

**High-resolution records of Oceanic Anoxic Event 2: Insights  
into the timing, duration and extent of environmental  
perturbations from the palaeo-South Pacific Ocean**

**Authors:** S.K. Gangl<sup>a,b,\*</sup>, C.M. Moy<sup>c</sup>, C.H. Stirling<sup>a,b</sup>, H.C. Jenkyns<sup>d</sup>,  
J.S. Crampton<sup>e,f</sup>, M.O. Clarkson<sup>a,b,1</sup>, C. Ohneiser<sup>c</sup>, D. Porcelli<sup>d</sup>

**Affiliations:**

<sup>a</sup> Department of Chemistry, University of Otago, PO Box 56, Dunedin, NZ

<sup>b</sup> Centre for Trace Element Analysis, University of Otago, PO Box 56, Dunedin, NZ

<sup>c</sup> Department of Geology, University of Otago, PO Box 56, Dunedin, NZ

<sup>d</sup> Department of Earth Sciences, University of Oxford, South Parks Road, Oxford OX1 3AN,  
UK

<sup>e</sup> GNS Science, PO Box 30368, Lower Hutt, NZ

<sup>f</sup> School of Geography, Environment and Earth Sciences, Victoria University of Wellington,  
PO Box 600, Wellington, NZ

<sup>1</sup> Now at: Department of Earth Sciences, ETH Zürich, Clausiusstrasse 25, 9092 Zürich, CH

\* Corresponding author: Sophie Gangl (sophie.gangl@otago.ac.nz)

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## Abstract

Oceanic Anoxic Event 2 (OAE 2), which took place around the Cenomanian–Turonian boundary (~94 Ma), is associated with extreme perturbations to the global carbon cycle, affected ocean basins worldwide and was associated with significant biological turnover. Although this event has been well studied in the northern hemisphere, the evolution and character of OAE 2, particularly in terms of the vertical and lateral extent of anoxia, is poorly constrained in the palaeo-Pacific Ocean. Furthermore, the precise timing, duration and character of this event, and the exact mechanisms driving OAE 2 environmental changes, are still being debated. Here, we present the first high-resolution records of carbon- isotopes, total organic carbon and magnetic susceptibility from the southern palaeo-Pacific Ocean during OAE 2, sampled at two sections in New Zealand. The carbon-isotope records from both localities reveal a ~2 ‰ positive excursion that represents the global change in the carbon cycle associated with OAE 2. When combined with a cyclostratigraphic age model, these new records constrain the duration of the OAE 2 carbon-isotope excursion to at least  $930 \pm 25$  ky and indicate a minimum duration of  $200 \pm 25$  ky for the ‘Plenus Cold Event’ that took place during OAE 2. The lithologies and low organic-carbon contents of the New Zealand sections imply that oxic conditions prevailed along, at least parts of, the margins of the palaeo-Pacific Ocean at mid- to high southern latitudes during OAE 2 while, contemporaneously, conditions were locally anoxic in the mid-water column of the equatorial Pacific Ocean. Despite these apparently oxic conditions in the New Zealand region, there was a partial collapse of benthic ecosystems leading up to, and during, OAE 2, suggesting environmental deterioration caused by intermittent oxygen deprivation, or other chemical or biological disturbances in the South Pacific region that remain to be elucidated.

47    **KEYWORDS:**

48    Oceanic Anoxic Event 2, Carbon-isotope excursion, Cretaceous, Organic carbon, Orbital  
49    forcing, Cenomanian–Turonian

50

## MAIN TEXT:

### 1. Introduction

The Oceanic Anoxic Events (OAEs) of the Mesozoic Era are well-known palaeoenvironmental phenomena that represent major disturbances of the global carbon cycle and were characterised by the burial of vast amounts of marine organic matter in all ocean basins, often preserved as ‘black shales’ (Schlanger and Jenkyns, 1976). OAEs were closely linked to the rapid influx of CO<sub>2</sub> into the atmosphere, which in turn drove abrupt global warming and increased weathering, thereby increasing the supply of nutrients to the oceans (e.g. Turgeon and Creaser, 2008; Du Vivier et al., 2014; Jenkyns et al., 2017; Clarkson et al., 2018). This process in turn stimulated plankton productivity and accentuated the flux of organic matter through the water column. The decomposition of some of this organic material by bacterial activity led to the enhanced consumption of oxygen in subsurface waters, resulting in widespread oceanic anoxia to euxinia (anoxic and sulfidic conditions) on regional to global scales (Jenkyns, 2010 and references therein). Due to extensive oxygen depletion in the oceans, OAEs are commonly associated with biotic turnover and extinction events. Constraining, in detail, the timing, sequence and magnitude of environmental, climate and biological disruptions during past OAEs provides a unique opportunity to deconvolve the complex interplay of the oceans and atmosphere during major climate perturbations. As such, OAEs can provide useful constraints for understanding the impact of modern anthropogenic activity on future climate.

Oceanic Anoxic Event 2 (OAE 2) spanning the Cenomanian–Turonian (C/T) boundary and occurring ~94 million years ago (Ma) (Obradovich et al., 1993; Gradstein et al., 2012; Meyers et al., 2012), was one of the most intense and spatially extensive of the Mesozoic OAEs (Jenkyns, 2010). As for all of the OAE’s, OAE 2 is characterised by the widespread



deposition of black shales, although the lateral extent of deposition is poorly known, particularly in the Indo-Pacific region. The global influence of OAE 2 is illustrated by abrupt changes in carbon-isotope ( $^{13}\text{C}/^{12}\text{C}$ ) stratigraphy, formulated as  $\delta^{13}\text{C}$ , across the C/T boundary from biostratigraphically well-dated sedimentary sections (e.g. Scholle and Arthur, 1980; Schlanger et al., 1987; Tsikos et al., 2004). Because photosynthetically produced carbon preferentially incorporates lighter  $^{12}\text{C}$  over heavier  $^{13}\text{C}$  isotopes, the anomalously high burial rate of organic matter led to a positive carbon-isotope excursion (CIE) in marine and atmospheric carbon reservoirs (Scholle and Arthur, 1980). The resultant 2–4 ‰ positive excursion, registered in both carbonate and organic-carbon archives, is widely referred to as the ‘OAE 2 CIE’ or ‘C/T CIE’ and provides an important chemostratigraphic tool for identifying the event in sedimentary records (e.g. Scholle and Arthur, 1980; Tsikos et al., 2004; Li et al., 2006; Jenkyns, 2010; Jenkyns et al., 2017).

The  $\delta^{13}\text{C}$  profiles from OAE 2 sites are often correlated to the biostratigraphically well-characterised and stratigraphically expanded section of semi-lithified chalk, originating from foraminiferal–nannofossil ooze, exposed in the cliffs at Eastbourne, in southern England (Fig. 1a) (Tsikos et al., 2004). General characteristics of the Eastbourne  $\delta^{13}\text{C}$  curve include a relatively sharp initial increase, a trough, a second peak, and an irregular plateau followed by a gradual decrease to pre-excursion values (Fig. 1b). The rise and fall of  $\delta^{13}\text{C}$  values has been interpreted to reflect an increase and decrease in global carbon burial, respectively, concomitant with the expansion and retreat of oceanic anoxia (e.g. Kuypers et al., 2002; Jenkyns et al., 2017). OAE 2 was probably triggered by a massive injection of  $\text{CO}_2$  into the ocean-atmosphere system through extensive volcanism (Turgeon and Creaser, 2008; Du Vivier et al., 2014; Scaife et al., 2017), which may have been superimposed on Earth’s natural  $\text{CO}_2$  cycles related to variations in Earth’s orbit around the sun (Fischer, 1986). The addition

of isotopically light C from volcanic sources may also have influenced the shape of the  $\delta^{13}\text{C}$  curve (Jenkyns, 2010; Clarkson et al., 2018).

The trough in the  $\delta^{13}\text{C}$  curve of OAE 2 correlates broadly with the so-called ‘Plenus Cold Event’ (PCE), which in Eastbourne is characterised by the southward invasion of cool-water boreal fauna into lower latitudes (Gale and Christensen, 1996). The occurrence of boreal fauna occurred as two pulses, also defined by a rise in bulk oxygen-isotope values, and is coincident with the location of the carbon-isotope trough. However, the stratigraphically lowest occurrence of these cool-water indicators marking the onset of the PCE occurs below the  $\delta^{13}\text{C}$  trough. The PCE or ‘Benthic Oxidic Event’, as it is known locally, is now widely recognised as a distinct period of cooling at a number of latitudes across the northern hemisphere (e.g. Sinninghe Damsté et al., 2010), which can be linked to widespread re-oxygenation and temporarily decreased silicate weathering. Evidence for re-oxygenation is both geochemical, in terms of abrupt changes in redox-sensitive trace-element concentrations and isotopic compositions (e.g. Dickson et al., 2016; Jenkyns et al., 2017; Clarkson et al., 2018; Sweere et al., 2018) and biological, in terms of sea-floor colonisation by benthic foraminifera (Eicher and Worstell, 1970; Kuhnt et al., 2005; Prokoph et al., 2013). Further, tentative evidence for decreased silicate weathering, against a backdrop of an overall increase over the OAE 2 interval, is derived from the lithium-isotope record from Eastbourne (Pogge von Strandmann et al., 2013; Clarkson et al., 2018). The PCE thus represents a short-term episode of re-oxygenation of the oceans with associated global cooling. This episode of re-oxygenation is credited with temporarily decreasing global carbon burial and caused less preferential extraction of  $^{12}\text{C}$  from the ocean-atmosphere system, possibly causing the trough in carbon-isotope values at this time, as registered in numerous northern hemisphere sections (e.g. Tsikos et al., 2004; Kuhnt et al., 2005; Sageman et al., 2006; Hasegawa et al., 2010; Jenkyns et al., 2017).

During OAE 2, the deposition of black shales probably started at marginal regions of the southern North Atlantic and the Western Interior Seaway (WIS) in North America and spread to the northern North Atlantic and eastward to the Tethys Ocean (Kuroda and Ohkouchi, 2006). Previous studies have therefore focused on sites in these northern hemisphere areas that show geochemical and sedimentological evidence for local and regional anoxia and euxinia (e.g. Schlanger et al., 1987; Kuypers et al., 2002; Tsikos et al., 2004; Kuhnt et al., 2005; Sageman et al., 2006; Jenkyns et al., 2007). However, carbon-isotope records that potentially record a global signal may be modified by local or regional processes such as upwelling, changes in the composition of planktonic biota, and diagenesis. Consequently, global coverage of  $\delta^{13}\text{C}$  stratigraphies, based on a multitude of records from different ocean basins, is needed to constrain how carbon cycle perturbations during OAE 2 were expressed globally across the entire ocean-atmosphere system. High-resolution  $\delta^{13}\text{C}$  stratigraphy is also fundamental for integrating palaeontological or more novel geochemical data into a global carbon cycle framework. Presently, there are only a few isolated records from the southern hemisphere, which compromises global-scale interpretations of OAE 2 (Li et al., 2006; Navarro-Ramirez et al., 2016, 2017; Dickson et al., 2017; Li et al., 2017).

Here, we build on previously reported low-resolution  $\delta^{13}\text{C}$  data from two palaeo-Pacific sections located in New Zealand (Hasegawa et al., 2013). We present highly resolved magnetic susceptibility (MS), total organic-carbon (TOC) and organic  $\delta^{13}\text{C}$  ( $\delta^{13}\text{C}_{\text{TOC}}$ ) stratigraphies of the OAE 2 interval, the last of which is largely terrestrial in origin and primarily monitors the carbon-isotope composition of atmospheric  $\text{CO}_2$ . Correlation of the carbon-isotope stratigraphies obtained here to a collection of globally dispersed carbon-isotope records indicates a well-mixed ocean-atmosphere carbon system during OAE 2. Detailed MS and TOC stratigraphies suggest dramatic changes in weathering regime and/or intensity and, at least locally, a predominance of generally oxic conditions in the south-

western palaeo-Pacific Ocean. Finally, we take advantage of the high temporal resolution of the  $\delta^{13}\text{C}_{\text{TOC}}$  and MS records to produce a cyclostratigraphic age model. Together, these datasets help constrain the timing, duration and character of OAE 2 from the unique perspective of a marginal, southern Pacific Ocean setting with millennial-scale resolution.

## 2. Stratigraphic Sections

In this study, sedimentary sections exposed at Sawpit Gully (north-eastern South Island) and Mangaotane Stream (eastern North Island) in New Zealand are examined (Fig. 1c). Both sites have well-exposed and well-characterised sections spanning OAE 2 (Crampton et al., 2001 and references therein; Hikuroa et al., 2009; Hasegawa et al., 2013). These OAE 2 strata were deposited at a palaeolatitude of  $\sim 70^\circ\text{S}$  (Sutherland, 1999) along the eastern continental margin of the proto-New Zealand landmass at about the time of its separation from Gondwanaland. Brief descriptions of the Sawpit Gully and Mangaotane sections are given below, while a detailed description of each section is provided in the Supplementary Material (SM).

### 2.1 *Sawpit Gully*

The Sawpit Gully section consists of non- or weakly calcareous siltstone, silty sandstone, and sandstone, and finely disseminated pyrite appears sporadically throughout the section. The OAE 2 sequence is mostly bioturbated. Although inoceramid bivalves are abundant below and above the sampled section, there is a lack of macrofossils in the sampled interval that indicates some disruption of the benthic ecosystem. The Sawpit Gully section contains dinoflagellates (Sch  ler and Crampton, 2014) but has not been examined further for microfossils. Calcareous microfaunas, if present, are likely to be poorly preserved (Hasegawa

et al., 2013). The depositional environment of the Sawpit Gully section is poorly constrained, but the lithology and stratigraphic context suggest mid- to outer shelf depths.

## 2.2 *Mangaotane B*

The Mangaotane B section at Mangaotane Stream is dominated by non- or weakly calcareous, indurated mudstone, with minor siltstone and very fine sandstone. The section is characterised by the presence of red shales. The red shales have previously been identified as ‘Cretaceous Oceanic Red Beds’ (CORBs) deposited under oxic conditions (Hikuroa et al., 2009). The red coloration is attributed to the presence of hematite ( $\text{Fe}_2\text{O}_3$ ), which may have developed diagenetically from hydroxide species formed in oxidising environments (e.g. Hu et al., 2005; Hikuroa et al., 2009). There is an absence of macrofossils in the Mangaotane B red shales, as is typical of CORBs, possibly due to their adaptation to preceding low-oxygen conditions (Wang et al., 2009). The thickest of these ‘red-beds’ (28.6–44.6 m) is well exposed in Mangaotane B (Fig. 2 and Fig. 3) and is not significantly disturbed by faulting. This major red-bed unit is interbedded with four ‘green beds’ of up to 43 cm thick that are interpreted to reflect anoxic conditions during deposition or early diagenesis (Lyle, 1983). The sequence is mostly bioturbated and mottled to varying degrees, but contains fine remnant millimetre-scale lamination in places and thin (<10 cm thick) laminated beds in the interval between 26 m and the base of the main red-bed. Although common below and above the section, macrofossils are lacking in the OAE 2 interval itself, indicating some disruption of the ecosystem. The environment of deposition for Mangaotane B was likely to have been at upper slope depths of ~300 m or greater (Crampton et al., 2001), which contrasts with the deeper pelagic depositional environments that are typical of red shales from other marine sites. The Mangaotane B section is located ~2 km northwest and downstream of the Mangaotane A section, which exposes the same lithological units spanning OAE 2 but is structurally

complicated over the interval of OAE 2 (Hasegawa et al., 2013).

### **3. Materials and Methods**

#### *3.1 Sample Collection*

Fieldwork was undertaken during three separate sampling trips between March, 2015 and February, 2016. Detailed stratigraphic sections spanning the OAE 2 interval were logged and sampled at both the Sawpit Gully and Mangaotane B sites. In total, 662 samples were taken from both exposures, with an average sampling resolution of 10 cm, using a concrete cutter to remove blocks of rock arrayed along bedding-perpendicular transects (see SM for details). This approach allowed for stratigraphically controlled, bulk sampling of relatively unweathered material. In Sawpit Gully, 477 samples were collected, while in Mangaotane B, 185 samples were taken, in both cases along four non- or slightly overlapping transects (A–D) (Fig. 4 and Fig. 5).

#### *3.2 Geochemical Analyses*

Values for TOC,  $\delta^{13}\text{C}_{\text{TOC}}$ , and MS were determined on every sample in the Sawpit Gully (n = 477) and Mangaotane B (n = 185) composite stratigraphies (see SM for details). Four samples dispersed throughout the Mangaotane B section (two grey shales and two red shales) were analysed for their lithological components by X-ray diffractometry (XRD). Rock-Eval pyrolysis was carried out on 6 samples from Sawpit Gully and 18 samples from Mangaotane B to obtain TOC and carbonate content, as well as the  $T_{\text{max}}$  (the temperature of maximum hydrocarbon generation from kerogens during pyrolysis) and HI

(hydrogen index; amount of hydrocarbons generated by pyrolysis at  $T_{\max}$  relative to TOC) organic geochemistry values.

### 3.3 *Time-Series Analyses*

An orbitally calibrated age model was developed for Sawpit Gully based on the frequency content of the  $\delta^{13}\text{C}_{\text{TOC}}$  and MS profiles constrained by key features of these records, including the start and end of the CIE (see SM for details).

## 4. **Results and Discussion**

### 4.1 *Origin of organic matter*

In a preceding study (Hasegawa et al., 2013),  $\delta^{13}\text{C}_{\text{TOC}}$  records for OAE 2 were derived for Sawpit Gully and Mangaotane B, albeit with low temporal resolution of ~50 ky. These records provide good first-order constraints on the character of OAE 2 along the palaeo-continental margin of New Zealand and show clear evidence for a positive CIE, but lack sufficient resolution to allow for global correlation and identification of shorter duration phenomena, such as the  $\delta^{13}\text{C}$  trough coincident with the PCE. The highly resolved sampling interval of 10 cm of this study corresponds to a temporal resolution of approximately 1–3 ky in Sawpit Gully and 4–8 ky in Mangaotane B, using the plausible 500–900 ky duration for OAE 2, defined as the entire CIE from the onset to return to pre-excursion values, as identified previously (Sageman et al., 2006; Meyers et al., 2012; Ma et al., 2014; Li et al., 2017; Jones et al. 2019).

In both the Sawpit Gully and Mangaotane B sections, TOC concentrations (<0.1 wt %) are approximately two orders of magnitude lower than typically observed in OAE 2 black shales of the northern hemisphere (Fig. 3 and Fig. 6), where wt % TOC values of up to 30–50 % have been documented that are indicative of locally persistent anoxic or euxinic conditions (Kuhnt et al., 2005; Kuypers et al., 2002; Monteiro et al., 2012; Tsikos et al., 2004). Such low TOC concentrations in the Sawpit Gully and Mangaotane B sections instead suggest relatively well-oxygenated depositional environments along the continental margin of New Zealand, in which preserved organic matter rarely exceeds 0.5 % in the sediment (e.g. Arthur and Sageman, 1994). The deposition of oceanic red beds, including the CORBs of Mangaotane B, and the presence of bioturbation in both New Zealand sections, also indicate mostly oxic conditions during deposition (e.g. Hu et al., 2005). It is possible that this is a consequence of shallow depositional environments of mid to outer shelf depths in the Sawpit Gully section. By contrast, the Mangaotane B section was deposited at depths of ~300 m or greater (Crampton et al., 2001), indicating well-oxygenated environments at intermediate depths in the palaeo-South Pacific Ocean during OAE 2.

Despite a dominantly oxic depositional setting, the minor green beds that occur irregularly within the major red-bed of Mangaotane B are inferred to have developed under reducing conditions at or below the sea floor, where red  $\text{Fe}^{3+}$  was reduced to green  $\text{Fe}^{2+}$  (Lyle, 1983), suggestive of intermittent low-oxygen conditions during deposition. Areas of remnant lamination throughout the section, but not in red shales, also suggest fleeting intervals of oxygen-deprived conditions during which bioturbators could not occupy the sediment. In summary, all indicators point toward generally oxic conditions with temporarily more oxygen-deprived phases during the deposition of OAE 2 sediments along the continental margin of New Zealand.



Organic geochemical indicators imply that the organic matter from the Sawpit Gully section is primarily of terrestrial origin (Fig. 7) and as such, is mostly derived from atmospheric CO<sub>2</sub> (Hasegawa et al., 2013). Therefore, the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  record is expected to represent carbon-isotope fluctuations of the atmospheric CO<sub>2</sub> reservoir and accurately record global changes in the carbon cycle. In addition, MS values are relatively uniform in the Sawpit Gully section (Fig. 6), with the exception of two prominent maxima that correspond with relatively low TOC. Magnetic Susceptibility is interpreted to track changes in weathering regime and/or intensity (e.g. Ellwood et al., 2000; Li et al., 2017) due to the association of magnetic minerals with terrestrial material. Therefore, the relatively invariant MS record for large parts of the Sawpit Gully section indicates that the character and/or flux of weathered material, carrying terrestrial carbon from the continent into the oceans, is likely to have been relatively consistent throughout OAE 2 at this location.

Although the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  record is expected to record carbon-isotope fluctuations of the global atmospheric CO<sub>2</sub> reservoir, interpretation of the Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy is more complex. Below the major red-bed (<28.6 m), TOC and MS values are similar to those for Sawpit Gully, and organic geochemical and petrographic markers indicate that organic matter from the lower part of the Mangaotane B section is primarily of terrestrial origin (Hasegawa et al., 2013). Therefore, the  $\delta^{13}\text{C}_{\text{TOC}}$  record for the interval leading up to red-bed deposition at Mangaotane B is interpreted to reflect global variability in  $\delta^{13}\text{C}$ , and changes in the isotopic composition of the atmospheric carbon reservoir.

However, low TOC coincident with persistently high MS values characterise the major red-bed interval (28.6–44.6 m), which produces a stratigraphy that differs from that of Sawpit Gully (Fig. 3). Higher MS values, coincident with the lowest concentrations of preserved organic matter, indicate that a higher proportion and/or a different type of

lithogenic material was transported to this location on the New Zealand margin. Hasegawa et al. (2013) observed a reduction in organic-carbon and terrestrial kerogen concentrations, coupled with a decline in  $\delta^{13}\text{C}_{\text{TOC}}$  values, in a correlative stratigraphic unit within the adjacent Mangaotane A section. These authors argued for a reduction in the proportion of organic matter sourced from terrestrial plants in the bulk sediment when TOC values are low, shifting the net  $\delta^{13}\text{C}_{\text{TOC}}$  signature towards the lighter compositions of marine organic matter. Although  $T_{\text{max}}$  and HI could not be measured at Mangaotane B because of exceedingly low TOC contents below instrumental detection limits, given the close proximity of the two Mangaotane sections, the organic geochemistry indices of the Mangaotane A red-bed are considered to be directly applicable to the red-bed of Mangaotane B. Thus, we infer that variable mixing of marine and terrestrially derived organic matter in the upper part of the Mangaotane B section within the major red-bed likely obscures the global carbon-isotope signature, which is superimposed on the changes in global carbon cycling.

#### 4.2 $\delta^{13}\text{C}_{\text{TOC}}$ stratigraphy for Sawpit Gully and Mangaotane B: Global versus local signatures

The character of the Sawpit Gully and Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphies differ considerably within the CIE interval defining the OAE 2 proper. While palynofloras support overall correlations to the C/T boundary (Hasegawa et al., 2013), they do not allow for detailed, independent correlations within the CIE interval between the Sawpit Gully and Mangaotane B sections. However, both sites show a clear 2 ‰ positive  $\delta^{13}\text{C}_{\text{TOC}}$  excursion at the start of the interval away from pre-excursion values of -25 ‰, followed by a return to pre-excursion values at the end of the CIE (Fig. 8). The onset of the CIE in both independent sections is abrupt, and the replication between the two sites suggests that stratigraphic condensation is minimal, but cannot be entirely ruled out. A trough superimposed on the

beginning of the positive CIE can be seen in both sections (red band in Fig. 8) but is more pronounced in Mangaotane B (Minimum 1) than in Sawpit Gully. When both New Zealand sections are compared to the Eastbourne  $\delta^{13}\text{C}_{\text{carb}}$  stratigraphy, the timing of this trough in the early phase of OAE 2 appears to coincide with that of the PCE, which previously has not been well documented in the southern hemisphere. The small rise in  $\delta^{13}\text{C}_{\text{TOC}}$  within the trough in the Sawpit Gully section possibly separates the two pulses of the PCE recorded in other proxies elsewhere in the northern hemisphere (Jenkyns et al., 2017). It is noteworthy that MS values increase in both sections just after the PCE, at the onset of the CIE plateau, possibly indicating a change in weathering regime and/or weathering intensity related to warmer climates after a period of cooling.

The remaining part of the Sawpit Gully stratigraphy above the stratigraphic expression of the PCE resembles the classically defined OAE 2 section in Eastbourne (Fig. 8). By contrast, the  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy for Mangaotane B shows two additional minima, both of large amplitude, displaying shifts of  $\sim 3$  and  $\sim 5$  ‰ towards more negative values for Minimum 2 and Minimum 3, respectively (Fig. 3). Such extreme negative shifts are not typical for other OAE 2 sections, although negative excursions in the ‘plateau’ phase have been recorded in  $\delta^{13}\text{C}_{\text{carb}}$  records from the Central European shelf sea, possibly related to the elevated presence of isotopically light carbon in bottom waters (Voigt et al., 2007, 2008). In the Mangaotane B stratigraphy, Minima 2 and 3 also correlate with anomalously low TOC contents (see SM for details). As discussed in section 4.1, these offsets are possibly caused by variable mixing of marine and terrestrially derived organic matter that obscures the global carbon-isotope signatures, as Cretaceous marine organic matter generally had a more negative carbon-isotope signature than contemporaneous terrestrial organic matter, likely due to elevated atmospheric  $\text{CO}_2$  levels relative to today (e.g. Dean et al., 1986; Hayes et al., 1999). Alternatively, these minima could be the result of variable mixing between different terrestrial

sources of organic matter with differing  $\delta^{13}\text{C}_{\text{TOC}}$ . Carbon-isotope values are variable within terrestrial higher plants (Gröcke, 2002), so that Minima 2 and 3 could be explained by an elevated proportion of terrestrial organic matter with a relatively low  $\delta^{13}\text{C}_{\text{TOC}}$  signature, as is the case for OAE 2 black shales in Italy (Farrimond et al., 1990; Kuroda et al., 2007). Regardless, the potential combination of changing organic matter provenance and diminished instrumental accuracy associated with low C concentrations during Minima 2 and 3 indicate that this part of the Mangaotane stratigraphy will likely mask global  $\delta^{13}\text{C}$  variations. To facilitate comparison of the Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy with globally distributed records, only samples with >0.1 wt % TOC were considered to have reliable carbon-isotope values. Filtering the  $\delta^{13}\text{C}_{\text{TOC}}$  dataset in this manner removes Minima 2 and 3 from the Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  record, but Minimum 1 is still visible (curve E in Fig. 8). Importantly, the screened  $\delta^{13}\text{C}_{\text{TOC}}$  record, including those for the lower part of the Mangaotane B section below the major red-bed, no longer show a systematic relationship with TOC content and the  $\delta^{13}\text{C}_{\text{TOC}}$  record resembles the Sawpit Gully curve. Thus, only the filtered carbon-isotope record for Mangaotane B is interpreted to represent global  $\delta^{13}\text{C}_{\text{TOC}}$  signatures and is considered for the remainder of the discussion. Following screening, the exact termination of the PCE trough and the beginning of the decrease to pre-excursion values cannot be determined with absolute certainty in the Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  record.

It is conceivable that the  $\delta^{13}\text{C}_{\text{TOC}}$  values for Sawpit Gully have also been influenced by local variations, including the source of organic matter, although to a much lesser extent than in Mangaotane B due to order-of-magnitude higher TOC contents. For example, the small positive excursion of  $\sim 0.5\text{‰}$  at the base of the Sawpit Gully section ( $\sim 1.0\text{ m}$ ) and the associated increase in wt % TOC could be the result of an elevated input of terrestrial organic matter with a relatively positive  $\delta^{13}\text{C}_{\text{TOC}}$ . Such minor fluctuations, however, have no significant effect on the general characteristics of the Sawpit Gully curve and do not influence

the overall interpretation of the  $\delta^{13}\text{C}_{\text{TOC}}$  record in the context of changes to the atmospheric  $\text{CO}_2$  and hence the global ocean–atmosphere carbon reservoir during OAE 2.

#### 4.3 $\delta^{13}\text{C}_{\text{TOC}}$ stratigraphy for Sawpit Gully and Mangaotane B: Low versus high resolution

Both the Sawpit Gully and Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  datasets of this study are in good agreement with the previous results for the same localities (Hasegawa et al., 2013) regarding their general OAE 2 profiles. However, the new Mangaotane B and Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  curves have greatly improved temporal resolution. Most importantly, when examined in detail, the high-resolution  $\delta^{13}\text{C}_{\text{TOC}}$  records presented here reveal additional higher frequency oscillations that are not apparent in the lower resolution chemostratigraphy. For example, the Hasegawa et al. (2013) curves do not capture the abruptness of the onset of the CIE, nor the trough that correlates with the PCE. These features demonstrate the importance of high-resolution sampling for capturing the full characteristics of the  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy, hence facilitating global correlation.

The carbon-isotope expression of the trough related to the PCE in both New Zealand sections is noteworthy (Fig. 3, Fig. 6 and Fig. 8). While the carbon-isotope expression of the PCE is readily identified in the northern hemisphere (Fig. 8), the New Zealand margin represents one of only three regions in the southern hemisphere where its signature can be identified, the others being the southern Tibet section of Gongzha formerly situated in the southern hemisphere at  $\sim 35^\circ\text{S}$  (curve H in Fig. 8; Li et al., 2017) and, tentatively, the Peruvian Western Platform (curve G in Fig. 8; Navarro-Ramirez et al., 2016, 2017). Previous studies have not, however, formally correlated the  $\delta^{13}\text{C}$  trough with this cooling and reoxygenation event. Identification of the PCE interval in both New Zealand sections confirms the assumption that these southern high latitudes record global changes in the carbon

cycle, but without accompanying palaeotemperature data, the occurrence of ocean–atmosphere cooling cannot be demonstrated.

#### 4.4 *Astronomical constraints on the timing, duration and character of OAE 2*

The  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy from Sawpit Gully has a sufficiently high resolution to reveal prominent high-frequency cycles that are superimposed on the OAE 2 CIE curve. Spectral analysis reveals a significant wavelength of 3.9 m with respect to stratigraphic thickness that exceeds the 95 % confidence level in the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  record (Fig. 9a; see SM for details). To this end, time-series analysis was conducted to characterise these cycles and determine whether astronomical forcing was a potential driver of the observed changes in  $\delta^{13}\text{C}_{\text{TOC}}$  and whether these cycles could be used to determine the duration of the CIE as recorded in the section.

The Sawpit Gully section does not contain any fauna or lithological indicators that can be used to independently constrain the timing and duration of OAE 2. Therefore, the  $\delta^{13}\text{C}_{\text{TOC}}$  dataset for Sawpit Gully was instead correlated to an equally resolved carbonate  $\delta^{13}\text{C}$  OAE 2 section from Gongzha, Tibet that was situated in the southern hemisphere at  $\sim 35^\circ\text{S}$  during the mid-Cretaceous (Li et al., 2017) (Fig. 6). For the Tibetan section, an orbital timescale was constructed from an eccentricity-paced MS record (Li et al., 2017). This correlation (Fig. 6) provides approximate ages for the onset (94.55 Ma at 4.71 m) and end (93.73 Ma at 37.89 m) of the CIE in the Sawpit Gully stratigraphy. These age constraints were used to create a linear age model and convert the  $\delta^{13}\text{C}_{\text{TOC}}$  record to the time domain. A simple calculation of the sediment thickness of the excursion (33.18 m) divided by the cycle length (3.9 m) indicates that 8.5 cycles can be accommodated within the excursion. We hypothesise that the 3.9 m cycles are paced with short eccentricity (100 ky), indicating a minimum duration of  $\sim 850$  ky for the  $\delta^{13}\text{C}_{\text{TOC}}$  excursion from the onset to return to pre-excursion values (8.5 cycles x

100 ky), which agrees with the upper estimates of  $820 \pm 25$  ky (Li et al., 2017) and  $866 \pm 19$  ky (Sageman et al., 2006) previously reported for the duration of the OAE 2 CIE (see SM for details).

To test the hypothesis that the 3.9 m cycle corresponds to eccentricity, the age-corrected  $\delta^{13}\text{C}_{\text{TOC}}$  record was band-pass filtered with a filter centred at 0.01/ky (100 ky) and a bandwidth of 0.003/ky (77–143 ky). The amplitude variations of the  $\delta^{13}\text{C}_{\text{TOC}}$  dataset were compared with the orbital reference record of eccentricity (Laskar et al., 2004), which was filtered to extract the short (100 ky) eccentricity cycle using the same parameters as the carbon-isotope record. To achieve the best correlation of the filtered  $\delta^{13}\text{C}_{\text{TOC}}$  record with the 100 ky eccentricity cycle, the age of the end of the CIE was adjusted from 93.73 Ma to 93.62 Ma (age-shift of 110 ky), which resulted in a one-to-one correlation with the filtered record. The weak short eccentricity cycles between 94.3 Ma and 93.9 Ma likely resulted in weak pacing of the carbon cycle at this time, which gave rise to the poorly developed cycles observed in the  $\delta^{13}\text{C}_{\text{TOC}}$  record. Uncertainties are estimated as one eighth of an eccentricity cycle (12.5 ky) per data point due to the error from comparison with the orbital reference record.

Spectral analyses of the MS record were also conducted to test the age model derived from  $\delta^{13}\text{C}_{\text{TOC}}$  data (see SM for details). As MS is interpreted to represent variations in the style and/or intensity of weathering input into the marine environment, it is also expected to record orbital cycles (Ellwood et al., 2008) that are more prominent in high-latitude records (Fischer, 1986), such as the palaeo-south Pacific region of our study sites. After applying the age model established from the  $\delta^{13}\text{C}_{\text{TOC}}$  record, spectral analysis revealed obliquity (38 ky) and precession (21 ky) cycles above the 95% threshold in the age-corrected MS dataset (Fig. 9b). Numerical calculations indicate that obliquity has slowed since the Cretaceous because of tidal dissipation, resulting in the observed 38 ky cycle compared to the modern-

day frequency of 41 ky (Laskar et al., 2004). The precise obliquity solution (i.e. temporal accuracy of cycles) is not accurate beyond 65 Ma because of the chaotic evolution of the planetary orbits, but the calculated obliquity frequency of cycles should be robust (Laskar et al., 2004, 2011).

The identification of eccentricity cycles in the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  record, as well as the appearance of obliquity and precession cycles in the MS record for the same section, gives confidence in the robustness of the resultant age model and its ability to constrain the timing and duration of specific events and excursions within OAE 2. On this basis, the position of the Cenomanian-Turonian boundary, nominally dated at 93.9 Ma; (Obradovich et al., 1993; Gradstein et al., 2012; Meyers et al., 2012), could be placed in the second half of the CIE plateau at 27.96 m in the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy (Fig. 6). This position contrasts with the placement of the C/T boundary based on available dinoflagellate biostratigraphy between ~1.9 m and ~16.5 m (Hasegawa et al., 2013). The orbitally tuned age model for the Sawpit Gully  $\delta^{13}\text{C}_{\text{TOC}}$  record also implies that the duration of the CIE associated with OAE 2 was at least  $930 \pm 25$  ky, which is only slightly longer than the upper estimate of  $866 \pm 19$  ky determined previously for the Western Interior Seaway of North America (Sageman et al., 2006). This duration corresponds to a temporal sampling resolution of ~3 ky in Sawpit Gully and ~9 ky in Mangaotane B. The Sawpit Gully record indicates that the onset of the carbon-isotope trough related to the PCE occurred approximately  $30 \pm 13$  ky after the onset of the CIE. This contrasts with the estimate of Li et al. (2017), who placed the trough  $110 \pm 25$  ky after the CIE onset. This apparent discrepancy could be the result of potential stratigraphic condensation of the Sawpit Gully record. The Sawpit Gully dataset further implies that the PCE carbon-isotope trough lasted for at least  $200 \pm 25$  ky and that the CIE plateau phase persisted for  $660 \pm 25$  ky, thereafter recovering to background  $\delta^{13}\text{C}_{\text{TOC}}$  values over  $40 \pm 25$  ky. These constraints contrast with those reported in Li et al. (2017), who estimated



467 durations of  $80 \pm 25$  ky for the carbon-isotope trough related to the PCE,  $370 \pm 25$  ky for the  
468 plateau phase and  $170 \pm 25$  ky for the subsequent recovery to background levels. If it is  
469 assumed that the age model presented here can be reliably extrapolated to the pre-OAE 2  
470 interval, then the observed shift towards more negative  $\delta^{13}\text{C}_{\text{TOC}} \sim 50$  ky prior to the onset of  
471 the event is broadly consistent with the onset of Large Igneous Province activity (Du Vivier et  
472 al., 2015) and might therefore be related to the injection of isotopically light C into the  
473 atmosphere.

474 Despite the warm equable climates of the greenhouse world during OAE 2, the global  
475 temperature regime most likely responded to changes in solar insolation driven by orbital  
476 cycles (e.g. Meyers et al., 2001; Dickson et al., 2017; Li et al., 2017). The concomitant  
477 changes in weathering and nutrient input would have led to varying rates of organic matter  
478 burial at low latitudes and corresponding global carbon-isotope variability forced by  
479 eccentricity. The detection of eccentricity cycles in the  $\delta^{13}\text{C}_{\text{TOC}}$  record in Sawpit Gully  
480 supports this inference. The recording of obliquity cycles in the MS data suggests that  
481 changes in insolation driven by astronomical forcing in the high latitudes of the south Pacific  
482 had a direct impact on changes in weathering regime and/or intensity along the New Zealand  
483 continental margin (Fischer, 1986).

#### 484 485 4.5 *Correlation of the Sawpit Gully and Mangaotane B $\delta^{13}\text{C}_{\text{TOC}}$ stratigraphy with other* 486 *records*

487 Fig. 8 shows the  $\delta^{13}\text{C}_{\text{TOC}}$  stratigraphy for Sawpit Gully and Mangaotane B in  
488 comparison to the C-isotope datasets of other well-defined OAE 2 sections from the former  
489 north European epicontinental pelagic shelf sea (Eastbourne, UK; Tsikos et al., 2004), eastern  
490 edge of the proto-Atlantic Ocean (Tarfaya Basin, Morocco; Kuhnt et al., 2005), Western

Interior Seaway (Colorado, USA; Sageman et al., 2006) and palaeo-Pacific Ocean (Tappu, Japan; Hasegawa et al., 2010) in the northern hemisphere, as well as sections deposited in the southern hemisphere (Western Platform, Peru; Navarro-Ramirez et al., 2016, 2017; and Gongzha, Tibet; Li et al., 2017). Correlation of the new Sawpit Gully and Mangaotane B  $\delta^{13}\text{C}_{\text{TOC}}$  datasets with other carbon-isotope stratigraphies is based on a combination of biostratigraphy, where applicable, and the location of prominent carbon-isotope shifts and  $\delta^{13}\text{C}_{\text{TOC}}$  maxima/minima that are present in the classically defined Eastbourne section (Tsikos et al., 2004).

All of the above carbon-isotope stratigraphies show the characteristic phases of OAE 2 enabling robust correlation of the New Zealand sections to OAE 2 sections from other parts of the world. The fact that the global carbon perturbation associated with OAE 2 is recorded in different archives, including terrestrial organic matter, marine organic matter and carbonates of both deep (pelagic) and shallow-water origin (Fig. 8), is a strong indication that the ocean–atmosphere inorganic carbon system was well mixed at this time. However, even though all of the above sections display similar general characteristics, each curve differs markedly in its higher frequency features. For example, while the curves from Tarfaya (Morocco), Colorado (USA) and the Western Platform (Peru) show C-isotope background levels consistent with those of marine organic matter ( $\delta^{13}\text{C}_{\text{TOC}}$  values of -27 to -29 ‰; Dean and Arthur, 1987), the records from Sawpit Gully, Mangaotane B and Tappu have background values approximately consistent with those of terrestrial organic matter ( $\delta^{13}\text{C}_{\text{TOC}}$  values of -23 to -25 ‰; Dean and Arthur, 1987). These differing carbon-isotope baseline values reflect local differences in their respective organic-carbon sources. Furthermore, the positive CIEs representing the OAE 2 proper vary in magnitude. This phenomenon might be due to local variations in the carbon-cycle response, induced by locality-dependent fluctuations in productivity, the source of organic matter, and/or diagenetic alteration.

The OAE 2 records from New Zealand indicate that oxic conditions were most likely prevalent at mid- to high latitudes, at least locally, along the southwest Pacific margin on the upper continental slope during OAE 2. Records of OAE 2 from Japan, USA and Peru indicate that similarly oxic conditions are likely to have been present in the mid- to high latitudes of the northwest and northeast Pacific and the sub-equatorial eastern Pacific, whereas conditions were likely anoxic in at least the mid-depth waters of the equatorial Pacific Ocean under areas of upwelling at this time, as shown by the organic-rich sedimentary record, particularly on submarine volcanic plateaus (Schlanger and Jenkyns, 1976; Arthur et al., 1987; Schlanger et al., 1987; Takashima et al., 2011). These observations are in agreement with an Earth system model, which suggests oxic conditions in the water column and seafloor of the south-western and north-western Pacific Oceans (Monteiro et al., 2012). Despite apparently oxygen-rich conditions, the partial collapse of benthic ecosystems in New Zealand and Japan indicates that major environmental perturbations occurred before and/or during OAE 2. Deconvolving the cause of this paradox will be the focus of future studies.

## **5. Conclusions**

We present the first high-resolution carbon-isotope stratigraphy from the high southern latitude palaeo-South Pacific Ocean during the warm, ‘super-greenhouse’ oceanic anoxic event spanning the Cenomanian/Turonian boundary at ~94 Ma (OAE 2). To this end,  $\delta^{13}\text{C}_{\text{TOC}}$  datasets have been acquired with an unprecedented temporal resolution of 3 ky and 9 ky for two outcrops at Sawpit Gully and Mangaotane B, respectively, in the New Zealand sector of the palaeo-Pacific Ocean. These new  $\delta^{13}\text{C}_{\text{TOC}}$  records are based primarily on terrestrial organic matter and hence are expected to represent carbon-isotope fluctuations of the atmospheric  $\text{CO}_2$  reservoir and accurately record global changes in the carbon cycle. Both sections show the characteristic chemostratigraphic phases of OAE 2, such as the initial ~2 ‰

excursion in  $\delta^{13}\text{C}$  marking the onset of the CIE, and the transient shift towards lower  $\delta^{13}\text{C}$  values that correlates with the PCE. These distinctive features enable robust correlation of the New Zealand sections to OAE 2 sections from other parts of the world, irrespective of the type of archive, which indicates a globally well-mixed ocean–atmosphere carbon system during OAE 2.

The high-resolution carbon-isotope stratigraphy for the Sawpit Gully section furthermore enables an astronomical age model to be established for this record based on cyclic oscillations, permitting the dating of specific perturbations of the carbon cycle within the  $\delta^{13}\text{C}_{\text{TOC}}$  record. An approach based on cyclic oscillations in the carbon-isotope dataset was used for orbital calibration. This astronomical age model suggests a duration of  $930 \pm 25$  ky for the primary CIE associated with OAE 2, agreeing with estimates from the Western Interior of North America. Further, this model implies a minimum duration of  $200 \pm 25$  ky for the PCE, beginning  $30 \pm 13$  ky after the initial onset of the CIE. The lithologies and ultra-low, sub-percent level TOC contents of the New Zealand sites implies that oxic conditions were developed, at least locally, along the palaeo-Pacific Ocean margins at mid- to high latitudes during OAE 2, while conditions were likely anoxic in mid-depth waters of the equatorial oceans at this time. The major red-bed present in the Mangaotane B section is inferred to record dramatic changes in weathering regime and/or intensity so that other environmental perturbations, in addition to changing redox-conditions, may have given rise to a unique environmental response to OAE 2 in the high-latitude south Pacific region.

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## FIGURE CAPTIONS:

**Fig. 1:** (a) Palaeogeographic reconstruction of the Cretaceous denoting section locations: Colorado, Western Platform, Tarfaya, Eastbourne, Gongzha, Tappu, Mangaotane B and Sawpit Gully (modified from Zhou et al., 2015). (b) Carbonate carbon-isotope stratigraphy from Eastbourne, UK (Tsikos et al., 2004). (c) Map of central New Zealand showing the sampling sites Mangaotane B and Sawpit Gully. Grey shading indicates Albian–Maastrichtian (113–66 Ma) sedimentary marine successions (modified from Hasegawa et al., 2013).

**Fig. 2:** Photograph, taken in a westerly direction, of the main red-bed in the Mangaotane B section, which has a stratigraphic thickness of ~29–36 m.

**Fig. 3:** Stratigraphy, magnetic susceptibility (MS), weight-per-cent total organic carbon (wt % TOC) and  $\delta^{13}\text{C}_{\text{TOC}}$  data relative to Vienna Pee Dee Belemnite (V-PDB) from Mangaotane B from samples collected in 2015. Error bars correspond to 1 SD. Coloured bands describe the carbon-isotope stages.

**Fig. 4:** Map of Sawpit Gully, Marlborough, showing the locations of the sampling transects. Inset map shows part of NZ Topo50 map sheet BS28, Kekerengu, and the 1 km NZ map grid, to facilitate location of the main figure.

**Fig. 5:** Map of Mangaotane B, Raukumara Peninsula, showing the locations of the sampling transects. Inset map shows part of the NZ Topo50 map sheet BE42, Haupoto, and the 1 km NZ map grid, to facilitate location of the main figure.

**Fig. 6:** Stratigraphy, magnetic susceptibility (MS), weight-per-cent total organic carbon (wt % TOC) and  $\delta^{13}\text{C}_{\text{TOC}}$  data relative to Vienna Pee Dee Belemnite (V-PDB) from Sawpit Gully (samples collected in 2015 and 2016) correlated with the  $\delta^{13}\text{C}_{\text{carb}}$  stratigraphy from Tibet (Li et al., 2017). Error bars correspond to 1 SD. Coloured bands describe the carbon-isotope stages, which include a relatively sharp initial increase (yellow), a trough (red), a plateau (green) and a gradual decrease to pre-excursion values with an average of ca. -25 ‰ (blue). The main OAE 2 stages shown as coloured bands result from correlation to the Eastbourne section, as discussed in Section 5.4. The Tibetan succession provides ages at the onset and end of the CIE used to create a linear age model for the Sawpit Gully record. See Fig. 3 for the remainder of the legend.

**Fig. 7:** Cross-plot of HI and  $T_{\text{max}}$  of selected samples from Sawpit Gully and Mangaotane B, whose contained organic matter is most likely of terrestrial origin.

**Fig. 8:** Comparison between the carbon-isotope stratigraphy of A) carbonate carbon from Eastbourne, UK (Tsikos et al., 2004); B) marine organic carbon from S75, Tarfaya Basin, SW Morocco (Kuhnt et al., 2005); C) marine organic carbon from Portland-1, Colorado, USA (Sageman et al., 2006); D) terrestrial organic carbon from Sawpit Gully, New Zealand (this study); E) marine and terrestrial organic carbon from Mangaotane B, New Zealand (this study; black data points derive from samples with TOC > 0.1 wt % representing global

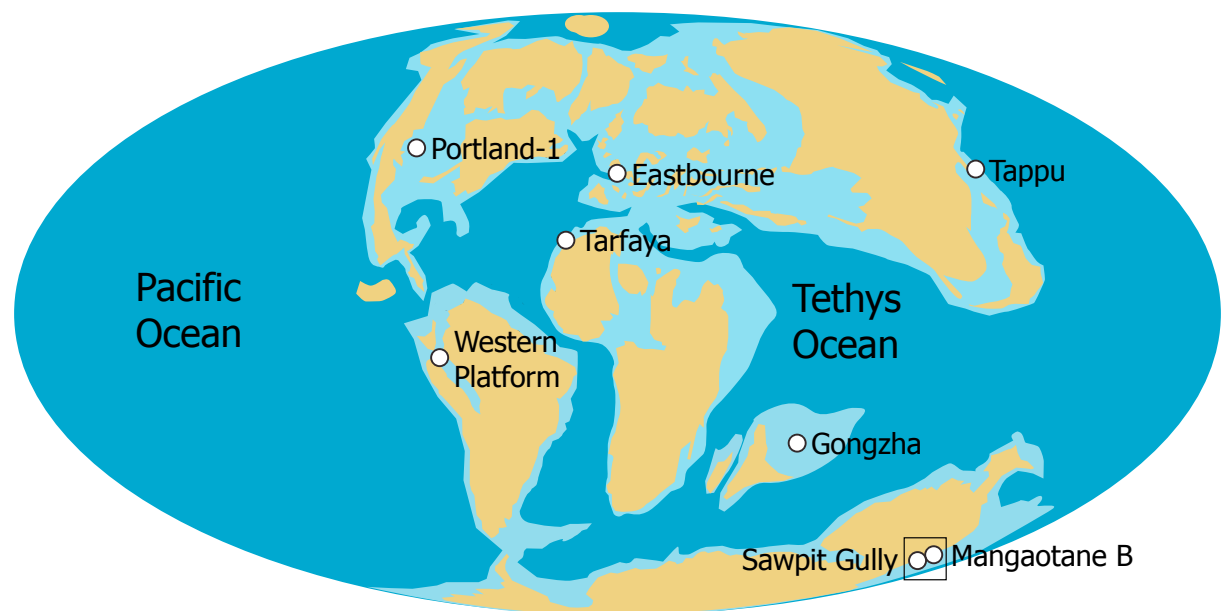
signatures; grey data points derive from samples with  $\text{TOC} \leq 0.1 \text{ wt } \%$ ); F) terrestrial organic carbon from Tappu, Hokkaido, Japan (Hasegawa et al., 2010; Nemoto and Hasegawa, 2011); G) marine organic carbon from Quebrada Chinchin and Piedra Parada, Western Platform, northern Peru (Navarro-Ramirez et al., 2016, 2017) and H) carbonate carbon from Gongzha, southern Tibet (Li et al., 2017). Squares in the Sawpit Gully and Mangaotane B sections show the data from Hasegawa et al. (2013). Coloured bands describe the carbon-isotope stages.

**Fig. 9:** (a) Spectral analysis of the main part of the  $\delta^{13}\text{C}_{\text{TOC}}$  excursion (CIE) across the OAE 2 interval from Sawpit Gully reveals three statistically significant cycles. Only the 3.9 m cycle is interpreted to be of relevance (see Supplementary Material for details) (b) Spectral analysis of the age-corrected MS data from Sawpit Gully identifies a 38 ky obliquity cycle and a 21 ky precession cycle.

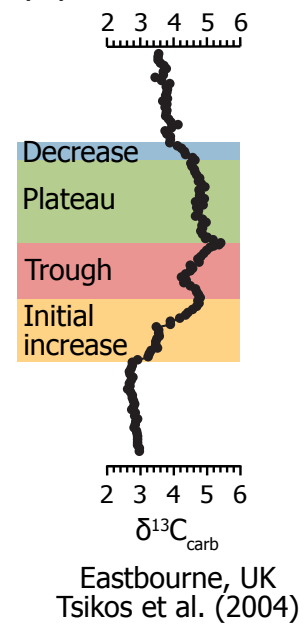
**Fig. 10:**  $\delta^{13}\text{C}_{\text{TOC}}$  age model for Sawpit Gully giving a duration of the CIE as a minimum of  $930 \pm 25 \text{ ky}$  and the position of the Cenomanian–Turonian boundary (dated at 93.9 Ma (Obradovich et al., 1993; Gradstein et al., 2012; Meyers et al., 2012)) in the second half of the plateau. A comparison of the filtered temporal  $\delta^{13}\text{C}_{\text{TOC}}$  record and the filtered eccentricity (Laskar et al., 2004) shows that the  $\delta^{13}\text{C}_{\text{TOC}}$  is in phase with the orbital record and has similar relative amplitude variations. The filtered and unfiltered eccentricity (Laskar et al., 2004) shows weak cycles from around 94.1 Ma to 93.9 Ma.



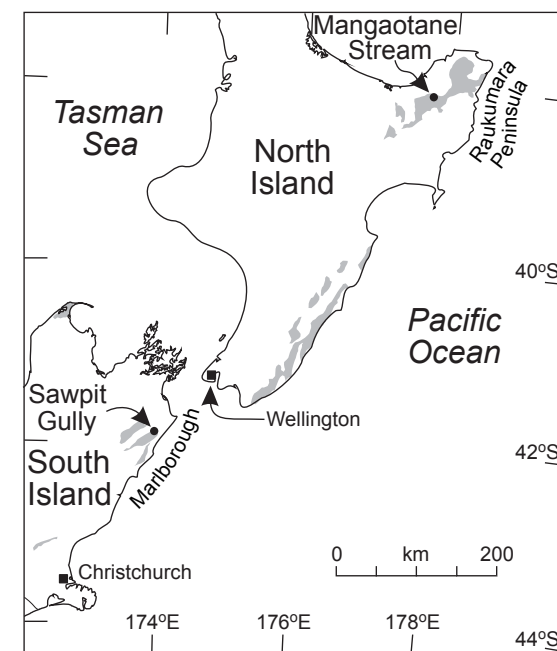
(a)



(b)



(c)

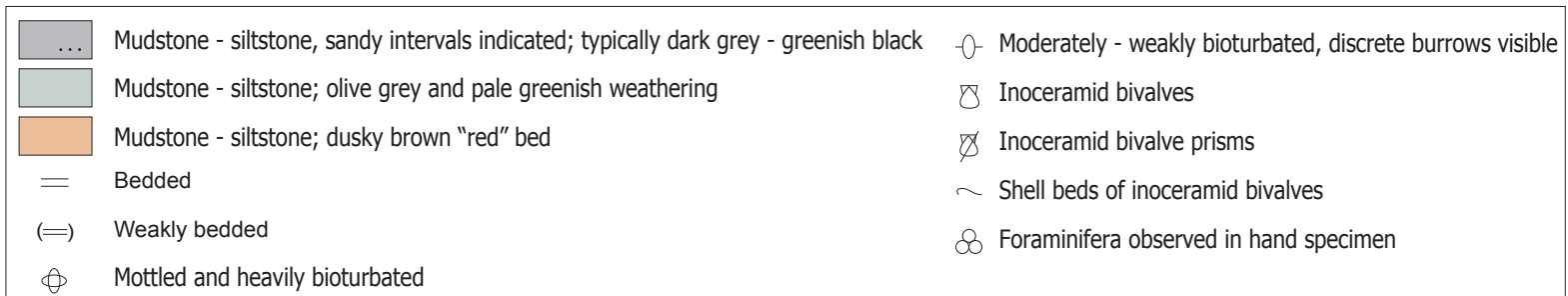
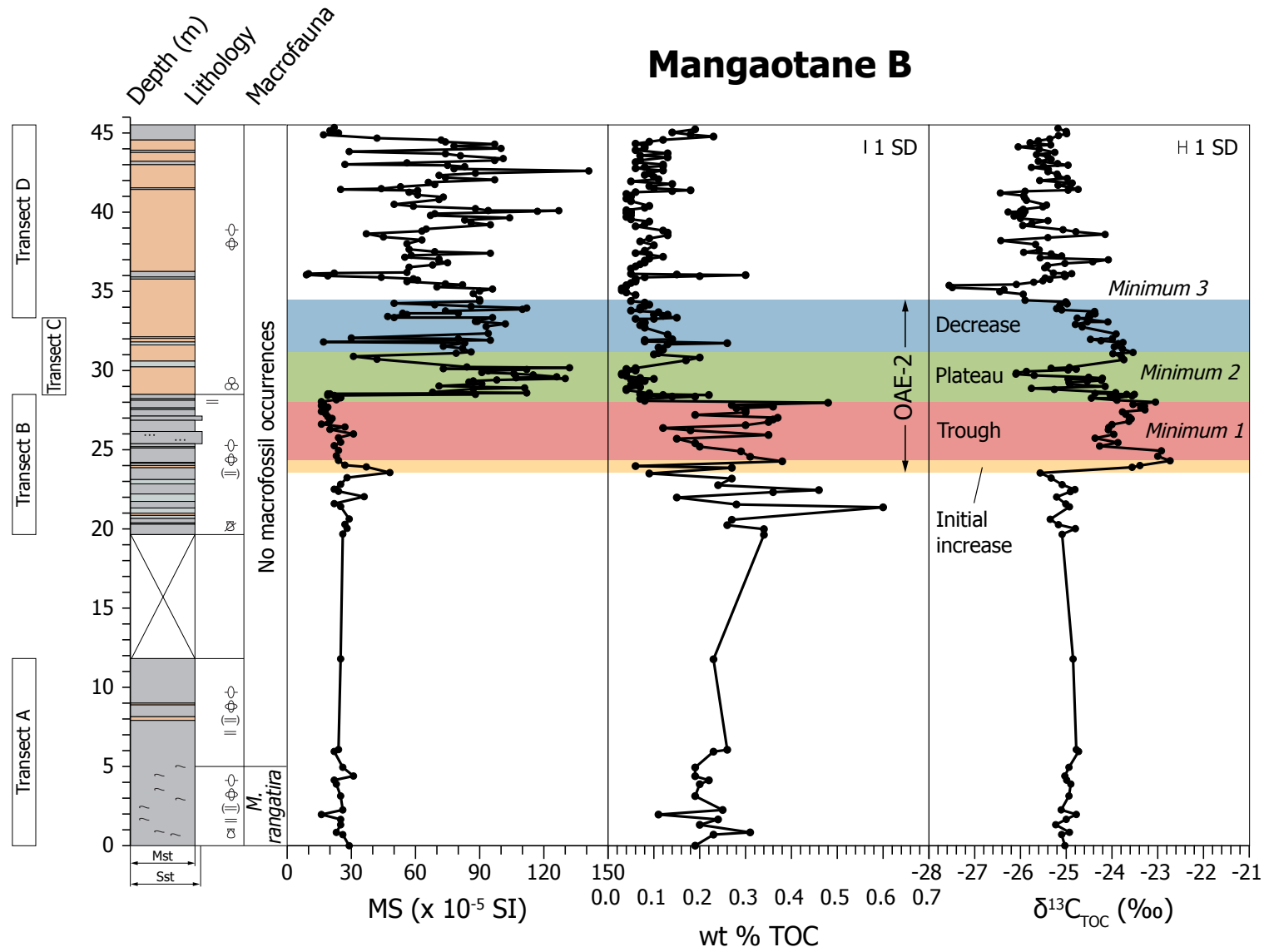


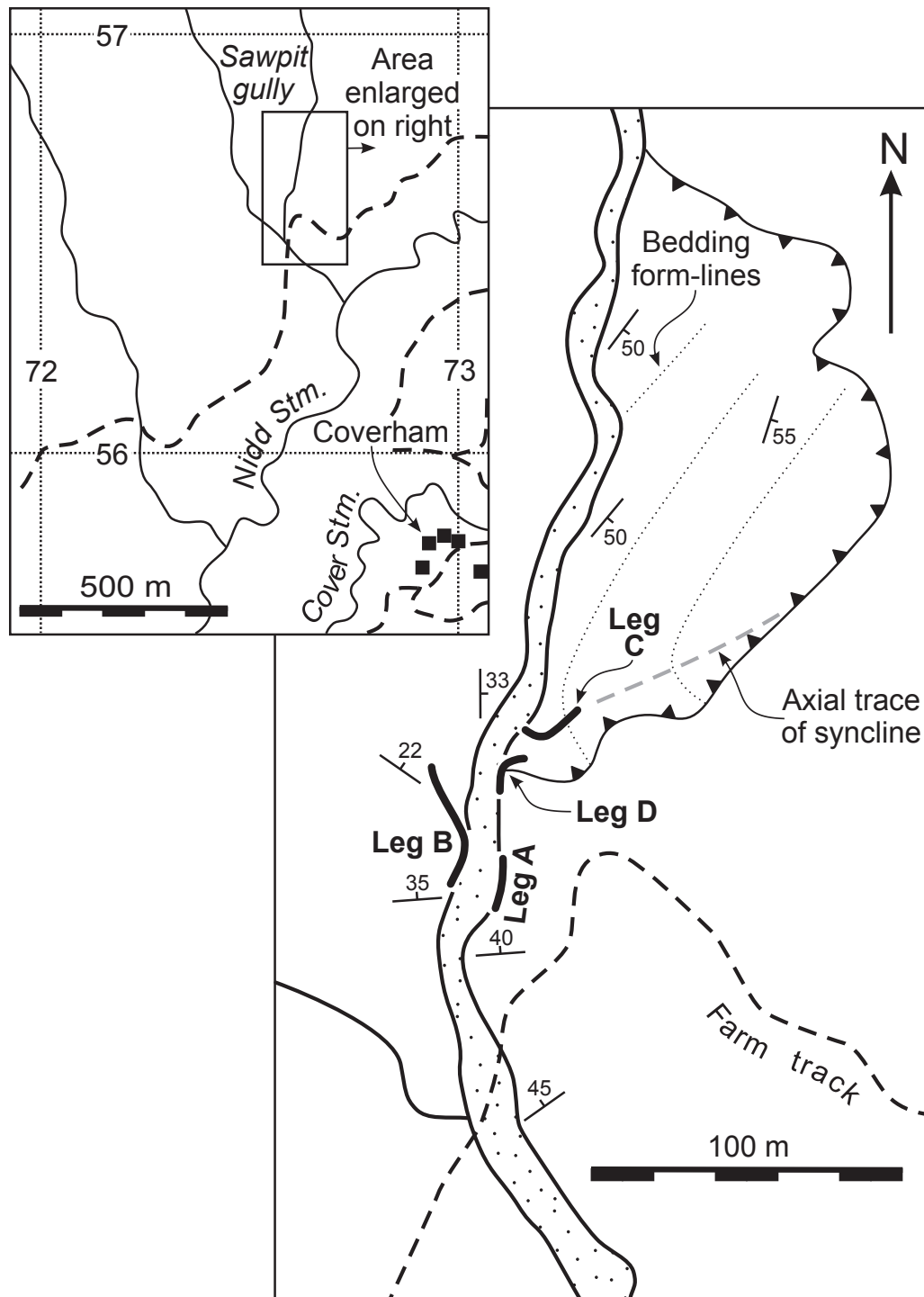


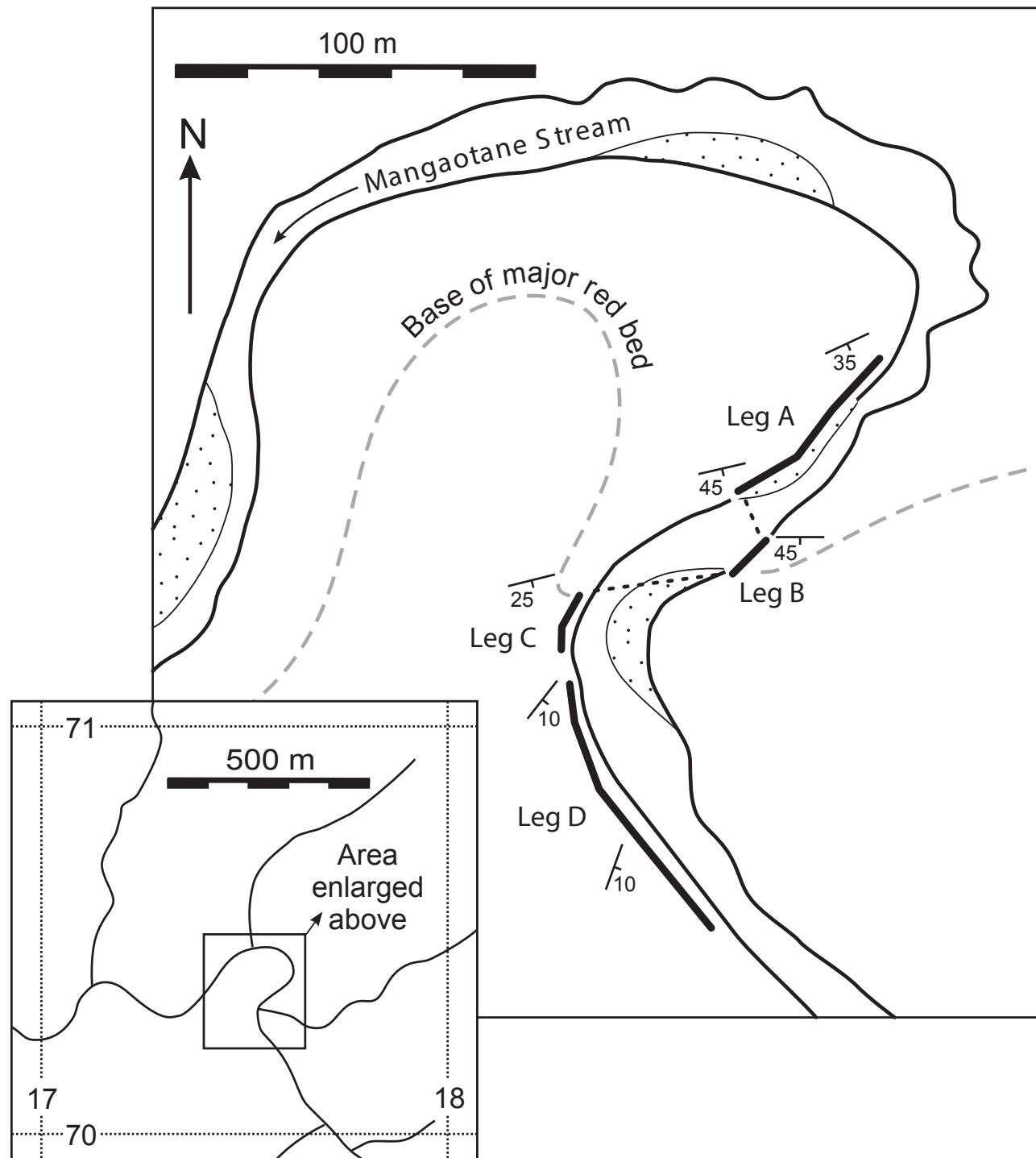




# Mangaotane B







# Sawpit Gully

