	—	LSX-21	Garnet (?)
8	58.93	28.45	4.68	6.03	—	1.91	—	—	LSX-23	Mica
9	42.21	13.69	27.23	4.64	—	—	1.97	10.26	LSX-23	Mica
10	100	—	—	—	—	—	—	—	LSX-23	Quartz