The Late Triassic Extinction at the Norian/Rhaetian boundary: Biotic evidence and geochemical signature

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ABSTRACT

The late Triassic was an interval of prolonged biotic extinction culminating in the end-Triassic Extinction (ETE). The ETE is now associated with a perturbation of the global carbon cycle just before the end of the Triassic that has been attributed to the extensive volcanism of the Circum-Atlantic Magmatic Province (CAMP). However, we attribute the onset of declining latest Triassic diversity to an older perturbation of the carbon cycle (δ¹³Corg) of global extent at or very close to the Norian/Rhaetian boundary (NRB). The NRB appears to be the culmination of stepwise biotic turnovers that characterize the latest Triassic and includes global extinctions of significant marine and terrestrial fossil groups. These biotic events across the NRB have been largely under-appreciated, yet together with a coeval disturbance of the carbon cycle were pivotal in the history of the Late Triassic. Here, we present new and published δ¹³Corg data from widespread sections (Italy, Greece, ODP, Australia, New Zealand, USA, Canada). These sections document a previously unknown perturbation in the carbon cycle of global extent that spanned the NRB. The onset of stepwise Late Triassic extinctions coincides with carbon perturbation (δ¹³Corg) at the NRB, indicating that a combination of climatic and environmental changes impacted the biota at a global scale. The NRB event may have been triggered either by gas emissions from the eruption of a large igneous province pre-dating the NRB, by a bolide impact of significant size or by some alternative source of greenhouse gas emissions. As yet, it has not been possible to clearly determine which of these trigger scenarios was responsible; the evidence is insufficient to decisively identify the causal mechanism and merits further study.

1. Introduction

The Triassic Period is unique among the geologic periods of the Phanerozoic because it is the only period constrained by two of the largest declines in biodiversity of the Phanerozoic: 1) the end-Permian mass extinction, which is the most extensive biotic extinction of the

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Phanerozoic (e.g., Benton and Twitchett, 2003; Erwin, 2006); and 2) the end-Triassic/Jurassic biotic decline, culminating in the extinctions across the Triassic/Jurassic boundary (TJB; e.g., Hallam, 2002; Tanner et al., 2004; Richoz et al., 2007; Lucas and Tanner, 2018).

Environmental instability is characteristic of the Triassic. During the Late Triassic in particular (ca. 237–201.3 Ma), Pangaea experienced a variable climate regime (e.g., Preto et al., 2010; Rigo et al., 2012a; Trotter et al., 2015) that was interrupted by several global events including humid episodes (e.g., Carnian Pluvial Episode, Late Triassic of Simms and Ruffell, 1989, 1990; Simms et al., 1995) and warm climates. The warm cycles (W1, W2, and W3) are documented by δ¹³C values of conodont biogenic apatite (Rigo et al., 2012a; Trotter et al., 2015) and supported by PCO2 reconstruction based on stable isotopes of pedogenic carbonates from the Newark Basin in the USA (Knobbe and Schaller, 2018). Additional evidence of climatic perturbations at the NRB has been suggested recently by evidence of a major mega-monsoon climate in Australian paleolatitudes. In fact, Zeng et al. (2019) describe a mega-monsoon climate for the Northwest Shelf of Australia that reached its peak during the Carnian Pluvial Episode and weakened at the NRB, so that the climate during the Rhaetian became non-seasonal. This change in seasonality also was expressed by changes in vegetation as indicated by certain plant biomarkers from the same region (e.g., cadalene and retene: Cesar and Grice, 2019). These climatic perturbations may have resulted from extreme volcanic activity, with consequent impacts on Earth’s biota (e.g., Raup and Sepkoski, 1982; McElwain et al., 1999; Hallam, 2002; Marzoli et al., 2004; Rigo et al., 2007; Rigo and Joachimski, 2010; McRoberts et al., 2008; Lucas, 2010; Oggi, 2012; Dal Corso et al., 2014; Trotter et al., 2015).

Important clues to the evolution of ocean water chemistry, oxygenation, and productivity of past marine environments, including those of the Triassic, are recorded by changes in the isotopic composition of sediments (organic and inorganic). In particular, perturbations observed in δ¹³C values are widely applied and interpreted as indicators of paleoclimatic and paleoenvironmental changes (e.g., Hayes et al., 1999; Veizer et al., 1999; Payne et al., 2004; Ward et al., 2004; Korte et al., 2005; Lucas, 2010; Muttoni et al., 2004, 2014; Galli et al., 2005, 2007; Mazza et al., 2010; Preto et al., 2010; Whiteside and Ward, 2011; Zaffani et al., 2017, 2018). The stable carbon isotope record available for the Triassic gives a general overview of the evolution of δ¹³C, but its paleoclimatic and paleoceanographic interpretation is somewhat uncertain due to the multiple ecological and geological controls that can drive changes in the stable carbon isotope system.

From a broad perspective, a pronounced negative δ¹³C excursion across the Permian/Triassic boundary marks the beginning of the Triassic Period. This severe perturbation of the global carbon cycle is followed by large isotopic variability during the earliest Triassic (e.g., Ogg, 2012; Dal Servedio, 1989; 1990; Simms et al., 1995). This positive carbon isotope excursion, which corresponds to the virtual extinction of the bivalve Monotis around the NRB (Ward et al., 2004; Wignall et al., 2007; Whiteside and Ward, 2011; Rigo et al., 2016; Bertinelli et al., 2016; Zaffani et al., 2017). This positive excursion has been interpreted as the possible result of reduced circulation of ocean waters (Sephton et al., 2002; Ward et al., 2004; Wignall et al., 2007), and it is preceded by a negative shift (Maron et al., 2015; Rigo et al., 2016; Bertinelli et al., 2016; Zaffani et al., 2017).

Relatively few δ¹³Corg records are available for the Rhaetian (ca. 205.7–201.3 ± 0.2 Ma; Williford et al., 2007; Schoene et al., 2010; Wotzlaw et al., 2014; Maron et al., 2015; Rigo et al., 2016), a stage marked by significant faunal turnovers in both the marine and continental realms, including taxa such as ammonites (Guex et al., 2004; Whiteside and Ward, 2011), conodonts (Kouz and Mock, 1991; Giordano et al., 2010; Karádi et al., 2019), bivalves (McRoberts and Newton, 1995), radiolarians (Carver, 1993; Ward et al., 2001; Carver and Hori, 2005; O’Dogherty et al., 2010), the coral reef community (Fügell and Kiessling, 2003), dinoflagellates and foraminifers (Hesselbo et al., 2002), calcareous nannofossils (van de Schootbrugge et al., 2007; Preto et al., 2012, 2013), theropod dinosaurs (Olsen et al., 2002), and terrestrial plants (McElwain et al., 1999, 2009; Kürschner et al., 2007; Bonis et al., 2009; Cesar and Grice, 2019). The Rhaetian experienced a series of biotic crises and turnovers that culminated in the crisis at the Triassic/Jurassic boundary (TJB), which supports the hypothesis of a step-like extinction pattern for the end-Triassic mass extinction (e.g., Hallam, 2002; Tanner et al., 2004; Whiteside and Ward, 2011; Lucas and Tanner, 2018; Karádi et al., 2019).

Moreover, the latest Rhaetian is widely accepted to have been affected by the extensive eruptive activity of the CAMP, which is thought to have triggered three major negative carbon isotope excursions (CIEs): the ‘main’ CIE at the TJB, preceded by the late Rhaetian ‘initial’ and ‘precursor’ CIEs; the latter two are commonly associated with two different eruptive phases of the Moroccan CAMP (Marzoli et al., 2004; Hesselbo et al., 2002, 2007; Deenen et al., 2016; Zaffani et al., 2018). Although much attention has been focused on δ¹³C excursion across the TJB, much less is known about the background carbon isotopic conditions from the Norian (aside from the aforementioned North American section) to the early-middle Rhaetian and their possible links to the faunal extinction patterns and/or climate events documented at this time.

Herein we present evidence of a significant δ¹³Corg excursion at the NRB. A key section representing this excursion is the Pignola-Abriola section (GSSP candidate for the Rhaetian Stage) (Lagonegro Basin, Southern Apennines, Italy; Fig. 1), which is a sedimentary NRB succession deposited in the western portion of the Tethys Ocean (Bazzucchi et al., 2005; Rigo et al., 2016; Bertinelli et al., 2016). We compare carbon isotope data from this site with data from other sections located
at varying latitudes and from both hemispheres, and we associate the δ¹³Corg perturbations across the NRB with heightened faunal turnovers, including ammonoids, conodonts, radiolarians, bivalves, coral reefs, and terrestrial and marine vertebrates.

2. Geological setting of the studied sections

We have examined and correlated multiple stratigraphic sections situated in the two hemispheres, on opposite sides of Pangaea, and located at different latitudes by using chemo- and biostratigraphic constraints (Fig. 1).

In particular, we studied three NRB sections that crop out in the Southern Apennines (Italy) in the Lagonegro Basin, which is considered a branch of the western Tethys Ocean and part of the Ionian Sea (Ciarapica and Passeri, 1998; Stampelli et al., 2003) (Fig. 1). The Lagonegro Basin, the Upper Triassic is represented by the Calcari con Selce (i.e., Cherty Limestones) and the overlying Scisti Silicei (e.g., radiolarites) formations. The Calcari con Selce Formation consists of limestones with cherts in nodules and beds, intercalated with marls and siltstones (rare calcarenites), and rich in conodonts, radiolarians and thin-shelled pelagic bivalves (e.g., Halobia, Monotis) (e.g. Amodeo, 1999; Bazzucchi et al., 2005; Bertinelli et al., 2005, 2016; Giordano et al., 2010; Rigo et al., 2012b, 2016) (Fig. 1). The three sections, Pignola-Abriola, Mt. Volturino and Madonna del Sirino, document a well-exposed and fairly complete NRB, and they have been investigated from stratigraphic, sedimentological and biostratigraphic points of view (e.g. Amodeo, 1999; Bertinelli et al., 2005, 2016; Bazzucchi et al., 2005; Reggiani et al., 2005; Rigo et al., 2005, 2012a, 2016; Giordano et al., 2010, 2011) and also for magnetostratigraphy (Pignola-Abriola, Maron et al., 2015, 2019), geochemistry (Amodeo, 1999; Casacci et al., 2016) and chemostratigraphy (Trotter et al., 2015; Zaffani et al., 2017) (Fig. 1). Another section from the western Tethys Ocean is the Kastelli section (Pindos Zone) that is exposed in the Peloponnese (Greece) (Degnan and Robertson, 1998), and belonging to the tectono-stratigraphic terranes of the Hellenides (Brunn, 1956). This section consists of cherty limestones alternating with shale, sometimes in bivalve coquina beds, belonging to the genus Halobia (Degnan and Robertson, 1998) (Fig. 1).

Both the Lagonegro Basin and Pindos Zone were located in the Northern Hemisphere during the Late Triassic (Fig. 2). The Kennecott Point (British Columbia) and the New York Canyon (Nevada) sections were also located in the Northern Hemisphere during the Norian-Rhaetian time interval, but on the western side of Pangaea, facing the Panthalassa Ocean (Fig. 2). The Kennecott Point section (Queen Charlotte Islands, British Columbia, Canada) is represented by black calcareous siltstone and shale, rich in fossils (e.g. bivalves, radiolarians), sometimes in bivalve coquina beds, belonging to the Peril Formation (e.g. Ward et al., 2001, 2004), and deposited in a back-arc basin (Cameron and Tipper, 1985; Carter, 1993). The New York Canyon section was also deposited in a back-arc basin before being accreted onto the western United States (Nevada), and it is represented by the two facies of the Nun Mine Member and Mount Hyatt Member of the Gabbs Formation, which are the offshore slope facies in the shale-and-carbonate succession, and a combination of nearshore facies, both
above and below fair-weather wave base in the carbonate-and-shale unit (Laws, 1982; Tackett and Bottjer, 2016) (Fig. 1, 2).

The studied sections from the North-West Shelf of Australia (Wombat and Carnarvon Basins) and New Zealand (Murihiku Terrane) were instead located in the Southern Hemisphere (southern Tethys) at that time, although at different paleolatitudes, i.e. 25–30° S for the North-West Shelf, Australia (Exon and von Rad, 1994) and close to the South Pole for the New Zealand section (Seton et al., 2012) (Figs. 1, 3).

The Triassic strata of Hole 761C (core R33 to R26) from the Wombat Basin are dominated by black clayey siltstone at the base, and layers of calcareous sediments above, representing a proximal slope and evolving into an inner shelf to reefal carbonate margin. These cores of the Ocean Drilling Program (ODP) have been constrained biostratigraphically with dinoflagellate cysts, forams and nannofossils (Bralower et al., 1992; Brenner et al., 1992; Zaninetti et al., 1992; Gardin et al., 2012) (Figs. 1, 3). The Northern Carnarvon Basin received instead a thick fill of fluviodeltaic sediments during a long-term regression. The predominant lithology corresponds to claystone and sandstone, with siltstone and some coal interbedded in the Mungaroo Formation (Norian), whereas claystone becomes more dominant in the overlying Brigadier Formation (Rhaetian), where the depositional environment had a major estuarine influence. Both formations are dated by palynomorphs (Woodside Energy Ltd., 1977; Hocking et al., 1987; Longley et al., 2002) (Figs. 1, 3). The Kiritehere section crops out on the western Tasman Sea coast of the central North Island (New Zealand), and it consists of a volcanioclastic sedimentary rock succession with slumps dominated by fine sandstone and siltstone rich in bivalve coquinas (Arawi Shellbeds Formation), grading into thin siltstones and shales (Ngutunui Formation), rich in bivalves (Grant-Mackie, 1981, 2013). This succession is interpreted to represent a mid-shelf environment that accumulated in an elongate forearc basin marginal to an active, subduction-related volcanic arc that lay to the west of Kiritehere, along the edge of Gondwana (Grant-Mackie, 2013) (Figs. 1, 3).

3. Geochemistry – materials and methods

3.1. Methods for δ13Corg of bulk organic matter

Prior to acidification, all samples were washed in high-purity water and selected to avoid sampling of unrepresentative portions (e.g.,
fracture-filling mineralization, bioturbation, diageneric alteration). A few grams of each sample were reduced to a fine powder using a Retsch RM0 grinder or manually, using an agate mortar, and dried overnight at 40 °C. All the pulverized rock samples were then acid-washed with 10% HCl overnight (at least 12 h). Successively, the solution was discarded after centrifuging. Samples were then neutralized in deionized water, dried at 40 °C overnight and wrapped in tin capsules.

The δ13Corg analyses were undertaken using a Delta V Advantage mass spectrometer connected to a Flash HT Elemental Analyzer. For every set of analyses, multiple blank capsules and isotope standards (IAEA CH-6 = -10.45‰, IAEA CH-7 = -32.15‰) (Coplen et al., 2006) were included. Raw data were corrected for the blank contribution and then were calibrated against IAEA CH-6 and IAEA CH-7 following the two-point calibration method. The standard deviation of the in-house standard (δ13Corg = -26.00‰) during the period of analyses was better than 0.2‰.

The δ13Corg analyses refer to the δ13C composition of bulk organic matter. Organic matter can be made up of a number of components, such as bacteria, phytoplankton, zooplankton, pollen and/or other terrestrial biomass. Each of these components has a characteristic value of δ13Corg. As a result, changes in relative contributions of these components could affect the δ13Corg record, without necessarily requiring changes in the isotopic composition of the ocean and/or the atmosphere (Van de Schootbrugge et al., 2008; Fio et al., 2010; Bartolini et al., 2012). Nevertheless, if similar coeval trends occur at distant locations, the robust correlation between δ13Corg excursions can indicate regional, or, if more extensive, global interpretation. In fact, the amplitude and absolute values of coeval δ13Corg changes may be amplified or dampened by local environmental conditions and source control, i.e., the contribution of marine relative to terrestrial organic matter, and profound changes in the composition of standing biomass, i.e., the terrestrial floras or phytoplankton communities (Van de Schootbrugge et al., 2008). Despite regional differences in absolute value, the similarity between two or more coeval trends in the δ13Corg profile likely indicates a regional or global history (Bartolini et al., 2012).

3.2. Compound specific isotope analysis (CSIA)

The aromatic compounds were isolated by small-scale silica gel liquid chromatography, dissolved in n-hexane (see also Maslen et al., 2011), and analysed by compound specific isotopic analyses. A Thermo Scientific Trace GC Ultra connected to a Thermo Scientific Delta V Advantage irMS via a GC Isolink and Conflo IV was used for the stable carbon compound specific isotope analysis (CSIA) by gas chromatography – isotope-ratio mass spectrometry (GC-IRMS). For each fraction, 1 µL of solution was injected into a split-splitless injector that operates in splitless mode, held at 280 °C. The compounds were separated chromatographically on a wall-coated open tubular (WCOT) fused silica capillary column (60 m X 0.25 mm internal diameter) with a 0.25 µm 5% phenyl-methyl-silicon stationary phase (DB-5). The temperature of the GC oven was programmed from 40 to 325 °C (at 3 °C/min) and was held isothermally for 45 minutes at 325 °C. Compound identification was achieved by comparing the mass spectra and the available literature. Helium was used as the carrier gas at a constant flow of 1.5 mL/min. GC column outflow passed through the GC Isolink combustion reactor (copper oxide and nickel oxide, held at 1000 °C), which combusted each peak to a separate peak of CO2. The δ13C values are expressed in parts per mil (%) relative to the International Vienna PeeDee
Belemnite (VPDB) standard and were calculated by Thermo Isodat software from the integration of the 44, 45 and 46 mass ion currents. For every two sample measurements, a mixture of standards with known δ13C values was analysed in order to ensure instrument accuracy. Peaks co-eluting, as well as those at very low concentrations, were not considered for our interpretations. Only target compounds with a standard deviation of less than 0.4‰ were taken into account.

Kerogen and total organic carbon (TOC) δ13Corg values were measured in a Delambre-1 well using a 20/20 ion ratio mass spectrometer and Roboprep preparation system, manufactured by Europa PDZ, UK. Ground samples were placed in a tin capsule and dropped onto a furnace at 1000 °C while in an oxygen atmosphere. Combustion products were passed through a second furnace (600 °C) where excess oxygen was absorbed, and nitrogen oxide was reduced to nitrogen. The samples were also passed over magnesium perchlorate to trap the water. The system was calibrated using known δ13C international laboratory standards (Grice et al., 2005).

4. Results

4.1. Lagonegro Basin, Pignola-Abriola section

A total of 14 new samples were prepared and analyzed to augment the organic carbon δ13C set across the NRB. The results are consistent with the samples analysed previously for the same time interval, fitting the δ13Corg curve presented in Zaffani et al. (2017) and illustrated in Fig. 2. Values for the Mt. Volturino and Madonna del Sirino sections are those reported in Zaffani et al. (2017) (Fig. 2) (Table S1).

4.2. Pindos Zone (Peloponnese, Greece), Kastelli section

Only 30 of the 99 samples collected from this section provided a δ13Corg signal; the carbon content of the remaining samples was too low for carbon isotope analysis. Therefore, the resolution of the δ13Corg profile is lower than expected, but some trends can be outlined and may be related to the biorstratigraphic data. The δ13Corg curve depicts three trends corresponding to portions of the three perturbations (S1, S2 and S3) described by Zaffani et al. (2017). These are, in ascending stratigraphic order: i) a decrease through the conodont Mockina bidentata Zone below the FO of Miskella hersteini Zone (ca. 6‰) that was interpreted as equivalent to S1 (Zaffani et al., 2017); ii) a positive trend within the Miskella hersteini Zone, representing the return to positive values of the S2 recorded in the Lagonegro Basin (Zaffani et al., 2017); and iii) a negative shift starting in the upper M. hersteini Zone and intersecting the FO of Miskella posthersteini, corresponding to the S3 negative shift (Zaffani et al., 2017) (Fig. 2) (Table S1).

4.3. Nevada – USA, New York Canyon section

Eighteen shale samples were analysed to describe the variation of the δ13Corg across the NRB in the New York Canyon section, though we note that the position of the NRB in the New York Canyon section is not tightly constrained by current biorstratigraphy. The resulting δ13Corg profile shows a negative CIE coinciding with the negative δ13Corg/Sr/Sr shift documented within the Member Hyatt Upper member of the Gabbins Formation (Tackett et al., 2014). These trends are similar to those recorded in other coeval sections (Korte et al., 2003). In the lower part of the section, a general trend towards more positive δ13Corg values is documented; this trend ends with a δ13C value of ca. -2.5‰. Subsequently, a δ13C depletion of ca. 2.5‰ is recorded, with a successive return to more positive values (Fig. 2) (Table S1).

4.4. Wombat Basin (northwestern Australia), ODP SITE 761C

A total of 25 shale samples (and rare shaley marls) were collected during the ODP were analysed. Despite the low core recovery at Hole 761C, a decrease in the δ13Corg profile of ca. 3.5‰ at the Norian-Rhaetian transition was detected and was placed between 32R and 33R by the occurrence of the typical Rhaetian nanofossil Euconusphaera zambrachensis in core 32R Section 3 (Gardin et al., 2012) (Fig. 3) (Table S1).

4.5. Murighiko Terrane – Zealandia (New Zealand), Kiritehere section

The 34-metre thick Kiritehere section was studied in great detail. Sixty-eight samples of shales and thin siltstones from the upper part and bivalve coquinas from the lower part were collected and analysed for organic stable carbon isotopes (δ13Corg). In the lower part of the section, a severe negative shift of ca. 5‰ was detected, preceded by a short positive peak with respect to background values of ca. -28‰. The NRB is placed between the δ13Corg negative shift and the disappearance of the bivalve Monotis. It is noteworthy that the δ13Corg curve after the NRB perturbation follows rapid positive-negative oscillations (Fig. 3) (Table S1).

4.6. Dampier sub-Basin, Carnarvon Basin, North-West Australia

A core from the Delambre-1 well demonstrates a trend of depleted isotopic values for the kerogen and total organic carbon, just below the NRB (Fig. 3). The recovery follows a positive excursion, and subsequently, the isotopic profiles become more stable. We also investigated the δ13C of select polycyclic aromatic hydrocarbons (PAHs) such as phenanthrene and source-specific PAHs, including fluoranthene and benzo[fluoranthenes], which are combustion products from land plants (Jiang et al., 1998; Grice et al., 2007; Cesar and Grice, 2017). Phenanthrene shows depleted δ13C values (comparable with the δ13C of kerogen and TOC) at the NRB in the Delambre-1 well, even though this signal might be overprinted by multiple sources (marine and terrestrial) of this compound. However, if we consider source specific compounds, i.e., fluoranthene and benzo[fluoranthenes], the negative isotope excursion is more evident in a core from the North Rankin-5 well, immediately above the NRB. In the North Rankin-5 well, these isotopic trends are corroborated by a record of anoxia (isoprenoids ratio Pristane/Phytane < 1) (Fig. 3) (Table S1).

5. Discussion

5.1. Extinction record across the NRB

Here we review the nature of extinctions that took place across the NRB and during the early-middle Rhaetian. Most of these extinctions were long conflated as a single mass extinction at the end of the Triassic, the so-called ETE at the TJB (e.g., Sepkoski Jr., 1982, 1996; Olsen and Sues, 1986; Olsen et al., 1987; Kiessling et al., 1999, 2007; Tomašových and Sibílk, 2007; Wignall et al., 2007; Vazquez and Clapham, 2017). However, recent developments in marine (e.g., McRoberts, 2010; O’Dougherty et al., 2010; Orchard, 2010; Rigo et al., 2018) and non-marine biorstratigraphy, magnetostratigraphy (e.g., Muttoni et al., 2010; Kent et al., 2017; Maron et al., 2019), radioisotopic dating (e.g., Wotzlaw et al., 2014; Davies et al., 2017) and other dating methods (e.g. Maron et al., 2015) have allowed a much more detailed sequencing of biotic events during the NRB and TJB intervals than was possible 20 years ago (cf. Hallam, 2002; Tanner et al., 2004; Lucas and Tanner, 2008, 2015, 2018). These developments enable the identification of significant extinctions of marine pelagic biota across the NRB and during the Rhaetian, and the interpretation of tetrapod extinctions on land across the same boundary.

Nevertheless, we add the caveat that better stratigraphic resolution is needed for many of the taxa across the NRB. Thus, some of the extinctions at the NRB identified here are likely subject to the compiled correlation effect, meaning that there may be within-late Norian or within-Rhaetian events that are condensed into an apparent, single NRB
extinction. Thus, further work is needed to better resolve the timing of some biotic events across the NRB. Nonetheless, we do believe that there is a strong signal, and that there were substantial extinction events across the NRB, here referred to as the NRB extinction.

5.1.1. Marine extinctions across the NRB.

In the marine realm, there were striking and extensive extinctions among several biotic groups across the NRB, almost all of which were members of marine pelagic communities. Most striking, and long known, are the ammonoid extinctions. It has long been clear that the largest extinction of Late Triassic ammonoids took place at the end of the Norian, not at the end of the Triassic (Kummel, 1957; House, 1963; Kennedy, 1977; Newell, 1967; Teichert, 1988; Whiteside and Ward, 2011; Lucas, 2018a). After this extinction, only a few ammonoid taxa populated the Rhaetian seas, limited to heteromorphs and some Arcestaceae and Clydonictaceae (Wiedmann, 1973). The Late Triassic ammonoid extinctions were a succession of diversity drops, with the last, most substantial drop at the end of the Norian, not at the end of the Triassic. Thus, ammonoid extinctions across the TJB are best described as stepwise (Wiedmann and Kullman, 1996; Lucas, 2018a).

A compilation of ammonoid global diversity at the stage level indicates that, after a Norian (mostly Aluphan) peak in diversity, the most substantial extinction of ammonoid families and genera took place across the NRB (Lucas, 2018a). This is best documented in the New York Canyon area of Nevada, USA, where Taylor et al. (2000, 2001), Guex et al. (2002, 2003), and Lucas et al. (2007) plotted ammonoid distribution based on decades of collecting and study. Their work documents a two-phase latest Triassic ammonoid extinction, one extinction phase in the late Norian followed by a phase of low diversity Rhaetian ammonoid fauna that became extinct by the end of the Triassic (also see Lucas and Tanner, 2008, 2018). Thus, the ammonoid extinction across the NRB is profound and evident in both abundance data and stratigraphic ranges.

Recent analysis of marine bivalve diversity across the TJB, based on a generic compilation at the stage level, shows that generic diversity peaked during the Norian and was followed by a sharp drop into the Rhaetian and Hettangian (Ros, 2009; Ros and Echevarria, 2011; Ros et al., 2011, 2012). Extinction rates were high during the Rhaetian, and origination rates were low. This is consistent with more detailed studies (local and regional) of Late Triassic marine bivalve stratigraphic distributions (e.g., Allaisinaz, 1992; McRoberts, 1994; McRoberts and Newton, 1995; McRoberts et al., 1995; Wignall et al., 2007). The latter studies identified multiple and selective bivalve extinction events within the Norian, Rhaetian, and across the TJB, with a particularly important extinction at the NRB. This extinction of bivalves included the virtual disappearance of the cosmopolitan and abundant pectinacean Monotis (Dagys and Dagys, 1994; Hallam and Wignall, 1997) including the two dwarf Rhaetian species (McRoberts, 2007; Krystyn et al., 2007; McRoberts et al., 2008). McRoberts’ (2007, 2010) summaries of the Late Triassic diversity dynamics of ‘flat clams’ (halobiids and monotids) indicate these organisms suffered their largest extinction at the NRB. Allaisinaz (1992) also drew attention to the end-Norian turnover of megalodontid bivalves and concluded that the marine bivalve extinction at the NRB was larger than the extinction at the TJB, during which megalodontids decreased in size before their extinction (Todaro et al., 2017, 2018).

A major turnover in radiolarians took place across the TJB, and this has been widely considered as an important component of a marine mass extinction. Like the ammonoid and bivalve extinctions, this ‘mass extinction’ also has important components that precede the TJB. Global compilations indicate a substantial drop in radiolarian generic diversity from the Norian into the Rhaetian (e.g., O’Dogherty et al., 2010; Kiessling and Danelian, 2011). At the best-studied and most complete radiolarian record across the TJB (Queen Charlotte Islands in western Canada, recently renamed Haidi Gwaii), Carter (1993, 1994) established the Proparvicingula moniliformis Zone and the Globolaxator tosomi Zone to encompass the lower and upper Rhaetian radiolarian assemblages. Over half of the 160 radiolarian species present at the base of the P. moniliformis Zone disappear by the top of this zone; this is a substantial within-Rhaetian radiolarian extinction (Longridge et al., 2007). Biostratigraphic ranges of 156 radiolarian species in the Triassic deep-sea sections from Japan show a dramatic increase in extinction rates in the end-middle Norian following the Manicouagan impact event (Hodych and Dunning, 1992; Ramezani et al., 2005), and during the late Rhaetian (Onoue et al., 2016).

Conodonts also underwent substantial extinctions across the NRB. Micropaleontologists have long known that the Late Triassic witnessed a stepwise decline in conodont diversity as extinction rates were relatively high and origination rates were low (e.g., Clark, 1980, 1981, 1983; Sweet, 1988; Kozur and Mock, 1991; Aldridge and Smith, 1993; De Renzi et al., 1996). The single largest Late Triassic extinction of conodonts took place during the Carnian (at the Julian/Tuvaluan boundary) when nearly all platform conodonts disappeared (Rigo et al., 2007, 2018; Rigo and Joachimski, 2010). Conodont diversity recovered somewhat through the Norian to decline again into the Rhaetian. Within the Rhaetian, nearly all conodont taxa disappeared before the TJB, with only one or two taxa found in the youngest Rhaetian conodont assemblages (Mostler et al., 1978; Kozur and Mock, 1991; Orchard, 2003, 2010; Orchard et al., 2007; Bertinelli et al., 2016; Rigo et al., 2016; Du et al., 2020). Karádi et al. (2019) show Rhaetian conodont diversity dynamics as a stepwise extinction of four genera.

In the aftermath of the end-Permian extinctions, the Triassic was a time when the marine fish fauna changed from the chondrichthyan-rich faunas of the late Paleozoic to the actinopterygian-dominated fish faunas of the Mesozoic-Cenozoic (e.g., Romano et al., 2013). The Late Triassic saw the diversification of neopterygian actinopterygians, and the origin of durophagous feeders. A global compilation by Romano et al. (2013) suggests that marine fish generic diversity dropped from the Norian into the Rhaetian, but the records of Norian fish fossils from Lagerstätten and the absence of similar deposits from the Rhaetian (cf. Tintori and Lombardo, 2018) make it difficult to determine if the diversity drop is real or an artifact of preservation and hence collection bias.

Similarly, marine reptiles show an ‘evolutionary bottleneck’ across the NRB (Renoesto and Dalla Vecchia, 2018). This is represented by a diversity crash among the ichthyosaurs and the extinction of the thalattosaurs and the tanystropheids. However, like the fish record, the marine reptile record is affected by biases, particularly the general reduction of shallow marine platforms during the Rhaetian, which were the preferred habitats and sites of fossil preservation of the marine reptiles. Thus, the evolutionary bottleneck of marine reptiles across the NRB may in part reflect this bias (Renoesto and Dalla Vecchia, 2018).

The extinctions in the reef community at the end of the Triassic are best documented in the Tethys, where the reef ecosystem collapsed at the end of the Triassic; carbonate sedimentation nearly ceased, and earliest Jurassic reefal facies are rare (see Lucas and Tanner, 2018 and references cited therein). However, whether this was a global event remains unclear. Notably, Stanley et al. (2018) presented a compelling analysis that shows a reduction in coralite integration across the TJB, which is a morphological simplification in response to environmental stress similar to that seen in the ammonoids and radiolarians across the TJB (cf. Guex, 2016). Clearly, the reef crisis in the Tethys began at the end of the Norian when the scleractinian coral reefs (which produce planktotrophic larvae today) reached their peak of diversity, structural complexity and distribution, to diminish through the Rhaetian to a sudden collapse at the TJB (e.g., Kiessling et al., 1999; Kiessling, 2001; Flügel, 2002; Flügel and Kiessling, 2002). There was also an extinction of most microbial reefs and of algal reefs across the NRB (Flügel and Senowbari-Daryan, 2001).

In conclusion, significant extinctions took place in the marine realm across the NRB, particularly among ammonoids, bivalves, radiolarians and conodonts. Similar extinctions likely took place among
actinopterygian fishes, marine reptiles and in the reef community, though their records suffer from problems of taphonomy, facies and provinciality that make these extinctions more difficult to interpret.

5.1.2. Terrestrial extinctions across the NRB.

Evaluating terrestrial extinctions across the NRB has long been confounded by the difficulty of identifying and correlating the base of the Rhaetian in non-marine sections (e.g., Lucas and Tanner, 2007; Kozur and Weems, 2010; Lucas et al., 2012; Lucas and Tanner, 2015). Nevertheless, the extinctions that took place across the NRB in the terrestrial realm do not appear to have been extensive or sudden. Land plants, both palynomorphs and megaflora, show no significant extinction across the NRB (e.g., Barbacka et al., 2017; Lucas and Tanner, 2018; Kustatscher et al., 2018), though there were diversity crashes of local and regional extent. These are best seen in the palynological record in Western Europe and in the Newark Supergroup in eastern North America where many sessicate palynomorphs disappear during the late Norian or at the NRB (Lucas and Tanner, 2007, 2015 and references cited therein). However, the broad significance of these events remains to be demonstrated, and we are not able to correlate them to any turnover in the megaflora.

There is no evidence of terrestrial arthropod extinctions across the NRB. Indeed, diverse analyses of the fossil record of insects detect no evidence of a diversity crash during the Late Triassic or across the TJB (Clapham and Karr, 2012; Condamine et al., 2016; Grimaldi and Engel, 2005; Karr and Clapham, 2015; Labandeira, 2005; Labandeira and Sepkoski Jr., 1993). Grimaldi and Engel (2005, p. 73) concluded, “there seems to have been little differentiation between insect faunas of the Late Triassic and Early Jurassic”.

The record of terrestrial tetrapods does indicate some turnover across the NRB. Most of the large temnospondyl amphibians were extinct by the end of the Norian, with capitosaurids disappearing just before the NRB (Lucas, 2018b). Among reptiles, two groups of herbivores – the rhynchosaurians and dicynodonts – that had been significant components of Middle Triassic-Carnian tetrapod communities, became extinct late in the Norian (Spielmann et al., 2013; Racki and Lucas, 2018). The traversodontid cynodonts, a diverse group of Gondwanan synapsid herbivores, also disappeared at or just before the NRB (Abdala and Gaetano, 2018).

The classic concept of a TJB tetrapod extinction was largely predicated on the disappearance of the ‘thecodonts,’ subsequently referred to as the crurotarsans and more specifically (during the Late Triassic) encompassing the rauisuchians, aetosaurs and phytosaurs. Rauisuchians became extinct during the late Norian, aetosaurs were of low diversity after the NRB and became extinct during the Rhaetian or at the TJB, and phytosaur diversity crashed across the NRB although they apparently survived at low diversity across the TJB (Maisch and Kapitzke, 2010; Lucas and Heckert, 2011; Lucas and Tanner, 2015; Lucas, 2018b). Therefore, we can only interpret some turnover in the terrestrial tetrapods, and mostly late Norian extinctions or NRB diversity crashes. These are parts of a stepwise extinction of tetrapod taxa across the TJB that presaged the dinosaur-dominated terrestrial communities of the Jurassic-Cretaceous (Lucas, 2018b).

In brief, the evidence for a terrestrial extinction across the NRB is limited. Some tetrapod groups went extinct between the late Norian and late Rhaetian, and there are some local turnover events in the palynoflora record across the NRB. The significance and synchronicity of these events merits further study, particularly to calibrate their timing more precisely.

5.2. Correlation and magnitude of CIE

In the uppermost Norian sections, multiple $\delta^{13}$C$_{org}$ perturbations are associated with increases in bulk sediment TOC at locations ranging from the western Tethys (Rigo et al., 2016; Zaffani et al., 2017) to the Panthalassa Ocean (Ward et al., 2001).

Immediately below the NRB, which is defined by the first occurrence of the conodont Misikella posthernsteini (Krystyn, 2010; Rigo et al., 2016; Bertinelli et al., 2016; Zaffani et al., 2017), a negative $\delta^{13}$C$_{org}$ excursion of up to 5% occurs at three localities in the western Tethys...
(Pignola-Abriola, Mt Volturino and Madonna del Sirino sections in the Lagonegro Basin). Intra-basinal δ13Corg correlations rely on integrating biostratigraphy (conodonts, radiolarians) to obtain time-constrained successions for comparisons. As a result, the δ13Corg decrease across the Norian-Rhaetian transition is a feature identified as common to all studied sections of the Lagonegro Basin (Fig. 4).

It is noteworthy that the three Lagonegro Basin sections document the δ13Corg negative shift at the NRB from different lithologies (Zaffani et al., 2017), as the Lagonegro Basin was progressively falling below the CCD during the latest Triassic to Early Jurassic (e.g., Amodeo, 1999; Bertinelli et al., 2005; Rigo et al., 2012b), thereby demonstrating that the excursions are not a diagenetic artifact (Jiang et al., 2019). Moreover, the δ13Corg negative trend documented at the NRB at these three Lagonegro Basin localities was also identified in another Tethyan site, the Kastelli section, which crops out in the Pindos Zone, on the Peloponnesian Peninsula of Greece (Kafoussia et al., 2011) (Fig. 4), thus extending the correlation within the western Tethys. Of greater importance, this event is not limited to the Tethys, but occurs also in western Canada at Kennebec Point (British Columbia) (Ward et al., 2001, 2004; Whiteside and Ward, 2011) and at the New York Canyon section, Nevada (USA), both of which were located on the far side of the Panthalassa Ocean during the Late Triassic (Fig. 4).

This CIE is also now documented in Southern Hemisphere sections, including the Dampier sub-Basin (Northern Carnarvon Basin) on the Northwest Shelf of Australia (Grice et al., 2005, 2007; Cesar and Grice, 2017) and for the first time in the Wombat Basin (NW Australia, ODP Site 761C, Core 33R-32R-31R) and the Kiritehere section (North Island of New Zealand), the latter of which was located at a high latitude during the Late Triassic (Fig. 4). In the Northern Carnarvon Basin, the most depleted δ13C values of kerogen and TOC are recorded just below the NRB in the Delambre-1 well (Fig. 4), followed by more positive δ13C values. This positive δ13Corg peak was also identified in British Columbia, close to the near extinction of the bivalve Monotis (Ward et al., 2004; Wignall et al., 2007; Whiteside and Ward, 2011), and interpreted as the result of a slowing of ocean circulation with a consequent decrease in oxygenation (Sephton et al., 2002; Ward et al., 2004). The δ13Corg values were measured in select polycyclic aromatic hydrocarbons (PAHs) from the Northern Carnarvon Basin, such as phenanthrene (Delambre-1 well), and source-specific PAHs, such as fluoranthene and benzo-fluoranthenes (North Rankin-5 well). Importantly, these compounds are produced by the combustion of land plants (Jiang et al., 1998; Grice et al., 2007; Cesar and Grice, 2017), and both exhibit a negative CIE excursion at or immediately above the NRB (Fig. 4) as defined by palynomorphs, thereby correlating the events from the marine and terrestrial realms. The amplitude and values of the δ13Corg curves differ among the different basins due to variations in the type of organic matter, but the similarity of the δ13Corg curves suggests an event of global extent (Fig. 4).

Based on the profiles presented here (e.g., Lagonegro Basin), a short positive excursion from background δ13Corg values is documented at the base of the radiolarian Proparvicingula moniliformis Zone (uppermost Norian), followed by the rapid onset of a significant negative shift that lasted ca. 1 Myr (Maron et al., 2015), and ended close to the base of the Rhaetian (= LO of conodont M. posthersteinii) (Rigo et al., 2016; Bertinelli et al., 2016; Zaffani et al., 2017). This negative shift in δ13Corg is also associated with the virtual disappearance of the cosmopolitan, standard-sized pelagic bivalve Monotis (Ward et al., 2004) (Fig. 4). Following this negative shift, the δ13Corg curve recovers quickly as a positive peak to near background for ca. 300 k.y. (Pignola-Abriola section), or slightly more negative values (e.g., Southern Hemisphere sections). This is followed by a chaotic interval that is characterized by short oscillations in δ13Corg and the presence of small (also referred to as dwarf) species of the bivalve Monotis. At the base of this chaotic interval, the standard-sized Monotis disappears, but the perturbation of the system persisted until the stabilization that is observed at the FO of the conodont M. ultima (Zaffani et al., 2017) (Fig. 4).

5.3. Potential causes of the CIE

The worldwide low δ13Corg interval associated with a strong faunal turnover at the NRB likely resulted from multiple mechanisms that may have included some combination of decreased primary productivity, enhanced magmatic activity and outgassing (including pyrogenic volatiles), dissociation of clathrates, thermal degradation of peatland, and/or input of 12C from an extraterrestrial object, any of which may potentially create a global signal (e.g., Kent et al., 2003; Jenkins, 2010; Schaller et al., 2012; Meyers, 2014; Zaffani et al., 2017; Clapham and Renne, 2018). We thus propose that a large volume of 12C-depleted CO2 entered the ocean-atmosphere system just before the NRB (grey area in Fig. 4).

Although the impact of a large extraterrestrial object has not been proposed as a cause of the NRB extinction, such an event could have produced the negative shift of δ13Corg, reflecting consequent decreased primary productivity (D’hondt et al., 1998). The largest impact known for the Late Triassic formed the 90-km diameter Manicouagan structure (Spray et al., 2010), but repeated dating of the structure by various methods has consistently yielded middle Norian ages (Hodych and Dunning, 1992; Ramezani et al., 2005; van Soest et al., 2011; Clutson et al., 2018), nearly 10 million years prior to the NRB (Fig. 5). The radioisotopically constrained U-Pb age of 205.7 Ma for the NRB (Wotzlaw et al., 2014) is close to radiometric ages from melt rock at the Rochechouart impact crater (< 5 km diameter) in south-central France; Schmiedel et al. (2010) provided an 40Ar/39Ar age of 203 ± 2 Ma (recalculated to the decay constants of Renne et al., 2011), and Cohen et al. (2017) presented an 40Ar/39Ar age of 206.92 ± 0.20/0.32 Ma (Fig. 5). However, geochemical evidence of the iron meteorite impactor (Tagle et al., 2009) combined with the lack of carbonate-rich target rocks (Lambert, 2010) suggests that the volume of climatically active gases (e.g., CO2) released from the impact site would not have had an appreciable environmental effect that could have triggered the NRB extinction. Moreover, there have been no reports of impact debris (shock-metamorphosed quartz, melt-glass spherules or anomalous levels of Ir) associated with NRB sections. Hence, we discount the likelihood of a bolide impact as the driver of the NRB environmental event.

The concentration of marine extinctions at the NRB in the pelagic realm (planktonic, nektonic, benthic) suggests that the negative δ13Corg excursion records decreased productivity. Although the largest known impacts (e.g., Chicxulub) are considered capable of disrupting the trophic system by blocking photosynthetically active radiation (PAR), the Rochechouart structure was produced by an impactor nearly an order of magnitude too small to have been effective on a global scale. Thus, eruption of a LIP, with resultant outgassing, seems a far more likely candidate. As has been suggested for the TJB, large-scale volcanic outgassing is capable of producing short-term temperature reductions, due to the effects of sulfuric acid (H2SO4) aerosols formed from outgassed sulphur dioxide (SO2), followed by longer-term warming forced by increased atmospheric pCO2 (Tanner et al., 2004; Lucas and Tanner, 2018).

Previous workers have documented global-scale drops in both the 87Sr/86Sr and 187Os/188Os ratios in the earliest Rhaetian (Kuroda et al., 2010; Callegaro et al., 2012; Onoue et al., 2018). Such excursions result from changes in the sedimentary input to the ocean, either through reduction of continental input due to cooling and decreased precipitation, or through increased input from the weathering of mafic igneous sources (Callegaro et al., 2012). Notably, emplacement of a LIP is capable of producing both effects. Acid fallout of H2SO4 aerosols, and outgassed chlorides and fluorides could produce abrupt short-term decreases in the pH of surface waters, impacting planktonic autotrophs and the entire marine trophic system (e.g. Hönisch et al., 2012; Greene et al., 2012). Aside from slowing ocean circulation, subsequent warming also can affect the marine realm in the longer term through enhanced chemical weathering that leads to increased delivery of
nutrients to the ocean, with an increase in %TOC, as is seen in the Lagonegro Basin (Rigo et al., 2016; Zaffani et al., 2017) and in British Columbia (Ward et al., 2001, 2004).

Therefore, a possible trigger for the Norian-Rhaetian conditions could have been vast emissions of greenhouse gases during large-scale volcanic activity, such as the emplacement of a Large Igneous Province (LIP). Two LIPs documented during the latest Triassic–earliest Jurassic are the CAMP (the Central Atlantic Magmatic Province) and the volcanics of the Angayucham Terrane. The CAMP is represented by the subaerial tholeiitic basalts and intrusive bodies (ca. 1.5–3 million km³), emplaced in the central portion of the supercontinent Pangaea, straddling the palaeoequator (e.g., Marzoli et al., 2018). The NRB has been astronomically intercalibrated (Maron et al., 2015) and radioisotopically dated (Wotzlaw et al., 2014) at 205.7 Ma, while the oldest known radiometrically dated CAMP igneous rocks (Kakoulima layered mafic intrusion, Guinea) yield an age of ca. 201.63 Ma (Davies et al., 2017). Therefore, CAMP is ca. 4 Ma too young to have caused the CIE across the NRB (Fig. 5).

The other known Late Triassic LIP is the Angayucham oceanic plateau and oceanic islands complex accreted onto the Brooks Range (Pallister et al., 1989), cropping out in Alaska, with an estimated age of 214 ± 7 Ma (Ernst and Buchan, 2001; Prokoph et al., 2013). The estimated total volume of the Angayucham LIP, evaluated from the areal extent of ophiolite outcrops, ranges between 225 and 450 × 10³ km³ (Ernst and Buchan, 2001; Prokoph et al., 2013). Considering the uncertainty of the age of the Angayucham LIP and the estimated age of ca. 207 Ma for the beginning of both the δ¹³Corg and the stepwise extinction, the Angayucham LIP seems a potential candidate for triggering the events recorded across the NRB (Fig. 5). It is noteworthy that the exposed Alaskan rocks likely greatly under-represent the original volume of volcanic basalts as a consequence of the obduction of the Angayucham terrane onto the Brooks Range continental margin (Pallister et al., 1989). Moreover, the radiometric age of 214 ± 7 Ma (Ernst and Buchan, 2001; Prokoph et al., 2013) probably represents only a portion of the entire LIP, and younger material is either not preserved or has not yet been radioisotopically calibrated.

The inconsistencies in age estimates for the onset of the Angayucham basalts lead to two possible scenarios for the perturbation of the δ¹³C curve around the NRB. In the first scenario, an as-yet undocumented LIP body was emplaced at ca. 207 Ma (post-Angayucham, pre-CAMP), close to the base of the Rhaetian, during the Sevatian (late Norian). The onset of this volcanism played the role of the main trigger for the NRB climatic and geochemical perturbation. This scenario is compatible with the stepwise overturn of major Triassic marine groups such as conodonts, radiolarians, ammonoids, and cosmopolitan bivalves (Tanner et al., 2004; Lucas and Tanner, 2018), starting dramatically around the NRB and culminating just before the TJB (Hallam, 2002; Tanner et al., 2004; Wignall et al., 2007)(Fig. 5). In the second scenario, undocumented late Norian volcanic activity was part of the geodynamic evolution of Pangaea during the Late Triassic-Early Jurassic, which included Central Atlantic rifting previously documented from the late Norian (Cleveland et al., 2008). This scenario does not involve the Angayucham or CAMP basalts, other than the role of the latter in the late Rhaetian ETE, as documented by radioisotopic age determinations (Blackham et al., 2013).

Lastly, we note that destabilization of clathrates has been cited as a contributing factor toward the negative CIE associated with the ETE (e.g., Pálfy et al., 2001; Van de Schootbrugge et al., 2008), which could

Fig. 5. Schematic diagram of events occurred close to the Norian/Rhaetian boundary (NRB), including faunal and floristic turnovers, impact craters, Large Igneous Provinces (LIPs), and simplified δ¹³Corg curve of the latest Triassic – earliest Jurassic (not in scale).
have been triggered variously by volcanic activity or a bolide impact (Fig. 5). Tanner et al. (2004) noted the difficulties associated with this hypothesis, however, not least amongst them the lack of supporting evidence. Moreover, the later part of the Norian stage (Sevatian) is generally considered an interval of overall warm climate based on the δ18O of conodont apatite (Trotter et al., 2015) and other sedimentary evidence (Tanner, 2018), which calls into question the depth and mass of clathrate available in the deep ocean for dissociation. More significantly, the abrupt positive shift of δ18O values early in the Rhaetian indicates a cooling episode, contra-indicating the sudden release of methane. At present, we find no evidence to support this mechanism as the driver of the negative CIE at the NRB.

6. Conclusions

In summary, our data document the global extent of a substantial perturbation to the carbon cycle that spanned the NRB. The CIE described herein extended across the Panthalassa Ocean to both sides of the Pangean supercontinent and is recorded in both the Northern and Southern Hemispheres. The onset of the stepwise Late Triassic extinctions coincided with this NRB carbon perturbation, indicating that the combination of climate and environmental changes impacted the biota at this time. We suggest that the most likely proximal cause of the negative shift in δ13Corg at the NRB was a large-volume emission of greenhouse gases from a large-scale volcanic event pre-dating the NRB. As no LIP dated to this time is yet known, we acknowledge that alternative sources of greenhouse gas emissions, or other mechanisms of carbon-cycle disruption, are possible. Further investigation is needed to better understand the dynamics that induced the disturbance of the carbon cycle during the Late Triassic.

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Appendix A. Supplementary data

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Declaration of Competing Interests

Authors declare no competing interests;

Data materials availability

All data available in the main text or the supplementary materials.

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