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Integrated Sr isotope variations and global environmental changes through the Late Permian to early Late Triassic

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Abstract

New $^{87}Sr/^{86}Sr$ data based on 127 well-preserved and well-dated conodont samples from South China were measured using a new technique based on single conodont albid crown analysis. These reveal a spectacular climb in seawater $^{87}Sr/^{86}Sr$ ratios during the Early Triassic that was the most rapid of the Phanerozoic. The rapid increase began in

Bed 25 of the Meishan section (GSSP of the Permian-Triassic boundary, PTB), and coincided closely with the latest Permian extinction. Modelling results indicate that the accelerated rise of ^{87}Sr /^{86}Sr ratios can be ascribed to a rapid increase (> 2.8×) of riverine flux of Sr caused by intensified weathering. This phenomenon could in turn be related to an intensification of warming-driven run-off and vegetation die-off. Continued rise of ^{87}Sr /^{86}Sr ratios in the Early Triassic indicates that continental weathering rates were enhanced > 1.9 times compared to that of the Late Permian. Continental weathering rates began to decline in the middle-late Spathian, which may have played a role in the decrease of oceanic anoxia and recovery of marine benthos. The ^{87}Sr /^{86}Sr values decline gradually into the Middle Triassic to an equilibrium values around 1.2 times those of the Late Permian level, suggesting that vegetation coverage did not attain pre-extinction levels thereby allowing higher run-off.

Keywords: strontium isotopes; Permian-Triassic extinction; Early Triassic; conodonts

1. Introduction

The Permian-Triassic (P-Tr) crisis eliminated over 90% of marine species (Erwin, 1993; Song et al., 2013), and also had a severe impact on terrestrial ecosystems (Benton and Newell, 2014). This disaster event is widely attributed to the eruption of the Siberian Traps and associated environmental effects (e.g. Wignall, 2001). After the P-Tr crisis, volcanically induced environmental perturbations, such as global warming, enhanced soil erosion, and ocean anoxia, are postulated to have lasted almost the entire Early Triassic (Wignall and

Twitchett, 2002; Payne et al., 2004; Retallack, 2005; Joachimski et al., 2012; Song et al., 2012; Sun et al., 2012). Unstable and stressful environments no doubt contributed to the fitful recovery of marine taxa in the aftermath of the P-Tr extinction (e.g. Brayard et al., 2009; Song et al., 2011). At present, it is difficult to uncover the interconnections between environmental factors on land and in the sea because few proxies can be directly related to both terrestrial and marine settings. Seawater $^{87}Sr/^{86}Sr$ ratios are an exception because they are primarily determined by fluxes from continental weathering via rivers and seafloor hydrothermal circulation at mid-ocean ridges (Broecker and Peng, 1982; Palmer and Edmond, 1989). Therefore, strontium isotopes associated with oxygen isotopes and trace elements derived from bio-apatite (conodonts) provide an opportunity to study the secular changes of terrestrial and marine environments.

Significant changes in seawater $^{87}Sr/^{86}Sr$ values during the P-Tr transition have already been reported in the previous studies (e.g. Veizer and Compston, 1974). Subsequent publications based on whole rock analyses of carbonate samples have added some details on P-Tr strontium isotopic evolution (e.g. Huang et al., 2008; Sedlacek et al., 2014). However, obtaining original seawater values from whole rock analyses is difficult due to common diagenetic alteration (Brand, 2004). Low-Mg calcitic and phosphatic shells are believed to represent the most reliable fossil material for measuring Sr isotope ratios in the geologic time (e.g. Korte et al., 2003; Brand et al., 2012). Thus, the Late Permian and Triassic Sr isotopic data are mainly derived from brachiopods and conodonts (e.g. Martin and Macdougall, 1995; Korte et al., 2003; 2006). The rarity of articulate brachiopods in the aftermath of the P-Tr extinction and the need for sufficient conodont elements for measurement has hindered the establishment of a detailed Early Triassic Sr isotopic curve; existing knowledge is based on conodont data from only a few levels (e.g. Martin and Macdougall, 1995; Korte et al., 2003; 2004).

In this study, a new technique of *in situ* Sr isotope measurement using LA-MC-ICPMS on single conodont albid crown is applied to provide a high precision seawater $^{87}Sr/^{86}Sr$ curve through the Late Permian to early Late Triassic. Samples come from the Meishan, Qingyan, and Guandao sections in South China, which are biostratigraphically well constrained and were previously investigated for changes in Th/U ratios, ΩCe, and $\delta^{18}O$ in conodont apatites by Song et al. (2012), Joachimski et al. (2012), and Sun et al. (2012), thus making it easy to directly compare changes between these different environmental proxies.

2. Geological setting

The South China Block was located near the equator in the eastern Palaeotethys Ocean during the P-Tr transition (Fig. 1A). The three sections sampled in this study, Meishan, Guandao, and Qingyan sections, have high-resolution conodont stratigraphic data (see Yin et al., 2001; Payne et al., 2004; Ji et al., 2011), providing accurate correlation and a high resolution age model (Fig. 1C).

Meishan was situated in the northeastern slope margin of Yangtze Platform during the P-Tr transition (Fig. 1B). As the Global Stratotype Section and Point (GSSP) of the P-Tr boundary (PTB), Meishan strata provide a complete Changhsingian and PTB record (Yin et al., 2001). The PTB is placed at the base of Bed 27c marked by the first occurrence of conodont *Hindeodus parvus* (Yin et al., 2001). During the Early-Middle Triassic, Guandao was located at the platform margin of the Great Bank of Guizhou, Nanpanjiang Basin. The location comprises two sections: the Upper Guandao section mainly exposes the Upper Permian to Anisian succession, and the Lower Guandao section consists of a Spathian to Carnian succession (Payne et al., 2004). The Qingyan section was located at the transition between Yangtze Platform and Nanpanjiang Basin (Fig. 1B) and exposes an uppermost Permian to Middle Triassic succession.

3. Materials and methods

A total of 127 conodont samples ranging in age from the Late Permian to the early Late Triassic were collected from South China. Among them, 33 samples are from the Changhsingian and lower Griesbachian succession in the Meishan sections, 64 samples are from the Griesbachian to lower Carnian succession in the Guandao section, and 30 samples are from the Griesbachian to middle Anisian succession in the Qingyan section (Fig. 2). Conodonts were extracted by dissolving 1 cm^3-sized fragments of limestone with 10% acetic acid, and the residue sieved with the 90-μm to 700-μm fraction. Conodont elements were hand-picked from insoluble residue under a binocular microscope. Only well-preserved single conodont elements with colour alteration index (CAI, an index to evaluate the preservation status of the conodonts) of $\leq$ 2 were selected for analysis.

Transmission electron microscopy investigations and parallel geochemical studies have shown that the albid crown of a conodont element is the apatite tissue most resistant to diagenetic alteration (Trotter and Eggins, 2006). Accordingly, conodont elements with well-preserved albid crown were washed in ultrapure water, affixed to double-sided adhesive carbon tape attached to a silica glass side, and placed within a sample cell. The strontium isotope composition of conodont samples were measured *in situ* by Laser Ablation Multi-collector Inductively Coupled Plasma Mass Spectrometer (LA-MC-ICP-MS) at the State Key Laboratory of Continental Dynamics of Northwest University (Xi'an, China). The MC-ICP-MS system is Nu Plasma HR from Nu Instrument Ltd. Static multi-collection in low-resolution mode was applied during measurement. The used Laser ablation system is a 193 nm ArF-excimer laser (GeoLas 2005). Single spots at the conodont albid crown were assayed using a laser beam size of 60 μm, laser energy of 80 mJ, and a repetition rate of 3 Hz. The operating procedures of these instruments and data reduction are similar to those described by Hobbs et al. (2005) and Yang et al. (2014). Strontium international standards (NIST SRM 987) were measured before the conodont were analyzed and yielded a weighted average $^{87}Sr/^{86}Sr$ value of 0.710248 ± 0.000027 (2σ, n = 19). This obtained value is equal to the NIST SRM 987 reference value of 0.710248 (McArthur et al., 1993). The strontium concentrations of conodont samples are mostly between 1000 and 2500 ppm, and the average value is 1803 ppm (see Supplementary Material). The errors of $^{87}Sr/^{86}Sr$ are concentrated on ~0.0001-0.0002 for the present dataset.

The Sr isotopic composition of seawater is mainly controlled by the proportion of riverine and mantle flux (Palmer and Edmond, 1989; Allègre et al., 2010). The global average $^{87}Sr/^{86}Sr$ ratio of riverine input is 0.7119 and hydrothermal $^{87}Sr/^{86}Sr$ ratio is 0.7035 (Palmer and Edmond, 1989). The residence time of Sr in the oceans is about 3×10^6 years

(Hodell et al., 1990).

A simple box model, based on a modification of Zachos et al.'s study (1999), was performeded in this study. Changes of ocean $^{87}Sr/^{86}Sr$ from t to t+1 were calculated as following equations:

$$Sr(t+1) = Sr(t) + [a/(1+Q) + b*Q/(1+Q) - Sr(t)]/r \qquad (1)$$

$Sr(t)$ is the Sr isotope value at time t; r is the residence time of Sr in the oceans, $r = 3 \times 10^6$ (Hodell et al., 1990); a is the global average $^{87}Sr/^{86}Sr$ ratio of riverine input, a = 0.7119 (Palmer and Edmond, 1989); b is the hydrothermal $^{87}Sr/^{86}Sr$ ratio, b = 0.7035 (Palmer and Edmond, 1989); Q is the ratio of riverine flux to mantle flux, $Q = F_R/F_M$; F_R is the riverine flux; F_M is the mantle flux.

Differentiating equation (1) with respect to t, we obtain the new equation as:

$$dSr(t)/dt = [a/(1+Q) + b* Q/(1+Q) - Sr(t)]/r \qquad (2)$$

Whence equation (3) follows immediately from equation (2):

$$Q = (a - b)/[\mathrm{r}* dSr(t)/dt + Sr(t) - b] - 1 \qquad (3)$$

4. Results

The temporal distribution of $^{87}Sr/^{86}Sr$ values of 127 conodont elements from the Meishan, Qingyan, and Guandao sections are shown in Figure 2. Sr isotopic change through the Late Permian to early Late Triassic can be divided into four stages. The $^{87}Sr/^{86}Sr$ ratios are about 0.70710 with slight variability in the Changhsingian. The second stage is characterized by a steep climb for $^{87}Sr/^{86}Sr$ ratios, from 0.70711 in the latest Changhsingian to 0.70836 in the middle Spathian. This increase started between *Clarkina yini* Zone and *Hindeodus changxingensis* Zone, from 0.70716 at Bed 24e to 0.70751 at Bed 27a in the Meishan section. It has already been noted that the steepest and greatest change in $^{87}Sr/^{86}Sr$ occurred during the Late Permian to Early Triassic period (e.g. McArthur et al., 2001). Here this study shows that the average increase rate of $^{87}Sr/^{86}Sr$ ratios in this interval is about 3.3×10^{-4} Myr^{-1} (0.00125 in 3.8 Myr), which is about seven times steeper than the next fastest rate of increase between the early Eocene and middle Miocene (Hess et al., 1986). The average increase rate of $^{87}Sr/^{86}Sr$ ratios through the early Eocene to middle Miocene is about 4.6×10^{-5} Myr^{-1} (0.00125 in 27 Myr).

In the third stage of our $^{87}Sr/^{86}Sr$ record, ratios decline from 0.70836 at the middle Spathian to ~0.7079 by the end of Anisian. The average descent rate of $^{87}Sr/^{86}Sr$ ratios in this stage is about 7.1×10^{-5} (0.00046 in 6.5 Myr). The $^{87}Sr/^{86}Sr$ ratios in the fourth/last stage from the Ladinian to the Carnian are stable, with values of ~0.7077 showing only modest fluctuations.

5. Discussion

5.1. Comparison of our data with published Sr isotopes through the Late Permian to early Late Triassic

It has long been recognized that the lowpoint of the $^{87}Sr/^{86}Sr$ Phanerozoic record was in the Middle Permian (Martin and Macdougall, 1995, Jasper, 1999; Korte et al., 2003) and was followed by gradual increase beginning near the Guadalupian-Lopingian boundary (Korte et al., 2006). Here, our data show that the Changhsingian $^{87}Sr/^{86}Sr$ ratios are characterized by a plateau with slightly fluctuations around 0.70710. The Changhsingian $^{87}Sr/^{86}Sr$ values of conodont crowns from the Meishan section, are clearly lower than several published Sr isotopic values from whole conodont samples from Pakistan (Martin and Macdougall, 1995) and Iran (Korte et al., 2003; 2004), but are similar with brachiopod values from the Southern Alps, northern Italy, Iran and South China (Korte et al., 2006; Brand et al., 2012). At Meishan, a marked increase in $^{87}Sr/^{86}Sr$ ratios began at the main extinction horizon (between *Clarkina yini* and *Hindeodus changxingensis* zones). The average $^{87}Sr/^{86}Sr$ values in the *Clarkina yini* Zone and *Hindeodus changxingensis* Zone-*Hindeodus parvus* Zone are 0.70710 (N = 9) and 0.70729 (N = 5), respectively. In contrast, both Twitchett (2007) and Korte et al. (2010) suggested there was a plateau of Sr isotope ratios during the PTB interval caused by the weathering of non-radiogenic basaltic rocks from the Siberian Traps. This trend is not seen in our data and it is likely that it is caused by inaccuracies of geochronological constrains across the PTB. Furthermore it is unclear that there would have been substantial basalt weathering because the eruptions occurred in paralic locations and large areas were subsequently buried and preserved beneath younger strata of the West Siberian Basin (Reichow et al., 2009). Finally the onset of basalt eruptions probably occurred at the same time as the extinction with most lava emplacement occurring after this crisis (Reichow et al. 2009).

Latest U-Pb dating reveals a very short duration for the PTB interval between *Clarkina yini* Zone and *Isarcicella isarcica* Zone: either 0.18 Myr using the ages obtained by Shen et al. (2011) or 0.061 Myr using Burgess et al.'s (2104) dates. A somewhat longer duration (0.58 myr) is obtained using the $^{40}Ar/^{39}Ar$ dating method (Reichow et al., 2009). Regardless of which dating technique is used, it is clear the steepest climb of $^{87}Sr/^{86}Sr$ values occurred during the PTB interval (see Figs. 2, 4).

Published strontium isotopic data from conodonts and brachiopod shells in the Early Triassic are much rarer than those from the Late Permian and the Middle Triassic (see Fig. 2). One reason for this is the extreme rarity of well-preserved brachiopod shells in the post-extinction successions. Early Triassic conodont Sr isotopic data have been reported in Pakistan and Italy (Martin and Macdougall, 1995; Korte et al., 2003; Twitchett, 2007). Here, Sr isotopic data obtained from conodonts in South China provide a high-solution $^{87}Sr/^{86}Sr$ curve for the Early Triassic that shows ratios continued to rise until the mid-Spathian before they began to gradually fall. This decreasing trend continued into the Anisian. This earliest Middle Triassic decrease is comparable to those obtained in both brachiopod shells and conodonts in Europe (see Korte et al., 2003) although some higher values are found in the early Anisian (Fig. 2). Ladinian $^{87}Sr/^{86}Sr$ are about 0.7077 with a moderate fluctuation, a slightly higher value than previously obtained from conodonts but slightly lower than those obtained from brachiopods (Korte et al., 2003).

5.2. Modeling $^{87}Sr/^{86}Sr$: causes for rapid increase of seawater $^{87}Sr/^{86}Sr$

The isotopic composition of Sr in the oceans is homogenous because the residence time of Sr is much longer (~3×10^{6} year) than the mixing time of the oceans (~10^{3} year; Broecker and Peng, 1982). The $^{87}Sr/^{86}Sr$ values of seawater depend primarily on two major sources: the riverine input of radiogenic Sr due to continental weathering and mantle input via hydrothermal circulation at mid-ocean ridges (Palmer and Edmond, 1989; Allègre et al., 2010). Thus, variation in seawater $^{87}Sr/^{86}Sr$ through time is useful for correlating and dating sedimentary rocks and investigating geologic processes such as continental weathering, global climate changes, tectonic uplifts, and hydrothermal circulation (Veizer and Compston, 1974; McArthur et al., 2001). The rivers supply more radiogenic Sr (average $^{87}Sr/^{86}Sr$ = 0.7119) to the oceans compared with hydrothermal inputs of 0.7035 (Palmer and Edmond, 1989). Recently, this simple interpretation has been challenged because hydrothermal flux estimates in modern oceans is much less that the flux required to balance the oceanic Sr budget (Allègre et al., 2010). To balance the marine Sr cycle, Berner (2006) argued that the present volcanic Sr weathering flux on land is ~2.7 times higher than that of basalt-seawater Sr exchange. Allègre and others (2010) re-assessed the Sr isotopic budget and documented that intensive weathering on volcanic islands represent ~60% of mantle Sr input to the oceans, with the remaining 40% supplied by ridge-crest hydrothermal activity and seafloor low-temperature alteration of basalts.

A simple model indicates that the rapid increase of $^{87}Sr/^{86}Sr$ ratios during the PTB interval could have resulted from either a rapid increase of continental weathering (Fig. 3) or a quick decline of mantle Sr flux. However, the eruptions of Siberian Traps near the PTB would have strengthened volcanic weathering in the Early Triassic, even if substantial parts of the lava flows were buried, increasing the mantle Sr flux. Therefore, it is likely that a progressive increase of $^{87}Sr/^{86}Sr$ ratios in the Early Triassic most likely resulted from enhanced continental weathering. On the basis of a simple box model, we present a F_R/F_M (the ratio of riverine flux to mantle flux) curve through the Late Permian to early Late Triassic (Fig. 4B). F_R/F_M in the PTB interval is ~2.8 times that of the Late Permian and the mean ratios in the Early Triassic has increased by ~1.9 times. If we consider the effects of Siberian Trap eruption which would be expected to drive down $^{87}Sr/^{86}Sr$ ratios, then the effect of increased $^{87}Sr/^{86}Sr$ ratios due to continental weathering would be even stronger. The basaltic flows of the Siberian Traps covered about 2.6×10^6 km^2 (Reichow et al., 2002), the volcanic island surface area in the Late Permian is unknown but the modern counterpart is about 4.8×10^6 km^2 (Allègre et al., 2010). In addition, PTB volcanic layers and rocks are widespread in the Tethys regions and western margin of the Panthalassa (Yin and Song, 2013, and references therein). Thus, Siberia Traps eruptions and contemporaneous volcanism around the world may have caused a significant increase of unradiogenic strontium entering the oceans (Twitchett, 2007; Korte et al., 2010). The rapid climb of F_R/F_M may have been buffered by the basalts and tuffs from the Siberian Traps and contemporaneous volcanisms. Accordingly, the continental weathering rates during the PTB interval and the Early Triassic may have increased by substantially more than 2.8 and 1.9 times, respectively (Fig. 4).

5.3. Controls of elevated continental weathering

An increase of continental weathering rates can be caused by several factors: regression, global warming and an enhanced hydrological cycle, and the demise of the land plants. Rapid regression in the latest Permian *Clarkina meishanensis* zone (Yin et al., 2014) could plausibly have involved a quick increase of continental erosion but this is short-term effect compared with the 4 Myr duration of Sr isotopic increase and sea-level rose rapidly following brief regression (Haq et al., 1987; Hallam and Wignall, 1999). Acid rain, as a byproduct of Siberian Traps eruptions, has been considered a cause of P-Tr terrestrial extinction (Sephton et al., 2005; Black et al., 2014) and is a potential source of increased chemical weathering. However, Early Triassic evidence from Germany and South Africa show that any acidification was not prolonged (Retallack et al., 1996).

Climatic warming began at the extinction horizon, persisted into the Early Triassic (Joachimski et al., 2012; Sun et al., 2012; Schobben et al 2014), and is thought to be an important contributory factor in the mass extinction.(Song et al., 2014). Both laboratory studies and field observations reveal a strong dependence of chemical weathering on temperature (Kump et al., 2000, and references therein). Therefore, the extreme high temperature found in the PTB interval and most of the Early Triassic (Joachimski et al., 2012; Sun et al., 2012; Schobben et al 2014) may have resulted in an accelerated chemical weathering in this period. In addition, Early Triassic global warming may have resulted in an amplified hydrological cycle and extreme seasonal rainfall along the margins of the Tethys Ocean (Schobben et al., 2014). As primary factors influencing chemical erosion rates (Kump et al., 2000), amplified hydrological cycle and intensified monsoonal activity may have played considerable roles on the enhanced continental weathering in the Early Triassic.

The accelerated weathering coincided with the die-off of vegetation (Retallack et al., 1996) and $^{87}Sr/^{86}Sr$ ratios only began to decline following the recovery of vegetation in the middle-late Spathian (Looy et al., 1999). Vegetation provides essential erosion resistance, hence soil erosion are sensitive to changes in vegetation abundance and composition. The plant ecosystems suffered a severe crisis during the P-Tr extinction, and this was followed by an Early Triassic coal gap (Retallack et al., 1996). The loss of land vegetation is likely to have resulted in extensive soil erosion. Earliest Triassic sepic pedoliths indicate severe and wide spread soil erosion associated with forest dieback at the P-Tr boundary (Retallack, 2005). The widespread transition from meandering to braided channels at this time has been attributed to the increased sediment delivery from vegetation-denuded hill slopes to channels lacking abundant rooted plants (Ward et al., 2000). The excessive supply of soil materials would have offered increased riverine strontium flux to the oceans. Renewed proliferation of conifer-forest did not occur until the Spathian (Looy et al., 1999). Accordingly, continental weathering pattern could be explained by vegetation gap in the Early Triassic especially if coincident with an increase of run-off (Looy et al., 1999; Korte et al., 2003; Schobben et al., 2014).

In sum, we conclude that a combination of climatic warming, intensified hydrological cycle, and vegetation die-off were the primary cuase of enhanced continental weathering, and increased radiogenic Sr flux to the oceans, during the PTB interval and most of the Early Triassic.

5.4. Effects of enhanced weathering on marine environments

These Sr isotopic data in this study derive from sections with a well-established redox and SST (sea surface temperature) data (see Joachimski et al., 2012; Song et al., 2012; Sun et al., 2012; Chen et al., 2013), allowing co-evaluation of these environmental parameters. Synchronous changes in $^{87}Sr/^{86}Sr$ ratios, temperature, redox states, and F_R/F_M imply causal relations among these parameters (Fig. 5). The loss of vegetation cover and enhanced weathering would have drastically increased terrestrial exports of organic and inorganic sediment to the oceans. Increased nutrient input is modelled as a key factor in generating enhanced productivity and consequent oxygen depletion in Early Triassic oceans (e.g. Winguth & Winguth 2012). The decline in $^{87}Sr/^{86}Sr$ ratios in the middle-late Spathian may reflect a decline in continental weathering. The recovery of land vegetation, and especially the reappearance of forests at this time (Looy et al., 1999), lends credence to this notion. Subdued weathering and falling global temperatures in the middle-late Spathian may also help to explain the coincident decline of anoxia in the ocean (Song et al., 2012) and the accelerated recovery of benthic ecosystems (Song et al., 2011).

During the Middle Triassic, the observation that $^{87}Sr/^{86}Sr$ values declined gradually and stabilised with $^{87}Sr/^{86}Sr$ ratios suggesting continental weathering was about 1.2 times that of the Late Permian levels (Fig. 5). The Early Triassic coal gap – a signal of low vegetation density – persisted into the Middle Triassic (Retallack et al., 1996; Sun et al., 2012) and this may explain why high levels of terrestrial run-off helped to maintain the elevated $^{87}Sr/^{86}Sr$ values for over 15 Myr after the P-Tr mass extinction.

6. Conclusions

New Sr isotopic measurements based on 127 well-preserved and well-dated conodont samples from South China show that the Late Permian-early Late Triassic $^{87}Sr/^{86}Sr$ is marked by two plateaus, a steep climb and a gradual decrease. The plateaus occur in the Changhsingian and Ladinian-Carnian, and are characterized by constant values around 0.70710 and 0.70770, respectively. The steep climb occurs at the main extinction horizon, reaching 0.70836 at the middle Spathian, followed by a gradual decrease from the late Spathian to the earliest Ladinian.

The rapid increase of $^{87}Sr/^{86}Sr$ ratios during the P-Tr transition is associated with, and probably caused by, a rapid increase of continental weathering. Modelling results indicate that the weathering rates during the PTB interval and the Early Triassic increased by >2.8 and >1.9 times, respectively. A combination of global warming, intensified hydrological cycle, and vegetation die-off probably contributed to the extremely high $^{87}Sr/^{86}Sr$ values and enhanced continental weathering in the Early Triassic. The loss of vegetation cover and enhanced weathering drastically increased the terrestrial export of nutrients to the oceans, which increased productivity and depleted oxygen. The combination of long-term intensified continental weathering and climate anomalies accord well the observed delay in recovery of benthic marine ecosystems following the P-Tr mass extinction.

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Figure captions

Fig. 1. Schematic maps of the study areas. A. Palaeogeography illustrating the position of South China during the end-Permian extinction (after Erwin, 1993). B. Map showing the studying sites (after Song et al., 2013). C. Conodont zones in the Meishan, Qingyan, and Guandao sections.

Fig. 2. Comparison of $^{87}Sr/^{86}Sr$ for conodonts from this study with literature data derived from conodonts and brachiopods. Time scale is constrained by conodont stratigraphic data. Radiometric dates are from Cohen et al. (2013). Changhs., Changhsingian; Griesba., Griesbachian; *C.*, *Clarkina*; *H.*, *Hindeodus*; *I.*, *Isarcicella*; *Ns.*, *Neospathodus*; *Ic.*, *Icriospathodus*; *Tr.*, *Triassospathodus*; *Cs.*, *Chiosella*; *Ni.*, *Nicoraella*; *Pg.*, *Paragondolella*; *Ng.*, *Neogondolella*; *Bv.*, *Budurovignathus*.

Fig. 3. Numerical model simulating the influence of the riverine flux to mantle flux (F_R/F_M) ratio on ocean $^{87}Sr/^{86}Sr$ values. The initial ocean $^{87}Sr/^{86}Sr$ value is set to the earliest Changhsingian value (the dark line at ~0.7071) where F_R/F_M is 0.73. A, a rapid increase of

F_R/F_M by 1.5×, 2.0×, 2.5×, and 3.0× times. B, a gradual increase of F_R/F_M by 1.5×, 2.0×, 2.5×, and 3.0× times.

Fig. 4. A. The LOWESS (locally weighted scatterplot smoothing) curve for smoothing of strontium isotopic data in this study. B, Numerical model showing F_R/F_M changes through the Late Permian to early Late Triassic.

Fig. 5. The variations of seawater $^{87}Sr/^{86}Sr$, SST (sea surface temperature), redox state, F_R/F_M, and sea level through the Late Permian to early Late Triassic. Conodont $\delta^{18}O$ data are from Joachimski et al. (2012), Sun et al. (2012), Chen et al. (2013), and Trotter et al. (2015). Conodont ΩCe data are from Song et al. (2012). Sea-level changes modified from Haq et al. (1987) and Yin et al. (2014).

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