Transient Permian-Triassic euxinia in the southern Panthalassa deep ocean

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ABSTRACT

Both the duration and severity of deep-water anoxic conditions across the Permian-Triassic mass extinction (PTME) are controversial. Panthalassa Ocean circulation models yield varying results, ranging from a well-ventilated deep ocean to rapidly developing northern-latitude, but not southern-latitude, anoxia in response to Siberian Traps–driven global warming. To address this uncertainty, we examined a southern-paleolatitude pelagic record. Trace metal and pyrite framboid data suggest bottom-water euxinic conditions developed in the southern Panthalassa Ocean at the PTME, coincident with enhanced volcanic activity indicated by Hg geochemistry. While a global ocean euxinic event at the PTME placed extraordinary stress on marine life, southern surface waters appear to have recovered more quickly as radiolarian populations returned several million years before they did in northern Panthalassa.

INTRODUCTION

The Permian-Triassic mass extinction (PTME) and prolonged Early Triassic ecological recovery (Chen and Benton, 2012) are thought to have been driven by recurrent episodes of marine anoxia related to global warming induced by Siberian Traps volcanism (e.g., Wignall and Twitchett, 2002). The record of marine anoxia, however, is largely restricted to northern-latitude continental-margin records from the Tethys and northwestern Pangea, where shallow-water anoxia would have placed the greatest extinction pressures on fauna. However, understanding the full ecologic stress that temperature-dependent hypoxia placed on marine life and models of its recovery require insight into total ocean O₂ loss (Penn et al., 2018). Uranium isotopes and cerium anomaly data reflecting mean global-ocean redox conditions suggest anoxic seafloor area expanded from ~0.2% to somewhere between 17% and 60% across the PTME (Elrick et al., 2017; Song et al., 2012; Zhang et al., 2018). While the extent of anoxia was substantial, these results imply 40%–83% of the seafloor remained oxic. Consistent with this, ocean circulation models suggest that deep waters became at most dysoxic (Winguth and Winguth, 2012), whereas Penn et al. (2018) suggested development of anoxic seafloor was restricted to northern abyssal plains.

Although the Panthalassa ocean covered 70% of Earth’s surface, records from it are rare because its ancient seafloor has been subducted. Surviving remnants of equatorial and northern-paleolatitude pelagic deposits, preserved in accretionary terranes in Japan and Canada, provide insight into deep-water anoxia based on geochemical proxies for local redox conditions. Isozaki (1997) and Kato et al. (2002) suggested anoxia developed in the late Permian and persisted until the Middle Triassic, although Fujisaki et al. (2019) argued for well-ventilated bottom waters throughout the late Permian. Others have suggested, based on pyrite framboid size distributions (Wignall et al., 2010) and redox-sensitive trace elements (Takahashi et al., 2014, 2015), that bottom-water euxinia developed only during the PTME interval. Southern-latitude deep-water anoxia was also suggested by Hori et al. (2007), although timing relative to the PTME is uncertain. In general, geochemical records suggest deep-water anoxia was much more widespread than ocean-circulation and uranium-isotope mass-balance models predict.

Uncertainty in the global extent, duration, and severity of deep-ocean anoxia limits our understanding of how global warming associated with the Siberian Traps affected global ocean O₂ levels. To address this knowledge gap, we examined a southern-hemisphere, mid-paleolatitude pelagic sequence utilizing a combined geochemical, petrographic, and sedimentologic approach.

STUDY AREA

We studied the intertidal section at Island Bay (36°46.131′S, 175°00.200′E; World Geodetic System 1984 [WGS84] datum), Waiheke Island, New Zealand (Fig. 1). Upper Permian through Lower Triassic mudstones and cherts (Kiripaka Formation) overlie oceanic crust that has been obducted in an accretionary prism (Spörli et al., 1989). Conodont biostatigraphy and organic carbon stable isotope (δ¹³Corg) data show that the continuously exposed strata span the Changhsingian to Anisian except for the lower Olenekian (Hori et al., 2011). Paleomagnetic data indicate a paleolatitude of ~34°S (Kodama et al., 2007).

METHODS

We measured a sedimentary log, with numbered lithologically distinct units (Fig. 2). Age assignments are from Hori et al. (2011). Samples were collected every 1–20 cm from fresh surfaces exposed by breaking off marine encrusters and weathered rims, then powdered by agate mortar and pestle. Major and trace elements were analyzed on digested powders in a 2:2:1:1 solution of H₂O-HF-HClO₄-HNO₃.
RESULTS

The stratigraphic section begins with red mudstones that pass into younger gray-green mudstone and then gray, laminated siliceous mudstone. In addition to three black shales (units 5, 7, and 9 in Fig. 2), siliceous mudstones persist across the Permain-Triassic boundary to the end Induan (4.2 m). Spathian conodonts occur immediately above the Induan-Olenekian boundary, suggesting that the Smithian here is either highly condensed or cut by layer-parallel faulting (Hori et al., 2011). The lowest Olenekian strata consist of 2-cm-thick interbeds of siliceous mudstone and mudstone; both lithologies are gray-green in the lower 80 cm (unit 11). Above unit 11, the siliceous mudstones become red while the interbedded mudstones remain gray-green (unit 12). Around the Olenekian-Anisian boundary and a short distance above a 30-cm-thick siliceous mudstone (unit 13), all beds become red regardless of lithology (Fig. 2). No radiolarians are present in the Changoxygenian strata, but they appear in the Lower Triassic siliceous mudstones, forming radiolarian-rich chert beds ~5–10 mm thick, interbedded with finely laminated intervals (Fig. 3). This style of interbedding persists into the Anisian strata.

The δ¹⁹⁸Hg values in units 1–3 vary from −28‰ to −26‰, falling in the latest Changoxygenian to −30‰ to −28‰ in unit 4, and then to −34‰ in the 20-cm-thick black shale of unit 5, before increasing to −30‰ up to the latest Induan. The δ¹⁹⁸Hg values further increase to −25‰ at the level where Spathian conodonts appear, and then are variable, but with a general increasing trend, through the Olenekian to a maximum of −22‰ (Fig. 2). The marked late Smithian negative δ¹⁹⁸Hg excursion (e.g., Grasby et al., 2013) is not present at Island Bay, supporting the Hori et al. (2011) conclusion that this level is absent.

In the basal and upper parts of the section, content of redox-sensitive element Mo (absolute and normalized to Al) are close to those of the average shale of Wedepohl (1995) (1.3 ppm and 0.15 ppm/wt%, respectively). Mo/Al values increase from 0.1 to 5.7 ppm/wt% in unit 4 and stay relatively elevated (>1.2 ppm/wt%) up to the top of the Induan (unit 10) before returning to values of <0.2 ppm/wt%. In unit 5, Mo/Al reaches 135 ppm/wt% (~900 × that of average shale) (Fig. 2). Both U and U/Al increase in unit 4, exceeding average shale values of 3.7 ppm and 0.42 ppm/wt%, respectively, and stay elevated to unit 10 before declining below average shale values (Fig. 2). In unit 5, U is 5.5–32.9 ppm and U/Al is as much as 31 × that of average shale. Th/U ratios are >2 at the base and top of the section but drop to <2 in unit 4 and remain low until unit 10. In unit 5, Th/U ranges from 0.15 to 0.42. Mn values are high in the late Permian (as high as 2900 ppm) and then progressively decline through the late Changhsingian to values <50 ppm in unit 5, maintaining these low values until the top of the Induan before steadily increasing up to the Anisian. Values of TOC range from 0.21 to 0.55 wt% except in unit 5 where TOC exceeds 9 wt%.

Pyrite is absent from the basal and uppermost lithologies but is present and locally abundant as frambooids in units 4 through 10 (Fig. 2). Of all the units that contain a statistically valid frambooid population, unit 4 has the largest mean and standard deviation of pyrite frambooid diameters (Fig. 4). In unit 5, frambooids become smaller (<5 μm mean) and less variable in size.

Through most of the section, Hg values range from 50 ppb to below the 1 ppb detection limit. A prominent Hg spike (7590 ppb) occurs at the PTME boundary, and then a lesser spike occurs in unit 9 (1230 ppb) (Fig. 2). Both spikes greatly exceed average Hg values (62.4 ppb) in sedimentary rock (Grasby et al., 2019). These spikes survive normalization to TOC, Al, and total sulfur, showing that they are not related to changes in Hg sequestration (Grasby et al., 2019) but represent excess Hg deposition. Background Hg concentrations were too low for stable isotope analyses, but the Hg spikes at the PTME horizon show δ²⁰₂Hg values ranging from −1.52‰ to −0.97‰ and Δ²⁰²Hg values that range from 0.04‰ to 0.09‰, with the exception of a single sample with a Δ²⁰²Hg value of −0.07‰ (Fig. 2).

DISCUSSION

The end-Permian negative carbon isotope excursion across the PTME at Island Bay and subsequent shift to higher values through the Early Triassic are consistent with carbonate (Payne et al., 2004) and organic carbon (Grasby et al., 2013) records, supporting the premise that Island Bay represents a deep-water record of the PTME.

Low values of Mo/Al and U/Al, high Th/U and Mn, and the absence of pyrite all indicate that the southern Panthalassa deep ocean wasoxic until the very end of the late Permian, similar to the trend demonstrated by equatorial deep-water sections (Wignall et al., 2010; Fushiki et al., 2019) and to the modern deep ocean. However, in the late Changhsingian, increasing

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¹Supplemental Material. Analytical data. Please visit https://doi.org/10.1130/GEOL.S.14319572 to access the supplemental material, and contact editing@geosociety.org with any questions.
Mo/Al and U/Al, a fall in Th/U to \(<2\), and the appearance of finely laminated bedding all indicate that deep-water dysoxia developed just prior to the PTME (gray zones in Fig. 2). This is supported by the appearance of pyrite frambooids with size ranges that plot in the dysoxic field (Bond and Wignall, 2010) (Fig. 4). The progressive shift to low Mn values suggests development of broader ocean anoxia at the same time (Frakes and Bolton, 1984). Declining oxygen levels prior to the PTME are consistent with other records and are thought to represent the initial emplacement of Siberian Traps magma prior to the main eruption phase that drove extinction (Wignall et al., 2010; Grasby et al., 2013; Burgess et al., 2017).

At the PTME and throughout unit 5, Mo/Al values of as much as \(900 \times\) average shale values indicate persistently euxinic bottom-water conditions (Lyons et al., 2009) (yellow bar in Fig. 2), as do Th/U ratios as low as 0.15 (Wignall and Twitchett, 1996). Euxinia is supported by pyrite frambooid populations with mean diameters \(<5\,\mu m\) (Fig. 4), which are characteristic of formation within a \(H_2S\)-rich water column (Bond and Wignall, 2010). This brief euxinic episode is marked by extraordinarily high TOC (\(>9\) wt\%) for deep-sea sediments.

The spike in Hg concentrations is similar to that observed at PTME sections globally and is
CONCLUSIONS

The southern Panthalassa Ocean was well ventilated during the late Permian until oxygen-poor conditions developed just prior to the PTME that transitioned to euxinic conditions across the PTME. While oxygenation slightly improved following the PTME, the deep ocean remained largely anoxic until more ventilated conditions returned in the Olenekian. The PTME deep-ocean euxinic event, developed in both the southern and northern hemispheres, reflects much greater net O2 loss than models predict and must have placed extraordinary stress on marine life. After this event, southern surface waters were more favorable for pelagic life, as testified by the intermittent reappearance of radiolarian populations several million years before they returned in northern Panthalassa. Although the models of Penn et al. (2018) predict anoxia in northern-latitude oceans, they do not replicate the complex ocean redox history presented in the rock record. Full understanding of oxygen stress and recovery requires incorporation of these transient, and potentially devastating, euxinic events into global ocean biogeochemical models.

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