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Integrated carbon, sulfur, and nitrogen isotope chemostratigraphy of the Ediacaran Lantian Formation in South China: spatial gradient, ocean redox oscillation, and fossil distribution

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Abstract

The Ediacaran Doushantuo Formation in South China is a prime target for geobiological investigation because it offers opportunities to integrate chemostratigraphic and paleobiological data.

Previous studies were mostly focused on successions in shallow-water shelf facies, but data from deep-water successions are needed to fully understand basinal redox structures. Here we report $\delta^{13}C_{carb}$, $\delta^{13}C_{org}$, $\delta^{34}S_{pyr}$, $\delta^{34}S_{CAS}$, and $\delta^{15}N_{sed}$ data from a drill core of the fossiliferous Lantian Formation, which is a deep-water equivalent of the Doushantuo Formation. Our data confirm a large (>10‰) spatial gradient in $\delta^{13}C_{carb}$ in the lower Doushantuo/Lantian formations, but this gradient is probably due to the greater sensitivity of carbonate-poor deep-water sediments to isotopic mixing with ^{13}C-depleted carbonate cements. A pronounced negative $\delta^{13}C_{carb}$ excursion (EN3) in the upper Doushantuo/Lantian formations, however, is spatially consistent and may be an equivalent of the Shuram excursion. $\delta^{34}S_{pyr}$ is more negative in deeper-water than shallow-water facies, particularly in the lower Doushantuo/Lantian formations, and this spatial pattern is interpreted as evidence for ocean redox stratification: pyrite precipitated in euxinic deep waters has lower $\delta^{34}S_{pyr}$ than that formed within shallow-water sediments.

The Lantian Formation was probably deposited in oscillating oxic and euxinic conditions. Euxinic black shales have higher TOC and TN contents, but lower $\delta^{34}S_{pyr}$ and $\delta^{15}N_{sed}$ values. In euxinic environments, pyrite was predominantly formed in the water column and organic nitrogen was predominantly derived from nitrogen fixation or NH_4^+ assimilation because of quantitative denitrification, resulting in lower $\delta^{34}S_{pyr}$ and $\delta^{15}N_{sed}$ values. Benthic macroalgae and putative animals occur exclusively in euxinic black shales. If preserved in-situ, these organisms must have lived in brief oxic episodes punctuating largely euxinic intervals, only to be decimated and preserved when the local environment switched back to euxinia again. Thus, taphonomy and ecology were the primary factors controlling the stratigraphic distribution of macrofossils in the Lantian Formation.

Key words: Ediacaran Period, Lantian Formation, South China, carbon isotopes, sulfur isotopes, nitrogen isotopes

INTRODUCTION

The co-evolution of life and environment is highlighted in the recent debate on the possible relationship between ocean oxygenation and animal evolution in the Neoproterozoic. Some argue that the progressive oxygenation of the Earth's surface in the late Neoproterozoic removed the final environmental hurdle to animal evolution (Planavsky *et al.*, 2014; Cole *et al.*, 2016), whereas others counter that atmospheric pO_2 levels were sufficiently high to support basal animal metabolsim in the Mesoproterozoic, long before the rise of animals themselves (Zhang *et al.*, 2015). In the latter scenario, it is the rise of animals to ecological importance that drove the the progressive oxygenation of deep oceans (Lenton *et al.*, 2014). Building upon previous work (Lyons *et al.*, 2009; Och & Shields-Zhou, 2012), recent attempts to test these competing hypotheses are focused on the compilations and meta-analyses of redox proxies including Fe speciation data (e.g., Sperling *et al.*, 2015b), redox-sensitive trace elements (e.g., fig. 4 of Sahoo *et al.*, 2016), and stable isotopes (e.g., Pogge von Strandmann *et al.*, 2015). While these meta-analyses offer a critical view of the big-picture redox trend through the Neoproterozoic, they also omit the more nuanced picture of spatial heterogeneity and often lack the stratigraphic resolution to appreciate the temporal dynamics of Neoproterozoic redox evolution. More importantly, these studies necessarily depend on global stratigraphic correlations to draw conclusions about the co-evolution of life and environment. Such correlations come with uncertainties that are often under appreciated (Xiao *et al.*, 2016). As a complementary approach, one can examine fossiliferous Neoproterozoic successions at high stratigraphic resolution in order to analyze the evolutionary, ecological, and taphonomic factors that control the local and regional distribution of Neoproterozoic organisms. This approach has been applied with success in the investigation of several Ediacaran fossiliferous successions (Fike *et al.*, 2006; McFadden *et al.*, 2008; Johnston *et al.*, 2013; Sperling *et al.*, 2015a; Wood *et al.*, 2015; Li *et al.*, 2015; Cui *et al.*, 2016a).

Ediacaran successions in South China, particularly the Doushantuo Formation, are highly fossiliferous (Xiao *et al.*, 2002; Liu *et al.*, 2014; Xiao *et al.*, 2014). Thus, they are ideal targets for focused and integrated geochemical analysis at high resolution in order to infer the impact of redox conditions on evolution, ecology, and taphonomy. Previous studies of the Doushantuo Formation have been focused on shelf sections in the Yangtze Gorges and surrounding areas, where the Doushantuo Formation is highly fossiliferous and consists of carbonates, shales, and phosphorites (McFadden *et al.*, 2008; Li *et al.*, 2010; Sawaki *et al.*, 2010; Cui *et al.*, 2015). To fully understand the Ediacaran redox structure of South China, it is necessary to gain insights from sections in deep-water slope and basinal facies. However, integrated studies of the Ediacaran successions in deep-water facies are few, and these tend to have limited stratigraphic coverage (Sahoo *et al.*, 2012), have limited stratigraphic resolution (Guo *et al.*, 2007; Och *et al.*, 2016), or come from non-fossiliferous sections (Sahoo *et al.*, 2016; Wang *et al.*, 2016). To address this problem, we carried out an integrated investigation of drill core samples (and supplementary outcrop samples) of the Lantian Formation, which is correlated with the Doushantuo Formation but was deposited in deeper waters and contains exceptionally preserved macroscopic fossils, including putative macrometazoans (Yuan *et al.*, 2011; Wan *et al.*, 2016). In this paper, we present carbonate carbon and oxygen ($\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$), organic carbon ($\delta^{13}C_{org}$), carbonate-associated sulfate sulfur ($\delta^{34}S_{CAS}$), pyrite sulfur ($\delta^{34}S_{pyr}$), and sedimentary nitrogen ($\delta^{15}N_{sed}$) isotopic compositions of the Lantian Formation. These data provide a first-order chemostratigraphic framework of this unit and shed new light on the redox history and its possible impact on the ecology and taphonomy of the Lantian biota.

GEOLOGICAL BACKGROUND

The South China Craton consists of the Yangtze and Cathaysia blocks that amalgamated along the Jiangnan orogen at ~830–820 Ma (Zhao & Cawood, 2012). A rift basin developed to the southeastern margin of the Yangtze block ca. 780 Ma. The rift-to-drift transition occurred during the deposition of Cryogenian sediments. The Ediacaran System on the Yangtze block was deposited on a passive continental margin that deepens to the southeast in present geographic orientation (Jiang *et al.*, 2011). Paleogeographic studies show that, during the Ediacaran Period, the Yangtze block consisted of a shelf to the northwest, a deep basin to the southeast, and a narrow slope in between (Fig. 1A) (Jiang *et al.*, 2011).

According to various paleogeographic maps (Zhu *et al.*, 2007; Jiang *et al.*, 2011), the Lantian Formation in southern Anhui Province was deposited in basinal facies, but it is also possible that it was deposited in a deep-water but somewhat restricted basin near the enigmatic Poyang Landmass (Fig. 1A). It overlies the terminal Cryogenian diamictite of the Leigongwu Formation, which is equivalent to the Nantuo Formation in the Yangtze Gorges area, and underlies the Piyuancun Formation ("PYC" in Fig. 1C), which contains macrofossils characteristic of terminal Ediacaran strata (Dong *et al.*, 2012). A drill core of the Lantian Formation was made near the Lantian village of Xiuning County, Anhui Province (Fig. 1B). In this drill core, the Lantian Formation begins with the cap dolostone or Member I, which is a 4-m-thick unit of light gray siliceous dolostone (Fig. 2A–B). This is overlain by Member II, a 90-m-thick unit consisting of a lower sub-unit of 21 m gray calcareous siltstone interbedded with argillaceous limestone and an upper sub-unit of 69 m black shale with rare argillaceous limestone interbeds (Fig. 2C–D). Carbonaceous compression macrofossils and pyritized fossils—including macroalgae (Fig. 2E–F), putative macrometazoans, the enigmatic macrofossil *Orbisiana* (Fig. 2G), and the discoidal fossil *Chuaria* which may be a multicellular eukaryote (Tang *et al.*, 2016)—occur more or less continually in the upper subunit of Member II (Yuan *et al.*, 1999; Yuan *et al.*, 2011; Wan *et al.*, 2014; Wan *et al.*, 2016). Further upsection, Member III is composed of ca. 30 m interbedded gray argillaceous dolostone and black shale, followed by ca. 50 m gray limestone (Fig. 2H–J). Member IV of the uppermost Lantian Formation consists of 10 m black shale (Fig. 2K–L), which is overlain by siliceous rocks of the Piyuancun Formation. Overall, the lithostratigraphic succession intersected by the drill core is similar to those on the outcrop (Wan *et al.*, 2014; Wan *et al.*, 2016), although there are variations in stratigraphic thickness of the four units. The succession is also similar to and can be easily correlated with the four lithostratigraphic members of the Doushantuo Formation in the Yangtze Gorges area (Yuan *et al.*, 2011). Unlike the uncertainty in global correlation of Ediacaran strata (Xiao *et al.*, 2016), the regional correlation between the Liantian and Doushantuo formation is rather straightforward, allowing us to use the radiometric ages from the Yangtze Gorges area to constrain the Lantian Formation between 635 Ma and 551 Ma (Condon *et al.*, 2005).

METHODS

Most samples were collected from the Lantian drill core (29°57.094′ N, 118°2.274′ E; Fig. 1B), but supplementary $\delta^{13}C_{carb}$–$\delta^{18}O_{carb}$ samples of Member III were also collected from outcrops near the drill site (29°55.724′ N, 118°5.424′ E). Sample surfaces were removed to avoid drilling fluid and weathering contaminations. Thin sections were made for petrographic observation. Guided by petrographic observations, powders were taken from polished slabs using a handheld micro-drill, avoiding calcite veins or other visible late diagenetic structures. For comparison and diagenetic evaluation, powders were also micro-drilled from calcite veins. These micro-drilled powders were used for $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ analyses. In addition, bulk samples were powdered using Retsch RS200 for $\delta^{13}C_{org}$, $\delta^{15}N_{sed}$, $\delta^{34}S_{CAS}$, and $\delta^{34}S_{pyr}$ analyses.

$\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ were measured at the Nanjing Institute of Geology and Palaeontology. An aliquot of 80–100 μg powder was allowed to react with orthophosphoric acid for 150–200 seconds at 72°C in a Kiel IV carbonate device connected to a MAT 253 mass spectrometer. CO_2 evolved from this reaction was introduced to the mass spectrometer for $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ determination. The standard GBW-04405, with a $\delta^{13}C$ value of 0.57 ± 0.03‰ and a $\delta^{18}O$ of –8.49 ± 0.13‰ (VPDB), was used for calibration during the analysis. Analytical precision is better than 0.08‰ for $\delta^{13}C_{carb}$ (VPDB) and 0.1‰ for $\delta^{18}O$ (VPDB).

$\delta^{13}C_{org}$, $\delta^{15}N_{sed}$, total organic carbon (TOC), and total nitrogen (TN) were analyzed at Indiana University. An aliquot of 1–3 g bulk-sample powder was decarbonated overnight in a Polypropylene centrifuge tube with 3% (V/V) HCl. The process was repeated at least three times to ensure complete removal of carbonate. The carbonate-free residue were then neutralized using Milli-Q water following a stepwise washing procedure, and then dried at 70°C. Samples were loaded into capsules and flash combusted at 1060°C in a Costech Elemental Analyzer (ECS4010) fitted with zero blank autosampler. The resulting CO_2 and N_2 gas were measured by thermal conductivity detector to obtain TOC (wt%) and TN (wt%), then transferred by continuous flow for isotopic ratio determination on a Delta plus XP Isotope Ratio Mass Spectrometer. C/N atomic ratios were calculated from TOC and TN measurements. $^{13}C/^{12}C$ and $^{15}N/^{14}N$ ratios are reported in the conventional delta notation as per mil deviation from the VPDB and AIR standard, respectively. Calibration of $\delta^{13}C$ and $\delta^{15}N$ values were accomplished using an acetanilide working standard ($\delta^{13}C$ = –27.6‰ and $\delta^{15}N$ = 1.61‰). Sample $\delta^{13}C$and $\delta^{15}N$ values were repeatable to a standard error of 0.1‰ and 0.32‰, respectively.

$\delta^{34}S_{CAS}$ and $\delta^{34}S_{pyr}$ were analyzed at Indiana University. To prepare for $\delta^{34}S_{CAS}$ analysis, ~200 g powder was rinsed with 10% NaCl and repeated three–five times until no sulfate was present in filtered solutions to remove soluble sulfates (e.g., from sulfide oxidation) and water-soluble sulfate minerals (e.g., anhydrite). CAS was obtained by dissolving the NaCl-rinsed powders in 2N HCl for 2 hours at room temperature. Following dissolution, samples were filtered to remove insoluble residues and an excess of 1 M $BaCl_2$ solution was added to precipitate $BaSO_4$ overnight for $\delta^{34}S_{CAS}$ analysis. To prepare for $\delta^{34}S_{pyr}$ analysis, 1–6 g powder was prepared for pyrite extraction using the chromium reduction method (Canfield *et al.*, 1986). Pyrite extraction was performed under N_2 gas by the addition of 6 N HCl and 0.4 M reduced chromium chloride. The reaction was allowed to proceed for at least 3 hours with the sulfide collected as silver sulfide after bubbling through a sodium citrate buffer and into a silver nitrate (0.1 M) trap.

Rinsed, filtered, and dried $BaSO_4$ and Ag_2S precipitates were then combined with an excess of V_2O_5 and analyzed for sulfur isotope composition on a Delta V advantage isotope ratio gas source mass spectrometer fitted with a peripheral Costech Elemental Analyzer (ECS4010) for on-line sample combustion. Sulfur isotope compositions are expressed in standard δ-notation as per mil (‰) deviation from VCDT, with an analytical error of <0.2‰, calculated from replicate analyses of samples and laboratory standards. Analyses were calibrated using four internal standards: silver sulfide (ERE Ag_2S: –4.3‰), chalcopyrite (EMR-CP: +0.9‰), pyrite (PQM2: –14.8‰), and barite (PQB-2: +38.0‰).

RESULTS

Chemostratigraphic data are presented in Table S1 and Fig. 3. The chemostratigraphic profile of $\delta^{13}C_{carb}$ is similar to those reported previously from outcrop samples at Shiyu (Zhao & Zheng, 2009; Wang *et al.*, 2014) and Jinlongshan in the Lantian area (Yuan *et al.*, 2011; see Fig. 1B for locations). Different from the Doushantuo Formation in the Yangtze Gorges area (McFadden *et al.*, 2008), $\delta^{13}C_{carb}$ values of the Lantian Formation range from 0.4‰ to –14.1‰, almost entirely negative values. $\delta^{13}C_{carb}$ of the cap carbonate (Member I) is around –5‰, decreases to lower values in Member II, and reaches to the lowest value of –14‰ in the fossiliferous upper Member II. Subsequently, $\delta^{13}C_{carb}$ rises and stabilizes at –3% in uppermost Member II and lower Member III, and drops again in dolostone and limestone of upper Member III to values around –10‰; the same pattern is observed at the supplementary outcrop section. $\delta^{13}C_{carb}$ values of calcite veins are often indistinguishable but are sometimes different from those of host rocks, with $\delta^{13}C_{host\ rock} - \delta^{13}C_{vein}$ ranging from –5.1‰ to 11.0‰ (average = 0.5‰, s.d. = 3.9‰, n = 23).

$\delta^{18}O_{carb}$ is mostly lower than –10‰, with the exception of lower Member III. Consistent with previous studies (Zhao & Zheng, 2013; Wang *et al.*, 2014), $\delta^{18}O_{carb}$ of calcite veins is significantly lower than that of host rocks (paired t-test, one-tail $p = 2.8 \times 10^{-6}$; $\delta^{18}O_{host\ rock} - \delta^{18}O_{vein}$ = 8.1‰ on average, ranging from –1.3‰ to 19.6‰, s.d. = 6.3‰, n = 22).

$\delta^{13}C_{org}$ shows significant variations between –27‰ and –35‰ in the cap carbonate (Member I) and lower Member II. Further upsection, $\delta^{13}C_{org}$ exhibits more stratigraphic consistency (or less stratigraphic scattering), shifting from a nadir of –35‰ at around 30 m to –30‰ around 70 m and then stabilizing mostly between –29‰ and –30‰ at 70–155 m in upper Member II and much of

Member III. There are more variations in uppermost Member III and Member IV, with values ranging from –35‰ to –26‰. Throughout the Lantian Formation, $\delta^{13}C_{org}$ does not seem to co-vary with $\delta^{13}C_{carb}$ (Fig. 4D; $R^2 = 0.2$), so that $\Delta\delta^{13}C_{carb–org}$ values range from 15.3‰ to 30.0‰.

$\delta^{34}S_{pyr}$ profile of the Lantian Formation is broadly similar to but display a more consistent stratigraphic pattern than a previous reconnaisance study (Shen *et al.*, 2008). It can be divided into six intervals, which are highlighted and marked to the right end of Figure 3: (A) mostly positive values in lithostratigraphic Member I and lower Member II (0 to *ca*. 22 m); (B) mostly negative values in the lower part of upper Member II (*ca*. 22–62 m), reaching a minimum of –31.5‰; (C) a short interval of positive values in middle part of upper Member II (ca. 62–71 m), reaching an apex of 36.5‰ and representing a shift of 68‰ from the minimum in Interval B; (D) mostly negative values in uppermost Member II and lower Member III (*ca*. 71–118 m); (E) mostly positive values in upper Member III (118–174.5 m); and (F) a decreasing trend toward negative values in Member IV (174.5–185 m). Iron speciation data from the same drill core indicate that the redox conditions of the first four intervals oscillate between oxic and euxinic conditions (as marked to the right end of Fig. 3) (Guan, 2014).

$\delta^{34}C_{CAS}$ data are not available for the lower 60 m of the Lantian Formation because of inappropriate lithology: the cap dolostone is highly siliceous and the lower Member II consist of mostly siltstone and black shale. For the rest of the Lantian Formation, $\delta^{34}C_{CAS}$ values were obtained from carbonate interbeds in upper Member II and lower Member III, as well as limestone in upper Member III. These values scatter from 2.2‰ to 32.0‰ (mean = 20.4‰, standard deviation = 6.8‰) without any clearly defined stratigraphic pattern. There is no significant correlation between $\delta^{34}C_{CAS}$ and $\delta^{34}S_{pyr}$ (Fig. 4E; R2 = 0.07).

$\delta^{15}N_{sed}$ values range from 2.9‰ to 7.9‰ in the Lantian Formation (average = 5.3‰, s.d. = 1.1‰, n =152). They vary around 4.5‰ in the first 60 m of the Lantian Formation (or $\delta^{34}S_{pyr}$ intervals A–B), followed by a positive excursion of over 7.0‰ at ca. 70–81 m, a decrease to 2.9‰ at ca. 106 m, another minor positive excursion of 6.2‰ at ca. 119 m, a broad drop to 3.3‰ at ca. 144 m, and a return to values around 6‰ in Member IV.

TOC and TN contents show similar variation patterns in the Lantian Formation (positive correlation, $R^2 = 0.5$; Fig. 4A). Atomic ratios of C/N average at 28.0 (s.d. = 28.6, n = 151), comparable to those reported from the Doushantuo Formation (Kikumoto *et al.*, 2014). As expected, black shales in upper Member II, lower Member III, and Member IV have elevated TOC and TN contents. In contrast, the cap dolostone, siltstone in lower Member II, limestone in upper Member III, as well as carbonate interbeds in upper Member II and lower Member III, have lower values. In general, high TOC and TN contents tend to match with low $\delta^{34}S_{pyr}$ values ($\delta^{34}S_{pyr}$ intervals B, D, F described above) and vice versa.

EVALUATION OF POST-DEPOSITIONAL ALTERATION

Carbon isotopes

$\delta^{13}C_{carb}$ can be variably altered during post-depositional interactions with meteoric, diagenetic, and metamorphic fluids (Knauth & Kennedy, 2009; Derry, 2010; Swart, 2015). Indeed, there are reasons to be cautious about the primary origin of the $\delta^{13}C_{carb}$ values of the carbonate interbeds in the organic-rich shales of upper Member II and lower Member III. First, the majority of Lantian samples have $\delta^{18}O_{carb}$ values less than −10‰ (Fig. 4B). Such low $\delta^{18}O_{carb}$ values are indicative of possible digenetic alteration of $\delta^{13}C_{carb}$ values (Kaufman & Knoll, 1995), even though $\delta^{13}C_{carb}$ are more rock-buffered than $\delta^{18}O_{carb}$ and as a result $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ are not positively correlated (Fig. 4B). Second, the presence of calcite veins in some intervals of the Lantian Formation suggests the possibility of interaction with diagenetic or hydrothermal fluids (Zhao & Zheng, 2013). Our data show that calcite veins in the Lantian Formation have consistently lower $\delta^{18}O_{carb}$ values than the host rock (Fig. 4B–C), suggesting that vein-forming fluids were exogenous. As such, $\delta^{13}C_{carb}$ and particularly $\delta^{18}O_{carb}$ of the host rock can be affected by interaction with vein-forming fluids, although it can be argued that $\delta^{13}C_{carb}$ is less susceptible than $\delta^{18}O_{carb}$ to diagenetic or hydrothermal alteration because of the buffering effect of carbonate rocks (Banner & Hanson, 1990). Third, the abundance of black shales in the Lantian Formation raises the possibility that ^{13}C-depleted authigenic or diagenetic carbonate cements, derived from anaerobic oxidation of organic carbon (e.g., Schrag *et al.*, 2013; Cui *et al.*, 2017), may be a major contributor to the negative $\delta^{13}C_{carb}$ values, particularly in upper Member II and lower Member III where TOC content is high. Thus, only $\delta^{13}C_{carb}$ values from the organic-lean units (i.e., Member I and upper Member III) are more likely to record primary values.

In contrast to $\delta^{13}C_{carb}$ data where organic-rich lithologies are more susceptible to post-depositional alteration, primary $\delta^{13}C_{org}$ values are likely recorded in organic-rich intervals, including black shales in upper Member II and lower Member III. This interpretation is supported by the observation that the $\delta^{13}C_{org}$ profile of the Lantian Formation exhibits the best stratigraphic consistency (or least scattering) in upper Member II and lower Member III where TOC content is generally greater than 0.5 wt%. In contrast, greater stratigraphic variations are observed in Member I, lower Member II, and upper Member III where TOC content is generally <0.5 wt%. When TOC content is low, $\delta^{13}C_{org}$ is more susceptible to terrigenous organic carbon contamination and diagenetic alteration (Jiang *et al.*, 2012; Johnston *et al.*, 2012). Thus, upper Member II and lower Member III likely record primary $\delta^{13}C_{org}$ values, and the strong $\delta^{13}C_{org}$ offset from –35‰ at 30 m to –30‰ at 70 m is a real signal related to primary production. The stratigraphic trends of $\Delta^{13}C_{carb-org}$, however, have limited chemostratigrpahic value because strata recording the most reliable $\delta^{13}C_{carb}$ values do not overlap with those recording the most reliable $\delta^{13}C_{org}$ values. As a consequence, there is no significant correlation between $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ (Fig. 4D) or between $\delta^{13}C_{org}$ and TOC (Fig. 4F).

Sulfur isotopes

Sedimentary pyrite in the Lantian Formation is mostly framboidal and cuboidal in morphology (Guan *et al.*, 2014; Guan *et al.*, 2016), but there is no evidence for significant sulfide mineralization related to vein-forming hydrothermal fluids. Additionally, the Lantian Formation contains relatively abundant pyrite (0–9.1%, average = 1.4%, s.d. = 1.6%; Table S1, Fig 3) and shows a stratigraphically consistent $\delta^{34}S_{pyr}$ profile. There is no significant correlation between pyrite content and $\delta^{34}S_{pyr}$ (Fig. 4G; R^2 = 0.04). Thus, $\delta^{34}S_{pyr}$ of the Lantian Formation likely records the sulfur isotope signatures of sedimentary pyrite, including both syngenetic pyrite formed in the water column and early authigenic pyrite precipitated in sediments.

$\delta^{34}C_{CAS}$ values, on the other hand, can be altered by pyrite oxidation (Marenco *et al.*, 2008) and contamination from atmospheric sulfates (Peng *et al.*, 2014), although meteoric diagenesis seems to have a limited impact (Gill *et al.*, 2008). Although $\delta^{34}C_{CAS}$ and $\delta^{34}C_{pyr}$ do not show significant positive correlation (Fig. 4E), we cannot exclude the possibility of contamination from atmospheric sulfates without ^{17}O data (Peng *et al.*, 2014). Because contamination from both atmospheric sulfates and pyrite oxidation tend to drive $\delta^{34}C_{CAS}$ to lower values, $\delta^{34}C_{CAS}$ values of the Lantian Formation (20.4 ± 6.8‰, n = 40) likely represent minimum estimates of seawater sulfate sulfur isotopic compositions. Indeed, $\delta^{34}C_{CAS}$ values of the Lantian Formation are statistically lower than previously reported $\delta^{34}C_{CAS}$ values from the equivalent Doushantuo Formation in South China (27.9 ± 8.3‰, Mann-Whitney test, $p = 2.0 \times 10^{-7}$) (Shields et al., 2004; McFadden *et al.*, 2008; Li *et al.*, 2010; Xiao *et al.*, 2012).

Nitrogen isotopes

Organic matter experiences chemical alteration due to degradation in the water column and sediments, and such alteration is expected to have an impact on $\delta^{15}N_{sed}$. In modern marine settings, $\delta^{15}N$ alteration in the water column and at the water-sediment interface is largely controlled by oxygen exposure time (Robinson *et al.*, 2012), but this factor likely played a weaker role in Ediacaran oceans that were dominated by anoxia. Alteration within marine sediments is largely driven by bacterial decomposition of organic matter, leading to ^{15}N-enrichment in residual organic matter because isotopically light NH_4^+ is released from organic matter to sediment porewater, and this process results in a few ‰ increase in $\delta^{15}N$ (Lourey, 2003; Thunell *et al.*, 2004; Gaye-Haake *et al.*, 2005; Robinson *et al.*, 2012). However, because pore-water NH_4^+ can be effectively scavenged by clay minerals, bulk sediment nitrogen isotopic compositions ($\delta^{15}N_{sed}$) can partially compensate for within-sediment alteration and provide a reliable proxy for $\delta^{15}N$ values of sedimentary organic matter (Higgins, 2012). In the Lantian Formation, TN is relatively abundant in upper Member II and lower Member III, and $\delta^{15}N_{sed}$ values in these intervals are better buffered against diagenetic alteration and more likely to preserve primary $\delta^{15}N$ values. However, because $\delta^{15}N_{sed}$ and TN do not show any significant correlation (Fig. 4H), it is possible that $\delta^{15}N_{sed}$ values are overall reliable.

SPATIAL PATTERNS

Carbon isotopes

$\delta^{13}C_{carb}$ of the Lantian Formation is almost entirely in the negative territory, which represents a significant departure from $\delta^{13}C_{carb}$ profiles of Ediacaran carbonate successions elsewhere in South China (e.g. Tahata *et al.*, 2013; Wang *et al.*, 2016) and on other continents (Xiao *et al.*, 2016). In particular, $\delta^{13}C_{carb}$ values in upper Member II and lower Member III are more than 10‰ lower than those of presumably equivalent units at Jiulongwan and other sections in shelf and upper-slope facies (Jiang *et al.*, 2011; Zhu *et al.*, 2013; Wang *et al.*, 2016). Indeed, all lower-slope and basinal sections of the Doushantuo Formation analyzed so far are lithologically dominated by organic-rich sediments (e.g., black shales) and characterized by almost exclusively negative $\delta^{13}C_{carb}$ (Jiang *et al.*, 2007; Wang *et al.*, 2014) (Fig. 5). This facies-dependent spatial pattern was interpreted as evidence for a strong depth-dependent $\delta^{13}C_{DIC}$ gradient in excess of >10‰ (Jiang *et al.*, 2007). However, to maintain such a strong gradient for upper Member II and lower Member III (likely representing >10^7 Myr) would require a strong biological pump to transport isotopically light carbon to the deep ocean and an exceptionally strong supply of phosphorus to sustain the biological pump over millions of years. The depth gradient of $\delta^{13}C_{DIC}$ in modern ocean is only ~2‰, and modeling results show that to sustain a 10‰ depth gradient requires a combination of biological pump 2× modern value (as measured by the e-folding depth, i.e., the depth by which 1/e or ~37% of the organic carbon exported from the surface ocean remains) and a phosphorus supply >5× modern value (Meyer *et al.*, 2016). Given that Ediacaran primary producers were likely dominated by non-skeletal and slow sinking organisms relative to their modern counterparts (e.g., diatoms, coccolithophorans, fecal pellets, and other large-celled eukaryotes), the biological pump in Ediacaran oceans was unlikely much stronger than its modern counterpart (Lenton *et al.*, 2014). Also, considering the negative feedbacks that stabilize the long-term phosphorus cycles (Van Cappellen & Ingall, 1996), it is difficult to envision a sustained phosphorus supply >5× modern level. Thus, we regard that a >10‰ depth gradient in $\delta^{13}C_{DIC}$ is unlikely to sustain for millions of years and we must seek alternative explanations to account for the spatial pattern of $\delta^{13}C_{carb}$ observed in South China.

As an alternative explanation, we explore the possibility that post-depositional calcite cements derived from anaerobic oxidation of organic carbon may be a significant contribution to the low $\delta^{13}C_{carb}$ values in the organic-rich and carbonate-poor sediments deposited in basinal facies (Schrag *et al.*, 2013; Cui et al., 2017). This hypothesis implies that carbonate-poor sediments such as shales and siltstones are more sensitive to isotopic mixing of ^{13}C-depleted cements, which is expected to be <50% of rock volume. The plausibility of this hypothesis can be evaluated using a simple mixing model that involves two endmembers (Johnston *et al.*, 2012): (1) a depositional component (i.e., sediments) with carbonate percentage of X_A and carbonate carbon isotopic composition of $\delta^{13}C_A$, and (2) a post-depositional component (i.e., pure calcite cement) with carbonate percentage of $X_B = 1$ and carbonate carbon isotopic composition of $\delta^{13}C_B$. When these two endmembers are mixed at a mass ratio of $f_A : f_B$ (where $f_A + f_B = 1$), the mixture would have a carbonate content (in percentage) of $X_M = f_A * X_A + f_B * X_B$. The carbonate carbon isotopic composition of the mixture ($\delta^{13}C_M$) would be defined by both the mixing ratio and the carbonate content of the two components: $\delta^{13}C_M = (\delta^{13}C_A * f_A * X_A + \delta^{13}C_B * f_B * X_B) / X_M$. We assume $\delta^{13}C_A = 0‰$ (considering the maximum $\delta^{13}C_{carb}$ value of 0.4‰ in upper Member II and lower Member III) and $\delta^{13}C_B = –30‰$ (considering the likely source of ^{13}C-depleted post-depositional calcite from the oxidation of organic carbon in Member II and lower Member III, which has $\delta^{13}C_{org}$ values mostly between –35‰ and –29‰). In this mixing model, calcareous black shales in upper Member II and lower Member III, which typically have TOC content of 1–10 wt%, carbonate content of <50 wt% (and mostly <10 wt%), and $\delta^{13}C_{carb}$ values between 0.4‰ and –13‰, can be mathematically accounted for when $f_B < 5\%$ and $X_A = 1–50\%$ (Fig. 6). In other words, only a minor amount of post-depositional cements (i.e., <5% of rock mass) is needed to explain the carbonate content and $\delta^{13}C_{carb}$ data of calcareous black shales in upper Member II and lower Member III. For the carbonate interbeds in upper Member II and lower Member III, a greater amount of cements (but still <20% of rock mass) is implied for the majority of samples (Fig. 6). These predictions make petrographic sense because cements are expected to be <50% of rock mass or volume. This hypothesis should be investigated in the future using petrographically guided microanalysis or SIMS analysis of $\delta^{13}C_{carb}$ (Drake *et al.*, 2015; Cui *et al.*, 2016b).

Mathematically, negative $\delta^{13}C_{carb}$ values in Member I (cap carbonate) and upper Member III can also be explained by the mixing model described above. However, $\delta^{13}C_{carb}$ values of Member I and upper Member III—unlike those of upper Member II and lower Member III—seem to show regional and global consistency; in other words, a >10‰ depth gradient is not observed in these units. For example, basal Ediacaran cap carbonates in South China and other continents have consistent $\delta^{13}C_{carb}$ values around –5‰ and this is also true for Member I in the Lantian Formation (Fig. 3). Similarly, the negative $\delta^{13}C_{carb}$ excursion in upper Member III has been known from many sections across the Yangtze block and has been interpreted as equivalent to the Shuram negative $\delta^{13}C_{carb}$ excursion (Wang *et al.*, 2014), and potential examples of the Shuram excursion have been reported from numerous Ediacaran successions around the world (Grotzinger *et al.*, 2011). These excursions all have nadir $\delta^{13}C_{carb}$ values around –10‰. Although it is possible that authigenic/diagenetic calcite may have played a role in the origin of the Shuram excursion (Macouin *et al.*, 2012; Cui *et al.*, 2016b, 2017; Furuyama *et al.*, 2016), the possibility that a globally synchronous diagenetic event

(Grotzinger *et al.*, 2011; Schrag *et al.*, 2013; Cui *et al.*, 2017) needs to be investigated in order to determine whether the Shuram excursion can be used as a chemostratigraphic marker (Xiao *et al.*, 2016).

The overall $\delta^{13}C_{org}$ profile of the Lantian Formation is broadly similar to that of the Doushantuo Formation at the inner shelf Northern Xiaofenghe section (Xiao *et al.*, 2012), intra-shelf Jiulongwan section (McFadden *et al.*, 2008), outer shelf Zhongling section (Li *et al.*, 2010; Cui *et al.*, 2015), and the upper slope Siduping section in South China (Wang *et al.*, 2016). Although the negative excursion of $\delta^{13}C_{org}$ in Member III around 160–170 m may be compared with those in Member IV of the Doushantuo Formation in the Yangtze Gorges area (McFadden *et al.*, 2008) or the Shuram Formation in Oman (Fike *et al.*, 2006; Lee *et al.*, 2015), the best preserved $\delta^{13}C_{org}$ values of the Lantian Formation are in the organic-rich sediments in upper Member II and lower Member III. There is a gradual negative shift from variable values in lower Member II to –35‰ in middle Member II, followed by a gradual positive shift to about –32‰ that terminates upward with an offset to about –30‰. Above this offset, there is a plateau of values between –28‰ and –30‰. A possible equivalent of this excursion to –35‰ is seen in the lower Doushantuo Formation at the upper slope Siduping section and perhaps also at the shelf margin Zhongling section, but not present in the inner shelf and intra-shelf sections at Northern Xiaofenghe and Jiulongwan (Fig. 7). Consistent $\delta^{13}C_{org}$ values between –28‰ and –30‰ are characteristic of lower-middle Doushantuo Formation at Northern Xiaofenghe and Jiulongwan sections, but slightly more variable and greater values characterize presumably equivalent strata at the Zhongling section (between –25‰ and –27‰) and Siduping section (between –26‰ and –30‰). Other Doushantuo sections that have been analyzed for $\delta^{13}C_{org}$ values tend to have relatively low stratigraphic resolution (Guo *et al.*, 2006; Guo *et al.*, 2007; Ader *et al.*, 2009) to warrant a reliable comparison with the Lantian Formation for analysis of spatial variations.

Sulfur isotopes

Spatial variations in $\delta^{34}S_{pyr}$ and $\delta^{34}S_{CAS}$ of Ediacaran successions in South China are shown in Fig. 8. Although $\delta^{34}S_{pyr}$ shows significant stratigraphic variations, there is a clear spatial trend that can be discerned from the data. The most important spatial pattern is an overall decrease in $\delta^{34}S_{pyr}$ values from shallow to deep facies, particularly in the lower Doushantuo and Lantian formations (Fig.

8). For example, $\delta^{34}S_{pyr}$ values of the Doushantuo Formation at the inner shelf section (Northern Xiaofenghe) and outer shelf sections (Zhongling) are almost entirely positive, mostly 20–40‰ in the lower Doushantuo but slide to 0–20‰ in the upper Doushantuo (Zhongling). In deeper intra-shelf facies (Jiulongwan), $\delta^{34}S_{pyr}$ values are mostly 0–40‰ in the lower Doushantuo Formation but decrease to mostly negative values (between 0‰ and –20‰) in the upper Doushantuo Formation. In still deeper sections in slope and basinal facies (e.g., Wuhe and Lantian sections), $\delta^{34}S_{pyr}$ values become increasingly negative.

There are not sufficient $\delta^{34}S_{CAS}$ data from slope and basinal sections of the Doushantuo Formation, because these sections tend to be dominated by black shales. Additionally, the issue of contamination might compromise the reliability of $\delta^{34}S_{CAS}$ values as a proxy for seawater sulfate (Marenco *et al.*, 2008; Peng *et al.*, 2014). Thus, the big picture spatial pattern of $\delta^{34}S_{CAS}$ is not discernable from currently available data. Nonetheless, $\delta^{34}S_{CAS}$ values of the Lantian Formation are broadly comparable to, although slightly lower than, those of the Doushantuo Formation at Northern Xiaofenghe, Jiulongwan, and Zhongling (Fig. 8), all with a large range between 0‰ and 40‰.

Following previous authors (Li *et al.*, 2010; Xiao *et al.*, 2012), the observed spatial pattern of $\delta^{34}S_{pyr}$ is best interpreted as evidence for oceanic redox stratification with euxinic deep waters and non-euxinic (and probably oxic) shallow waters. When pyrite precipitation occurs in euxinic deep waters, lower $\delta^{34}S_{pyr}$ values are expected because sulfate reducing microbes have full access to marine sulfate, which was already a few mM in Ediacaran oceans (Hurtgen *et al.*, 2002), and maximum sulfur isotope fractionation is possible. In contrast, when pyrite precipitation occurs within sediments in shallow-water facies, sulfate availability is limited by diffusion and high $\delta^{34}S_{pyr}$ values are expected. Following this simplistic model, it is inferred that Doushantuo sediments in inner and outer shelf facies were mostly deposited in non-euxinic (and probably oxic) conditions, whereas euxinic conditions frequented deeper waters in intra shelf, slope, and basinal facies (Jiang *et al.*, 2011; Muscente *et al.*, 2015). This simple model offers a first-order framework for us to interpret the spatial (and also temporal; see below) variation in $\delta^{34}S_{pyr}$ and redox conditions.

Of course, the reality is likely more complex because many other factors also control the apparent fractionation of sulfur isotopes between sulfate and pyrite. These factors include, among others, primary production and availability of organic matter (Leavitt *et al.*, 2013), marine sulfate concentration (Gomes and Hurtgen, 2015; Bradley *et al.*, 2016), cell-specific sulfate reduction rate

(Sim *et al.*, 2012; Bradley *et al.*, 2016), and oxidative recycling of sulfide (Fike *et al.*, 2015). Importantly, Gomes and Hurtgen (2015) have shown that, in modern euxinic environments, the relationship between $\delta^{34}S_{sulfide}$ and sulfate concentration is not a uniform one: the reservoir effect dominates when sulfate concentration is low (<5 mM), whereas sulfate reduction rates and location of pyrite formation in a euxinic water column become more important when sulfate concentration is high. Given that the Ediacaran oceans may have had a sulfate concentration of about 5 mM (Hurtgen *et al.*, 2002) and may have had a euxinic wedge (Li et al., 2010), these factors are relevant and may have been important in affecting the $\delta^{34}S_{pyr}$ record of the Lantian Formation. Although we regard pyrite formation in a euxinic water column vs. in sediments as a primary control because the congruence between $\delta^{34}S_{pyr}$ values and redox conditions, we do recognize that this is a simple model and other factors must be considered to explain the second-order variation in the $\delta^{34}S_{pyr}$ record of the Lantian Formation.

Nitrogen isotopes

Biologically available nitrogen in modern oceans has a rather short residence time relative to oceanic circulation (Gruber, 2008). Thus, $\delta^{15}N$ of nitrate, organic carbon, and seafloor bulk sediments in modern global oceans show significant spatial variations over 10‰ (Gaye-Haake *et al.*, 2005; Tesdal *et al.*, 2013). Given these variations, it is remarkable that $\delta^{15}N_{sed}$ values of the Lantian Formation are broadly consistent with those of the Doushantuo Formation in South China (Ader *et al.*, 2014; Kikumoto *et al.*, 2014) (Fig. 9). Although the Lantian Formation (average $\delta^{15}N_{sed}$ = 5.3‰) has statistically higher values than the Doushantuo Formation at Yangjiaping (average $\delta^{15}N_{sed}$ = 4.4‰) and Yangtze Gorges area of South China (average $\delta^{15}N_{sed}$ = 4.8‰), the differences in average $\delta^{15}N_{sed}$ values are within 1‰. In terms of $\delta^{15}N_{sed}$ values, the Doushantuo and Lantian formations are broadly similar to some Ediacaran successions (e.g., $^{15}N_{sed}$ = 5.1 ± 0.5‰ in Svalbard, and 5.1 ± 2.0‰ in Brazil) but sharply different from others (e.g., $\delta^{15}N_{sed}$ = 2.9 ± 0.6‰ in northwestern Canada) (Ader *et al.*, 2014; Spangenberg *et al.*, 2014). They are also distinct from Cambrian successions in South China ($\delta^{15}N_{sed}$ = 1.1 ± 2.1‰) (Cremonese *et al.*, 2013; Wang *et al.*, 2013; Cremonese *et al.*, 2014; Wang *et al.*, 2015) and Cryogenian successions of South China ($\delta^{15}N_{sed}$ = 6.6 ± 0.7‰) (Wei *et al.*, 2016). This overall similarity in $\delta^{15}N_{sed}$ values of the Doushantuo and Lantian formations in South China, if confirmed with additional data in the future, suggests that basinal rather than local factors controlled the nitrogen cycle.

TEMPORAL REDOX OSCILLATION

On the basis of integrated petrographic and geochemical data, we propose that the Lantian Formation records oscillating redox conditions between likely euxinic and oxic environments. Evidence in support of this hypothesis comes from iron speciation data from the same drill core, which show $\delta^{34}S_{pyr}$ intervals A and C in Fig. 3 were likely deposited in oxic conditions, whereas intervals B and D in euxinic conditions (Guan, 2014). This interpretation is supplemented by the congruence of lithological, TOC, TN, and $\delta^{34}S_{pyr}$ data. Euxinic intervals with organic-rich shales (B, D, and possibly F in Fig. 3) are expected to have high TOC and TN contents, but low and negative $\delta^{34}S_{pyr}$ values. In contrast, oxic intervals with organic-lean carbonates and siltstones (A, C, and possibly E in Fig. 3) are expected to have low TOC and TN contents, but high and positive $\delta^{34}S_{pyr}$ values. Other independent evidence supportive of this model comes from framboidal pyrite data. Pyrite framboids in black shales of intervals B and D are largely 4–8 μm in size (Guan *et al.*, 2014), consistent with syngenetic pyrite precipitation in euxinic water column (Wilkin *et al.*, 1996), particularly considering that the Lantian data probably represent a maximum estimate of framboid size because of possible pyrite overgrowth (Wacey *et al.*, 2015). Thus, it is possible that oscillating euxinic and oxic conditions are characteristic of the Lantian Formation (Fig. 3), although the redox condition of interval E–F remains to be confirmed with Fe speciation data. In this oscillating redox model, pyrite with high and positive $\delta^{34}S_{pyr}$ values was largely precipitated in sediments during oxic intervals A, C, and E, whereas that with low and negative $\delta^{34}S_{pyr}$ values was mostly formed in euxinic water column during intervals B, D, and F. The oscillating euxinic and oxic conditions in the Lantian Formation are somewhat similar to those recorded in the time-equivalent Doushantuo Formation in deep-water basinal facies (Sahoo *et al.*, 2016). However, $\delta^{34}S_{pyr}$ seems to respond differently in the Doushantuo Formation, where more negative $\delta^{34}S_{pyr}$ values occur in oxic events and more positive $\delta^{34}S_{pyr}$ values in euxinic intervals—a pattern that was interpreted in terms of sulfate reservoir effect (Sahoo *et al.*, 2016), which would be most prominent when seawater sulfate concentration was less than 5 mM (Gomes and Hurtgen, 2015). Nonetheless, the oscillating redox conditions documented in the Lantian and Doushantuo formations are intriguing, although an exact correlation of redox conditions between these two formations is premature.

Redox oscillation in the Lantian Formation can be evaluated using other geochemical data. Due to post-depositional alteration and inappropriate lithologies, reliable $\delta^{13}C_{carb}$ data only available in the cap carbonate and upper Member III, and $\delta^{34}S_{CAS}$ data in carbonate interbeds in Member II–III. Thus, they do not have sufficient stratigraphic coverage to allow in-depth evaluation of the redox oscillation model. $\delta^{15}N_{sed}$ and $\delta^{13}C_{org}$ data, on the other hand, do have a relatively continuous record in the Lantian Formation, particularly upper Member II and lower Member III (or roughly $\delta^{34}S_{pyr}$ intervals B–D). It is interesting to note that, in intervals B–D, statistically lower $\delta^{15}N_{sed}$ values tend to occur in the euxinic intervals B and D, whereas higher $\delta^{15}N_{sed}$ values in the oxic interval C. A similar correlation may be present in $\delta^{13}C_{org}$ variations, although this is more dubious because $\delta^{13}C_{org}$ in interval C is not particularly higher than in interval D. Below we explore the possibility that variations in $\delta^{15}N_{sed}$ and $\delta^{13}C_{org}$ in intervals B–D may reflect redox fluctuation.

The most important control on $\delta^{15}N_{sed}$ at a global scale is denitrification, which anaerobically reduces NO_3^- to N_2 and involves a significant isotopic fractionation (ε = 20–30‰) (Altabet, 2006). Globally, denitrification needs to be complemented by nitrification, which aerobically oxidizes NH_4^+ to NO_3^-. Coupled nitrification-denitrification has been established probably since the Archean (Godfrey & Falkowski, 2009), and we assume that it was also part of the nitrogen cycle in Ediacaran as in modern oceans. In modern oceans, $\delta^{15}N$ of NO_3^- is elevated in the oxygen minimum zone (OMZ) where denitrification preferentially removes isotopically light NO_3^- to form N_2. As denitrification proceeds, $\delta^{15}N$ of NO_3^- (and hence $\delta^{15}N$ of microorganisms that assimilate NO_3^-) becomes progressively higher as $[NO_3^-]$ decreases (Altabet & Francois, 1994), and reaches the highest (e.g., 18.7‰) in OMZ under the influence of strong oceanic upwelling and with dissolved oxygen content below 1–2 μM (i.e., sub-oxic to hypoxic conditions) (Voss *et al.*, 2001). However, in euxinic basins where denitrification is quantitative and NO_3^- is exhausted at the chemocline, denitrification does not have an isotopic influence on $\delta^{15}N$ of organic matter because NH_4^+ and nitrogen fixation, rather than NO_3^-, become the main sources of organic nitrogen. For example, $\delta^{15}N$ of seafloor sediments in the euxinic Cariaco Basin is only 3.5‰ (Thunell *et al.*, 2004). Thus, $\delta^{15}N_{sed}$ is expected to have a complex relationship with redox conditions (Ader *et al.*, 2014). For euxinic condition to develop, NO_3^- must have been exhausted to allow microbial sulfate reduction to prevail (Canfield, 2006). Thus, $\delta^{15}N_{sed}$ of enxinic environments is expected to be low because of NO_3^- exhaustion and hence minimal influenced of denitrification on $\delta^{15}N_{sed}$ (Thunell *et al.*, 2004). $\delta^{15}N_{sed}$

of suboxic environments, on the other hand, would have highly elevated values (>10‰) because of ^{15}N-enrichment related to incomplete denitrification (Quan *et al.*, 2008). However, $\delta^{15}N_{sed}$ of oxic environments in OMZ margins and open oceans would approach the modal $\delta^{15}N_{sed}$ value of 6‰ of modern marine sediments (Tesdal *et al.*, 2013).

Viewed in this light, it is possible to explain the $\delta^{15}N_{sed}$ profile of the Lantian Formation under the redox oscillation model. The most reliable $\delta^{15}N_{sed}$ values are preserved in upper Member II and lower Member III where TN content is relatively high. Low $\delta^{15}N_{sed}$ values (3–5‰) occur in $\delta^{34}S_{pyr}$ interval B and the upper part of interval D, which are independently interpreted as representing euxinic environments based on congruent TOC, TN, $\delta^{34}S_{pyr}$, and Fe speciation data. Such values are similar to $\delta^{15}N$ of modern euxinic basins such as the Cariaco Basin (Thunell *et al.*, 2004). The highest $\delta^{15}N_{sed}$ values occur in $\delta^{34}S_{pyr}$ interval C and the lower part of interval D. Interval C is independently interpreted as representing oxic environments, and its $\delta^{15}N_{sed}$ values are similar to those of modern open marine sediments (Tesdal *et al.*, 2013). The high $\delta^{15}N_{sed}$ values in the lower part of interval D are enigmatic, but they could represent partial denitrification in a suboxic environment at the transition from an oxic condition (interval C) to a euxinic condition (interval D). On the other hand, the Lantian Formation lacks strongly positive $\delta^{15}N_{sed}$ values characteristic of suboxic upwelling zones in modern oceans (Voss *et al.*, 2001; Gaye-Haake *et al.*, 2005; Tesdal *et al.*, 2013), possibly due to generally low [NO_3^-] levels in Ediacaran oceans so that nearly quantitative denitrification may have occurred even in suboxic conditions.

$\delta^{15}N_{sed}$ values of intervals A, E, and F do not seem to be compatible with the simple redox model presented above. However, intervals A and E have relatively low TN contents and interval F has rather low stratigraphic sampling intensity. Thus, it is possible that $\delta^{15}N_{sed}$ values of intervals A, E, and F are not reliable, that the inference of local redox conditions based on Fe speciation and $\delta^{34}S_{pyr}$ data is not reliable, or that the model about the relationship between $\delta^{15}N_{sed}$ and redox conditions is too simplistic. These possibilities need to be investigated in the future.

The relationship between $\delta^{13}C_{org}$ and redox fluctuation in intervals B–D is less clear. However, it is possible that the lower $\delta^{13}C_{org}$ values in euxinic intervals, particularly interval B, were related to recycling of organic carbon in the system. In other words, photosynthetic organic carbon can be anaerobically oxidized (e.g., by sulfate reduction bacteria) and reutilized by chemoautotrophs. Consequently, both photosynthetic and chemosynthetic processes may have made contributions to

TOC. This scenario can be tested in the future by comparing $\delta^{13}C_{org}$ of TOC and photosynthetic macroalgae in the Lantian Formation.

What may have caused the oscillating redox conditions in the Lantian Formation? It is tempting to speculate that global redox changes may have placed an overarching control on the local redox conditions as recorded in the Lantian Formation. Indeed, Sahoo *et al.* (2016) proposed on the basis of redox-sensitive trace metals that global Ediacaran deep oceans were predominantly anoxic/euxinic, with oceanic oxygenation events at 635–630 Ma (e.g., basal Doushantuo Formation), ~580 Ma (e.g., lower Member III of the Doushantuo Formation), and ~560 Ma (Member IV of Doushantuo Formation). The absolute age of these oxygenation events can be a matter of debate, and a correlation of Sahoo et al.'s (2016) global oxygenation events with the oxic intervals in the Lantian Formation is also an open question. For example, although oxic interval A in the Lantian Formation is likely correlated with the basal Ediacaran oxygenation event in Member I and lower Member II of the Doushantuo Formation, it is uncertain whether the oxic interval C in upper Member II of the Lantian Formation is correlated with the oxygenation event recorded in lower Member III in the Doushantuo Formation (Sahoo *et al.*, 2016).

Regional and local factors may have also played a role in modulating the redox condition in the Lantian Formation. Certainly, regional sea level changes, continental weathering, hydrographic factors, and oceanic upwelling may have influenced local productivity, which is a primary factor controlling redox conditions, chemocline depth, sulfate reduction, and denitrification (and hence $\delta^{13}C_{org}$, $\delta^{34}S_{pyr}$, and $\delta^{15}N_{sed}$). Similarly, varying sedimentation rates may be a factor controlling TOC and TN contents. The observation that oxic conditions largely occur in organic-lean carbonates and siltstones whereas euxinic conditions in organic-rich shales suggests that local and regional factors must have modulated the redox conditions in the Lantian Formation. Because of these local controls, it is not always straightforward to correlate redox history among different basins—for example between South China (Sahoo *et al.*, 2016) and northwestern Canada (Sperling *et al.*, 2015a), or even within the same basin—for example between the Member IV of the Doushantuo Formation at Miaohe and Jiulongwan (Li *et al.*, 2015) or between the Doushantuo Formation and the Lantian Formation (Sahoo *et al.*, 2016; this paper).

FOSSIL PRESERVATION

Macroscopic eukaryote fossils occur mostly in the black shales of upper Member II, or in euxinic intervals B and D (Fig. 3). These fossils include pyrite molds of *Chuaria* (Yuan *et al.*, 2001; Guan *et al.*, 2016) as well as a variety of carbonaceous compressions including the modular fossil *Orbisiana* (Wan *et al.*, 2014), macroalgal fossils such as *Flabellophyton* and *Doushantuophyton* (Yuan *et al.*, 1999), and putative animal fossils such as *Lantianella* and *Xiuningella* (Yuan *et al.*, 2011; Wan *et al.*, 2016). So far, no fossils have been discovered from other units of the Lantian Formation, and attempts to recover microfossils using acid digestion techniques have been futile. Thus, fossils appear to be restricted largely, if not exclusively, to euxinic black shales, although not all euxinic black shales contain fossils.

What controls the stratigraphic distribution of fossils in the Lantian Formation? In principle, evolutionary, ecological, and taphonomic factors may provide the answer. However, evolutionary factors (i.e., fossil taxa of the Lantian Formation are evolutionarily restricted to intervals B and D) can be ruled out. For example, several Lantian genera (i.e., *Doushantuophyton* and *Enteromorphites*) range into the younger Miaohe biota in Member IV of the Doushantuo Formation (Xiao *et al.*, 2002) but so far have not been discovered in members III-IV of the Lantian Formation. This absence in members III-IV of the Lantian Formation, if confirmed in future investigation, must be related to ecological and taphonomic factors.

Both ecological and taphonomic factors probably played an important role in controlling the stratigraphic distribution of Lantian fossils, although it is difficult to tease apart these factors because both are related to depositional environments. The lack of fossils in carbonates and siltstones of the Lantian Formation could be simply due to unfavorable environments or water depths for the colonization or preservation of Lantian fossils. However, the lack of macrofossils in black shales of Member IV is enigmatic. It is possible that black shales in Member II and Member IV represent different redox conditions despite their similar appearance. For example, Li *et al.* (2015) and Och *et al.* (2016) have shown, on the basis of Fe speciation and trace metal data, that Member IV black shales of the Doushantuo Formation at sections separated by a few kilometers are characterized by differing redox conditions (euxinic vs. ferruginous). Whether black shales in Member II and Member IV of the Lantian Formation represent different redox conditions need to be tested with Fe speciation and redox-sensitive trace metal data.

If the Lantian fossils were preserved in-situ (Yuan *et al.*, 2011; Wan *et al.*, 2014), then the presence of macroalgae and putative animals in upper Member II of the Lantian Formation implies oxic conditions at least at ecological time scales. To reconcile these oxygen-breathing organisms and their preservation in euxinic black shales in intervals B and D, Yuan *et al.* (2011) speculated that these black shales were deposited in largely euxinic environments punctuated by brief oxic episodes, when macroalgae and putative animals thrived, only to be subsequently killed and preserved by frequent switch-back to euxinic conditions. This hypothesis is extremely difficult to test, given that each chemostratigraphic sample likely integrates geochemical signals over several centimeters of strata, thus lacking the stratigraphic resolution to detect redox changes at ecological time scales (Sperling *et al.*, 2015a). To make the task more challenging, it is possible that a euxinic water column would likely overprint the geochemistry of sediments deposited in a preceding oxic episode. These processes result in geochemical time averaging, analogous to taphonomic time averaging in fossil assemblages (Walker & Bambach, 1971). Currently available chemostratigraphic techniques cannot overcome the geochemical time averaging problems in order to resolve such rapidly changing redox conditions.

Regardless, the restriction of Lantian fossils in intervals B and D indicates that euxinia may have played an important role in fossil preservation. The taphonomic modes of Lantian fossils include both carbonaceous compression and pyritization (Yuan *et al.*, 2001; Guan *et al.*, 2016; Wan *et al.*, 2016). Carbonaceous compression fossils in the Lantian Formation sometimes show rusty colors (Wan *et al.*, 2016), likely resulting from pyrite weathering and suggesting a continuum between the two taphonomic modes in the Lantian Formation as is also observed in other Ediacaran Lagerstätten (Cai *et al.*, 2012; Hall et al., 2013; Schiffbauer *et al.*, 2014). We hypothesize that anoxic and particularly euxinic waters allowed the precipitation of thin veneer of pyrite on degrading carcasses at or just beneath the water-sediment interface, thus providing a mechanism to replicate soft-bodied organisms in the Lantian Formation. Early authigenic pyrite precipitation may turn out to be a common taphonomic process responsible for various types of fossil preservation in the Ediacaran Period (Liu, 2016).

CONCLUSIONS

High-resolution chemostratigraphic analysis of integrated geochemical data—including $\delta^{13}C_{carb}$, $\delta^{13}C_{org}$, $\delta^{34}S_{pyr}$, $\delta^{34}S_{CAS}$, and $\delta^{15}N_{sed}$—from the Lantian Formation deposited in deep-water environments in South China allows a better resolution of the spatial patterns and temporal variations of Ediacaran oceanic redox conditions. Comparison with other geochemical datasets from South China, particularly those from the time-equivalent Doushantuo Formation, makes it possible to determine spatial patterns. A large spatial gradient of $\delta^{13}C_{carb}$, with a magnitude of >10‰, is confirmed to be present in the lower Doushantuo/Lantian Formation. Unlike previous authors, we argue that this gradient unlikely represents a >10‰ depth gradient of $\delta^{13}C_{DIC}$; instead, we propose that this gradient is largely due to the influence of ^{13}C-depleted authigenic/diagenetic calcite cements in organic-rich and carbonate-poor sediments deposited in deep water facies. A pronounced negative $\delta^{13}C_{carb}$ excursion in the upper Doushantuo/Lantian Formation, however, is spatially consistent and does not display a >10‰ depth gradient. This negative excursion has been previously described as EN3 (Ediacaran Negative excursion 3) in South China and regarded as equivalent to the Shuram excursion in Oman. The consistency of this excursion at a basinal scale may have chemostratigraphic significance, but its origin remains to be resolved. Unlike $\delta^{13}C_{carb}$, the $\delta^{13}C_{org}$ record shows better basinal scale consistency, particularly if only the more reliable $\delta^{13}C_{org}$ values of organic-rich sediments are considered.

The $\delta^{34}S_{pyr}$ record seems to show a spatial pattern, with more negative $\delta^{34}S_{pyr}$ values in deeper water facies, but the $\delta^{34}S_{CAS}$ record (and hence the $\Delta\delta^{34}S$ record) is insufficient in both stratigraphic and geographic coverage to allow the determination of spatial patterns. The spatial pattern of $\delta^{34}S_{pyr}$ is interpreted as evidence for a chemically stratified basin, with predominantly euxinic deep waters where pyrite precipitated in the water column has lower $\delta^{34}S_{pyr}$ values, and predominantly non-euxinic shallow waters where pyrite formed within sediments has higher $\delta^{34}S_{pyr}$ values. This first-order redox structure is superimposed by local and temporal variations.

Relative to the spatial variations of $\delta^{15}N_{sed}$ in modern oceans, the $\delta^{15}N_{sed}$ record of the Lantian and Doushantuo formations is remarkably consistent. Overall, the $\delta^{15}N_{sed}$ data seem to suggest nitrate limitation so that denitrification is often quantitative. The Lantian and Doushantuo formations do not preserve highly positive $\delta^{15}N_{sed}$ values characteristic of partial denitrification in suboxic environments of modern oxygen minimum zone, although the $\delta^{15}N_{sed}$ database is currently rather limited in spatial and stratigraphic coverage.

Integrated TOC, TN, $\delta^{34}S_{pyr}$, and Fe speciation data suggest that oscillation between oxic and euxinic conditions is characteristic of the Lantian Formation, at least in Member I to lower Member

III. As in the spatial model, pyrite in euxinic intervals was precipitated in the water column and has lower $\delta^{34}S_{pyr}$ values, whereas pyrite in oxic intervals was precipitated within sediments and has higher $\delta^{34}S_{pyr}$ values due to diffusion-related sulfate limitation. The redox oscillation model can explain some aspects of the $\delta^{15}N_{sed}$ profile of the Lantian Formation.

Macrofossils in the Lantian Formation include morphologically complex macroalgae and putative animals. Stratigraphic distribution of these fossils is largely restricted to euxinic black shales, although not all black shales preserve macrofossils. In reconciling the preservation of oxygen-breathing organisms in euxinic black shales, it is hypothesized that the euxinic environments were punctuated by brief oxygenation events when macroalgae and putative animals thrived, but were subsequently decimated and preserved when local environment switched to euxinic conditions. If true, then ecological and taphonomic factors were the main controls on fossil distribution in the Lantian Formation (and perhaps in other Ediacaran successions). This hypothesis, however, is difficult to evaluate, owing to the short ecological time scales (compounded by geochemical time-averaging) over which redox variation is presumed to occur.

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REFERENCES

Ader M, Macouin M, Trindade RIF, Hadrien M-H, Yang Z, Sun Z, Besse J (2009) A multilayered water column in the Ediacaran Yangtze platform? Insights from carbonate and organic matter paired $\delta^{13}C$. *Earth and Planetary Science Letters,* **288**, 213–227.

Ader M, Sansjofre P, Halverson GP, Busigny V, Trindade RIF, Kunzmann M, Nogueira ACR (2014) Ocean redox structure across the Late Neoproterozoic Oxygenation Event: A nitrogen isotope perspective. *Earth and Planetary Science Letters,* **396**, 1–13.

Altabet MA (2006) Isotopic tracers of the marine nitrogen cycle: present and past. In: *Marine Organic Matter: Biomarkers, Isotopes and DNA* (ed Volkman JK). Springer-Verlag, Berlin, pp. 251–293.

Altabet MA, Francois R (1994) Sedimentary nitrogen isotopic ratio as a recorder for surface ocean nitrate utilization. *Global Biogeochemical Cycles,* **8**, 103–116.

Banner JL, Hanson GN (1990) Calculation of simultaneous isotopic and trace element variations during water-rock interaction with applications to carbonate diagenesis. *Geochimica et Cosmochimica Acta,* **54**, 3123–3137.

Bradley AS, Leavitt WD, Schmidt M, Knoll AH, Girguis PR, Johnston DT (2016) Patterns of sulfur isotope fractionation during microbial sulfate reduction. *Geobiology,* **14**, 91–101.

Cai Y, Schiffbauer JD, Hua H, Xiao S (2012) Preservational modes in the Ediacaran Gaojiashan Lagerstätte: Pyritization, aluminosilicification, and carbonaceous compression. *Palaeogeography Palaeoclimatology Palaeoecology,* **326–328**, 109–117.

Canfield DE (2006) Models of oxic respiration, denitrification and sulfate reduction in zones of coastal upwelling. *Geochimica et Cosmochimica Acta,* **70**, 5753–5765.

Canfield DE, Raiswell R, Westrich JT, Reaves CM, Berner RA (1986) The use of chromium reduction in the analysis of reduced inorganic sulfur in sediments and shale. *Chemical Geology,* **54**, 149–155.

Cole DB, Reinhard CT, Wang X, Gueguen B, Halverson GP, Gibson T, Hodgskiss MSW, Mckenzie NR, Lyons TW, Planavsky NJ (2016) A shale-hosted Cr isotope record of low atmospheric oxygen during the Proterozoic. *Geology,* **7**, 555–558.

Condon D, Zhu M, Bowring S, Wang W, Yang A, Jin Y (2005) U-Pb ages from the Neoproterozoic Doushantuo Formation, China. *Science,* **308**, 95–98.

Cremonese L, Shields-Zhou G, Struck U, Ling H, Och L, Chen X, Li D (2013) Marine biogeochemical cycling during the early Cambrian constrained by a nitrogen and organic carbon isotope study of the Xiaotan section, South China. *Precambrian Research,* **225**, 148–165.

Cremonese L, Shields-Zhou GA, Struck U, Ling H-F, Och LM (2014) Nitrogen and organic carbon isotope stratigraphy of the Yangtze Platform during the Ediacaran-Cambrian transition in South China. *Palaeogeography Palaeoclimatology Palaeoecology,* **398**, 165–186.

Cui H, Kaufman AJ, Xiao S, Zhou C (2017) Methane-derived authigenic carbonates from the uppermost Doushantuo Formation in South China: Was the Ediacaran Shuram Excursion a globally synchronized early diagenetic event? *Chemical Geology*, 10.1016/j.chemgeo.2016.1012.1010.

Cui H, Kaufman AJ, Xiao S, Zhu M, Zhou C, Liu X-M (2015) Redox architecture of an Ediacaran ocean margin: integrated chemostratigraphic ($\delta^{13}C-\delta^{34}S-{}^{87}Sr/{}^{86}Sr-Ce/Ce^{*}$) correlation of the Doushantuo Formation. *Chemical Geology,* **405**, 48–62.

Cui H, Kaufman AJ, Xiao S, Peek S, Cao H, Min X, Cai Y, Siegel Z, Liu XM, Peng Y, Schiffbauer JD, Martin AJ (2016a) Environmental context for the terminal Ediacaran biomineralization of animals. Geobiology, **14**, 344–363.

Cui H, Xiao S, Zhou C, Peng Y, Kaufman AJ, Plummer RE (2016b) Phosphogenesis associated with the Shuram Excursion: Petrographic and geochemical observations from the Ediacaran Doushantuo Formation of South China. *Sedimentary Geology,* **341**, 134–146.

Dong L, Song W, Xiao S, Yuan X, Chen Z, Zhou C (2012) Micro- and macrofossils from the Piyuancun Formation and their implications for the Ediacaran-Cambrian boundary in southern Anhui. *Journal of Stratigraphy,* **36**, 600–610.

Drake H, Åström ME, Heim C, Broman C, Åström J, Whitehouse M, Ivarsson M, Siljeström S, Sjövall P (2015) Extreme ^{13}C depletion of carbonates formed during oxidation of biogenic methane in fractured granite. *Nature Communications,* **6**, 7020.

Fike DA, Bradley AS, Rose CV (2015) Rethinking the ancient sulfur cycle. *Annual Review of Earth and Planetary Sciences,* **43**, 593-622.

Fike DA, Grotzinger JP, Pratt LM, Summons RE (2006) Oxidation of the Ediacaran ocean. *Nature,* **444**, 744–747.

Furuyama S, Kano A, Kunimitsu Y, Ishikawa T, Wang W (2016) Diagenetic overprint to a negative carbon isotope anomaly associatedwith the Gaskiers glaciation of the Ediacaran Doushantuo Formationin South China. *Precambrian Research,* **276**, 110–122.

Gaye-Haake B, Lahajnar N, Emeis KC, Unger D, Rixen T, Suthhof A, Ramaswamy V, Schulz H, Paropkari AL, Guptha MVS, Ittekkot V (2005) Stable nitrogen isotopic ratios of sinking particles and sediments from the northern Indian Ocean. *Marine Chemistry,* **96**, 243–255.

Gill BC, Lyons TW, Frank TD (2008) Behavior of carbonate-associated sulfate during meteoric diagenesis and implications for the sulfur isotope paleoproxy. *Geochimica et Cosmochimica Acta*, **72**, 4699–4711.

Godfrey LV, Falkowski PG (2009) The cycling and redox state of nitrogen in the Archaean ocean. *Nature Geoscience*, **2**, 725–729.

Gomes ML, Hurtgen MT (2015) Sulfur isotope fractionation in modern euxinic systems: Implications for paleoenvironmental reconstructions of paired sulfate-sulfide isotope records. *Geochimica et Cosmochimica Acta*, **157**, 39-55.

Grotzinger JP, Fike DA, Fischer WW (2011) Enigmatic origin of the largest-known carbon isotope excursion in Earth's history. *Nature Geoscience*, **4**, 285–292.

Gruber N (2008) The marine nitrogen cycle: overview and challenges. In: *Nitrogen in the Marine Environment (Second Edition)* (eds Capone DG, Bronk DA, Mulholland MR, Carpenter EJ). Academic Press, Cambridge, UK, pp. 1–50.

Guan C, Wang W, Zhou C, Muscente AD, Wan B, Chen X, Yuan X, Chen Z, Ouyang Q (2016) Controls on fossil pyritization: Redox conditions, sedimentary organic matter content, and Chuaria preservation in the Ediacaran Lantian Biota. *Palaeogeography Palaeoclimatology Palaeoecology*, 10.1016/j.palaeo.2016.1005.1013.

Guan C, Zhou C, Wang W, Wan B, Yuan X, Chen Z (2014) Fluctuation of shelf basin redox conditions in the early Ediacaran: Evidence from Lantian Formation black shales in South China. *Precambrian Research*, **245**, 1–12.

Guan CG (2014) Geochemical studies of Ediacaran Lantian Formation in southern Anhui Province and their paleoenvironmental implications (in Chinese with English abstract). *University of Chinese Academy of Sciences (PhD Dissertation)*, 1–138.

Guo Q, Liu C, Strauss H, Goldberg T, Zhu M, Pi D, Wang J (2006) Organic carbon isotope geochemistry of the Neoproterozoic Doushantuo Formation, South China. *Acta Geologica Sinica*, **80**, 670–683.

Guo Q, Strauss H, Liu C, Goldberg T, Zhu M, Pi D, Heubeck C, Vernhet E, Yang X, Fu P (2007) Carbon isotopic evolution of the terminal Neoproterozoic and early Cambrian: Evidence from the Yangtze Platform, South China. *Palaeogeography Palaeoclimatology Palaeoecology*, **254**, 140–157.

Hall M, Kaufman AJ, Vickers-Rich P, Ivantsov A, Trusler P, Linnemann U, Hofmann M, Elliott D, Cui H, Fedonkin M, Hoffmann K-H, Wilson SA, Schneider G, Smith J (2013) Stratigraphy, palaeontology and geochemistry of the late Neoproterozoic Aar Member, southwest Namibia: Reflecting environmental controls on Ediacara fossil preservation during the terminal Proterozoic in African Gondwana. *Precambrian Research*, **238**, 214–232.

Higgins MD, Robinson RS, Husson JM, Carter SJ, Pearson A (2012) Dominant eukaryotic export production during ocean anoxic events reflects the importance of recycled NH_4^+. *Proceedings of the National Academy of Sciences of the United State of America*, **109**, 2269–2274.

Hurtgen MT, Arthur MA, Suits NS, Kaufman AJ (2002) The sulfur isotopic composition of Neoproterozoic seawater sulfate: implications for a snowball Earth? *Earth and Planetary Science Letters*, **203**, 413–429.

Jiang G, Kaufman AJ, Christie-Blick N, Zhang S, Wu H (2007) Carbon isotope variability across the Ediacaran Yangtze platform in South China: Implications for a large surface-to-deep ocean $\delta^{13}C$ gradient. *Earth and Planetary Science Letters*, **261**, 303–320.

Jiang G, Shi X, Zhang S, Wang Y, Xiao S (2011) Stratigraphy and paleogeography of the Ediacaran Doushantuo Formation (ca. 635-551 Ma) in South China. *Gondwana Research*, **19**, 831–849.

Jiang G, Wang X, Shi X, Xiao S (2012) The origin of decoupled carbonate and organic carbon isotope signatures in the early Cambrian (ca. 542–520 Ma) Yangtze platform. *Earth and Planetary Science Letters*, **317–318**, 96–110.

Johnston DT, Macdonald FA, Gill BC, Hoffman PF, Schrag DP (2012) Uncovering the Neoproterozoic carbon cycle. *Nature*, **483**, 320–323.

Johnston DT, Poulton SW, Tosca NJ, O'brien T, Halverson GP, Schrag DP, Macdonald FA (2013) Searching for an oxygenation event in the fossiliferous Ediacaran of northwestern Canada. *Chemical Geology*, **362**, 273–286.

Kaufman AJ, Knoll AH (1995) Neoproterozoic variations in the C-isotope composition of sea water: Stratigraphic and biogeochemical implications. *Precambrian Research*, **73**, 27–49.

Kikumoto R, Tahata M, Nishizawa M, Sawaki Y, Maruyama S, Shu D, Han J, Komiya T, Takai K, Ueno Y (2014) Nitrogen isotope chemostratigraphy of the Ediacaran and early Cambrian platform sequence at Three Gorges, South China. *Gondwana Research*, **25**, 1057–1069.

Knauth LP, Kennedy MJ (2009) The late Precambrian greening of the Earth. *Nature*, **460**, 728–732.

Leavitt WD, Halevy I, Bradley AS, Johnston DT (2013) Influence of sulfate reduction rates on the Phanerozoic sulfur isotope record. *Proceedings of the National Academy of Sciences of the United State of America*, **110**, 11244-11249.

Lenton TM, Boyle RA, Poulton SW, Shields-Zhou GA, Butterfield NJ (2014) Co-evolution of eukaryotes and ocean oxygenation in the Neoproterozoic era. *Nature Geoscience,* **7**, 257–265.

Lee C, Love GD, Fischer WW, Grotzinger JP, Halverson GP (2015) Marine organic matter cycling during the Ediacaran Shuram excursion. *Geology,* **43**, 1103–1106.

Li C, Love GD, Lyons TW, Fike DA, Sessions AL, Chu X (2010) A stratified redox model for the Ediacaran ocean. *Science*, **328**, 80–83.

Li C, Planavsky NJ, Shi W, Zhang Z, Zhou C, Cheng M, Tarhan LG, Luo G, Xie S (2015) Ediacaran marine redox heterogeneity and early animal ecosystems. *Scientific Reports,* **5**, 17097.

Liu AG (2016) Framboidal pyrite shroud confirms the ‘death mask' model for moldic preservation of Ediacaran soft-bodied organisms. *Palaios*, **31**, 259–274.

Liu P, Xiao S, Yin C, Chen S, Zhou C, Li M (2014) Ediacaran acanthomorphic acritarchs and other microfossils from chert nodules of the upper Doushantuo Formation in the Yangtze Gorges area, South China. *Journal of Paleontology,* **72 (supplement to No 1)**, 1–139.

Lourey MJ, Trull TW,Sigman DM. (2003) Sensitivity of ^{15}N of nitrate, surface suspended and deep sinking particulate nitrogen to seasonal nitrate depletion in the Southern Ocean. *Global Biogeochemical Cycles*, **17**.

Lyons TW, Anbar AD, Severmann S, Scott C, Gill BC (2009) Tracking euxinia in the ancient ocean: A multiproxy perspective and Proterozoic case study. *Annual Review of Earth and Planetary Sciences*, **37**, 507–534.

Macouin M, Ader M, Moreau MG, Charlespoitou, Yang Z, Sun Z (2012) Deciphering the impact of diagenesis overprint on negative δ^{13}C excursions using rock magnetism: Case study of Ediacaran carbonates, Yangjiaping section, South China. *Earth and Planetary Science Letters,* **351–352**, 281–294.

Marenco PJ, Corsetti FA, Hammond DE, Kaufman AJ, Bottjer DJ (2008) Oxidation of pyrite during extraction of carbonate associated sulfate. *Chemical Geology,* **247**, 124–132.

Mcfadden KA, Huang J, Chu X, Jiang G, Kaufman AJ, Zhou C, Yuan X, Xiao S (2008) Pulsed oxygenation and biological evolution in the Ediacaran Doushantuo Formation. *Proceedings* *of the National Academy of Sciences, USA*, **105**, 3197–3202.

Meyer KM, Ridgwell A, Payne JL (2016) The influence of the biological pump on ocean chemistry: implications for long-term trends in marine redox chemistry, the global carbon cycle, and marine animal ecosystems. *Geobiology*, **14**, 207–219.

Muscente AD, Hawkins AD, Xiao S (2015) Fossil preservation through phosphatization and silicification in the Ediacaran Doushantuo Formation (South China): A comparative synthesis. *Palaeogeography Palaeoclimatology Palaeoecology*, **434**, 46–62.

Och LM, Cremonese L, Shields-Zhou GA, Poulton SW, Struck U, Ling H, Li D, Chen X, Manning C, Thirlwall M, Strauss H, Zhu M (2016) Palaeoceanographic controls on spatial redox distribution over the Yangtze Platform during the Ediacaran–Cambrian transition. *Sedimentology*, **63**, 378–410.

Och LM, Shields-Zhou GA (2012) The Neoproterozoic oxygenation event: Environmental perturbations and biogeochemical cycling. *Earth Science Reviews*, **110**, 26–57.

Peng Y, Bao H, Pratt LM, Kaufman AJ, Jiang G, Boyd D, Wang Q, Zhou C, Yuan X, Xiao S, Loyd S (2014) Widespread contamination of carbonate-associated sulfate by present-day secondary atmospheric sulfate: Evidence from triple oxygen isotopes. *Geology*, **42**, 815–818.

Planavsky NJ, Reinhard CT, Wang X, Thomson D, Mcgoldrick P, Rainbird RH, Johnson T, Fischer WW, Lyons TW (2014) Low mid-Proterozoic atmospheric oxygen levels and the delayed rise of animals. *Science*, **346**, 635–638.

Pogge Von Strandmann PaE, Stueken EE, Elliott T, Poulton SW, Dehler CM, Canfield DE, Catling DC (2015) Selenium isotope evidence for progressive oxidation of the Neoproterozoic biosphere. *Nature Communications*, **6**, 10157.

Quan TM, Schootbrugge BVD, Field MP, Rosenthal Y, Falkowski PG (2008) Nitrogen isotope and trace metal analyses from the Mingolsheim core (Germany): Evidence for redox variations across the Triassic-Jurassic boundary. *Global Biogeochemical Cycles*, **22**, GB2014.

Robinson RS, Kienast M, Albuquerque AL, Altabet M, Contreras S, Holz RDP, Dubois N, Francois R, Galbraith E, Hsu TC, Ivanochko T, Jaccard S, Kao S-J, Kiefer T, Kienast S, Lehmann M, Martinez P, Mccarthy M, Möbius J, Pedersen T, Quan TM, Ryabenko E, Schmittner A, Schneider R, Schneider-Mor A, Shigemitsu M, Sinclair D, Somes C, Studer A, Thunell R, Yang JY (2012) A review of nitrogen isotopic alteration in marine sediments.

Paleoceanography, **27**, PA4203.

Sahoo SK, Planavsky NJ, Jiang G, Kendall B, Owens JD, Wang X, Shi X, Anbar AD, Lyons TW (2016) Oceanic oxygenation events in the anoxic Ediacaran ocean. *Geobiology*, **14**, 457–468.

Sahoo SK, Planavsky NJ, Kendall B, Wang X, Shi X, Scott C, Anbar AD, Lyons TW, Jiang G (2012) Ocean oxygenation in the wake of the Marinoan glaciation. *Nature*, **489**, 546–549.

Sawaki Y, Ohno T, Tahata M, Komiya T, Hirata T, Maruyama S, Windley BF, Han J, Shu D, Li Y (2010) The Ediacaran radiogenic Sr isotope excursion in the Doushantuo Formation in the Three Gorges area, South China. *Precambrian Research*, **176**, 46–64.

Schiffbauer JD, Xiao S, Cai Y, Wallace AF, Hua H, Hunter J, Xu H, Peng Y, Kaufman AJ (2014) A unifying model for Neoproterozoic–Palaeozoic exceptional fossil preservation through pyritization and carbonaceous compression. *Nature Communications*, **5**, 5754 (doi: 5710.1038/ncomms6754).

Schrag DP, Higgins JA, Macdonald FA, Johnston DT (2013) Authigenic carbonate and the history of the global carbon cycle. *Science*, **339**, 540–543.

Shen Y, Zhang T, Hoffman PF (2008) On the co-evolution of Ediacaran oceans and animals. *Proceedings of the National Academy of Sciences, USA*, **105**, 7376–7381.

Shields G, Kimura H, Yang JD, Gammon P (2004) Sulphur isotopic evolution of Neoproterozoic-Cambrian seawater: new francolite-bound sulphate $\delta^{34}S$ data and a critical appraisal of the existing record. *Chemical Geology*, **204**, 163–182.

Sim MS, Ono S, Bosak T (2012) Effects of iron and nitrogen limitation on sulfur isotope fractionation during microbial sulfate reduction. *Applied and Environmental Microbiology*, **78**, 8368–8376.

Spangenberg JE, Bagnoud-Velásquez M, Boggiani PC, Gaucher C (2014) Redox variations and bioproductivity in the Ediacaran: Evidence from inorganic and organic geochemistry of the Corumbá Group, Brazil. *Gondwana Research*, **26**, 1186–1207.

Sperling EA, Carbone C, Strauss JV, Johnston DT, Narbonne GM, Macdonald FA (2015a) Oxygen, facies, and secular controls on the appearance of Cryogenian and Ediacaran body and trace fossils in the Mackenzie Mountains of northwestern Canada. *Geological Society of America Bulletin*, **128**, 558–575.

Sperling EA, Wolock CJ, Morgan AS, Gill BC, Kunzmann M, Halverson GP, Macdonald FA, Knoll

AH, Johnston DT (2015b) Statistical analysis of iron geochemical data suggests limited late Proterozoic oxygenation. *Nature*, **523**, 451–454.

Swart PK (2015) The geochemistry of carbonate diagenesis: The past, present and future. *Sedimentology*, **62**, 1233–1304.

Tahata M, Ueno Y, Ishikawa T, Sawaki Y, Murakami K, Han J, Shu D, Li Y, Guo J, Yoshida N, Komiya T (2013) Carbon and oxygen isotope chemostratigraphies of the Yangtze platform, South China: Decoding temperature and environmental changes through the Ediacaran. *Gondwana Research*, **23**, 333–353.

Tang Q, Pang K, Yuan X, Xiao S (2017) Electron microscopy reveals evidence for simple multicellularity in the Proterozoic fossil *Chuaria*. *Geology*, 10.1130/G38680.38681.

Tesdal JE, Galbraith ED, Kienast M (2013) Nitrogen isotopes in bulk marine sediment: linking seafloor observations with subseafloor records. *Biogeosciences*, **10**, 101–118.

Thunell RC, Sigman DM, Muller-Karger F, Astor Y, Varela R (2004) Nitrogen isotope dynamics of the Cariaco Basin, Venezuela. *Global Biogeochemical Cycles*, **18**, GB3001.

Van Cappellen P, Ingall ED (1996) Redox stabilization of the atmosphere and oceans by phosphorus-limited marine productivity. *Science*, **271**, 493–496.

Voss M, Dippner JW, Montoya JP (2001) Nitrogen isotope patterns in the oxygen-deficient waters of the eastern tropical North Pacific ocean. *Deep Sea Research Part I: Oceanographic Research Papers*, **48**, 1905–1921.

Wacey D, Kilburn MR, Saunders M, Cliff JB, Kong C, Liu AG, Matthews JJ, Brasier MD (2015) Uncovering framboidal pyrite biogenicity using nano-scale CN_{org} mapping. *Geology*, **43**, 27–30.

Walker KR, Bambach RK (1971) The significance of fossil assemblages from fine-grained sediments: time-averaged communities. *Geological Society of America Abstracts with Programs*, **3**, 783–784.

Wan B, Xiao S, Yuan X, Chen Z, Pang K, Tang Q, Guan C, Maisano JA (2014) *Orbisiana* linearis from the early Ediacaran Lantian Formation of SouthChina and its taphonomic and ecological implications. *Precambrian Research*, **255**, 266–275.

Wan B, Yuan X, Chen Z, Guan C, Pang K, Tang Q, Xiao S (2016) Systematic description of putative animal fossils from the early Ediacaran Lantian Formation of South China. *Palaeontology*, **59**, 515–532.

Wang D, Struck U, Ling H-F, Guo Q-J, Shields-Zhou GA, Zhu M-Y, Yao S-P (2015) Marine redox variations and nitrogen cycle of the early Cambriansouthern margin of the Yangtze Platform, South China: Evidence from nitrogen and organic carbon isotopes. *Precambrian Research*, **267**, 209–226.

Wang W, Zhou C, Guan C, Yuan X, Chen Z, Wan B (2014) An integrated carbon, oxygen, and strontium isotopic studies of the Lantian Formation in South China with implications for the Shuram anomaly. *Chemical Geology*, **373**, 10–26.

Wang W, Zhou CM, Yuan XL, Chen Z, Guan CG (2011) Variations of the carbonate carbon isotope in Ediacaran Doushantuo ocean of South China (in Chinese with English abstract). *Journal of Stratigraphy*, **35**, 349–360.

Wang X, Jiang G, Shi X, Xiao S (2016) Paired carbonate and organic carbon isotope variations of the Ediacaran Doushantuo Formation from an upper slope section at Siduping, South China. *Precambrian Research*, **273**, 53–66.

Wang X, Shi X, Tang D, Zhang W (2013) Nitrogen isotope evidence for redox variations at the Ediacaran-Cambrian transition in South China. *Journal of Geology*, **121**, 489–502.

Wei W, Wang D, Li D, Ling H, Chen X, Wei G, Zhang F, Zhu X, Yan B (2016) The marine redox change and nitrogen cycle in the early Cryogenian interglacial time: evidence from nitrogen isotopes and Mo contents of the basal Datangpo Formation, northeastern Guizhou, South China. *Journal of Earth Science*, **27**, 233–241.

Wilkin RT, Barnes HL, Brantley SL (1996) The size distribution of framboidal pyrite in modern sediments: An indicator of redox conditions. *Geochimica et Cosmochimica Acta*, **60**, 3897–3912.

Wood RA, Poulton SW, Prave AR, Hoffmann KH, Clarkson MO, Guilbaud R, Lyne JW, Tostevin R, Bowyer F, Penny AM, Curtis A, Kasemann SA (2015) Dynamic redox conditions control late Ediacaran metazoan ecosystems in the Nama Group, Namibia. *Precambrian Research*, **261**, 252–271.

Xiao S, Mcfadden KA, Peek S, Kaufman AJ, Zhou C, Jiang G, Hu J (2012) Integrated chemostratigraphy of the Doushantuo Formation at the northern Xiaofenghe section (Yangtze Gorges, South China) and its implication for Ediacaran stratigraphic correlation and ocean redox models. *Precambrian Research*, **192-195**, 125–141.

Xiao S, Muscente AD, Chen L, Zhou C, Schiffbauer JD, Wood AD, Polys NF, Yuan X (2014) The Weng'an biota and the Ediacaran radiation of multicellular eukaryotes. *National Science Review*, **1**, 498–520.

Xiao S, Narbonne GM, Zhou C, Laflamme M, Grazhdankin DV, Moczydłowska-Vidal M, Cui H (2016) Toward an Ediacaran time scale: problems, protocols, and prospects. *Episodes*, **39**, 540–555.

Xiao S, Yuan X, Steiner M, Knoll AH (2002) Macroscopic carbonaceous compressions in a terminal Proterozoic shale: A systematic reassessment of the Miaohe biota, South China. *Journal of Paleontology*, **76**, 347–376.

Yuan X, Chen Z, Xiao S, Zhou C, Hua H (2011) An early Ediacaran assemblage of macroscopic and morphologically differentiated eukaryotes. *Nature*, **470**, 390–393.

Yuan X, Li J, Cao R (1999) A diverse metaphyte assemblage from the Neoproterozoic black shales of South China. *Lethaia*, **32**, 143–155.

Yuan X, Xiao S, Li J, Yin L, Cao R (2001) Pyritized chuarids with excystment structures from the late Neoproterozoic Lantian Formation in Anhui, South China. *Precambrian Research*, **107**, 253–263.

Zhang S, Wang X, Wang H, Bjerrum CJ, Hammarlund EU, Costa MM, Connelly JN, Zhang B, Su J, Canfield DE (2015) Sufficient oxygen for animal respiration 1,400 million years ago. *Proceedings of the National Academy of Sciences of the United State of America*, **113**, 1731–1736.

Zhao G, Cawood PA (2012) Precambrian geology of China. *Precambrian Research*, **222–223**, 13–54.

Zhao YY, Zheng YF (2009) Stable isotope evidence for involvement of deglacial meltwater in Ediacaran carbonates in South China. *Chemical Geology*, **271**, 86–100.

Zhao YY, Zheng YF (2013) Geochemical constraints on the origin of post-depositional fluids in sedimentary carbonates of the Ediacaran system in South China. *Precambrian Research*, **224**, 34–363.

Zhu M, Lu M, Zhang J, Zhao F, Li G, Yang A, Zhao X, Zhao M (2013) Carbon isotope chemostratigraphy and sedimentary facies evolution of the Ediacaran Doushantuo Formation in western Hubei, South China. *Precambrian Research*, **225**, 7–28.

Zhu M, Zhang J, Yang A (2007) Integrated Ediacaran (Sinian) chronostratigraphy of South China. *Palaeogeography Palaeoclimatology Palaeoecology*, **254**, 7–61.

Figure captions

Fig. 1. Geological maps and stratigraphic column. (A) Generalized paleogeographic map of the Yangtze Block during the early–middle Ediacaran Period, showing the approximate location of shelf, slope, and basinal facies. Numbered triangles indicate location of Ediacaran sections mentioned in the text, with the Lantian area highlighted in red. (B) Geological map of the Lantian area, showing the location of the Lantian drill core analyzed in this study. Supplementary outcrop samples were collected near the drilling site. The Shiyu and Jinlongshan sections examined in previous studies are also marked in the map. (C) Simplified lithostratigraphic column of the Lantian drill core. PYC: Piyuancun Formation.

Fig. 2. Petrographic photographs of the Lantian drill core (A–D, H–L) and macrofossils from outcrops (E–G). (A) Drill core segment of cap dolostone (Member I). (B) Cross-polarized light photomicrograph of cap dolostone thin section, showing dolomicrite (arrow) and siliceous cement with first-order gray birefringence color. (C) Drill core segment of Member II black shale. (D) Plane-polarized light photomicrograph of Member II black shale, showing silica cements and a compressed *Chuaria* fossil (outlined by red ellipse) surrounded with diagenetic clays. (E–F) The macroalgal fossil *Flabellophyton* from Member II. Specimens illustrated in (E) are preserved with greater three dimensionality, due to pyritization, whereas specimen in (F) is preserved as two-dimensional carbonaceous compression. (G) The problematic fossil *Orbisiana* from Member II. (H) Plane polarized light photomicrograph of argillaceous limestone in upper Member II, showing a mixture of calcite, dolomite, clay, and siliceous cement. (I) Drill core segment of upper Member III limestone. (J) Cross-polarized light photomicrograph of upper Member III limestone, showing calcitic microspar, silty detritus, and siliceous cement. (K) Drill core segment of Member IV black shale. (L) Plane-polarized light photomicrograph of Member IV black shale, showing organic-rich matrix and siliceous cement. Arrows in core segments (A, C, I, K) mark stratigraphic up direction.

Fig. 3 (in landscape format). Chemostratigraphic profiles of the Lantian Formation. The $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ profiles include data from both drill core and outcrop samples, with blue, red, and black symbols representing measurements of drill core, supplementary outcrop, and calcite vein samples, respectively. Calcite veins were also sampled from the Lantian drill core. Data in all other profiles are based on data from drill core samples. Solid symbols denote likely primary signatures based our evaluation of diagenetic alteration. Redox interpretation for intervals A–D is based on Fe speciation from the same drill core (Guan, 2014).

Fig. 4. Geochemical cross-plots. Note that Fig. 4B plots $\delta^{13}C_{carb}$ vs. $\delta^{18}O_{carb}$, whereas Fig. 4C plots $\delta^{13}C_{carb\text{-}vein}$ vs. $\delta^{13}C_{carb\text{-}host}$ and $\delta^{18}O_{carb\text{-}vein}$ vs. $\delta^{18}O_{carb\text{-}host}$.

Fig. 5 (in landscape format). Spatial variation in $\delta^{13}C_{carb}$ of the Doushantuo/Lantian formations. Dashed lines represent top and bottom boundaries of the Doushantuo/Lantian formations. Gray shade highlights the negative $\delta^{13}C_{carb}$ excursion EN3 that is regionally consistent and is widely regarded as equivalent to the Shuram excursion. Approximate paleobathymetric location of sections is marked on a simplified cross-section of the Yangtze Platform (Jiang *et al.*, 2011). See Fig. 1A for exact geographic location of sections. Data source: Jiulongwan (Jiang *et al.*, 2007), Siduping (Wang *et al.*, 2016), Taoying (Wang *et al.*, 2011), Yuanling (Jiang *et al.*, 2007), Yanwutan (Guo *et al.*, 2007), Lantian drill core and supplementary outcrop (this study), Shiyu (Wang *et al.*, 2014), and Jinlongshan (Yuan *et al.*, 2011). DY: Dengying Formation; PYC: Piyuancun Formation. EN: Ediacaran negative $\delta^{13}C_{carb}$ excursion; EP: Ediacaran positive $\delta^{13}C_{carb}$ excursion.

Fig. 6. Mixing model showing that variation in $\delta^{13}C_{carb}$ could be due to variable mixing of (1) a depositional component with carbonate mass fraction X_A = 1–100% and carbonate carbon isotopic composition $\delta^{13}C_A$ = 0‰, and (2) a post-depositional component of pure calcite cement with carbonate mass fraction $X_B = 1$ and $\delta^{13}C_B$ = –30‰. When these two endmembers are mixed at a mass ratio of $f_A : f_B$ (where $f_A + f_B = 1$), the mixture has a carbonate content of $X_M = f_A * X_A + f_B * X_B$, and $\delta^{13}C_M = (\delta^{13}C_A * f_A * X_A + \delta^{13}C_B * f_B * X_B) / X_M$. $\delta^{13}C_{carb}$ values of carbonate-poor samples from upper Member II and lower Member III of the Lantian Formation (solid blue dots) can be accounted for by a relatively small amount of post-depositional cement (f_B < 5%), whereas those

of carbonate interbeds (open circles) require greater amount of cements (but still <20% for most samples).

Fig. 7 (in landscape format). Spatial variation in $\delta^{13}C_{org}$ of the Doushantuo/Lantian formations. Dashed lines represent top and bottom boundaries of the Doushantuo/Lantian formations. Approximate paleobathymetric location of sections is marked on a simplified cross-section of the Yangtze Platform (Jiang *et al.*, 2011). See Fig. 1A for exact geographic location of sections. Data source: Northern Xiaofenghe (NXF) (Xiao *et al.*, 2012), Jiulongwan (JLW) (McFadden *et al.*, 2008), Zhongling (ZL) (Cui *et al.*, 2015), Siduping (SDP) (Wang *et al.*, 2016), and Lantian (LT) (this study). DY: Dengying Formation; PYC: Piyuancun Formation.

Fig. 8 (in landscape format). Spatial variation in $\delta^{34}S_{pyr}$ and $\delta^{34}S_{CAS}$ of the Doushantuo/Lantian formations. Dashed lines represent top and bottom boundaries of the Doushantuo/Lantian formations. Approximate paleobathymetric location of sections is marked on a simplified cross-section of the Yangtze Platform (Jiang *et al.*, 2011). See Fig. 1A for exact geographic location of sections. Data source: Northern Xiaofenghe (NXF) (Xiao *et al.*, 2012), Jiulongwan (JLW) (McFadden *et al.*, 2008), Zhongling (ZL) (Cui *et al.*, 2015), Wuhe (WH) (Sahoo *et al.*, 2016), and Lantian (LT) (this study). DY: Dengying Formation; PYC: Piyuancun Formation.

Fig. 9. Spatial variation in $\delta^{15}N_{sed}$ of the Doushantuo/Lantian formations. Dashed lines represent top and bottom boundaries of the Doushantuo/Lantian formations. Approximate paleobathymetric location of sections is marked on a simplified cross-section of the Yangtze Platform (Jiang *et al.*, 2011). See Fig. 1A for exact geographic location of sections. Data source: Yangtze Gorges (Kikumoto *et al.*, 2014), Yangjiaping (YJP) (Ader *et al.*, 2014), and Lantian (LT) (this study). DY: Dengying Formation; PYC: Piyuancun Formation.