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Title: Progressive development of ocean anoxia in the end-Permian pelagic Panthalassa

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Abstract

The end-Permian mass extinction (EPME) has been linked with the widespread development of oxygen-poor oceanic conditions. However, information on the spatial extent of anoxia in the Panthalassa super-ocean has been limited. This study reports oceanic redox records from a deep-sea chert succession (the Waiheke 1 section, WHK 1, New Zealand) that was located in southern mid-latitudes of Panthalassa. High-resolution carbon isotope ($\delta^{13}C$) correlation between Waiheke and the Permian-Triassic boundary (PTB) type section indicates that the EPME is recorded in a thin black claystone interbedded between siliceous mudstone beds at WHK 1. Pyrite-dominated enrichment in highly reactive iron, coupled with elevated U/Al and Mo/Al ratios, are prevalent through this black claystone bed and the overlying Permo-Triassic transition strata, suggesting the development of euxinic water column conditions. Similar redox variations across the EPME horizon have been reported from other Panthalassic deep-sea PTB sections. Comparison with these PTB sections indicates that euxinic conditions were widespread in low-latitude regions of the Panthalassan ocean, and

such conditions developed earlier than in mid-latitude settings, up to 100,000 years before the EPME. This suggests there was a gradual expansion of ocean anoxia from low to middle-high latitude regions during the Permo-Triassic transition. The extent of ocean anoxia resulted in a decrease in the seawater inventory of redox sensitive trace metals (e.g., Mo), which is evident in the earliest Triassic strata of the studied section and other PTB sections. Panthalassic anoxia during the EPME coincides with extreme climate warming and the associated effects (e.g., changes in ocean circulation, marine eutrophication intensified by terrestrial weathering) were likely critical triggers for ocean deoxygenation.

1. Introduction

The most significant biotic catastrophe of the Phanerozoic, the end-Permian mass extinction event (EPME), was associated with the widespread development of oceanic anoxia (e.g., Wignall and Twitchett, 1996; Isozaki, 1997). Previous research has demonstrated that many shallow water and deep oceanic regions of Late Permian and Early Triassic age experienced anoxic conditions (e.g., Wignall and Twitchett, 2002; Kaiho et al., 2016). Organic molecular fossils (biomarkers) of anaerobic green sulphur bacteria have shown that euxinia (anoxic, sulphidic conditions) extended into the photic zone in the paleo-Tethys (e.g., Grice et al., 2005, locality 3 in Figure 1A) and Panthalassan oceans (Hays et al., 2007, locality 5 in Figure 1A). Enrichments in pyrite and redox-sensitive trace elements have also been reported from PTB sections of paleo-Tethys (e.g., Lei et al., 2017, Xiang et al., 2020, Shen et al. 2016, Clarkson et al., 2016), pelagic Panthalassa (locality 2 in Figure 1A; Algeo et al., 2011; Takahashi et al., 2014; Onoue et al., 2021), and the North Pangean margin (e.g., locality 4 in Figure 1A; Grasby et al., 2009; Schobben et al., 2020), indicating the widespread development of anoxia/euxinia. The extent of oxygen-poor conditions may have caused a decrease in both seawater sulphate concentrations (Song et al., 2014; Schobben et al., 2017) and trace elements such as Mo (e.g., Takahashi et al., 2014; Xiang et al., 2020), due to enhanced pyrite burial and trace metal drawdown (Takahashi et al., 2014; Schobben et al., 2015). The extent of ocean anoxia also changed the U isotope ratio of seawater, as recorded in marine carbonate ($\delta^{238}U_{carbonate}$) (Brennecka et al., 2011, Lau et al., 2016; Zhang et al., 2018ab, 2020).

While anoxia was clearly widespread, there remains a significant absence of environmental records from the Panthalassa (Fig. 1A). Since most of the sedimentary record of Permian-Triassic Panthalassic seafloor has been subducted, limited material is available to allow reconstruction of redox conditions, particularly in terms of revealing temporal and spatial variability in oceanic redox during the EMPE. Previous research has, however, focussed on Panthalassic PTB sections from Japanese accretionary complexes. These formed in low latitude central Panthalassa (locality 2 in Figure 1A), based on measured paleo-latitudes of Middle Triassic chert from the Mino Belt of central Japan (cf. paleomagnetic studies by Ando et al., 2001 and Oda and Suzuki, 2000). It was at least one hundred million years before these sediments were accreted in the Middle to Late Jurassic (Mastuda and Isozaki, 1991; Takahashi et al., 2014). Since oceanic plates migrate at a rate of several cm/yr, this indicates that the Japanese PTB sections were originally thousands of kilometres from land.

According to reports from Japan (e.g., Isozaki, 2009, Muto et al., 2020), the EPME horizon occurs at the transition from Permian radiolarian chert to siliceous claystone and black claystone layers. Based on the concentrations of redox-sensitive trace elements (Mo, U and V) and framboidal pyrite populations in the PTB sections, a shift to oxygen-poor conditions during the EPME has been suggested, with the specific development of either anoxic through to euxinic bottom waters, or an oxygen minimum zone setting (Wignall et al., 2010; Algeo et al., 2011; Takahashi et al., 2014; Onoue et al., 2021).

Southern Panthalassa PTB sections have also been documented in the accretionary complex on the North Island of New Zealand (Spörli et al., 2007a; Hori et al., 2011; Grasby et al., 2021), where data from Lower Triassic chert suggests a paleolatitude of ca. 34°S (Kodama et al., 2007). The Permian-Triassic deposits accreted during the Late Triassic, suggesting the original depositional location was close to Gondwana (Spörli et al, 2007b; locality 1 in Figure 1A). Two sections from Arrow Rocks and Waiheke Island (Fig.1B) indicate that black, bedded chert and claystone occur around the EPME, similar to their occurrence in the Japanese sequences. Hori et al. (2007) measured trace element enrichments (Mo and U) in the Griesbachian black chert layers of the Arrow Rocks (Unit 2) and identified an anoxic water column event they termed “OAE α” (they described an “OAEβ” in the Dienerian). However, this PTB section has minor faults around the basal part of the black chert layers, making the continuous redox history in relation to the EPME and PTB unclear.

The PTB section on Waiheke Island (Waiheke section) has a continuous succession across the PTB, which has been dated using conodonts (Hori et al., 2011). Organic carbon isotope, sulphide sulphur isotope, and trace element profiles through this section suggest the development of anoxic-sulphidic water column conditions during the EPME (Hori et al., 2007; Takahashi et al., 2013; Grasby et al., 2021). Here, we present new, high-resolution paleoenvironmental records for the Waiheke section, incorporating multiple paleoredox proxies, including iron speciation and redox-sensitive trace element systematics. We aim to reveal how ocean anoxia developed prior to and after the EPME, as well as the associated changes in seawater chemistry. In addition, we present a compilation of pelagic Panthalassic PTB records to investigate the spatial and temporal development of redox conditions in the Permian-Triassic Panthalassa.

2. Geological setting

The Waiheke PTB section is an intertidal platform section in Island Bay, north-western Waiheke Island, North Island, New Zealand (Fig. 1B). It belongs to the Kiripaka Formation of the Waipapa composite terrain (Schofield, 1974, 1979), which is the accretionary complex forming part of the basement of New Zealand. The sedimentary sequence comprising the Waiheke section is approximately 10 m thick and divided into six litho-units (Unit A to F; Hori et al., 2011). This study focuses on the lowest part of these units (Unit A), in the Waiheke 1 section (abbreviated as “WHK 1” hereafter). It mainly consists of pale green tuffaceous claystone and mudstone towards the base, followed by black and yellowish-black chert with black claystone (Fig. 2).

The pale green siliceous and tuffaceous claystone and mudstone in the lower half of Unit A are composed of mm- to cm-scale alternations of siliceous microfossil enriched layers (radiolarians and sponges) and muddy layers (Fig. 3A). The black claystone consists of fine-grained clay materials and organic matter (Fig. 3B). Although siliceous microfossils are observed in the black claystone, fossils are rare and generally small (up to several μm in diameter). Black and black-yellow cherts in the upper part of WHK 1 also show mm- to cm-scale microfossil enriched layers and muddy layers (Figs 3C, 3D). The size of visible microfossils are similar to those from the pale green beds in the lower part of the study section. 88 single beds have been identified and labelled as Beds 1 to 88 (Hori et al., 2011; Takahashi et al., 2013). The conodont occurrences indicate that the section ranges from the uppermost Permian to the lowermost Triassic (Figs. 2 and 4; Hori et al., 2011). *Hindeodus* sp. and *Neogondolella carinata* were obtained from Bed 29 of the black-yellow chert. Above this horizon, *Ng. carinata* and *Sweetospathodus* cf. *kummeli* occur from Bed 40. These index fossils indicate that Bed 29 is Induan, and Bed 40 extends into the lower to upper part of Induan (uppermost Griesbachian or even the Dienerian). Ten previously published organic carbon isotope ($\delta^{13}C_{org}$) analyses demonstrate a negative excursion from Bed 14 to Bed 22, indicating the uppermost Permian carbon isotopic signal corresponding to the end-Permian mass extinction event (Hori et al., 2011). Based on this information, we place the Permian-Triassic transitional zone at this level. Here, we provide further $\delta^{13}C_{org}$ data and discuss the timescale issue in more detail.

3. Methods

3.1. Organic carbon isotope analysis

We examined organic carbon isotope ratios of bulk chert and claystone samples from 28 horizons of WHK 1, using samples with no visible veinlets or staining. After washing with Ellix® water, all samples were ground into fine powder using an agate mill. The powdered samples were decarbonated with 10% HCl. Organic carbon isotopic compositions and organic carbon contents were measured using an elemental analyser (Flash EA 1112) coupled with a Thermo-Finnigan Delta plus Advantage isotope ratio mass spectrometer at the Center for Advanced Marine Core Research, Kochi University, Japan. Isotopic measurements were repeated up to four times for each horizon to check reproducibility. The carbon contents were calculated using standard sulfanilamide, and the precision of organic carbon isotope ($\delta^{13}C_{org}$) analyses was better than 0.1‰. Organic carbon isotope measurements were calibrated using histidine and alanine obtained from Sugito Ltd. Lab. All isotopic results are reported in conventional delta (δ) notation, defined as per mil (‰) deviation from the Pee Dee belemnite (PDB) standard value. Here we use the average values calculated from repeat (up to 4 times) measurements of the same sample.

3.2. Major and trace element analysis

Major and trace elements were measured by inductively coupled plasma atomic emission spectrometry (ICP-AES; SPS3500, SII nanotechnology) and inductively coupled plasma mass spectrometry (ICP-MS; iCAPQc, Thermo scientific) at the Department of Earth and Planetary Science, University of Tokyo. For these analyses, 0.5 g of sample powder was treated with 4 mL of $HClO_3$ + HNO_3 (1:1 mixture) in PTFE beakers for 45 min at 120°C, followed by two treatments with 4 mL of $HClO_3$ + HF (1:1 mixture) for 30 min at 120°C. The residues from these acid treatments were re-dissolved with 5 mL HNO_3 and diluted to 1:2000 using ultra pure water. Indium (In) and bismuth (Bi) were used as internal standards for the ICP-MS analyses. Working standards were prepared from a series of SPEX Multi-Element Plasma Standards supplied by SPEX Industries (SPEX). Repeated analysis on the reference material (GSJ-JR-1; rhyolite reference material supplied by Geological Survey of

Japan) confirms that values are reproducible to within 1.8% for Al, 0.8% for Fe, 12% for U, and 9% for Mo, respectively (as relative standard deviation: RSD).

3.3. Normalization

To assess the enrichment or depletion of redox sensitive elements relative to typical detrital material, enrichment factors (X_{EF}) were calculated as follows (e.g., Tribovillard et al., 2006):

$$X/Al = X_{sample}\ (ppm)\ /Al_{sample}\ (wt\ \%) \qquad (1)$$

$$X_{EF} = (X_{sample}\ /\ Al_{sample})\ /\ (X_{PAAS}\ /\ Al_{PAAS}) \qquad (2)$$

Here, X_{sample}, Al_{sample}, X_{PAAS} and Al_{PAAS} are the weight concentrations of element X or Al in the sample or in PAAS (the post-Archean average shale; Tayler and McLennan, 1985). To compare our geochemical data with other PTB sections, Fe/Al, U/Al and Mo/Al ratios were also calculated and compared with literature data (Algeo et al., 2011; Takahashi et al., 2014).

3.4. Iron speciation

A sequential extraction targeting different operationally defined Fe pools, including carbonate ($Fe_{carbonate}$), ferric (oxyhydr)oxides (Fe_{OX}) and magnetite ($Fe_{magnetite}$), was performed on a split of 0.07–0.09 g of powdered sample (Poulton and Canfield, 2005; Poulton, 2021). $Fe_{carbonate}$ was first extracted using Na-acetate at pH 4.5 and 50 °C for 48 h, followed by Fe_{OX} using Na-dithionite for 2 hours at room temperature, and finally $Fe_{magnetite}$ was determined with an ammonium oxalate extraction for 6 h. Dissolved iron concentrations (for $Fe_{carbonate}$, Fe_{OX} and $Fe_{magnetite}$ pools) were measured by atomic absorption spectrometry (AAS; Thermo ice 3000 at the School of Earth and Environment, University of Leeds). Replicate extractions of an international reference material (WHIT; Alcott et al., 2021) gave a RSD of <5% for each Fe pool.

Pyrite iron (Fe_{pyrite}) was determined stoichiometrically based on Ag_2S precipitates formed after a boiling HCl and chromous chloride distillation (Canfield et al., 1986). All samples were tested for the presence of acid volatile sulphide (AVS; predominantly FeS) via an initial 6 M HCl distillation, but in all cases AVS was below detection (<0.001 wt %). Replicate pyrite extractions gave a RSD of <5%. The sum of these measured Fe pools defines the total concentration of highly reactive Fe (Fe_{HR}):

$$Fe_{HR} = Fe_{carbonate} + Fe_{OX} + Fe_{magnetite} + Fe_{pyrite} \qquad (1)$$

A 1 h 0.5 N HCl extraction was also conducted on ~0.5_g splits of powdered samples to assess for potential oxidation during weathering (see Matthews et al., 2017). This treatment targets poorly crystalline ferric oxide minerals ($Fe(III)_{HCl}$), such as ferrihydrite, in addition to reduced Fe phases such as Fe_{AVS} and potentially some ferrous carbonate/phosphate phases. Since Fe_{AVS} was not present in our samples (see above), this phase can be discounted. Thus, this extraction targets the remaining non-sulphidized Fe(II) minerals (termed $Fe(II)_{HCl}$), as well as $Fe(III)_{HCl}$. Poorly crystalline $Fe(III)_{HCl}$ minerals such as ferrihydrite are not stable on geological timescales, and hence the $Fe(III)_{HCl}$ pool can be considered to dominantly result from more recent weathering of reduced minerals (particularly pyrite). Thus, quantification of this pool provides a maximum estimate of the extent of post-depositional pyrite oxidation (e.g., Matthews et al., 2017).

The $Fe(II)_{HCl}$ pool was determined immediately by spectrophotometer using the ferrozine assay (Stookey, 1970). The total amount of Fe dissolved by the 0.5 N HCl extraction was determined following reduction of Fe(III) to Fe(II) using hydroxylamine hydrochloride, followed by analysis by spectrophotometer using the ferrozine assay. Subtraction of $Fe(II)_{HCl}$ from this Fe pool gives $Fe(III)_{HCl}$. Corrected Fe pools were calculated as:

$$\text{Corrected } Fe_{carbonate}\ (Fe^{*}_{carbonate}) = Fe_{carbonate} - Fe(III)_{HCl} \tag{2}$$

$$\text{Corrected } Fe_{pyrite}\ (Fe^{*}_{pyrite}) = Fe_{pyrite} + Fe(III)_{HCl} \tag{3}$$

4. Results

All geochemical data are shown in Table S1 and Figure 2.

4.1. Organic carbon isotopes

Organic carbon isotope ratios in WHK1 vary between −32.8‰ and −25.4‰, but with distinct trends up-section. Specifically, $\delta^{13}C_{org}$ values in Beds 1-7 vary between −27.5‰ and −25.4‰ (Fig. 2). Values then decrease to −28.6‰ in Bed 8, followed by an increase to-26.1‰ in Bed 14. The most pronounced decrease in $\delta^{13}C_{org}$ then occurs between beds 14 and 17, reaching the minimum value of -32.8 ‰. Low $\delta^{13}C_{org}$ values persist up to Bed 28, before a gradually increasing trend, interrupted by a low value in Bed 48 (-29.2‰), is established up to bed 50. Above Bed 50, $\delta^{13}C_{org}$ values are relatively stable around -26.9‰, with variability within 0.6‰.

4.2. Major and trace elements

Concentrations of Al and Fe range between 0.74–5.2 wt % and 0.26–6.4 wt %, respectively (Table S1). Beds 35 and 38 have low values of 0.74 and 0.75 wt % for Al, and 0.42 and 0.40 wt % for Fe, respectively. The Al content is relatively high in the tuffaceous claystone and mudstone (3.3-5.2 wt %) and the black claystone (1.7-4.2 wt %), but is low in chert beds (0.9-2.7 wt %). Iron is high in the black claystone (more than 4 wt %) and the black chert of Beds 29 (6.4 wt %).

Concentrations of U and Mo vary between 1.0-39.1 ppm and 0.8-194.3 ppm, respectively. Normalized U/Al ratios (Fig. 2) and U_{EF} (Table S1) values range between 0.84-13.5 and 2.7-43.6, respectively. Notably, high U/Al ratios (>5) occur in Beds 17-35 and Bed 44. Mo/Al (Fig. 2) and Mo_{EF} (Table S1) vary between 0.3-61.1 and 2.4-611.0, respectively, with particularly high Mo/Al ratios (>5) occurring in Beds 17-24 and Bed 29.

4.3. Iron speciation

Fe speciation profiles are shown in Figure 2. Relatively high concentrations of $Fe(III)_{HCl}$ (more than 0.7 wt %) were detected in the black claystone samples of Beds 22, 24 and 48, as well as the black chert of Beds 23 and 29, which for these samples has a significant effect on corrected $Fe_{carbonate}$ and Fe_{pyrite} concentrations. However, this correction for oxidative weathering of pyrite brings the Fe speciation data more broadly in line with the trends evident in Mo/Al ratios (Fig. 2). Indeed, both Fe_{pyrite}/Fe_{HR} and Mo/Al reflect the availability of sulphide in the water column and during early diagenesis, and as such, the general similarity in the position of peaks in Fe^*_{pyrite}/Fe_{HR} and Mo/Al (Fig. 2) provides strong support that our approach is robust.

We distinguish the Fe speciation results for Beds 31, 35 and 50 (open circles in Figure 2) from the remainder of the data, as these samples contain less than 0.5 wt % total Fe, and as such should be viewed with caution when interpreting Fe speciation data (Clarkson et al., 2016). We thus do not discuss the Fe speciation characteristics of these samples further. For the remaining samples, Fe_{HR}/Fe_{total} ratios range from 0.15 to 1.0. Fe_{HR}/Fe_{total} ratios are higher than 0.38, which represents the conventional boundary for distinguishing samples deposited from an oxic or anoxic water column (Raiswell and Canfield, 1998; Poulton and Canfield, 2011), across most horizons, with the exception of Beds 46 and 58. Fe^*_{pyrite}/Fe_{HR} ratios range from 0.11 to 0.88, with low values (less than 0.22) in the siliceous siltstone and claystone beds of the lower part of the section (Beds 1, 6, 7, 8, and 13). Fe^*_{pyrite}/Fe_{HR} ratios exceed 0.6 in the black claystone and chert beds in the middle and upper parts of the section, except for Beds 20 and 28 which show moderate values (0.48 and 0.42, respectively).

5. Discussion

5.1. Carbon isotope correlation

The Upper Permian strata of WHK1 exhibits a two-step decrease in $\delta^{13}C_{org}$ values, resulting in an overall decrease of ca. 8‰. The first decrease occurs in Beds 7-8, followed by a larger decrease in Beds 14-17 (Figs. 2 and 4). The minimum $\delta^{13}C_{org}$ value occurs in the thin black claystone bed (Bed 17) sandwiched between siliceous claystone beds. Then, $\delta^{13}C_{org}$ shows an increasing trend towards the overlying Lower Triassic black chert beds.

Inorganic carbonate carbon isotope ($\delta^{13}C_{carb}$) values from low-latitude carbonate PTB sections show similar trends. One of the best documented examples is from the Meishan section, the type section for the Permian-Triassic boundary (Fig. 4; GSSP: Global strata-type section and point; Yin et al., 2001). The Meishan section also displays a two-step decrease in $\delta^{13}C_{carb}$ in the uppermost Permian strata below the base of the Triassic (Beds 23-25 in Fig 4; Jin et al., 2000; Cao et al., 2002; Kaiho et al., 2009). We label these two decreases in $\delta^{13}C$ as "the first $\delta^{13}C$ decrease" and "the second $\delta^{13}C$ decrease", respectively (marked by green and blue horizontal bars in Figure 4).

Based on U-Pb ages from zircon minerals in volcanic ash beds at the GSSP, these two isotopic steps are estimated to have occurred at 251.999 ± 0.039 Ma and 251.941 ± 0.037 Ma, respectively (Burgess et al., 2014), suggesting a 58 kyr separation. The first and second decreases are correlated with the two negative drops in $\delta^{13}C_{org}$ observed at WHK 1 (Fig. 4). The second decrease in $\delta^{13}C_{carb}$ at Meishan corresponds to the EPME horizon, and is associated with the loss of many marine taxa (Jin et al., 2000; Song et al., 2012). The Meishan section also shows a gradually increasing trend in $\delta^{13}C_{carb}$ above the EPME horizon up to the Permian-Triassic boundary and into the overlying lowest Triassic strata, similar to the increasing trend seen in the strata of Beds 17-28 below the first Triassic conodont fossil occurrence from Bed 29 in WHK 1(Figs. 2 and 4). This increasing $\delta^{13}C$ trend above the EPME horizon provides a further correlative interval between PTB sections (marked by the pink horizontal bar in Figure 4). Therefore, based on this carbon isotope correlation within the biostratigraphic framework for the Permian-Triassic transition, defined at the GSSP Meishan section, we place the EPME horizon at the base of Bed 17 in WHK 1, which is a thin black claystone bed within the uppermost pale-green siliceous mudstone beds (Figs. 2

and 4). Another deep-sea PTB section from New Zealand (Arrow Rocks) also has a published $\delta^{13}C_{org}$ profile across the PTB (Fig. 4; Hori et al., 2007). This PTB section also shows decreasing trends in $\delta^{13}C_{org}$ between a chert bed with the Changhsingian *Neogondolella prechangxingensis* in the lower part of the section (at -7 m, not shown in the Figure 4) and a Triassic black chert beds which yield conodonts of *Neogondolella carinata* and *Hindeodus* sp. of Triassic affinity (Yamakita et al., 2007). As the number of $\delta^{13}C$ data from the lower half of the section are limited, the exact positions of the first and second $\delta^{13}C$ decreases and their minimum troughs are uncertain. However, the $\delta^{13}C_{org}$ profile in the Permian chert and siliceous mudstone beds shows a decrease at -0.6 to -0.3 m, and then a further decrease across the fault dividing a siliceous claystone unit (Unit 2a; Hori et al., 2007) and overlying black chert unit (Unit 2b; Hori et al., 2007). These $\delta^{13}C$ trends are interpreted to correspond to the first $\delta^{13}C$ decrease and the negative trough above the second $\delta^{13}C$ decrease observed in the WHK 1 section (Fig. 4).

The two-step $\delta^{13}C_{org}$ decrease during the Permian-Triassic transition, followed by a gradual increase toward the PTB, are also observed in the low latitude, deep-sea PTB sections of the Canadian Ursula Creek section (Wang et al., 1994), as well as the Japanese Akkamori-2 (Takahashi et al., 2010; Fig. 4), Ubara (Kaiho et al., 2012, Fig. 4) and NF1212C (Sano et al., 2012) sections. In the Japanese PTB sections, located in central Panthalassa, black claystone commonly occurs instead of siliceous claystone and bedded chert during the second $\delta^{13}C$ decrease. This lithologic change reflects a significant decrease in siliceous microfossils (Takahashi et al., 2009). The black claystone facies in the Japanese PTB section continues during the Olenekian period, over a million years later (Takahashi et al., 2009). By contrast, WHK 1 has bedded chert beds just above the EPME horizon. The dominant black claystone facies of WHK 1 appears above these chert beds, with a return to chert beds below the PTB. This discrepancy in the timing of bedded chert deposition indicates a relatively short duration for the radiolarian chert gap in the southern mid-latitudes of Panthalassa.

5.2. Redox evaluation inferred from Fe speciation and trace elements in the Waiheke 1 section

5.2.1. Fe speciation systematics

Fe-speciation is a widely used proxy for reconstructing regional water column redox conditions (Poulton and Canfield, 2011; Poulton, 2021). Enrichments in Fe_{HR} (Fe_{HR}/Fe_{total} > 0.38) commonly characterise deposition under anoxic water column conditions, and for samples that show such enrichment, the extent of pyritization of the Fe_{HR} pool provides an indication of whether the water column was ferruginous ($Fe_{pyrite}/Fe_{HR} < 0.6$-$0.8$) or euxinic ($Fe_{pyrite}/Fe_{HR} > 0.6$-$0.8$). The samples that record a euxinic water column obtain syngenetic pyrite from the water column, in addition to possible continued pyritization of Fe_{HR} during diagenesis. The threshold values for Fe_{HR}/Fe_{total} and Fe_{pyrite}/Fe_{HR} are not absolute, and may be influenced by high sedimentation rates (particularly during turbidite deposition; e.g., Canfield et al., 1996) or transfer of Fe_{HR} to clay minerals (such as glauconite) during diagenesis, the latter of which may be particularly prevalent in organic-rich, non-sulphidic sediments (e.g., Poulton and Raiswell, 2002; Poulton et al., 2010). However, we found no evidence for glauconite in the samples from WHK 1, based on thin section observations, which would be consistent with the relatively high pyrite contents of many of the samples. Regarding sedimentation rates, deep sea cherts in general have slow sedimentation rates (order of mm/kyr; e.g., Mastuda and Isozaki, 1991). Indeed, the calculated sedimentation rate for WHK 1 (3.8 mm/kyr) roughly corresponds to 0.001 g cm^{-2}/yr (= 0.38 cm/kyr × 2.7g/cm^3), based on dating for Beds 7 and 17 (see above), which would not be expected to impact Fe speciation systematics. However, the tuffaceous sedimentary layers in this interval may potentially have masked Fe_{HR} enrichments due to rapid sedimentation (see Poulton et al., 2004).

5.2.2. Uranium and molybdenum systematics

Uranium and Mo exist as soluble oxyanions in oxic seawater (e.g., Tribovillard et al., 2006), with a relatively homogenous distribution because of their longer residence times relative to the mixing time of the ocean. Under reducing conditions, U transforms to tetravalent U(IV), whereas in the presence of dissolved sulphide, Mo converts to thiomolybdate. This results in enrichments in these elements, with Mo removal primarily occurring under euxinic conditions (but a degree of drawdown can also occur in open system sulphidic porewaters), whereas U uptake occurs within anoxic sediments (McManus et al., 2006). Based on observations from modern environments, Mo enrichments may begin to occur when transformation to thiomolybdates occurs (>11μM H_2S; Helz et al., 2011). Scott and Lyons (2012) suggest that sediment Mo concentrations >25 ppm commonly indicate at least intermittent water column euxinia, whereas concentrations >100 ppm likely indicate a persistent, strongly euxinic water column. Bennett and Canfield (2020) recommend using U/Al and Mo/Al ratios as the most objective normalization approach for minimising influence from detrital inputs. Furthermore, based on a compilation of modern settings, Bennett and Canfield (2020) suggest that euxinic conditions are indicated when U/Al and Mo/Al ratios exceed ~5 (μg/g), and oxygen-poor sea floor sediments ($O_2 \geq 10$ μM) are indicated when U/Al ratios are >1 (μg/g).

Consistent with the emerging perspective that local baseline values for redox proxy thresholds should be assigned wherever possible (Algeo and Li, 2020; Poulton, 2021), we apply different threshold values to the very different lithologies that comprise the upper cherty (partly muddy) strata (Beds 19-88) and the lower tuffaceous strata (Beds 1-18) (see Figure 2). For the upper cherty sediments, two samples (Beds 46 and 58) that give a clear oxic signal based on all of our applied redox proxies (with Fe_{HR}/Fe_{total} <0.22, low U/Al (0.90 and 0.84; average = 0.87) and low Mo/Al (0.93 and 0.77; average = 0.83), can be used to define the oxic baseline for these parameters. By contrast, the tuffaceous strata in the lower part of WHK 1 would have a different baseline value, due to dilution by tuffaceous materials with a very different chemical composition, as well as the potential for masking of enrichments due to rapid sedimentation (see above). Minimum U/Al and Mo/Al ratios are observed in Beds 1 and 6, respectively. We use these beds to define oxic baseline ratios of 0.84 for U/Al, and 0.30 for Mo/Al. Although the elemental composition of volcanic ash may depend on multiple factors (including magma source and crystallization differentiation), U is commonly enriched in tuff, whereas Mo is relatively depleted (Zielinski, 1983; Pichavant et al., 1987). For example, the Macusani tuff in southeast Peru has a U/Al ratio of at least 0.79, and a Mo/Al ratio of 0.03 (Pichavant et al., 1987). Furthermore, Bennet and Canfield (2020) show that oxic marine sediments have U/Al ratios in the range of 0.2-1.0, while Mo/Al ratios are in the range of 0.08-0.4. Thus, our defined oxic baseline for U/Al is comparable to these values, whereas our Mo/Al oxic baseline value is only slightly higher than these values. In addition, our approach to defining local redox thresholds provides a high level of consistency when interpreting the different redox proxies we apply. Thus, taken together, these observations provide strong support that our approach is robust.

5.2.3. Redox evaluation of the Waiheke 1 section

The lower part of WHK 1 (Beds 1-15) shows slightly elevated Fe_{HR}/Fe_{total} and U/Al ratios, both of which show an overall progressive increase through the Upper Permian strata (Fig. 2). This consistent behaviour between two independent proxies suggests a progressive decline of bottom-water oxygen levels through this interval. Low Fe^*_{pyrite}/Fe_{HR} and Mo/Al ratios indicate that the water column did not become euxinic at this time. However, during deposition of the thin black claystone bed at the EPME horizon (Bed 17), enrichments in Fe_{HR}/Fe_{total} and U/Al, combined with high Fe^{*}_{pyrite}/Fe_{HR} and Mo/Al (Fig. 2), strongly support the onset of euxinia. This redox state was dominant until deposition of the earliest Triassic bedded cherts, which contain the first occurrence of Triassic conodont fossils (Bed 29), although two intervals of low Fe^{*}_{pyrite}/Fe_{HR} and Mo/Al in Beds 19 and 28, combined with elevated Fe_{HR}/Fe_{total} and U/Al, suggest intermittent development of ferruginous conditions (Fig. 2).

Above these strata, up until around Bed 44, Fe_{HR}/Fe_{total} and U/Al ratios remain high (although U/Al ratios are somewhat lower in this interval, relative to the underlying euxinic sediments; Fig. 2), suggesting persistent anoxia (as discussed above, the low Fe_{total} data shown by open circles are not considered reliable). The Fe^{*}_{pyrite}/Fe_{HR} ratios are elevated, suggesting water column euxinia, while Mo/Al ratios are also slightly elevated, but at a lower level than in the underlying euxinic strata. On a local scale, a reduction in the extent of Mo drawdown could simply be a consequence of less intense euxinia in the water column (Nägler et al., 2011). However, since U also shows a reduction in the extent of drawdown, and since U responds to anoxia rather than specifically euxinia, less intense euxinia does not appear to provide a valid interpretation for the muted trace metal enrichments we observe. Alternatively, however, if euxinia was widespread, then the muted enrichments we observe under euxinic conditions would more likely reflect a significant decrease in the seawater trace metal inventory (e.g., Algeo, 2004; Takahashi et al., 2014; Goldberg et al., 2016), which we consider in more detail below.

Above Bed 44, the limited available data show evidence for fluctuating oxic and anoxic redox conditions, with the possible development of euxinic conditions in some horizons (e.g., Bed 71). When the redox state appears to have been euxinic (i.e., Bed 71), low trace metal drawdown (Fig. 2) may indicate prolonged seawater trace metal limitation through the earliest Triassic.

5.3. Temporal relationship of end-Permian oceanic anoxia in the low latitude and south mid-latitude Panthalassa

To investigate the development of anoxia across the Panthalassic ocean, we have compiled redox proxy records for deep-sea PTB sections across the Permian to Triassic transition (Fig. 5). Available conodont-dated, Panthalassic deep-sea PTB sections carbon isotope records and redox sensitive trace element data include Arrow Rocks in New Zealand (Hori et al., 2007), and the Akkamori 2 (Takahashi et al., 2014) and Ubara sections (Algeo et al., 2011) in Japan. In Figure 5, U/Al and Mo/Al ratios for WHK 1 and these additional PTB sections are compared based on δ^{13}C chemostratigraphic correlation (Fig. 4).

WHK 1 and Arrow Rocks, which were deposited in the south mid-latitude of the Panthalassa Ocean, might be expected to have similar redox characteristics. During the first δ^{13}C decrease in the late Changhsingian, WHK 1 consists of pale green-coloured tuffaceous sedimentary rocks. Arrow Rocks also consist of similar tuffaceous strata in the late Changhsingian (Fig. 5). Given this lithologic similarity, we use the baseline values for U/Al and Mo/Al defined for WHK 1 to guide interpretation of the Arrow Rocks redox data. However, U concentrations were below detection (<1.5 ppm) in the lower part of the Arrow Rocks succession (Hori et al., 2007), preventing a direct redox interpretation. Furthermore, while Mo/Al ratios are relatively high, total sulphur concentrations are below detection (<1.8 ppm), and thus it is difficult to provide a robust redox interpretation for this part of the succession. However, in the overlying Permian-Triassic transitional section at Arrow Rocks, enrichment of U/Al is evident, alongside very high peaks in Mo/Al (up to 670). This coincides with U and Mo enrichment in WHK 1 (Fig. 5), supporting the notion of widespread development of euxinic water column conditions in the south mid-latitude Panthalassa, at least from the EPME to the PTB.

Geochemical records from low latitude Panthalassa, represented by the Akkamori 2 and Ubara sections, provide further insight into temporal and spatial redox dynamics. To evaluate the redox records for these PTB sections, we again use the minimum measured U/Al (Akkamori = 0.55, Ubara = 0.11) and Mo/Al (Akkamori = 1.13, Ubara = 1.01) ratios to define oxic baseline values. Both sites show evidence for generally persistent anoxia (elevated U/Al) throughout the entire section, while elevated Mo/Al ratios suggest that the water column may have become euxinic prior to the first δ^{13}C decrease (Fig. 5). Furthermore, while there was some fluctuation in the precise chemistry of the anoxic water column after the EPME, euxinia was a prevalent feature immediately after the EPME at the Akkamori site, and potentially also at the Ubara site.

Our interpretation of the redox dynamics at the Ubara site based on Al-normalized values of U and Mo contrast with those of Algeo et al. (2011), who considered suboxic conditions to have been prevalent. In their study, the elevated bulk concentrations of U and Mo at Ubara site are interpreted as being due to the slower sedimentation rate of Permian-Triassic black claystone relative to the Permian chert beds resulting in enrichment. There thus remains some uncertainty in the precise redox state of the Ubara site. However, we note that our interpretation is more consistent with the redox conditions that are much more compellingly demonstrated by highly elevated enrichments in Mo/Al at the Akkamori site. Thus, taken together, it appears likely that the ocean oxygenation state of the Akkamori and Ubara sites in the low latitude Panthalassa deteriorated earlier, relative to the Waiheke site in the south mid-latitude Panthalassa.

5.4. Covariation between U and Mo, and Mo drawdown after the Permian-Triassic boundary

The bulk concentrations of Mo and U in the PTB sections are recast as enrichment factors (Mo_{EF} and U_{EF}) and shown as a cross plot in Figure 6. Algeo and Tribovillard (2009) and Tribovillard et al. (2011) provide a framework for utilizing Mo_{EF} and U_{EF} as a means to identify three enrichment patterns. Firstly, co-increases in Mo_{EF} and U_{EF} may occur following a deterioration in redox conditions from suboxic to euxinic in open marine settings (Fig. 6). The second case occurs where the Mo_{EF} is significantly elevated relative to the U_{EF}, which may be a consequence of active Mo transportation from the water column to the sediment, for example via Fe-Mn (oxyhydr)oxide shuttling. The third pattern occurs when smaller and relatively constant Mo_{EF} are combined with high U_{EF}. This scenario is considered to represent the case where extensive Mo-depleted seawater arises because of a persistent sulphidic deep water mass, as may occur in a relatively restricted setting.

In the PTB sections of the low latitude Panthalassa during the Late Permian (Changhsingian), before the $\delta^{13}C$ decrease, Mo_{EF} values are high relative to U_{EF} values (Fig. 6-1). This supports the redox observations based on U/Al and Mo/Al ratios (Fig. 5), and suggests anoxic, non-sulphidic water column conditions across this interval, with elevated Mo drawdown due to uptake and sequestration by Fe-Mn (oxyhydr)oxide minerals. At the WHK 1 site, the Mo_{EF} and U_{EF} data (Fig. 6-1) document the slower development of anoxia in the south mid-latitude Panthalassa, with evidence for the very beginning of water column de-oxygenation, consistent with the Fe speciation, U/Al and Mo/Al (Fig. 2).

Between the two step decreases of $\delta^{13}C$, highly elevated Mo_{EF}, and to a lesser extent U_{EF}, are evident for the low latitude Panthalassa localities (Fig. 6-2). These values support the continued deterioration of water column redox conditions, with sporadic development of euxinic conditions (as documented in Figure 5) further enhancing trace metal drawdown (particularly Mo). In contrast, at the WHK 1 site in the south mid-latitude Panthalassa, U_{EF} and Mo_{EF} values document a later, progressive increase (Fig. 6-2). Across this interval, the gradual co-increase in U_{EF} and Mo_{EF} values at the WHK 1 site suggests a classic redox transition from suboxic, through anoxic, to euxinic water column conditions (Fig. 6; Algeo et al., 2009; Tribovillard et al., 2012). This is generally consistent with the trends evident in the Fe speciation data (Fig. 2), supporting a progressive deterioration in the redox state of the water column and enhanced drawdown of U and Mo into the sediments.

During the interval from the EPME to the PTB, high U_{EF} and Mo_{EF} values are recorded in both the south middle latitude and low latitude Panthalassa (Fig. 6-3). However, data from the WHK 1 site plot closer to the typical trend for euxinic settings, which supports the suggestion from trends in U/Al and Mo/Al (Figs. 2 and 5), that while euxinia was likely widespread, this redox state may have become more prevalent and persistent in south middle latitude regions, relative to the low latitude Panthalassa.

After the peaks across the EPME-PTB interval, U_{EF} and Mo_{EF} shift to generally lower values during the early Triassic (Fig. 6-4). Indeed, enrichment patterns do not return to the open-marine redox evolution trend. Instead, while the data show a wide range, Mo_{EF} values in particular are generally lower than during the EPME-PTB interval. Thus, we suggest that the development of widespread anoxia, with the common occurrence of euxinia, during the EPME-PTB interval subsequently resulted in limited Mo sequestration in the sediments due to a significant decrease in the seawater Mo inventory caused by intense Mo-bearing sulphide deposition (e.g., Algeo, 2004; Takahashi et al., 2014; Goldberg et al., 2016). Although euxinic water condition enhances uptake of both Mo and U into the sediment, relatively stronger accumulation of Mo results in significantly decreased residence time of Mo in the seawater and depletion of Mo relative to U. Alternatively, such a signal could be due to lower sulphide availability in the water column, which might result from seawater sulphate drawdown after EPME anoxia (Song et al., 2014; Schobben et al., 2017). However, the fact that several intervals have Fe_{HR}/Fe_{total} and Fe^{*}_{pyrite}/Fe_{HR} ratios suggestive of euxinia, implies that Mo depletion from the water column, rather than low sulphate availability, is a more likely explanation for the relatively subdued Mo enrichments evident in the earliest Triassic.

5.5. Possible drivers of Panthalassic anoxia

This study indicates that the pelagic Panthalassa became anoxic, and was often euxinic, in south middle latitude and low latitude regions during the EPME. However, the development of anoxia occurred earlier at low latitudes. The progressive development of ocean anoxia recorded by our data is in agreement with a decreasing trend in $\delta^{238}U_{carbonate}$ recorded in multiple carbonate PTB sections (Fig. 7; Zhang et al., 2020). The $\delta^{238}U_{carbonate}$ signal begins to decrease during the first $\delta^{13}C$ decrease of the Changhsingian, which continues across the EPME, resulting in a negative trough in the earliest Triassic. This $\delta^{238}U_{carbonate}$ variability is consistent with drawdown of U into Panthalassic anoxic deep-sea sediments through the Late Permian and Early Triassic.

In Figure 7, data that can promote marine anoxia are compiled for the EPME. The Siberian Traps were likely the main trigger for environmental change across the Permo-Triassic boundary (e.g., Wignall., 2001). A mercury (Hg) anomaly, taken as an indicator of enhanced volcanic activity, has been detected in late Changhsingian horizons in the Akkamori 2 and Ubara sections in the low latitude Panthalassa (Fig. 7; Shen et al., 2019), and at WHK 1 (Grasby et al., 2021), demonstrating a coincidence of major volcanism with the onset of euxinia. Identified Hg anomalies appear below the EPME horizon at the low latitude sites, but just above the EPME horizon at the south mid-latitude WHK 1site (Fig. 7). One possible reason for this discrepancy is the south mid-latitude location was too far from the Siberian Traps to receive enough Hg before the EPME. The low sampling resolution across the WHK 1 section could be another possibility for this discrepancy. Grasby et al. (2021) did not measure Hg in the thin black claystone bed at the EPME (Bed 17 of this study) and adjacent layers.

The development of ocean anoxia prior to the EPME and persistence into the earliest Triassic coincides with extreme climate warming (Fig. 7; Joachimski et al., 2012, 2019). This warming has been linked to massive release of CO_2 from heated sediments and upper crusts by Siberian Traps sills (Svensen et al., 2009). Subsequently, seawater warming would contribute to marine deoxygenation in these oceanic regions through the decreasing solubility of dissolved O_2. Furthermore, thermohaline circulation driven by subduction of warm saline surface water formed in mid and low latitudes would potentially contribute to the development of a stratified ocean structure with anoxic deep waters (Kidder and Worsley, 2004).

The development of anoxia during the EPME coincides the cessation of radiolarian chert deposition (Fig. 7). Our $\delta^{13}C$ correlation of PTB sections reveals that this phenomenon occurred in the south mid-latitude Panthalassa around ten kiloyears later than in the low latitude Panthalassa (Fig. 7). This may be due to a decreased biogenic silica burial flux (declining radiolarian population and/or degraded preservation of siliceous tests) and dilution of biogenic silica by increased clastic material (Takahashi et al., 2009; Muto et al., 2018, 2020). The environmental conditions during the EPME, including ocean anoxia and seawater warming may have decreased the siliceous faunal population and increased sedimentation of clay-rich material relative to biotic silica. If warming is a major factor limiting deposition of biotic silica, then the different timing for the disappearance of radiolarian chert in the low latitude and south mid-latitude ocean is noteworthy. For example, the higher latitude ocean would warm more slowly, thus resulting in the observed delayed onset of both ocean anoxia and the decrease in biotic silica deposition. The subsequent return to lower temperatures favourable for the radiolarian chert deposition would also arrive earlier in mid-latitude regions.

6. Conclusions

This study provides new geochemical records across the end-Permian mass extinction event in the south middle latitude pelagic Panthalassa. Fe-speciation and redox-sensitive elements (U and Mo) show the progressive development of ocean anoxia, with euxinia occurring during the EPME. Comparison with previously reported paleo redox records from the low latitude Panthalassa indicates that anoxia developed earlier in low latitude regions. The pelagic Panthalassa PTB sections show Mo drawdown in the earliest Triassic, after the EPME likely attributable to high Mo burial flux into anoxic marine sediments. The Late Permian anoxia that occurred in the low latitude Panthalassa, and the progressive development of anoxia during the run-up to the EPME in the south middle latitude Panthalassa, coincides with Siberian Traps volcanic activity and de-vegetation trends on land. Furthermore, the persistent anoxia that developed in both latitudinal regions of Panthalassa coincide with severe seawater warming.

The results of this study shed light on spatiotemporal changes in open ocean redox conditions across the Permian-Triassic boundary and show that the redox history of this super-ocean varied according to latitude. Conditions for radiolarians were persistent and harsh in equatorial latitudes relative to the mid-latitude site discussed here, indicating that high temperatures were a major factor controlling their absence. Environmental records for mid-high latitudes and pelagic regions of the ocean remain highly limited, and this should be a prime target for further investigation.

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Figure captions

Figure 1. Location map of the studied Permian-Triassic section (Waiheke 1 section, New Zealand). (A) Paleo-geographical map of the Permian-Triassic showing the depositional position of pelagic deep-sea Permian-Triassic boundary sections. Labeled numbers (1-7) indicate the following localities: (1) pelagic sections from New Zealand accretionary complexes, such as the Waiheke section (Spörli et al., 2007); (2) pelagic sections from Japanese accretionary complexes; (3) shallow marine sections such as Meishan, South China; (4) sections in Sverdrup Basin, arctic Canada (Grasby and Beauchamp, 2011); (5) sections in Peace River Basin, western Canada (Hays et al., 2007). (B) Locality of the Waiheke-1 section (WHK1) in the North Island of New Zealand. Distribution of the Triassic-Jurassic accretionary complex is based on Black (1994) and Begg and Johnson (2000). Locality of Arrow Rocks is also shown.

Figure 2. Lithologic column of the Waiheke 1 section, including organic carbon isotopes ($\delta^{13}C_{org}$), iron speciation pools (Fe_{pools}), highly reactive iron normalised to total iron (Fe_{HR}/Fe_{total}), corrected pyrite iron as a function of highly reactive iron (Fe^{*}_{pyrite}/Fe_{HR}), U, U/Al and Mo/Al. The error ranges on the organic carbon isotope values ($\delta^{13}C_{org}$) are based on repetitions of carbon isotope measurements (Table S1). Open circles in the plots of Fe_{pools}, Fe_{HR}/Fe_{total} and Fe^{*}_{pyrite}/Fe_{HR} mean the samples with low Fe content (less than 0.5 wt % Fe_{total}). We did not use these samples for paleoenvironmental reconstruction. Local background values of U/Al and Mo/Al for tuffaceous strata (Beds 1-18) and cherty strata (Beds 19-88) are shown as vertical dashed lines with “baseline” notation. The values are based on the minimum values in the tuffaceous strata and the values in the cherty strata with oxic Fe_{HR}/Fe_{total} signals.

Figure 3. Thin section photographs from the Waiheke 1 section. (A) tuffaceous siliceous claystone of Bed 8; (B) black claystone of Bed 22; (C) bedded chert of Bed 50; (D) bedded chert from around Bed 60.

Figure 4. Stratigraphic correlation of the Permian-Triassic boundary sections of the Meishan section, Waiheke 1 section, Arrow Rocks section, Akkamori 2 section and Ubara section. Compiled data of the Meishan section are based on Jin et al. (2000), Cao et al. (2002), Kaiho et al. (2009) and Burgess et al. (2014). Data for Arrow Rocks, Akkamori and Ubara sections are modified from Hori et al. (2007), Takahashi et al. (2010) and Kaiho et al. (2012), respectively. The horizontal bars indicate correlative stratigraphic zones around the Permian-Triassic boundary. Green: the first decrease in carbon isotope ratio ($\delta^{13}C$) in the latest Changhsingian, Blue: the second decrease in $\delta^{13}C$ in the latest Changhsingian coinciding with the EPME, Pink: the interval from the $\delta^{13}C$ minimum during the EPME to the $\delta^{13}C$ increase toward the PTB.

Figure 5. Vertical plots of U/Al and (Mo/Al) for the Waiheke 1 and other pelagic deep-sea sections. The data for Arrow Rocks, Akkamori 2 and Ubara are modified from Hori et al. (2007), Takahashi et al. (2014) and Algeo et al. (2011), respectively. The local background value (baseline) for U/Al and Mo/Al are based on the WHK 1 baseline values for Arrow Rocks, and are based on the minimum values for the Akkamori and Ubara sections.

Figure 6. Cross plots of molybdenum (Mo) and uranium (U) enrichment factors (Mo_{EF} and U_{EF}, respectively) for Panthalassic PTB sections from Waiheke 1, Arrow Rocks, Akkamori and Ubara. The data for Arrow Rocks, Akkamori 2 and Ubara are modified from Hori et al. (2007), Takahashi et al. (2014) and Algeo et al. (2011), respectively. The data are plotted in four different panels representing the time intervals based on the carbon isotope correlation demonstrated in Figures 4 and 5. 1: Late Permian (Changhsingian) before the $\delta^{13}C$ decrease; 2: between the first and second steps decreases of $\delta^{13}C$ in the latest Permian; 3: the interval from the $\delta^{13}C$ minimum at the EPME to the $\delta^{13}C$ increase toward the PTB; 4: Earliest Triassic (Induan) after the PTB. Superimposed grey-coloured areas representing modern oceanic sediments are from Algeo and Tribovillard (2009) and Tribovillard et al. (2012). The U/Mo_{SW} is the weight ratio in modern seawater (SW), which ranges from 3.0 to 3.1 (modern Pacific and Atlantic, respectively; Anderson, 1987; Bruland, 1983; Chen et al., 1986; Emerson and Huested, 1991).

Figure 7. Compilation of oceanic environmental records from Panthalassa, seawater temperature for the paleo-Tethys reconstructed by oxygen isotope ratios in conodont apatite ($\delta^{18}O_{apatite}$; Joachimski et al., 2009, 2019), global anoxic water extent referred from uranium isotopes in carbonate ($\delta^{238}U_{carbonate}$; Zhang et al., 2020). Time scale through Permian and Triassic is from Burgess et al. (2014). Information on oceanic environmental records for Panthalassa is based on litho- and chemo-stratigraphy by this study, Hori et al. (2007, 2011), Algeo et al., (2011), Takahashi et al. (2014), Kaiho et al. (2016), Shen et al., (2019) and Grasby et al., (2021). Thin lines indicate the range of each Permian-Triassic boundary section. The bars coloured black, orange, purple and grey refer to Hg anomaly (Hg/TOC > 100), radiolarian chert occurrence, geochemical signals of euxinia (high Fe_{HR}/Fe_{total}, Fe_{pyrite}/Fe_{HR} and/or Mo/Al and U/Al), and seawater Mo drawdown (lower Mo/Al and Mo_{EF} despite high Fe_{HR}/Fe_{total} and Fe_{pyrite}/Fe_{HR} and/or U/Al and U_{EF}), respectively.