Background Earth system state amplified Carnian (Late Triassic) environmental changes

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Abstract

Past major biological turnovers are coeval to large injections of CO_2 into the atmosphere–ocean system that are often linked to the emplacement of Large Igneous Provinces. The impact of these CO_2 pulses on ecosystems is however different at different times, and this difference is contingent on the initial boundary conditions. Here, we show how delayed vegetation recovery from the Permo–Triassic mass extinction (ca. 252 Ma) and continuing amalgamation of Pangaea, determined the style of the Late Triassic Carnian Pluvial Episode (ca. 233 Ma), a global climate change event linked to a major biological turnover and to the emplacement of the Wrangellia LIP. Our new biogeochemical modelling can reproduce changes in C and Sr isotopes only when we assume that high terrestrial productivity recovers in the Carnian, ~15 Myrs after the Permo–Triassic crisis, consistent with the re-appearance of large coal swamps. In the model, this early

Carnian expansion of the terrestrial organic C reservoir results in a drop of atmospheric pCO_2 and global temperatures. Furthermore, climate modelling shows how the resulting low pCO_2 conditions, coupled to the aridity of Pangaea, made the Carnian Earth system more susceptible to hydrological cycle enhancements following C inputs from coeval volcanism, thus explaining the nature of the Carnian Pluvial Episode.

1. Introduction

The Carnian witnessed major changes in marine and terrestrial ecosystems. Many marine groups experienced high extinction rates, while key herbivores on land disappeared (Dal Corso et al., 2020a). Concurrently, other groups—including dinosaurs, crocodiles, turtles, mammals, and modern conifer families on land, and Scleractinia coral reefs and dinoflagellates in the sea—rapidly diversified or first appeared, and formed new ecosystems (Dal Corso et al., 2020a). The biological turnover occurred at the time of the Carnian Pluvial Episode (CPE; Figure 1), a climate change event associated with negative C-isotope excursions (NCIEs; Figure 2), global warming, and the eruption of the Wrangellia large igneous province (LIP), a vast oceanic plateau (e.g., Dal Corso et al., 2020a, 2012; Hornung et al., 2007; Miller et al., 2017; Simms and Ruffell, 1989; Sun et al., 2016; Tomimatsu et al., 2021). The CPE was marked by a strong enhancement of the hydrological cycle, with evidence for generally more humid conditions recorded from Pangaea to deep water Panthalassa (Dal Corso et al., 2020a; Ruffell et al., 2016; Simms and Ruffell, 1989). As a consequence, continental runoff rose, rivers and lakes expanded, and the siliciclastic flux to marine depositional environments increased (Arche and López-Gómez, 2014; Dal Corso et al., 2020a, 2018; Klausen et al., 2019; Lu et al., 2021; Mancuso et al., 2020; Nakada et al., 2014; Ruffell et al., 2016; Simms and Ruffell, 1989).

Proxies give different atmospheric pCO_2 estimates for the Carnian (Figure 1). While all proxies give similar pCO_2 levels for the Early–earliest Late Triassic, pedogenic carbonate data overall suggest higher Carnian–Norian atmospheric pCO_2 values (ca. 3000 ppm) than stomatal index and phytane (<1000 ppm) (Figure 1; Foster et al., 2017; Retallack and Conde, 2020; Schaller et al., 2015; Witkowski et al., 2018). This discrepancy hinders mass-balance calculations of C fluxes during the CPE based on C-isotope data (Miller et al., 2017), and limits our understanding of the peculiar hydrological changes.

The CPE came ~18 Myrs after the Permo–Triassic mass extinction (PTME), the Phanerozoic's most severe biological crisis (e.g., Fan et al., 2020), and can be seen as the culmination of the slow recovery from this earlier catastrophe (Payne et al., 2004). An impressive example of this long recovery is the Early–Middle Triassic "coal gap" (Retallack et al., 1996) (Figure 1). Paleozoic vegetation collapsed at the PTME (e.g., Feng et al., 2020; Retallack et al., 1996), and for the following ~15 Myrs (i.e., until the end of the Middle Triassic) no significant coal deposits formed: Thin coal levels are found in the Middle Triassic, but thick and extensive coal measures do not re-appear until the Carnian (Retallack et al., 1996; Figure 1).

The Middle–early Late Triassic land plants' recovery could have played a central role in modulating global biogeochemical cycles (Figure 1). A positive Ladinian–Carnian CIE is recorded by terrestrial and marine organic matter, and marine carbonates (Figure 1 and 2), and has been previously interpreted as the effect of increasing terrestrial C burial linked to the Middle–Late Triassic re-expansion of coal swamps (Figure 1; Dal Corso et al., 2011; Korte et al., 2005). Similarly, Sr isotopes measured on brachiopod calcite and conodont apatite show a marked decline in the Middle Triassic (Figure 1), interpreted as the consequences of reducing continental weathering and transport of radiogenic Sr due to the recovery of vegetation, acting as a protective cover and stabilizing the land surface (Korte et al., 2003).

Here we use modelling to investigate the long-term biogeochemical changes in the PTME–early Late Triassic interval, and to understand their legacy on the short-term, volcanically-induced environmental perturbations of the CPE. Constraining the boundary environmental conditions created during the Middle–early Late Triassic is crucial because they determined the style of the CPE and, consequently, the direction that evolution took in the Carnian.

2. Methods

2.1 Geochemical data compilation

C and Sr isotope geochemical records ($\delta^{13}C$ and $^{87}Sr/^{86}Sr$), and atmospheric pCO_2 estimates from stomatal index, phytane, and palaeosol carbonates (Figure 1) were compiled from published studies. We expanded the ISOORG database (Nordt et al., 2016) for terrestrial $\delta^{13}C$ (plant remains and soil) including recently published data for the Triassic (see dataset in the Supplementary Materials). Brachiopod and bulk carbonate $\delta^{13}C$ are from (Korte et al., 2005). pCO_2 estimates are from stomatal index (Retallack and Conde, 2020), phytane (Witkowski et al., 2018) and palaeosol carbonates (compiled by Foster et al., 2017). We revised the ages of the compiled $\delta^{13}C$, $^{87}Sr/^{86}Sr$ and pCO_2 data according to the most recent International Chronostratigraphic Chart (2020) and up-to-date stratigraphic constraints. We grouped $\delta^{13}C$ and stomatal pCO_2 data in time bins corresponding to their ammonoid Zone/Subzone or Substage, depending on the available stratigraphic information (bio- and litho-stratigraphy, and absolute ages). The $\delta^{13}C$ data has been compiled with the aim of showing long-term geochemical trends, to the inevitable detriment of the brief anomalies that punctuate the studied interval, especially those of the Early Triassic (Payne et al., 2004). $\delta^{13}C$ and stomata pCO_2 data with revised ages are available in the Supplementary Materials, along with biostratigraphic information and references.

2.2 Model description

Our biogeochemical box model is based on a previous Hg-C cycle model which was applied to the PTME (Figure 3; Dal Corso et al., 2020b). The ocean is split into three boxes ('surface', 'high-latitude' and 'deep') in order to represent a thermohaline circulation and ocean–atmosphere exchange, and thus simulate perturbations over the geologically-short timescales relevant to rapid CO_2 release. In this paper the Hg cycle is switched off, but we add a Sr cycle, which is based on the single-box biogeochemical model COPSE (Figure 3; Lenton et al., 2018; Mills et al., 2014). All hydrosphere-crust fluxes follow COPSE exactly and we allow Sr to circulate based on its concentration and model water mass movement. Granite and basalt weathering, which contribute to the Sr composition of riverine input, are linked to overall silicate weathering by assuming a constant fraction of 'basaltic' silicate weathering at 30%, as in the present day, and sediment weathering is linked to carbonate weathering. Mantle input is linked to the model degassing rate, and seafloor weathering (sfw) is linked to the deep ocean Sr concentration. The degassing and emplacement of the Siberian Traps may well have had a complex short-term effect on the Sr cycle inputs and outputs, but we limit the analysis here to the clearer longer-term trends driven by erosion, overall silicate weathering and long-term degassing. C cycle equations and parameters are described in (Dal Corso et al., 2020b). Sr cycle equations and parameters can be found in the Supplementary Materials. Full model code is available at bjwmills.com.

As in Dal Corso et al. (2020b) we assume a pre-PTME $\delta^{13}C$ composition of seawater of around 4 ‰, which we achieve in the model by raising terrestrial productivity above the present-day steady state value, and setting the isotopic composition of weathered material to reflect burial of isotopically-heavy carbon over long timescales. We achieve a pre-PTME strontium isotope composition of seawater by altering the relative contribution of radiogenic and unradiogenic lithologies within reasonable bounds (Mills et al., 2019). We also increase the volcanic degassing rate above the present-day value to achieve a pre-PTME CO_2 level of around 1000 ppm, again following Dal Corso et al. (2020b) and consistent with long-term reconstructions of this rate (Mills et al., 2019). Model outputs include atmospheric pCO_2 and global average surface temperature, as well as the changes in carbon and strontium isotopes. It is these C and Sr isotopic changes which are chiefly used to evaluate the model behavior against the geological record, given that both the CO_2 and temperature records are more uncertain.

2.3 Model scenarios

Model simulations begin with the PTME experiment from (Dal Corso et al., 2020b) where a large CO_2 input from Siberian Traps was imposed on the model for 50 kyr, and a collapse of the terrestrial biosphere was represented by a permanent reduction in organic C burial as well as a 100-year pulse of organic C oxidation, which drove a ~50 kyr reduction in $\delta^{13}C$ with a superimposed NCIE at the time of the C oxidation pulse (Dal Corso et al., 2020b). In this work we retain the PTME event, but also consider how the simulation evolves over the following 20 Myrs.

We tested three end-member scenarios. In scenario 1, we consider only volcanic CO_2 inputs and the previously modelled soil oxidation event at the PTME (Dal Corso et al., 2020b), and assume terrestrial productivity remained constant over the studied interval, as well as the global erosion rates. In scenario 2, we also added changes to global erosion rates, assuming they increase sharply at the PTME and are then gradually reduced, but again maintain constant terrestrial productivity. Scenarios 1 and 2 are incomplete, as the geologic record suggests that a substantial reduction in terrestrial productivity likely occurred in the Early–Middle Triassic as evidenced by plant diversity estimates record and the “coal gap” (Figure 1). However, the extent and duration of terrestrial productivity loss throughout the post-PTME interval is not well understood, and scenarios 1 and 2 serve as the baseline to test the effect of changing terrestrial biosphere on the long-term geochemical changes. These scenarios are plotted in the SI.

In scenario 3, which represents our overall view of the system changes, a sustained collapse of the terrestrial biosphere is considered alongside changes in erosion rates, soil oxidation and carbon inputs, following previous interpretation of geochemical data (Figure 1; e.g., Korte et al., 2005, 2003). Recovery of terrestrial productivity is modelled in three different ways: a) high terrestrial productivity recovering at 247 Ma (beginning of the Anisian); b) gradual increase of terrestrial productivity during the Middle Triassic; c) high terrestrial productivity recovering at 237 Ma (Ladinian–Carnian boundary). Scenario 3 assumes that high erosion rates which followed the PTME return to pre-event values by the Carnian.

In all of these scenarios we do not consider the complex C cycle perturbations of the Early Triassic (Payne et al., 2004) as the aim of the study is to understand long-term vegetation recovery dynamics. Each scenario includes a further CO_2 release from the Wrangellia LIP at 234 Ma, which we model as a sequence of short pulses. Additionally, to explore the effects of the modelled Wrangellia CO_2 release on changes to continental runoff we used outputs from the Fast Ocean Atmosphere Model (FOAM) (Donnadieu et al., 2006; Goddéris et al., 2014).

3. Results and discussion

3.1 15 Myrs of low terrestrial productivity after the PTME

In the model, scenarios 1 (volcanic CO_2 + PTME soil oxidation) and 2 (volcanic CO_2 + PTME soil oxidation + erosion) fail to reproduce the geochemical data. Supplementary Figure S1 shows the model results for scenario 1, with a spike in CO_2 concentration at the PTME and a further succession of CO_2 spikes associated with Wrangellia. The PTME produces a transient increase in global temperature and $^{87}Sr/^{86}Sr$ ratios, and a decrease in carbonate $\delta^{13}C$. Wrangellia produces the same behavior on a smaller scale. However, the large long-term excursions in $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ between the events (PTME and CPE) are not captured. In Supplementary Figure S2, we add changes to global erosion rates (scenario 2), assuming they increase at the PTME and are then gradually reduced. This alteration drives large changes in the model. The PTME is now followed by a period of lower pCO_2 and global temperatures (due to enhanced weathering), alongside elevated $^{87}Sr/^{86}Sr$ and lower $\delta^{13}C$ (again due to enhanced weathering and burial of carbonates; Shields and Mills, 2017). The high erosion rate increases the magnitude of the NCIE at the PTME as the burial of carbonate is increased relative to the burial of organics. However, scenario 2 does not reproduce sustained low carbonate $\delta^{13}C$ during the Middle Triassic, despite being able to match the shape of the $^{87}Sr/^{86}Sr$ excursion (Supplementary Figure S2).

Figure 4 shows the combined scenario (scenario 3) where a sustained collapse of the terrestrial biosphere is considered alongside changes in erosion rates, soil oxidation and carbon inputs. Three paths for the terrestrial productivity recovery have been tested: a) Anisian recovery, b) gradual Middle Triassic recovery, c) Carnian recovery. A full recovery of terrestrial productivity in the Anisian drives a positive $\delta^{13}C$ shift in marine carbonates at 247 Ma coupled to a sudden lowering of $^{87}Sr/^{86}Sr$ that are not consistent with $\delta^{13}C$ data from brachiopod calcite and bulk carbonates, and with $^{87}Sr/^{86}Sr$ data from conodont apatite (Figure 4). A gradual recovery during the Middle Triassic drives equally gradual changes in $\delta^{13}C$ and $^{87}Sr/^{86}Sr$: in this scenario, while model $^{87}Sr/^{86}Sr$ values can reproduce the general trend of $^{87}Sr/^{86}Sr$ data from conodont apatite, model $\delta^{13}C$ fails to reproduce the low Middle Triassic $\delta^{13}C$ values recorded by marine carbonates and terrestrial organic matter (Figure 1 and 4).

A full recovery of terrestrial productivity in the Carnian—evidenced in the geologic record by the re-appearance of thick coal measures (Retallack et al., 1996)—is the scenario most consistent with proxy data (Figure 1 and 4). Reduced terrestrial productivity coupled to increased rates of erosion during ~15 Myrs after the PTME (Early–Middle Triassic) is consistent with reduced $\delta^{13}C$, and higher $^{87}Sr/^{86}Sr$ values due to further enhancement of terrestrial weathering when climate is warm. Due to limited terrestrial productivity, atmospheric pCO_2 levels and global temperature are increased during the Early–Middle Triassic relative to scenario 2. The magnitude of the strontium isotope excursion is not reproduced in any of these scenarios, but part of this isotopic change could be due to the system relaxing towards higher $^{87}Sr/^{86}Sr$ values following the initial weathering of the Siberian Traps, which delivered unradiogenic Sr.

An increase of terrestrial productivity in the Carnian causes a rise of $\delta^{13}C$ in the model shallow-ocean, which is in agreement with data from brachiopod calcite and bulk carbonate (Korte et al., 2005), and decreases ocean $^{87}Sr/^{86}Sr$ values, reproducing the overall trend shown by Sr isotope values measured in conodont apatite (Korte et al., 2003; Song et al., 2015). We note that scenario 3b (gradual Middle Triassic recovery), appears to slightly better match the $^{87}Sr/^{86}Sr$ record than scenario 3c (Carnian recovery), suggesting that terrestrial productivity must have indeed started to rise in the Anisian sufficiently to limit erosion rates and to produce thin coal layers (Figure 1): hence, the post-PTME Middle Triassic vegetation recovery, with the development of a protective terrestrial plants cover, likely had a gradual effect on continental weathering (Korte et al., 2003). However, model and proxy $\delta^{13}C$ strongly indicate that, in the Middle Triassic, C burial on land was negligible, and increased later in the Carnian, when thicker coal deposits formed (Korte et al., 2005; Retallack et al., 1996). In general, a full re-establishment of terrestrial productivity in the Carnian is consistent with $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ data, confirming previous interpretations of the geochemical records (Korte et al., 2005, 2003).

The model includes also pCO_2 and T, which can be compared to proxy data (Figure 4). The increase of terrestrial productivity at the beginning of the Carnian causes a drop of atmospheric pCO_2 (Figure 4). Lower model atmospheric pCO_2 in the Carnian is consistent with low-resolution values calculated from stomatal index and a few pedogenic carbonates (Figure 1). However, the model suggests that most of the Carnian atmospheric pCO_2 estimates from pedogenic carbonates could be overestimated, particularly when one considers the temperature changes suggested by conodont apatite oxygen isotopes, which do not follow the paleosol CO_2 trajectory. Interestingly, stomatal index and one pedogenic carbonate data give lower pCO_2 levels than model pCO_2 in the earliest Carnian possibly suggesting that terrestrial productivity might have been higher than before the PTME, as in the model a stronger return of the vegetation would further lower pCO_2.

Alongside a drop of pCO_2, model Carnian global average temperature decreases (Figure 4). A cooling is indeed also apparent in global temperature reconstructions (Scotese et al., 2021) and temperature drop around the Ladinian–Carnian boundary, followed by a generally cooler earliest Carnian, is suggested by O-isotope ($\delta^{18}O$) values of conodont apatite (Trotter et al., 2015) (Figure 4). However, the Early Triassic super-greenhouse temperatures are not captured by the model. This Early Triassic hothouse could have been the result of a breakdown of the silicate weathering thermostat (Kump, 2018), which we do not consider.

Summarizing, the model-data comparison supports the hypothesis that the Carnian saw the full re-establishment of a productive terrestrial biosphere, with the re-appearance of large coal swamps, and stabilization of erosion rates, ~15 Myrs after the PTME. This resulted in early Carnian lower atmospheric pCO_2 levels and temperature.

3.2 Wrangellia volcanism and the CPE

Within the modelled early Carnian background climate state (lower pCO_2 and T), we imposed CO_2 releases from Wrangellia LIP at 234 Ma as a sequence of short pulses. Wrangellia erupted during the Carnian in the middle of tropical Panthalassa (Greene et al., 2010; Tomimatsu et al., 2021). Its estimated volume of ca. 1 million Km^3 is based on outcrops found in today's northwest American continent (Dal Corso et al., 2020a; Greene et al., 2010). However, similar coeval intraplate basalts, possibly belonging to the same LIP, are found in Japan and Russia (Tomimatsu et al., 2021).

A complex perturbation of the C-cycle associated with the eruption of Wrangellia marks the CPE (Figure 2). Multiple NCIEs are recorded by carbonates and organic matter in many marine and terrestrial successions (Figure 2), at the peak of the longer-term positive late Ladinian–early Carnian $\delta^{13}C$ shift (e.g., Baranyi et al., 2019; Dal Corso et al., 2018; Miller et al., 2017; Sun et al., 2016; Tomimatsu et al., 2021). The NCIEs indicate injections of isotopically light C into the atmosphere–ocean system, and are linked to global warming and biological changes, and are coeval to the emplacement of the Wrangellia LIP (Dal Corso et al., 2020a, 2012; Sun et al., 2016; Tomimatsu et al., 2021). In the marine marginal successions of the

Western Tethys and lacustrine series of North China (Jiyuan Basin), sedimentary Hg concentrations and Hg/TOC ratios increase in coincidence with the NCIEs that mark the CPE (Figure 2), supporting the hypothesis of four discrete injections of volcanic-related CO_2 into the Carnian atmosphere–ocean system (Lu et al., 2021; Mazaheri-Johari et al., 2021).

Hence, on the basis of the C-isotope and Hg records, on the temporal link with Wrangellia, and on the estimated volume of Wrangellia, we modelled volcanic CO_2 injections as four equally-spaced pulses during the CPE interval, whose entire duration is estimated to have been ~1.2–1.6 Myrs (Bernardi et al., 2018; Lu et al., 2021; Miller et al., 2017; Zhang et al., 2015). The total amount of degassed volcanic CO_2 is set at 5000 GtC (each pulse = 1250 GtC), which was calculated from the minimum volume of Wrangellia basalts (Dal Corso et al., 2012), and the $\delta^{13}C$ of volcanic CO_2 is set at -5‰.

The model can reproduce multiple NCIEs during the CPE (Dal Corso et al., 2018), but each modelled negative $\delta^{13}C$ shift has a magnitude of ≤1‰, which is smaller than the magnitudes (>2‰) inferred by $\delta^{13}C$ measurements of marine bulk carbonates (e.g., Sun et al., 2016). Release of additional C from a more ^{13}C-depleted reservoir (e.g., ocean floor clathrates) or an isotopically lighter volcanic C—which we did not model in this study—would drive larger NCIEs as recorded in sediments (Figure 4).

The model predicts sharp increases (approximately a doubling) of atmospheric pCO_2 for each LIP pulse during the CPE (Figure 4). These modelled pCO_2 spikes are also coupled to increases of global mean surface temperature of about 5°C (Figure 4). Model temperature agrees with the CPE temperature estimates from $\delta^{18}O$ of conodont apatite, which does indicate an initial warming of 4°C at the onset of the CPE and a longer-term warming of 7°C in the late Carnian

(Hornung et al., 2007; Sun et al., 2016). However, data resolution is too low to show other shifts of conodont apatite $\delta^{18}O$.

To understand the effects of the volcanic CO_2 injections into a lower-pCO_2 atmosphere on changes to continental runoff, we plot outputs from the FOAM general circulation climate model (Figure 5), which have been previously compiled for the GEOCLIM carbon-climate model (Donnadieu et al., 2006; Goddéris et al., 2014). These simulations have been run for a range of different CO_2 concentrations and paleogeographies, and Figure 5A shows the relative continental runoff rates in FOAM for three continental configurations in the Late Permian, Anisian and early Norian when the difference between a high CO_2 (1400 ppm) and low CO_2 (560 ppm) run are considered. These are the closest time points in the FOAM data compilation to our Wrangellia scenario.

Two things become apparent from these outputs. Firstly, no matter what the continental configuration, the effects of an increase in pCO_2 on the volume of continental runoff are greater when starting from low pCO_2 levels (Figure 5A). This is because the effect of pCO_2 on global surface temperature (and thus on evaporation rates and eventually precipitation) is logarithmic. Secondly, Earth's continental configuration becomes more susceptible to changes in runoff as we move from 260 to 220 Ma (Figure 5B). This appears to be due to the denudation of Pangaea's equatorial mountain belt and associated decrease in global precipitation and runoff—i.e., the direct temperature-humidity effect on global runoff is more pronounced in the absence of major mountain belts, and when beginning from a more arid global climate.

Acknowledging that a specific model configuration for the Carnian is still required, FOAM outputs nevertheless suggest that a combination of lower background atmospheric pCO_2 and continuing amalgamation of Pangaea meant that CO_2 inputs from Wrangellia and consequent rises of atmospheric pCO_2 could result in large changes in the hydrological cycle, thus explaining the nature of the CPE.

Strong enhancement of the hydrological cycle is indeed a striking characteristic of the CPE (e.g., Dal Corso et al., 2020a; Simms and Ruffell, 1989): Evidences of a strong hydrological cycle are observed in the Carnian sedimentary records, from terrestrial successions of Pangaea to deep water sequences of Panthalassa (e.g., Arche and López-Gómez, 2014; Dal Corso et al., 2018; Mancuso et al., 2020; Nakada et al., 2014). This rise of rainfall, likely linked to an increase of Triassic megamonsoon activity (Zeng et al., 2019), led to the formation of large riverine and lake systems, the development of soils typical of tropical humid climates and hygrophytic vegetation, and an increase in continental runoff, with the deposition of thick siliciclastic units in the basins, this being evident especially along the margins of the western Tethys (e.g., Arche and López-Gómez, 2014; Dal Corso et al., 2018; Klausen et al., 2019; Roghi et al., 2010; Ruffell et al., 2016). In many sedimentary successions it is possible to untangle distinct "humid" pulses linked to the NCIEs, each marked by an increased loading of siliciclastic material in marine depositional settings (Dal Corso et al., 2018). Similarly, in the lacustrine succession of North China (Jiyuan Basin), the NCIEs are linked to spikes of nutrient input to the lake (Lu et al., 2021). The magnitude of hydrological cycle enhancement appears to be among the largest in the geologic record (see for example precipitation estimates in Retallack, 2009), and the scale of these environmental transformations likely contributed to the Carnian extinctions and radiations. Furthermore, a more humid climate could force the expansion of forests and freshwater environments (Dal Corso et al., 2020a), thus providing new ecological niches for the rise of terrestrial groups such as the dinosaurs (Bernardi et al., 2018), turtles (Reolid et al., 2018), and metoposaurids (Lucas, 2021).

4. Conclusions

Our modelling, tied to published geochemical data, indicates that the Carnian Pluvial Episode (CPE) occurred in a time of relatively lower background atmospheric pCO_2 levels and temperatures, and during a continental configuration which amplified changes to global runoff rates.

In particular, biogeochemical modelling shows that an exceptionally long vegetation recovery after the Permo–Triassic mass extinction and re-establishment of a highly-productive terrestrial biosphere in the Carnian is consistent with the long-term Early–early Late Triassic changes recorded by the $\delta^{13}C$ of marine carbonates and terrestrial organic matter, and by conodont $^{87}Sr/^{86}Sr$ values, confirming previous interpretations of the geochemical records. In the model, the increased burial of terrestrial C causes a drop of atmospheric pCO_2 and global average T.

Within this early Carnian Earth system state, inputs of volcanic C from the Wrangellia LIP—which are modelled as a series of four distinct pulses on the basis of the C-isotope and Hg data—result in discrete negative shifts in $\delta^{13}C$, as observed in several sedimentary records of the CPE, and synchronous sharp spikes (doublings) of atmospheric pCO_2. Outputs from the FOAM general circulation climate model indicate that low background pCO_2 levels of the early Late Triassic, alongside Pangea's continental configuration, very likely made the Carnian Earth system more prone to a strong enhancement of the hydrological cycle upon an increase of atmospheric pCO_2.

Hence, the specific evolution of long-term processes (terrestrial C burial and paleogeography) and the ways in which they tuned Earth's boundary conditions seem to be at the root of the Carnian Pluvial Episode, meaning that volcanic CO_2 input events from Wrangellia LIP could indeed drive massive changes in the Carnian hydrological cycle.

Declaration of Competing Interest

The authors declare that they have no known competing interests.

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Figure captions

Figure 1. Long-term Early – early Late Triassic key geochemical and biological changes. Early–Late Triassic plants' diversity (Rees, 2002) and combined average thickness of coal seams in eastern Australia (Chern, 2004; Retallack et al., 1996) are shown alongside a set of geochemical data. Conodont apatite $^{87}Sr/^{86}Sr$ (Korte et al., 2003; Song et al., 2015) and $\delta^{13}C$ records from terrestrial organic matter (expanded ISOORG database; (Nordt et al., 2016)) and marine carbonates (Korte et al., 2005) were interpreted as the effect of continental weathering, and vegetation collapse at the PTME and following recovery (Korte et al., 2005, 2003). Stomatal index (Retallack and Conde, 2020), phytane $\delta^{13}C$ (Witkowski et al., 2018) and pedogenic carbonates (compiled by (Foster et al., 2017)) give similar pCO_2 estimates for the Early–early Late Triassic, but some pedogenic carbonate data suggest higher Carnian–Norian atmospheric pCO_2 values. Temperature changes that marked the Early–early Late Triassic are inferred from the conodont apatite $\delta^{18}O$ record (Trotter et al., 2015).

Figure 2. Reference C-isotope and Hg records across the CPE. Records from Western Tethys, Panthalassa and continental Pangea show a complex perturbation of the C-cycle during the CPE (blue bars), with multiple negative C-isotope excursions recorded by terrestrial and marine organic matter and bulk carbonates. These excursions overlap with the emplacement of Wrangellia LIP. Records from the Northwestern Tethys and North China show increases of Hg concentrations in correspondence with the negative C-isotope excursions, suggesting that injections of isotopically light CO_2 into the Carnian atmosphere–ocean system were linked to pulses of Wrangellia LIP. Data are from Northwestern Tethys (Dal Corso et al., 2018; Mazaheri-

Johari et al., 2021); North China (Lu et al., 2021); Devon, UK (Miller et al., 2017); Guizhou, China (Sun et al., 2016); Japan (Tomimatsu et al., 2021).

Figure 3. C-Sr biogeochemical box model. Modelled species are represented in the atmosphere (a), surface ocean (s), high-latitude ocean (h) and deep ocean (d). Exchange between boxes via air–sea exchange, circulation and mixing shown as dashed arrows. Biogeochemical fluxes between the hydrosphere and continents/sediments are shown as solid arrows. A. Carbon cycle (Dal Corso et al., 2020b). B Strontium cycle.

Figure 4. Biogeochemical modelling results. Model is driven by changes in the input of volcanic CO_2 (A), a soil oxidation pulse at the PTME, a spike in erosion rates between the PTME and mid-Triassic and a long-lived collapse of the terrestrial vegetation between the PTME and Carnian (B). Model results for shallow ocean carbonate $\delta^{13}C$ (C), marine $^{87}Sr/^{86}Sr$ ratio (D), atmospheric CO_2 concentration (E) are plotted against the data from Figure 1. We superimposed to model global average surface temperature (F) the conodont $\delta^{18}O$ composite record of Trotter et al. (2015) to compare model and proxy trends.

Figure 5. Late Permian – Late Triassic Climate modelling outputs. A) Total global runoff for different topographies (panel B) and CO_2 concentrations. FOAM climate model output (Donnadieu et al., 2006; Goddéris et al., 2014) interpolated onto a standard grid of CO_2 concentrations (Mills et al., 2021). The 220 Ma continental configuration is the most sensitive to changing CO_2 concentration, and runoff enhancement is greater at low CO_2 concentrations. B) Outputs of FOAM climate model (Donnadieu et al., 2006; Goddéris et al., 2014) showing

changes in continental runoff for different continental configurations in the Late Permian, Anisian and Norian, the closest values in the FOAM data compilation to the Carnian scenario. Panels a–c show simplified topography, and panels d–f show modelled change in runoff (mm/yr) as CO_2 increases from 560 ppm to 1400 ppm. The potential change in runoff becomes larger as we move from Late Permian to Late Triassic.