Climate warming, euxinia and carbon isotope perturbations during the Carnian (Triassic) Crisis in South China

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Abstract

The Carnian Humid Episode (CHE), also known as the Carnian Pluvial Event, and associated biotic changes are major enigmas of the Mesozoic record in western Tethys. We show that the CHE also occurred in eastern Tethys (South China), suggestive of a much more widespread and probably global climate perturbation. Oxygen isotope records from conodont apatite indicate a double-pulse warming event. The CHE coincided precisely with an initial warming of 4°C. This was followed by a transient cooling period and then a prolonged ~7°C warming in the later Carnian (Tuvalian 2). Carbon isotope perturbations associated with the CHE of western Tethys occurred contemporaneously in South China, and mark the start of a prolonged period of carbon cycle instability that persisted until the late Carnian. The dry-wet transition during the CHE coincides precisely with the negative carbon isotope excursion and the temperature rise, pointing to an intensification of hydrologic cycle activities due to climatic warming. While carbonate platform shutdown in western Tethys is associated with an influx of siliciclastic sediment, the eastern Tethyan carbonate platforms are overlain by deep-water anoxic facies. The transition from oxygenated to euxinic facies was via a condensed, manganiferous carbonate (MnO content up to 15.1 wt%), that records an intense Mn shuttle operating in the basin. Significant siliciclastic influx in South China only occurred after the CHE climatic changes and was probably due to foreland basin development at the onset of the Indosinian Orogeny. The mid-Carnian biotic crisis thus coincided with several phenomena associated with major extinction events: a carbonate production crisis, climate warming, $\delta^{13}C$ oscillations, marine anoxia, biotic turnover and flood basalt eruptions (of the Wrangellia Province).

Key words: Carnian Humid Episode, carbon isotopes, marine anoxia, climate warming, Wrangellia flood basalt, large igneous provinces

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1. Introduction

Environmental and biotic changes during the Carnian Stage (~237-227 Ma) of the Triassic are amongst the most enigmatic of the geological record (Simms and Ruffell, 1989; Ogg, 2015). Prominent amongst these was the Carnian Humid Episode (CHE, after Ruffel et al., 2015, but also known as Carnian Pluvial Event or Carnian Pluvial Phase), a short-lived phase (<1 Myr, Zhang et al., 2015) of increased rainfall. The CHE has been attributed to an intensification of a mega-monsoonal climate (Parrish, 1993; Kozur and Bachmann, 2010), although climate change driven by voluminous volcanism and greenhouse gas exhalation has also been suggested (Dal Corso et al., 2012). A contemporaneous major increase in siliciclastic influx to marine areas has been implicated in the shutdown of carbonate platforms and the demise of reef ecosystems in western Tethys and the peri-Gondwana margin (e.g., Hornung et al., 2007a). The CHE also coincided with extinctions amongst crinoids and scallops, while nekton such as ammonoids and conodonts also suffered losses (Simms and Ruffell, 1989; Rigo et al., 2007), suggesting a broad-reaching marine crisis. Terrestrial floras record a shift towards hygrophytic forms in the late Julian substage (early Carnian, ~237-233 Ma) before returning to xerophytic associations by the early Tuvalian substage (late Carnian, ~233-227 Ma) (Roghi et al., 2010, Mueller et al., 2016); a change that may have been responsible for elevated extinction rates reported amongst herbivorous tetrapods (Benton, 1986).

Despite the magnitude of the changes, the majority of evidence for the Carnian Crisis has come from western Europe (Fig. 1) and major questions remain as to the global significance and cause(s) of this event. We evaluate the global extent of the CHE and associated environmental changes by investigating palaeotemperature, $\delta^{13}C$, sedimentary facies and redox changes in South China, a region far removed (~13,000 km distant) from the better-studied western Tethyan

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locations. Our results indicate that the environmental changes in the Carnian were truly global and associated with a carbonate production crisis, double-pulse climate warming, intense oceanic anoxia and large $\delta^{13}C$ oscillations associated with flood basalt eruptions.

2. Geological setting

The studied section at Long Chang (25°26.565'N, 105°28.791'E) outcrops in a road cut ~2 km south of the town of Long Chang (Figs. 2 and 3). In comparison to the well-known section of Enos et al. (1998) in the nearby area (Fig. 2C), our section is more condensed and has been chosen for its excellent outcrop situation and ease of access. The Long Chang area was located in a transitional zone between the Yangtze Platform and Nanpanjiang Basin in eastern Tethys at ~15°N during the Carnian (Enos et al., 1998; Golonka, 2007). The deep-waters of the Nanpanjiang Basin opened to the east into the Panthalassa Ocean (Fig. 1). This platform-and-basin topography later evolved into a foreland basin in the latest Carnian due to collision between the North and South China blocks that saw carbonate platforms swamped by siliciclastic input (Enos et al., 1998). Therefore, while in western Tethys, the mid-Carnian is marked by a transition from carbonate to siliciclastic deposition, and a re-establishment of carbonate production in the aftermath of the CHE (e.g., Hornung et al., 2007b), in South China renewed carbonate deposition was short-lived because foreland basin siliciclastics were soon established. Our interpretation of climatic and environmental history of this region during the Carnian must therefore be viewed in the context of an on-going, and as yet poorly understood major tectonic change.

3. Methods

The 50 m-thick Long Chang section was logged and sampled at ~20 cm intervals yielding 190 samples for thin sections, conodont biostratigraphy, pyrite framboid and geochemical analyses. Conodont samples were crushed to small chips and dissolved by using 6%-10% acetic

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acid. The acid solution was buffered with tricalcium phosphate and exchanged every 48 hours until the samples were fully dissolved. Residuals were wet sieved, dried at 50°C and separated for heavy fractions using sodium polytungstate. Conodont elements were picked under a binocular microscope. For carbonate carbon isotope analyses ($\delta^{13}C_{carb}$), sample powders were reacted with 100% phosphoric acid at 70°C in a Gasbench II gas preparation and introduction system connected online to a Thermo Delta Plus mass spectrometer. Reproducibility, monitored by replicate analyses of laboratory standards calibrated to NBS 19 and LSVEC, was ±0.09‰ (1σ). Carbon isotopes were determined from total organic carbon ($\delta^{13}C_{org}$) using decarbonatized samples in an elemental analyzer (CE 1110) connected online to a Thermo Delta Plus mass spectrometer. Reproducibility, monitored by replicate analyses of the standard USGS 40, was ±0.06‰ (1σ). All carbon isotope values are given in per mille relative to VPDB. Oxygen isotopes were measured on conodont apatite ($\delta^{18}O_{PO4}$) following the method described in Joachimski et al. (2009) and are given in per mille relative to VSMOW. Analyses were performed using a TC-EA coupled online to a Thermo Delta Plus mass spectrometer. Reproducibility, monitored by replicate analyses of the standard NBS 120c (21.7‰), was ±0.14‰ (1σ). Pyrite framboid analysis was performed on 30 samples in the form of small polished chips that were examined at 2500x magnification using a Zeiss scanning electron microscope under backscattered electron mode. The minimum, maximum and mean diameter of framboids in each sample, and their standard deviation were calculated. Where possible, >100 individual framboids were measured in each sample. Mo, Mn, U, Th and Al contents were measured on powdered samples digested in a 2:2:1:1 acid solution of H_2O-HF-$HClO_4$-HNO_3. Solutions were analyzed using a PerkinElmer ICP-MS with a reproducibility better than ±2% (2σ). XRF analyses on high-Mn samples were further performed on an AMETEK Spectro XEPOS XRF analyzer for cross comparisons. Data were calibrated to the limestone

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standard JLs-1. Ten major element contents (CaO, P_2O_5, SiO_2, Al_2O_3, MgO, Fe_2O_3, Na_2O, K_2O, MnO and TiO_2) were quantitatively determined. Reproducibility was better than ±0.06 wt% (1σ). Mn concentrations were calculated from the MnO content by multiplication with a coefficient 0.775 (=molecular weight [Mn]/ molecular weight [MnO]).

4. Results

4.1 Biostratigraphy

The global Carnian conodont biostratigraphy is complex, and no consensus has been reached on a single biostratigraphic scheme (e.g., Kozur, 2003; Orchard, 2010). Here we present the first detailed attempt at an approach combing conodont zonation, bivalve and ammonoid biostratigraphy to define Carnian substage boundaries in South China. The Zhuganpo and Wayao Formations (Fm.) are likely regionally diachronous lithological units that require site-specific age diagnosis, and we have studied a part of the Zhuganpo Fm. that is time-equivalent to only the upper part of Enos et al.'s (1998) Zhuganpo Fm. Since most Late Triassic conodont species are long-ranging, we use assemblage zones for a better age constraint. The strata are biostratigraphically dated to the Julian 1 to Tuvalian 2 and record continuous sedimentation through most of the Carnian Stage (Fig. 4). The Long Chang conodont fauna is dominated by the genus *Quadralella* (*Q.*) with less abundant *Paragondolella* (*P.*), *Norigondolella* and exceptionally rare *Gladigondolella*. Three conodont zones are recognized. They are, in ascending order, the *P. foliata*-*Q. polygnathiformis* assemblage zone, *Q. polygnathiformis noah* zone and *Q. carpathica* assemblage zone. These zones can be correlated with the western Tethys zonation (e.g., Hornung et al., 2007b; Kozur, 2003; Fig. 5).

Conodonts from Long Chang display some features endemic to the region (Fig. 6). In comparison to western Tethyan and North American faunas, the South China conodonts generally

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have more elongated and less up-turned platforms. Adult specimens are locally rather rare in the Tuvalian strata. Some diagnostic features such as the presence of a bifurcated keel are less developed in the *Q. carpathica* group.

A few ammonoid and bivalve body fossils were collected from the uppermost Zhuganpo Fm. and overlying Wayao Fm. (Fig. 7). The ammonoid *Austrotrachyceras* ex gr. *A. austriacum* at ~26 m level (sample LC 34) indicates basal Julian 2 (i.e., Julian 2/I a sensu Gallet et al., 1994). The bivalve *Halobia* cf. *zitteli* at ~30 m level (sample LC 41) is suggestive of a Julian age.

Collectively, the Julian 1-Julian 2 boundary is assigned below sample LC 34. The Julian-Tuvalian boundary is tentatively assigned above sample LC 41 in which the youngest age-diagnostic Julian fossil is last seen. The Tuvalian 1-Tuvalian 2 boundary is defined at sample LC 53 (~38 m level) where *Q.* ex. gr. *carpathica* occurs.

4.2 Sedimentology

The Long Chang section records major facies changes. The basal 13 m lie within the early Carnian Zhuganpo Fm. (Julian 1) and consist of bioclastic wackestones. Thin-shelled bivalves (filaments) and holothurian sclerites are abundant; calcispheres are locally common while *Tubiphyte*s, ostracods and benthic foraminifers are rare. The succeeding 8 m consist of intraclastic limestones in which the matrix, of filamentous limestone, surrounds clasts of darker better-cemented patches of the matrix lithology (Fig. 3F). The clasts are generally well rounded and often rotated so that the alignment of the bivalve filaments is at an angle to the filaments in the matrix. The topmost metre of the Zhuganpo Fm. (Julian 2) sees the development of two distinct manganese-rich limestone beds: a lower bed consisting of nodular limestone with abundant ammonoids and an upper bed consisting of thinly bedded mudstones with calcispheres.

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The overlying Wayao Fm. (Julian 2-Tuvalian 2) consists mostly of black shales with interbeds of filamentous packstones composed of valves of the bivalve *Halobia* (Fig. 3E). The *Halobia* packstones range from 2 - 30 cm thick, although a metre-thick example is found around 45 m above the base of the section (Fig. 4). Ammonoids, crinoids and radiolarians are common, together with calcispheres (Fig. 3G-I). At one level, 28 m above the base of the section, calcipsheres ranging in size from 20 - 200 μm, dominate in thin section. Towards the top of the section (beginning at 45.5 m above the base of the section) thin interbeds of sharp-based, laminated calcisiltites appear.

The Wayao Fm. is overlain by a thick development of amalgamated sandstone turbidite beds belonging to the Laishike Fm. (possibly Tuvalian 3) that marks the start of flysch-style sedimentation (Enos et al., 1998), but note that black shale facies persist into the lowermost metres of the Laishike Fm. The boundary between Wayao Fm. and Laishike Fm. is defined regionally by the first occurrence of a fine-grained sandstone/mudstone (normally yellow-brownish in colour).

4.3 Redox Proxies

Pyrite framboid analysis is a widely used petrographic palaeoredox proxy (Wilkin et al., 1996). In general, populations of tiny framboids (mean 3-6 μm in diameter) indicate anoxic-euxinic conditions, and populations of larger framboids (mean 6-10 μm in diameter, and typically a larger standard deviation) indicate dysoxic conditions **(**Bond and Wignall, 2010). Pyrite crystals are absent from much of the Zhuganpo Fm. Only in the topmost metre does euhedral pyrite become common, although there are no framboids at this level. In contrast, tiny pyrite framboids are prolifically abundant throughout the Wayao Fm. Average diameters are small (mean ~4 μm, with maximum framboid diameter [MFD] ~8 μm) in the lower part of the Wayao Fm., and somewhat larger in the upper part, where MFD increase to ~20 μm (Fig. 4).

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Trace metals provide additional information on depositional conditions since many trace metals are insoluble in reducing environments and consequently adsorbed to sediments under an anoxic water column. Redox-controlled trace metal concentrations (Mn, Mo, U) show marked fluctuations at Long Chang. Manganese values are low (~200-2,500 ppm) through most of the section, but increase notably by two orders of magnitude to values ranging from ~0.8 to 11.2 wt% (8,000-112,000 ppm) in a 2 m interval at the Zhuganpo-Wayao transition (Julian 2) (Table 1; Fig. 3C; Fig. 4). Mo/Al ratios show a different trend. Mo/Al is ≤1.0 throughout the Zhuganpo Fm., before the ratio rises sharply at the base of the Wayao Fm. from where it remains between ~3 and ~10 up-section. U/Th ratios are ≤1.0 throughout most of the Zhuganpo Fm., and rise to ~2 in the uppermost Zhuganpo Fm. from where they remain between 1 and 2 up-section.

4.4 Stable isotopes

The $\delta^{13}C_{carb}$ values are stable at ~2.6‰ through the lower part of the section, then show an abrupt ~4.5‰ negative excursion starting at the latest Julian 1 (Fig. 4). A low point of -2‰ is reached at the Julian-Tuvalian boundary before values rise again to ~1.5‰ in the middle of the Tuvalian 1. A second, short-lived negative shift of ~2‰ is seen in the upper Tuvalian 1, before values stabilize up-section at ~2.5‰. $\delta^{13}C_{org}$ data show good covariance with $\delta^{13}C_{carb}$. Thus, $\delta^{13}C_{org}$ values decrease from ~-26‰ in the Julian 1 to ~-29‰ at the Julian-Tuvalian boundary, before stabilizing at pre-excursion values in the Tuvalian 2. However, the double negative spike observed in the $\delta^{13}C_{carb}$ record is not well developed in the $\delta^{13}C_{org}$ record, although there is a small increase in $\delta^{13}C_{org}$ values between the two negative shifts (Fig. 4).

The $\delta^{18}O_{PO4}$ record contains two negative excursions. A ~1‰ negative shift occurs in the Julian 2 when $\delta^{18}O_{PO4}$ decreased from ~20.0 to ~19.0‰. This was followed by a positive shift to

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pre-excursion $\delta^{18}O_{PO4}$ values. A second, more gradual ~1.5% negative shift occurs throughout Tuvalian 2, when $\delta^{18}O_{PO4}$ decreased from ~20.0 to ~18.5‰ (Fig. 4).

5. Discussion

The Long Chang section records many of the attributes of the Carnian Crisis and CHE that have been reported from far removed sections of western Tethys, indicating that those changes were manifest over a large (global?) geographical area (Fig. 1B). Significantly, the $\delta^{13}C_{org}$ trends known from western Tethys (e.g., Dal Corso et al., 2015) are replicated eastern Tethys where they also occur in the $\delta^{13}C_{carb}$ record. The $\delta^{18}O_{PO4}$ record reveals that the Carnian Crisis was associated with two phases of warming that correlate with the dry-wet climatic transition.

5.1 Towards an age model

Establishing the timing of the Carnian Crisis and the CHE in South China and elsewhere is a crucial step in the evaluation of various causal mechanisms that have been postulated for the inferred environmental changes. However, the age of Late Triassic strata in Guizhou is a subject of intense debate. Unfortunately, many age-diagnostic taxa exhibit regional differences both in terms of individual morphologies and in terms of temporal range during the Carnian (Fig. 5) and so thorough dating of individual sections is essential when developing a regional biostratigraphy for South China. Our approach combines conodont interval and assemblage zones (based on a combination of conodont species) with ammonoid/bivalve ranges to define the biostratigraphic scheme, therefore excluding problems caused by regional differences in the first and last occurrences of individual conodont species. Our biostratigraphy supports conclusions of a recent ammonoid study (Zou et al., 2015) that assigned the upper part of Zhuganpo Fm. (studied in this paper) and the overlying Wayao Fm. to the Carnian. However, our biostratigraphic age assignment for the Long Chang sequence differs from that inferred by magnetostratigraphy and

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cyclostratigraphy performed on similar lithological units (Zhang et al., 2015). Zhang et al. (2015) place the Zhuganpo, Wayao, Laishike and Bana formations entirely within the Carnian, thus constraining the older formations to the early part of the Carnian. It is possible that the discrepancy between bio- and magneto- and cyclostratigraphy is due to the diachronous nature of these rather loosely defined local lithological units. Thus, while other interpretations might be valid, our detailed ammonoid, bivalve and conodont sampling provides the first detailed biostratigraphy for the area, and it is independently verified by our carbon isotope chemostratigraphy that is correlatable to well-dated western Tethyan sections (Fig. 8).

5.2 A carbonate production crisis

A decrease in carbonate productivity is indicated by the facies changes at Long Chang. The basal bivalve-holothurian wackestones of the Zhuganpo Fm. represent a rather unusual variety of platform carbonate that probably formed in deeper waters (compared to the underlying Yangliujing cyclic limestones and dolostones, Fig. 2C). The lack (or rarity) of typical shallow-water Carnian components such as calcareous algae, *Tubiphytes* and crinoids is probably due to the location of Long Chang at the transition between the Yangtze platform and an adjacent basin. The overlying intraclastic limestones show a high degree of early cementation followed by re-orientation of the clasts. Enos et al. (1998) argued that the intraclasts did not undergo significant transport because they are of the same lithology as the matrix. While this observation is pertinent we note that frequent re-orientation of the clasts suggests some degree of movement, perhaps during debris flow emplacement. Carbonate productivity declined at the Julian 1-Julian 2 transition when a manganiferous, ammonoid wackestone developed—a facies typical of deep-water, condensed limestones such as the Jurassic Ammonitico Rosso (Jenkyns, 1991). This was followed by deposition of deep-water black shales and *Halobia* packstones of the Wayao Fm., indicating

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further base-level deepening. The *Halobia* packstones consist of horizontally laminated, densely packed thin valves, probably formed under sediment-starved conditions (Fig. 3G). This facies persists up to 42 m above the base of the studied section, suggesting that the sedimentation rates were probably locally still low in the early Tuvalian 2 substage.

The drowning of the carbonate platform did not result in a complete shutdown of carbonate productivity because 1) though carbonates were relatively rare, *Halobia* packstones occur at several levels in the Wayao Fm. (Fig. 3D, G); 2) in the adjacent Guanling area ~60 km away from Long Chang, carbonate production (though briefly punctuated by black shale deposition) continued well into the Tuvalian 3 substage (Wang et al., 2008); and 3) further afield, on the northwestern Yangtze Platform, carbonate platforms were replaced by ramp deposits and sponge reefs in Tuvalian and carbonate production persisted until the early Norian in Sichuan Province at which time they were inundated by flysch sediments (Wu, 1989; Li et al., 2014). The transition from an extensional platform-and-basin setting to a foreland basin setting in South China probably occurred between the latest Carnian (~Tuvalian 3 at Long Chang) and earliest Norian (in NW Sichuan). In this scenario the loss of carbonate production late in the Julian cannot be the result of foreland basin development because it slightly predates this tectonic change. At Long Chang, it seems that sedimentation rates were still rather low during the Julian 2 (uppermost Zhuganpo Fm.) and the Tuvalian 1-2 substages (lower-middle Wayao Fm.) as suggested by the occurrence of nodular ammonoid wackestones and *Halobia* packstones. Indications of foreland basin development such as a substantial increase in sedimentation rates and occurrence of basin infill only occurred in the Laishike Fm. (probably Tuvalian 3 and later).

5.3 Marine anoxia

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The inferred increase in water depth at the top of the Zhuganpo Fm. at Long Chang saw the establishment of intense euxinia, although only after the development of Mn-rich limestones (MnO content is generally>1.0 wt%, up to 15.1 wt%). The U/Th ratio increases at the same level, suggesting that redox is one of the main controls on Mn scavenging. The increase in the U/Th ratio slightly predates the increases of Mo/Al, probably a function of the greater enrichment of U in dysoxic waters relative to Mo, which only enriches in intensely anoxic sulfidic water (Algeo and Tribovillard, 2009). Unlike most metal sulfides, manganese sulfide is soluble in anoxic waters, but in oxygenated waters Mn is rapidly precipitated in a carbonate form or as particulate oxides (Brewer and Spencer, 1974). In many euxinic basins Mn is shuttled from sulfidic, deeper basinal waters to the redox boundary where manganese-rich carbonates form a "bath-tub ring" (Frakes and Bolton, 1984). Manganese-rich deposits in effect herald the arrival of upward-expanding euxinic waters (Jenkyns et al., 1991). Thus, the Mn enrichment in the uppermost Zhuganpo Fm. suggests deeper water euxinia had already developed prior to the appearance of euxinic facies at Long Chang.

Pyrite framboids in the Wayao Fm. are even smaller than those raining to the seafloor in the modern euxinic Black Sea (Wilkin et al., 1996), indicating intense water column stratification (Bond and Wignall, 2010) developing later in the Julian 2 interval. This conclusion is supported by the exceptionally high Mo/Al values and low biodiversity in the Wayao Fm., which lacks benthos, and is dominated by the bivalve *Halobia* and the pseudoplanktonic crinoid *Traumatocrinus* which attached itself to driftwood. The ecology of *Halobia* has been debated – some regard it as pelagic while others suggest it is a low-oxygen specialist able to survive during short periods of elevated benthic oxygen levels (Allison et al., 1995). Thus occurrences of *Halobia* coquinas in the Wayao Fm. indicate recurring dysoxic conditions in an overall anoxic environment.

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Our pyrite framboid data, which show a clear trend from no pyrite to euhedral pyrite and then abundant small framboids up-section, help to exclude the possibility that the enrichment of redox sensitive elements in the uppermost Zhuganpo Fm. reflects a post-depositional exchange with the overlying anoxic Wayao black shales because although trace metals can migrate after deposition, pyrite crystals cannot – they record a primary signal of oxygen depletion.

An alternative interpretation of the low biodiversity of the Wayao Fm. lies in the poisoning of benthos due to the excessive supply of terrestrial nutrients in the local basin. However, the lack of apparent faunal changes, the lack of evidence for increase in bioerosion towards the termination of platform, and the lack of phosphate enrichment in the pelagic and condensed intervals at Long Chang and adjacent areas (Enos et al., 1998) do not support a substantial increase of nutrient level. Instead, contemporaneous oxygen depletion recorded in the Italian Alps, North Africa, Sicily, and South China provides compelling evidence for a supra-regional anoxic event (Bellanca et al., 1995; Keim et al., 2006; Hornung et al., 2007b; Soua, 2014; Minzoni et al., 2015).

5.4 Carbon isotope and sea-surface temperature (SST) trends

The late Julian 1 negative $\delta^{13}C$ excursion at Long Chang matches that seen in the western Tethyan marine and terrestrial organic carbon isotope records (Dal Corso et al., 2012, 2015) indicating that it is a Tethys-wide, and probably global signal. The Long Chang $\delta^{13}C_{carb}$ profile reveals several further negative shifts, which are not yet known from elsewhere. This probably reflects a lack of detailed sampling at this level – the Tuvalian has received much less attention than the Julian (Fig. 1B; Fig. 8). In contrast, the $\delta^{13}C_{org}$ profile in South China suggests a single, prolonged negative excursion between the Julian 2 and Tuvalian 2 substages (Fig. 2). Despite this minor discrepancy between the $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ records, there was clearly a prolonged perturbation of the carbon cycle in the later Carnian, coincident with several contemporaneous

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volcanic eruptions (Walderhaug et al., 2005; Maury et al., 2008; Greene et al., 2010; summarized in Fig. 1C). The Wayao Fm. is known for the excellent preservation of large pieces of driftwood that were encrusted with crinoids (Wang et al., 2008). It is possible that the negative excursion seen in the $\delta^{13}C_{org}$ record is an artefact of such terrestrial organic carbon input. However, this is unlikely because in general, $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ co-vary at Long Chang, and terrestrial organic carbon input cannot account for the scale of changes seen in the $\delta^{13}C_{carb}$ record. However, $\delta^{13}C_{org}$ records from Long Chang and Luntz basin (Mueller et al., 2016) as well as within Northern Calcareous Alps (Dal Corso et al., 2015 v.s. Mueller et al., 2016) show different variation patterns in the Julian 2 (Fig. 8). Such differences likely suggest a major disturbance and compositional variation in regional organic carbon pools. The $\delta^{13}C_{org}$ signal may reflect local effects (e.g., terrestrial/marine organic carbon mixing) and (potentially) the unreliability of measuring $\delta^{13}C_{org}$ on a very small amounts of carbon (e.g., on sandstone, claystone etc.).

Calculating absolute seawater temperatures from $\delta^{18}O_{PO4}$ using Pucéat et al.'s (2010) equation and assuming an ice-free world in the Late Triassic ($\delta^{18}O_{seawater}$=-1‰), yields warm subtropical SSTs (~26°C-33°C). These temperatures correspond to modern equatorial SSTs and are similar to temperatures calculated for the Middle Triassic of the same region (Sun et al., 2012). The two negative $\delta^{18}O_{PO4}$ excursions indicate increased warmth with temperatures increasing by 4°C from ~27°C to ~31°C in the Julian 2 substage, followed by a short-lived cooling and then a second rise of ~7°C from ~26 to ~33°C, beginning in the Tuvalian 1 substage and persisting until the later Carnian. Despite differences in analytical methods, sampling resolution and regional $\delta^{18}O_{seawater}$, a comparable warming trend of ~6-8°C from the Julian to the late Tuvalian is also reported from contemporaneous sections in the Northern Calcareous Alps and Southern Alps

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(Hornung et al., 2007b; Trotter et al., 2015), suggesting that this is a true, global signature (Fig. 9).

5.5 Precipitation changes during the CHE

One of the most intriguing features of the CHE is that the prevailing Late Triassic dry climate was interrupted by a transient interval of enhanced rainfall (e.g., Simms and Ruffell, 1990). In western Tethys this caused the sudden shutdown of carbonate platforms by enhanced terrestrial siliciclastic input (sands and muds, etc.) due to elevated run-off rates (e.g., Kozur and Bachmann, 2010). High run-off is also invoked in North America where mid-Carnian sandstone occasionally contains kaolinite, suggesting a humid climate (Simms and Ruffell, 1990). The arid-humid transition in the Carnian is best constrained by palynological studies — the Carnian in general was very arid, but a humid pulse began at the Julian 1- Julian 2 boundary and ended in the early Tuvalian (Roghi et al., 2010; Mueller et al., 2016).

The ultimate cause(s) for the mid-Carnian climate change is a matter of intense debate. Enhanced precipitation is has long been postulated the result of mega-monsoon activities due to the palaeogeographic configuration of Pangea (e.g., Parrish, 1993). However, climate changes triggered by tectonic movements are generally gradual and long-lasting, in contrast to the sudden and transient nature of the CHE. The global climate change from arid to humid conditions in the Julian 2 (Mueller et al., 2016) coincided with the negative carbon isotope excursion and temperature rise seen at Long Chang and elsewhere (Fig. 4; Hornung et al., 2007b), which suggests instead a warm-climate-driven model.

Volcanic activity, both silicic and basaltic, was geographically widespread and intense during the Carnian (e.g., Walderhaug et al., 2005; Maury et al., 2008; Greene et al., 2010; Moix et al., 2013). The Wrangellia Large Igneous Province (LIP) and contemporaneous volcanic activity

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elsewhere (Fig. 1) likely injected copious light carbon to the atmosphere and initiated the first warming event seen in the Julian 2 substage. An intensification of the hydrologic cycle and an increase in precipitation is a predictable consequence of climate warming. Thus, widely-distributed wind-blown dust found in the Carnian Panthalassa Ocean points to strong atmospheric convection (Nakada et al., 2014).

Our combined biostratigraphy and carbon isotope records provide a temporal framework for the succession of intra-Carnian events, facilitating the detailed reconstruction of the sequence of environmental changes. The late Julian was marked by carbonate platform drowning (possibly correlating to sea level rise) and the development of intense basin anoxia. This coincided with the onset of a negative carbon isotope excursion and a warming trend that is correlative with the CHE and similar $\delta^{13}C$ trends in western Tethys. Platform drowning may have been related in part to the development of deep-water euxinia, which could have "poisoned" carbonate production leading initially to the development of condensed, manganiferous carbonates and ultimately to the loss of all but filamentous limestone production as the area of black shale deposition expanded. High Mn^{2+} concentrations in euxinic waters may have been another contributing (if not the key) factor to rarity of carbonates in the Wayao Fm. because Mn^{2+} inhibits carbonate precipitation (Dromgoole and Walter, 1990).

Once established, black shale and filamentous limestone deposition persisted into the Tuvalian when $\delta^{13}C$ and $\delta^{18}O_{PO4}$ fluctuations reveal substantial further perturbations to both the carbon cycle and SSTs: at least one more prolonged, warm episode and two further negative $\delta^{13}C$ excursions. These phenomena have not been reported from western Tethys, probably due to the less intense study of the post-CHE interval there (Fig. 1B). Coincident with the Wrangellia LIP

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and other coeval substantial volcanic activity, the CHE likely represents another case of volcanism-induced global climatic changes.

6. Conclusions

Major Carnian climatic and environmental changes were widespread throughout the Tethyan ocean region: they have the hallmarks of many mass extinction events including warming, carbon isotope oscillations, transgression and the spread of deep-water anoxic facies. The coincidence with the Wrangellia LIP and several Tethyan volcanism strengthens the similarity of the Carnian Crisis to major Permian and Triassic extinctions, and suggests the Carnian event could rank alongside those in terms of environmental perturbations.

The dry-wet transition occurred during the CHE and coincided with a -4‰ negative carbon isotope excursion and a 4°C warming, arguing for a volcanism-induced, warm-climate-driven intensification of hydrologic cycle activities.

A decrease in carbonate productivity during the CHE is likely a global phenomenon triggered by different mechanisms. The complete shutdown of carbonate production by siliciclastic influx seen in western Europe only occurred at a regional scale. Though inhibited, carbonate precipitation persisted until at least the end of the Julian 2 in many deeper water localities (e.g., Lukeneder et al., 2012; Soua, 2014; this study). The carbonate production decrease in South China (Julian 2) predated foreland basin development (latest Tuvalian 2 to Tuvalian 3) at the onset of the Indosinian Orogeny and was probably the result of platform drowning, euxinia and high Mn^{2+} concentrations.

Though more studies on other regions will provide a better understanding of this intriguing interval in Earth history, our work in South China supports a scenario in which environmental changes were both geographically and temporally widespread during the Carnian. The most

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harmful interval for marine life is likely to have occurred during the early Julian, coincident with the CHE, when a warming trend, increasing precipitation and intense marine anoxia were established, resulting in widespread, but as yet poorly constrained, faunal losses.

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References

Algeo, T.J., Tribovillard, N., 2009. Environmental analysis of paleoceanographic systems based on molybdenum–uranium covariation. Chem. Geol. 268, 211-225.

Allison, P. A., Wignall, P. B., Brett, C. E., 1995, Palaeo-oxygenation: effects and recognition. Geol. Soc. Spec. Pub. 83, 97-112.

Balini, M., Lucas, S.G., Jenks, J.F., Spielmann, J.A., 2010. Triassic ammonoid biostratigraphy: an overview. Geol. Soc. Spec. Pub. 334, 221-262.

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Bellanca, A., Di Stefano, P., Neri, R., 1995. Sedimentology and isotope geochemistry of Carnian deep-water marl/limestone deposits from the Sicani Mountains, Sicily: Environmental implications and evidence for a planktonic source of lime mud. Palaeogeogr. Palaeoclimatol. Palaeoecol. 114, 111-129.

Benton, M. J., 1986. More than one event in the late Triassic mass extinction. Nature 321, 857-861.

Bond, D.P.G., Wignall, P.B., 2010. Pyrite framboid study of marine Permian–Triassic boundary sections: A complex anoxic event and its relationship to contemporaneous mass extinction. Geol. Soc. Am. Bull. 122, 1265–1279.

Brewer, P. G., Spencer, D. W., 1974. Distribution of some trace elements in the Black Sea and their flux between dissolved and particulate phases, *in* Degens, E. T., and Ross, D. A., eds., In The Black Sea: Geology, chemistry and biology. AAPG Memoir, p. 137-143.

Dal Corso, J., Mietto, P., Newton, R.J., Pancost, R.D., Preto, N., Roghi, G., Wignall, P.B., 2012. Discovery of a major negative $\delta^{13}C$ spike in the Carnian (Late Triassic) linked to the eruption of Wrangellia flood basalts. Geology 40, 79-82.

Dal Corso, J., Gianolla, P., Newton, R. J., Franceschi, M., Roghi, G., Caggiati, M., Raucsik, B., Budai, T., Haas, J., Preto, N., 2015. Carbon isotope records reveal synchronicity between carbon cycle perturbation and the “Carnian Pluvial Event” in the Tethys realm (Late Triassic). Global and Planet. Change 127, 79-90.

Dromgoole, E.L., Walter, L.M., 1990. Inhibition of calcite growth rates by Mn^{2+} in $CaCl_2$ solutions at 10, 25, and 50°C. Geochim. Cosmochim. Acta 54, 2991-3000.

Enos, P., Wei, J., Lehrmann, D. J., 1998. Death in Guizhou — Late Triassic drowning of the Yangtze carbonate platform. Sediment. Geol. 118, 55-76.

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Frakes, L.A., Bolton, B.R., 1984. Origin of manganese giants: Sea-level change and anoxic-oxic history. Geology 12, 83-86.

Gallet, Y., Besse, J., Krystyn, L., Théveniaut, H., Marcoux, J., 1994. Magnetostratigraphy of the Mayerling section (Austria) and Erenkolu Mezarlik (Turkey) section: Improvement of the Carnian (late Triassic) magnetic polarity time scale. Earth Planet. Sci. Lett. 125, 173-191.

Golonka, J., 2007. Late Triassic and Early Jurassic palaeogeography of the world. Palaeogeogr. Palaeoclimatol. Palaeoecol. 244, 297-307.

Greene, A. R., Scoates, J. S., Weis, D., Katvala, E. C., Israel, S., Nixon, G. T., 2010. The architecture of oceanic plateaus revealed by the volcanic stratigraphy of the accreted Wrangellia oceanic plateau. Geosphere 6, 47-73.

Hornung, T., Krystyn, L., Brandner, R., 2007a. A Tethys-wide mid-Carnian (Upper Triassic) carbonate productivity crisis: Evidence for the Alpine Reingraben Event from Spiti (Indian Himalaya)? J. Asian Earth Sci. 30, 285-302.

Hornung, T., Brandner, R., Krystyn, L., Joachimskimo, M. M., Keim, L., 2007b. Multistratigraphic Constraints on the NW Tethyan "Carnian Crisis". The Global Triassic. New Mexico Museum of Natural Histroy and Science Bulletin 41, 59-67.

Jenkyns, H. C., 1991. Origin of red nodular limestones (Ammonitico Rosso, Knollenkalke) in the Mediterranean Jurassic: a diagenetic model: Pelagic Sediments. on Land and Under the Sea (IAS Spec. Publ. no. 1), p. 249-271.

Jenkyns, H. C., Géczy, B., Marshall, J. D., 1991. Jurassic Manganese Carbonates of Central Europe and the Early Toarcian Anoxic Event. J. Geol. 99, 137-149.

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Joachimski, M. M., Breisig, S., Buggisch, W., Talent, J. A., Mawson, R., Gereke, M., Morrow, J. R., Day, J., Weddige, K., 2009. Devonian climate and reef evolution: Insights from oxygen isotopes in apatite. Earth Planet. Sci. Lett. 284, 599-609.

Keim, L., Spöti, C., Brandner, R., 2006. The aftermath of the Carnian carbonate platform demise: a basinal perspective (Dolomites, Southern Alps). Sedimentology 53, 361-386.

Konstantinov, A. G., 2012. A revision of the Early Carnian Trachyceratidae (Ammonoidea) of Northeastern Asia. Paleontol. J. 46, 453-460.

Kozur, H., 2003. Integrated ammonoid, conodont and radiolarian zonation of the triassic and some remarks to Stage/Substage subdivision and the numeric age of the triassic stages. Albertiana 28, 57-74.

Kozur, H. W., Bachmann, G. H., 2010. The Middle Carnian Wet Intermezzo of the Stuttgart Formation (Schilfsandstein), Germanic Basin. Palaeogeogr. Palaeoclimatol. Palaeoecol., 290, 107-119.

Li, Y., Li, H., Zhou, R., Su, D., Yan, L., Yan, Z., 2014. Crustal thickening or isostatic rebound of orogenic wedge deduced from tectonostratigraphic units in Indosinian foreland basin, Longmen Shan, China. Tectonophysics 619–620, 1-12.

Lukeneder, S., Lukeneder, A., Harzhauser, M., İslamoğlu, Y., Krystyn, L., Lein, R., 2012. A delayed carbonate factory breakdown during the Tethyan-wide Carnian Pluvial Episode along the Cimmerian terranes (Taurus, Turkey). Facies 58, 279-296.

Maury, R. C., Lapierre, H., Bosch, D., Marcoux, J., Krystyn, L., Cotten, J., Bussy, F., Brunet, P., Senebier, F., 2008. The alkaline intraplate volcanism of the Antalya nappes (Turkey): a Late Triassic remnant of the Neotethys. Bulletin de la Societe Geologique de France 179, 397-410.

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Minzoni, M., Lehrmann, D.J., Dezoeten, E., Enos, P., Montgomery, P., Berry, A., Qin, Y., Meiyi, Y., Ellwood, B.B., Payne, J.L., 2015. Drowning of the Triassic Yangtze Platform, South China, By Tectonic Subsidence Into Toxic Deep Waters of An Anoxic Basin. J. Sediment. Res. 85, 419-444.

Moix, P., Vachard, D., Allibon, J., Martini, R., Wernli, R., Kozur, H.W., Stanpfli, G.M., 2013. Paleotethyan, Neotethyan and Huğlu-Pindos Series in the Lycian nappes (SW Turkey): Geodynamical implications, in: Tanner, L.H., Spielmann, J.A., Lucas, S.G. (Eds.), The Triassic System. New Mexico Museum of Natural History and Science Bulletin, 61, 401-444.

Mueller, S., Krystyn, L., Kürschner, W.M., 2016. Climate variability during the Carnian Pluvial Phase — A quantitative palynological study of the Carnian sedimentary succession at Lunz am See, Northern Calcareous Alps, Austria. Palaeogeogr. Palaeoclimatol. Palaeoecol. 441, 198-211.

Nakada, R., Ogawa, K., Suzuki, N., Takahashi, S., Takahashi, Y., 2014. Late Triassic compositional changes of aeolian dusts in the pelagic Panthalassa: Response to the continental climatic change. Palaeogeogr. Palaeoclimatol. Palaeoecol. 393, 61-75.

Ogg, J. G., 2015. The mysterious Mid-Carnian "Wet Intermezzo" global event. J. Earth Science 26, 181-191.

Orchard, M.J., 2010. Triassic conodonts and their role in stage boundary definition. Geol. Soc. Spec. Pub. 334, 139-161.

Parrish, J. T., 1993. Climate of the supercontinent Pangea. J. Geol. 101, 215-233.

Pucéat, E., Joachimski, M. M., Bouilloux, A., Monna, F., Bonin, A., Motreuil, S., Morinière, P., Hénard, S., Mourin, J., Dera, G., Quesne, D., 2010. Revised phosphate-water fractionation

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equation reassessing paleotemperatures derived from biogenic apatite. Earth Planet. Sci. Lett. 298, 135-142.

Rigo, M., Preto, N., Roghi, G., Tateo, F., Mietto, P., 2007. A rise in the Carbonate Compensation Depth of western Tethys in the Carnian (Late Triassic): Deep-water evidence for the Carnian Pluvial Event. Palaeogeogr. Palaeoclimatol. Palaeoecol. 246, 188-205.

Rigo, M., Trotter, J.A., Preto, N., Williams, I.S., 2012. Oxygen isotopic evidence for Late Triassic monsoonal upwelling in the northwestern Tethys. Geology 40, 515-518.

Roghi, G., Gianolla, P., Minarelli, L., Pilati, C., Preto, N., 2010. Palynological correlation of Carnian humid pulses throughout western Tethys. Palaeogeogr. Palaeoclimatol. Palaeoecol. 290, 89-106.

Ruffell, A., Simms, M.J., Wignall, P.B., 2015. The Carnian Humid Episode of the late Triassic: a review. Geol. Mag. 153, 271-284.

Simms, M. J., and Ruffell, A. H., 1989. Synchroneity of climatic change and extinctions in the Late Triassic. Geology 17, 265-268.

Simms, M.J., Ruffell, A.H., 1990. Climatic and biotic change in the late Triassic. J. Geol. Soc. 147, 321-327.

Soua, M., 2014. Early Carnian anoxic event as recorded in the southern Tethyan margin, Tunisia: an overview. Int. Geol. Rev. 56, 1884-1905.

Sun, Y.D., Joachimski, M.M., Wignall, P.B., Yan, C.B., Chen, Y.L., Jiang, H.S., Wang, L.N., Lai, X.L., 2012. Lethally Hot Temperatures During the Early Triassic Greenhouse. Science 338, 366-370.

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Trotter, J.A., Williams, I.S., Nicora, A., Mazza, M., Rigo, M., 2015. Long-term cycles of Triassic climate change: a new $\delta^{18}O$ record from conodont apatite. Earth Planet. Sci. Lett. 415, 165-174.

Walderhaug, H. J., Eide, E. A., Scott, R. A., Inger, S., Golionko, E. G., 2005. Palaeomagnetism and $^{40}Ar/^{39}Ar$ geochronology from the South Taimyr igneous complex, Arctic Russia: a Middle-Late Triassic magmatic pulse after Siberian flood-basalt volcanism. Geophys. J. Int. 163, 501-517.

Wang, X., Bachmann, G.H., Hagdorn, H., Sander, P.M., Cuny, G., Chen, X., Wang, C., Chen, L., Cheng, L., Meng, F., Xu, G., 2008. The Late Triassic Black Shales of the Guanling Area, Guizhou Province, South-West China: A Unique Marine Reptile and Pelagic Crinoid Fossil Lagerstätte. Palaeontology 51, 27-61.

Wilkin, R. T., Barnes, H. L., and Brantley, S. L., 1996. The size distribution of framboidal pyrite in modern sediments: An indicator of redox conditions. Geochim. Cosmochim. Acta 60, 3897-3912.

Wu, X., 1989. Carnian (upper triassic) sponge mounds of the Northwestern Sichuan Basin, China: Stratigraphy, facies and paleoecology. Facies 21, 171-187.

Zhang, Y., Li, M., Ogg, J.G., Montgomery, P., Huang, C., Chen, Z.-Q., Shi, Z., Enos, P., Lehrmann, D.J., 2015. Cycle-calibrated magnetostratigraphy of middle Carnian from South China: Implications for Late Triassic time scale and termination of the Yangtze Platform. Palaeogeogr. Palaeoclimatol. Palaeoecol. 436, 135-166.

Zou, X., Balini, M., Jiang, D.-Y., Tintori, A., Sun, Z.-Y., Sun, Y.-L., 2015. Ammonoids from the Zhuganpo Member of the Falang Formation at Nimaigu and their relevance for dating the

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Xingyi Fossil-Lagerstaette (Late Ladinian, Guizhou, China). Rivista Italiana di Paleontologia e Stratigrafia 121, 135-161.

FIGURE CAPTIONS

Figure 1. A) Palaeogeographic reconstruction of the Carnian world (revised after Golonka, 2007); B) summary of previous studies on the Carnian Humid Episode. Gray shading represents the short-lived CHE interval; C) (inset) biostratigraphic and radiometric ages of contemporaneous volcanic activity (Walderhaug et al., 2005; Maury et al., 2008; Greene et al., 2010; Moix et al., 2013).

Figure 2. Maps of the study area and summary of local Triassic stratigraphy (simplified from Enos et al., 1998). A) Palaeogeographic reconstruction of South China during the Triassic period; B) Geological map of the Long Chang area showing the location of the study section approximately 2 km southwest of Long Chang town; C) Summary of the Middle-Late Triassic strata in the study area. Formation abbreviations: YLJ., Yangliujing; ZGP., Zhuganpo; WY., Wayao; LSK., Laishike; BN., Banna; HBC., Huobachong.

Figure 3. Field photographs of the Long Chang section (A-E) and photomicrographs of samples (F-I). A) The middle and upper part of the Long Chang section, showing three lithological units. Yellow pentagons indicate carbonate beds within the black shale-dominated Wayao Fm. Motorcycle and passengers for scale (approximately 1.5 m high); B) The thickly bedded lower Zhuganpo Fm., C.B. Yan (~1.8 m high) for scale; C) The uppermost Zhuganpo Fm., showing a transition from thickly bedded limestone to ammonoid-bearing nodular limestone to Mn-enriched thin-bedded limestone. Manganese concentration increased by two orders of magnitude through the interval marked by the white arrow. Yellow field notebook (~15 cm long) for scale; D) Highly condensed *Halobia* limestone intercalated in the middle part of the Wayao Fm. Ruler for scale; E)

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crinoid *Traumatocrinus* stems (white arrows) at the Zhuganpo-Wayao transition (~27 m above the base of the section); F) Intraclast wacketone from the upper Zhuganpo Fm. (sample LC 24, 21 m above base of section), showing dark, better cemented intraclasts surrounded by a filamentous matrix. The clasts are generally well rounded and often rotated so that the alignment of the bivalve filaments is at an angle to the filaments in the matrix; G) *Halobia* filamenatous floatstone from the Wayao Fm. (sample LC 53, ~38.5 m above base of section). *Halobia* valves are densely packed, and this facies probably formed under sediment starved conditions. A crinoid (Cr) is also visible in this sample; H) Ammonoid wackestone from the Zhuganpo Fm. (sample LC 18, ~18 m above base of section), showing an ammonoid surrounded by aligned filaments in the matrix; I) Calciphere (Ca) and radiolarian (R) packstone from the Wayao Fm. (sample LC 46, ~33 m above base of section).

Figure 4. Log of Long Chang section with pyrite framboid box-and-whisker plots (the 'box' depicts the 25th and 75th percentile of framboid distributions, the 'whiskers' depict the minimum and maximum framboid diameters, and the central line depicts the median), inferred precipitation changes, conodont oxygen and carbon isotopes records and trace metal variations. Precipitation conditions are based on palynological studies (Roghi et al., 2010; Kozur and Bachmann, 2010; Mueller et al., 2016), and reflect a general (and potentially global) trend observed in the western Tethys. Mo/Al and Mn are plotted using a log scale. The black ammonoid symbol represents a level yielding Julian 2 ammonoid *Austrotrachyceras* ex gr. *A. austriacum* (Mojsisovics). The bivalve symbol represents the youngest occurrence of age-diagnostic Julian fossil *Halobia* cf. *zitteli* Lindstroem. The crinoid symbols represent levels yielding abundant pseudoplanktonic crinoid *Traumatocrinus*. A. Z., assemblage zone. CPE (Carnian Pluvial Event) here follows the

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definition of Mueller et al. (2016) as a sudden sedimentary turnover from carbonates to siliciclastics.

Figure 5. Correlation of ammonoid and conodont zonations of western Tethys, North America and South China. Data are collated from Kozur (2003), Hornung et al. (2007b), Balini et al. (2010), Orchard (2010) and this study.

Figure 6. Scanning electron microscope images of conodonts from the Long Chang section. Scale bars are 100 μm long. 'a' = upper view, 'b' = lateral view, and 'c' = back view. **1, 8, *Paragondolella foliata*** Budurov, 1, from sample LC30_02; 8, LC31_04. **2, *Paragondolella inclinata*** Orchard, LC19_01. **3, *Metapolynathus tadpole*** (Hayashi) LC19_03. **4, 6, 9, *Quadralella polygnathiformis*** (Budurov & Stefanov), 4, LC7.5_04; 6, LC34_02; 9, LC48_04. **5, 7, 14, *Quadralella* sp.** 5, LC2_02; **7**, LC40_04; 14, LC42_01. **10, *Quadralella* cf. *carpathica*** (Mock, 1979), LC53_19. **11, *Quadralella carpathica*** (Mock, 1979), LC53_21. **12, 13, *Quadralella polygnathiformis noah*** Hayashi, 12, LC46_04; 13, LC46_06.

Figure 7. Ammonoids (A and B) and bivalves (C and D) recovered from the studied section. The ammonoids *Austrotrachyceras* ex gr. *A. austriacum* (Mojsisovics) were from the nodular limestone of the uppermost Zhuganpo Fm (sample LC 34 at ~26 m level shown in Fig. 2C), indicating the basal Julian 2 (i.e., Julian 2/I a). Suture lines of respective specimens are included in the right side of the photos. Well-preserved thin-shelled *Halobia* cf. *zitteli* Lindstroem in the lower Wayao Fm. (sample LC 41, ~30 m above the base of section) indicates a Julian age.

Figure 8. Comparison of carbon isotope records from South China (this study) and western Tethyan sections from Europe (Dal Corso et al., 2012, 2015; Mueller et al., 2016). Note the regional difference in organic carbon records in the Julian 2 may reflect compositional variations in regional organic carbon pools. NCA, Northern Calcareous Alps.

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Figure 9. Summary of intra-Carnian palaeotemperature variations in South China (this study) and the Lagonegro Basin (Trotter et al., 2015) and Northern Calcareous Alps (Hornung et al., 2007b). Data are derived from conodont oxygen isotope thermometry. Pink shades indicate warming intervals. Despite differences in analytical methods, sampling resolution and regional $\delta^{18}O$ of seawater, all three studies show a long term temperature rise of 6-8°C from the Julian to the late Tuvalian.

Table 1. XRF data of major element variations in the manganiferous limestone from the uppermost Zhuganpo Fm. and lowermost Wayao Fm., showing unusually high MnO contents. Mn concentrations can be calculated from the MnO contents by multiplying with a coefficient 0.775 (molecular weight [Mn]/ molecular weight [MnO]). LOI, loss on ignition. Numbers are in wt%.

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