Orbital forcing, ice-volume and CO₂ across the Oligocene-Miocene Transition

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Keywords

OMT, Miocene, CO₂, Antarctica, cryosphere, boron isotopes

Key points

• CO₂ levels were relatively low (~265 ppm; 2σ ± 166 ppm) and comparatively stable in the 500 kyrs prior to and during the glaciation.
• CO₂ increased by ~65 ppm during the OMT deglaciation consistent with the latest generation of ice sheet models.
• The timing of the OMT glaciation is most likely controlled by both changes in CO₂ and favourable orbital forcing.

Abstract

Paleoclimate records suggest that a rapid major transient Antarctic glaciation occurred across the Oligocene-Miocene transition (OMT; ca. 23 Ma; ~50 m sea level equivalent in 200-300 kyrs). Orbital forcing has long been cited as an important factor determining the timing of the OMT glacial event. A similar orbital configuration occurred 1.2 million years prior to the OMT, however, and was not associated with a major climate event, suggesting that additional mechanisms play an important role in ice sheet growth and decay. To improve our understanding of the OMT, we present a boron isotope-based CO₂ record between 22 and 24 Ma. This new record shows that δ¹¹B/CO₂ was comparatively stable in the million years prior to the OMT glaciation.
and decreased by 0.7 ‰ (equivalent to a CO₂ increase of ~65 ppm) over ~300 kyrs during the subsequent deglaciation. More data are needed but we propose that the OMT glaciation was triggered by the same forces that initiated sustained Antarctic glaciation at the Eocene-Oligocene transition; long-term decline in CO₂ to a critical threshold and a superimposed orbital configuration favourable to glaciation (an eccentricity minimum and low-amplitude obliquity change). When comparing the reconstructed CO₂ increase with estimates of δ¹⁸Osw during the deglaciation phase of the OMT, we find that the sensitivity of the cryosphere to CO₂ forcing is consistent with recent ice sheet modelling studies that incorporate retreat into subglacial basins via ice cliff collapse with modest CO₂ increase, with clear implications for future sea level rise.

1. Introduction

Over the last 55 million years Earth’s climate has gradually cooled but superimposed upon this long-term evolution are numerous intervals of more rapid change (Zachos et al., 2008). One such example of rapid change is the glaciation that coincides with the Oligocene-Miocene stratigraphic boundary (terminology of Miller et al., 1991; ca. 23 Ma, see Fig. 1). This transient cooling event is evident in the oxygen isotope record as a two-step increase in benthic foraminiferal δ¹⁸O over 200-300 thousand years. The magnitude of this change has typically been estimated to be approximately 1 ‰, and interpreted to represent a temporary expansion in continental ice volume of between 30 and 90 m sea level equivalent (s.l.e) (Liebrand et al., 2011; Mawbey and Lear, 2013; Miller et al., 1991; Pälike et al., 2006a; Pälike et al., 2006b; Paul et al., 2000; Pekar et al., 2002). However, a recent re-evaluation of stacked benthic δ¹⁸O records (Mudelsee et al., 2014), alongside a new oxygen isotope record from IODP Site U1334 in the equatorial Pacific (Beddow et al., 2016), suggests that the excursion is smaller (~0.6 ‰) and that previous work placed too much emphasis on the extremes in the interpretation of the individual records published across the interval. Assuming the same δ¹⁸O to sea level relationship as the late Pleistocene, the re-evaluation of the oxygen isotope excursion suggests a sea level change of up to ~50 m (Beddow et al., 2016). Previous work has suggested δ¹⁸Oice may be less enriched in ¹₆O when ice sheets are smaller (e.g. Langebroek et al., 2010), which would lead to an increase in the sea level change inferred from a δ¹⁸Osw excursion (e.g. Edgar et al., 2007),
however, this effect is likely to be a relatively minor component (15-28%) of the total δ¹⁸O change during the Neogene (Gasson et al., 2016a; Gasson et al., 2016b; Langebroek et al., 2010). Slightly higher ice volume changes are estimated in a study by Liebrand et al., (2017), which uses the benthic δ¹⁸O record from Site 1264 and assumptions about bottom water temperature. That study estimates that the OMT was associated with a change in the East Antarctic ice sheet from near-fully deglaciated to one as large as the modern day. While it is not possible to discount a northern hemisphere contribution to the continental ice budget of the OMT, despite the uncertainties in total ice volume change, Antarctica is likely to have been the main locus of ice growth at this time (DeConto et al., 2008; Naish et al., 2001).

Existing studies have shown that orbital forcing plays a key role in OMT glaciation because its timing is closely associated with the 1.2 Myr minimum in the modulation of the Earth’s orbit and axial tilt (an obliquity ‘node’), as well as a minimum in the 400 kyr long eccentricity cycle (i.e. a very circular orbit), both of which reduce seasonal extremes and increase the chances of winter snowfall surviving the summer ablation season (Coxall et al., 2005; Pälike et al., 2006a; Zachos et al., 2001b) (Fig. 1). However, obliquity nodes and eccentricity minima occur regularly throughout the late Oligocene (Laskar et al., 2004) and the amplitude of the preceding node at 24.4 Ma is more extreme than the one associated with the Oligocene-Miocene transition (Pälike et al., 2006a). Consequently, despite a clear orbital pacing to the OMT glaciation, changes in other boundary conditions are required to fully explain this climate perturbation (Liebrand et al., 2017).

Records of deep-ocean cooling and ice sheet expansion/retreat associated with the OMT glaciation exhibit a number of orbitally paced steps (Lear et al., 2004; Liebrand et al., 2011; Liebrand et al., 2017; Mawbey and Lear, 2013; Naish et al., 2001; Pälike et al., 2006a; Pälike et al., 2006b; Zachos et al., 2001b). There is a ~100 kyr periodicity throughout the OMT in a number of benthic oxygen isotope records, as well as in δ¹⁸Osw (calculated from paired benthic δ¹⁸O and Mg/Ca measurements), which is expressed particularly clearly following the main glaciation (Beddow et al., 2016; Liebrand et al., 2011; Mawbey and Lear, 2013; Zachos et al., 2001b). Statistical analysis of the benthic δ¹⁸O record from ODP Site 1264 across the Oligocene-Miocene suggests that the symmetry of ~100 ky glacial-interglacial cycles changes
across the OMT with a switch to more asymmetric cycles, indicative of longer-lived
ice sheets that survive deeper into insolation maxima (increased ice sheet hysteresis)
together with more abrupt glacial terminations after ~23 Ma (Liebrand et al., 2017).

It has also been suggested that OMT glaciation was associated with a perturbation of
the carbon cycle (Mawbey and Lear, 2013; Paul et al., 2000; Zachos et al., 1997).
Modelling studies (DeConto and Pollard, 2003; Gasson et al., 2012) and proxy
reconstructions (e.g. Foster et al., 2012; Foster and Rohling, 2013; Greenop et al.,
2014; Martínez-Boti et al., 2015; Pagani et al., 2011; Pearson et al., 2009) both
suggest that CO2 plays an important role in controlling the timing of ice sheet
expansion and retreat throughout the Cenozoic. The long-term increase of 0.8‰ in
carbon isotopes from 24 to 22.9 Ma, alongside an increase in benthic foraminiferal
U/Ca has been attributed to an increase in global organic carbon burial and the
associated reduction in atmospheric CO2 (Fig. 1) (Mawbey and Lear, 2013; Paul et
al., 2000; Stewart et al., 2017; Zachos et al., 1997). On the basis of deep-ocean
CaCO3 preservation indicators and estimates of deep-ocean CO3²⁻, an increase in CO2
has also been implicated as one of the driving forces of the deglaciation that followed
the glacial maximum at 23 Ma (Mawbey and Lear, 2013). Yet, published CO2 records
are not of sufficient temporal resolution to test these hypotheses or evaluate the
presence of a CO2 decline that would be expected to accompany an increase in
organic carbon burial prior to OMT glaciation (Fig. 1).

The overall OMT glaciation-deglaciation event as seen in the δ¹⁸O record shows a
duration of about one million years and is largely symmetrical, with little evidence of
ice sheet hysteresis (Beddow et al., 2016; Liebrand et al., 2011; Mawbey and Lear,
2013; Zachos et al., 2001b). While the first generation of Antarctic ice sheet models
suggested that the CO2 threshold for retreat of a major ice sheet was high (>1000
ppm) (Pollard and DeConto, 2005), more recent studies suggest that it is possible to
simulate a more dynamic ice sheet by (i) incorporating an atmospheric component to
the model to account for ice sheet–climate feedbacks, (ii) allowing for ice sheet
retreat into subglacial basins via ice cliff collapse and, (iii) accounting for changes in
the oxygen isotope composition of the ice-sheet (Gasson et al., 2016b; Pollard et al.,
2015). Based on modelling experiments for the early to mid-Miocene Antarctic ice
sheet, a seawater oxygen isotope change of 0.52-0.66 ‰, can be simulated by
changing atmospheric CO₂ between 280 and 500 ppm together with applying an
astronomical configuration favorable for Antarctic deglaciation (Gasson et al.,
2016b). To assess the controls on ice sheet dynamics and the potential applicability of
this new generation of ice sheet models to the OMT glaciation, CO₂ data are required
at substantially higher resolution than is currently available (1 sample per ~500 kyr;
Fig 1). Here, we present a new boron isotope record with an average 50 kyr
resolution across the OMT glaciation and use published δ¹⁸O records to explore the
relationship between ice volume and CO₂ across this interval.

2. Methods and Site information

2.1 Site Location and Information:
We utilize sediments from two open ocean drill site holes: Ocean Drilling Program
(ODP) Hole 926B from Ceara Rise (3°43′N, 42°54′W; 3598 m water depth) in the
Equatorial Atlantic Ocean and ODP Hole 872C situated in the tropical north Pacific
gyre on the sedimentary cap of a flat-topped seamount (10°05.62′N, 162°52.002′E,
water depth of 1082 m). Both sites are currently located in regions where surface
water is close to equilibrium (+/- 25 ppm) with the atmosphere with respect to CO₂
(Fig. 2; (Takahashi et al., 2009)). Age models for Site 926 and Site 872 are from
Pälike et al., (2006a) (and references therein) and Sosdian et al., (2018) updated to
GTS2012 (Gradstein et al., 2012) respectively. Samples from ODP Site 926 were
taken from between 469 and 522 mcd and between 110 and 117 mcd at ODP Site 872.

2.2 Boron isotope measurements
Trace element and boron isotope (described in delta notation as δ¹¹B – permil
variation from the boric acid standard SRM 951; Catanzaro et al., 1970)
measurements were made on the CaCO₃ shells of the mixed-layer dwelling
foraminifera *Globigerina praebulloides* (250-300 μm) at Site 926. At Site 872, mixed
layer dwelling foraminifera *Trilobatus trilobus* (300-355 μm) was analysed. The
foraminifera were cleaned following the oxidative cleaning methodology of Barker et
al., (2003) before dissolution by incremental addition of 0.5 M HNO₃. Trace element
analysis was then conducted on a small aliquot of the dissolved sample at the
University of Southampton using a ThermoFisher Scientific Element XR to measure
Mg/Ca for ocean temperature estimates and Al/Ca to assess the competency of the
sample cleaning. For boron isotope analysis the boron was first separated from the Ca
(and other trace elements) matrix using the boron specific resin Amberlite IRA 743
(Foster, 2008; Foster et al., 2013). The boron isotopic composition was then
determined using a sample-standard bracketing routine on a ThermoFisher Scientific
Neptune multicollector inductively coupled plasma mass spectrometer (MC-ICPMS)
at the University of Southampton (closely following Foster et al., 2013). The
uncertainty in $\delta^{11}\text{B}$ is determined from the long-term reproducibility of Japanese

2.3 Determining pH from $\delta^{11}\text{B}$
The relationship between $\delta^{11}\text{B}_{\text{calcite}}$ and pH is very closely approximated by the
following equation:

\[
\text{pH} = pK_B^* - \log \left( \frac{\delta^{11}\text{B}_{\text{SW}} - \delta^{11}\text{B}_{\text{calcite}}}{\delta^{11}\text{B}_{\text{SW}} - \alpha_B \cdot \delta^{11}\text{B}_{\text{calcite}} - 1000 \cdot (\alpha_B - 1)} \right)
\] (1)

where $pK_B^*$ is the equilibrium constant, dependent on salinity, pressure, temperature
and seawater major ion composition (i.e. $[\text{Ca}]_{\text{sw}}$ and $[\text{Mg}]_{\text{sw}}$), $\alpha_B$ is the fractionation
factor between the two boron species (1.0272; Klochko et al., 2006) and $\delta^{11}\text{B}_{\text{sw}}$ is the
boron isotope composition of seawater. In the absence of changes in the local
hydrography, variations of atmospheric CO$_2$ have a dominant influence on pH and
$[\text{CO}_2]_{\text{aq}}$ in the surface water.

2.3.1 Vital effects
Although the $\delta^{11}\text{B}$ of foraminifera correlates well with pH and $[\text{CO}_2]_{\text{aq}}$ the $\delta^{11}\text{B}_{\text{calcite}}$ is
often not exactly equal to $\delta^{11}\text{B}_{\text{borate}}$ (e.g. Foster, 2008; Henehan et al., 2013; Sanyal et
al., 2001). For instance, while the pH sensitivity of $\delta^{11}\text{B}$ in modern $G.\text{bulloides}$ is
similar to the pH sensitivity of $\delta^{11}\text{B}$ in borate ion, the relationship between pH and
$\delta^{11}\text{B}$ falls below the theoretical $\delta^{11}\text{B}_{\text{borate}}$-pH line (Martínez-Botí et al., 2015) (i.e a
lower $\delta^{11}\text{B}$ for a given pH). This effect has been attributed to the dominance, in this
asymbiotic foraminifer, of respiration and calcification on the foraminifer's
microenvironment, which both act to drive down local pH (Hönisch et al., 2003;
Zeebe et al., 2003). In contrast, photosynthetic processes in symbiont-bearing
foraminifera can cause the pH of the micro-environment to be elevated above that of
the ambient seawater (Henehan et al., 2013) and the magnitude of the pH elevation
determines the offset between $\delta^{11}B_{\text{borate}}$ and $\delta^{11}B_{\text{calcite}}$, which is expressed in a species-specific calibration (Henehan et al., 2016; Hönsich et al., 2003; Zeebe et al., 2003). In order to use modern calibrations further back in time, when the foraminifera were growing under different $\delta^{11}B_{\text{sw}}$, it is necessary to also correct the calibration for the $\delta^{11}B_{\text{sw}}$ to avoid overcorrecting for vital effects (see Supp. Fig. 1, 2). Here we adjust the modern calibration intercept using:

$$c_{\delta^{11}B_{\text{sw}}} = c_0 + \Delta \delta^{11}B_{\text{sw}} (m_0 - 1) \quad (2)$$

where $c_0$ and $m_0$ are the intercept and slope of the calibration at modern $\delta^{11}B_{\text{sw}}$ and $\Delta \delta^{11}B_{\text{sw}}$ is the difference in $\delta^{11}B_{\text{sw}}$ between modern $\delta^{11}B_{\text{sw}}$ and the $\delta^{11}B_{\text{sw}}$ of interest (calculated from the mid-point in the OMT $\delta^{11}B_{\text{sw}}$ range; see below). Using the calibration corrected for OMT $\delta^{11}B_{\text{sw}}$ leads to a marginally higher calculated $\delta^{11}B_{\text{borate}}$ (~0.25 %o and hence lower pCO$_2$) compared to the modern calibration.

At Site 872 we measure $T.\ trilobus$ from the 300-350 $\mu$m size fraction and use the calibration of Sanyal et al., (2001) with a modified intercept so that it passes through the core top value for the related $T.\ sacculifer$ (300–355 $\mu$m) from ODP 999A (Seki et al., 2010) to correct for vital effects (Sosdian et al., 2018):

$$\delta^{11}B_{\text{borate}} = (\delta^{11}B_{T.\ trilobus} - 2.69)/0.833 \quad (3)$$

At Site 926 $G.\ praebulloides$ was measured from the 250-300 $\mu$m size fraction. Studies based on the change in $\delta^{13}C$ and $\delta^{18}O$ with size fraction have shown that at the Oligocene-Miocene transition $G.\ praebulloides$ appears to be symbiotic (Pearson and Wade, 2009), in contrast to the asymbiotic modern $G.\ bulloides$ that is considered to be its nearest living relative. Consequently the modern $\delta^{11}B$-pH calibration of $G.\ bulloides$ (Martinez-Botí et al., 2015) is not applicable. Instead we use the calibration for the symbiotic foraminifera $T.\ sacculifer$. In the absence of a $T.\ sacculifer$ calibration for the 250-300 $\mu$m size fraction, we apply the same calibration as at Site 872 from Sosdian et al., (2018). We then use the close temporal overlap between the data from our two sites and with the different species to examine the validity of these vital effect assumptions.

2.3.2 Parameters for calculating pK$_B$
Temperature changes across the Miocene-Oligocene boundary are assessed here using Mg/Ca derived temperatures. SSTs are calculated from tandem Mg/Ca analyses using the generic Mg/Ca temperature calibration of Anand et al., (2003). Adjustments were made for changes in Mg/Ca$_{sw}$ using the records of Brennan et al., (2013) and Horita et al., (2002) and correcting for changes in dependence on Mg/Ca$_{sw}$ following Evans and Muller, (2012) using $H$ = 0.42 calculated from T. sacculifer (Delany et al., 1985; Evans and Muller, 2012; Hasiuk and Lohmann, 2010). We apply a conservative estimate of uncertainty in Mg/Ca-SST of ± 3°C (2σ), to account for analytical and calibration uncertainty, as well as uncertainty in the magnitude of the Mg/Ca$_{sw}$ correction. The temperature effect on CO$_2$ calculated from δ$^{11}$B is ~ 10-15 ppm/°C, consequently uncertainty in SSTs does not significantly contribute to the final pH and CO$_2$ uncertainty. We assume salinity values of the same as modern day at both sites and apply a conservative estimate of ± 3 psu to account for any changes in this parameter through time. Salinity has little effect on CO$_2$ uncertainty calculated using δ$^{11}$B (± 3-14 ppm for a ± 3 ‰). We use the MyAMI Specific Ion Interaction Model (Hain et al., 2015) to adjust pK$_{a}$ for changing Mg/Ca$_{sw}$ based on the [Mg]$_{sw}$ and [Ca]$_{sw}$ reconstructions of Brennan et al., (2013) and Horita et al., (2002) (Supp. Fig. 3).

2.3.3 The boron isotopic composition of seawater (δ$^{11}$B$_{sw}$)

The long residence time of boron in the oceans (~ 10 to 20 Myrs) ensures that major changes in δ$^{11}$B$_{sw}$ during our 2 Myr-long study interval are unlikely (Lemarchand et al., 2000) but it is probable that δ$^{11}$B$_{sw}$ has shifted from its present value of 39.61 ‰ over the past 24 million years. The δ$^{11}$B$_{sw}$ during the Oligo-Miocene is therefore a large source of uncertainty and can have a significant effect on the absolute CO$_2$. For instance, Greenop et al., (2017) showed that the various records of δ$^{11}$B$_{sw}$ diverge significantly in the early Miocene leading to large uncertainties in absolute CO$_2$ estimates across this interval (Sosdian et al., 2018). Here we apply a flat probability for δ$^{11}$B$_{sw}$ in the range of 37.17 to 39.73‰ to encompass the different estimates. The minimum of this range is set to the lower 1σ uncertainty of the smoothed Greenop et al., (2017) record between 22.6 and 23.1 Ma calculated from paired planktic-benthic foraminiferal δ$^{11}$B and δ$^{13}$C analyses. The maximum extent is the average upper 1σ uncertainty of the δ$^{11}$B$_{sw}$ estimates between 21.7 Ma and 24.4 Ma from Raitzsch and Hönnisch, (2013) calculated from the δ$^{11}$B of benthic foraminifera, coupled to
assumptions in past changes in CO₂, using a $\alpha_B$ of 1.0272 (Klochko et al., 2006). This range also encompasses the geochemical modeling estimates of $\delta^{11}B_{sw}$ from Lemarchand et al., (2000) and estimates based on the non-linear relationship between $\delta^{11}B$ and pH alongside estimates of surface to thermocline pH gradients (Palmer et al., 1998; Pearson and Palmer, 2000) from the same time interval (Supp. Fig. 3).

2.4 Estimating absolute CO₂
To define atmospheric CO₂, a second carbonate system parameter, in addition to pH, is required. We use the regression of the Neogene DIC estimates from Sosdian et al., (2018), where deep-ocean DIC is calculated from benthic $\delta^{11}B$ derived estimates of bottom water pH and deep-ocean carbonate ion concentration ([CO₃²⁻]) constrained by the calcite compensation depth (CCD) and [Ca]sw. A linear regression is fitted through the deep-ocean DIC estimates and used to estimate changes in surface DIC relative to the modern value of 2000µmol/kg (Supp. Fig. 3). The major source of uncertainty in the DIC estimates is the $\delta^{11}B_{sw}$ record used to calculate bottom water pH (Sosdian et al., 2018). For instance, the three $\delta^{11}B_{sw}$ record used in Sosdian et al., (2018) results in a wide range of calculated DIC estimates (e.g. 1430 to 1940 µmol/kg at 21.2 Ma). Consequently to incorporate this uncertainty we calculate absolute CO₂ using the DIC regressions determined from the three $\delta^{11}B_{sw}$ records (Sosdian et al., 2018). We undertake a full error propagation of CO₂ using a Monte Carlo simulation (n=10000) by perturbing each data point within the 2σ uncertainty limits in the $\delta^{11}B$ measurement (±0.16-0.85 %), SST (± 3 °C), SSS (± 3 psu), $\delta^{11}B$ seawater (flat probability estimate between 37.15 to 39.51‰) and DIC (±378-502 µmol/kg). We then combine all the Monte Carlo simulations of CO₂ calculated using the three different DIC regressions (n=30000) to determine the mean and 2σ of the final CO₂ estimate (Supp. Fig. 4). By using this approach the final CO₂ estimate (and associated uncertainty) reflects the full spread of DIC estimates while utilizing the overlap in the DIC estimates calculated using different $\delta^{11}B_{sw}$ records to increase the certainty in our CO₂ estimates. This approach results in a slight decrease in the 2σ uncertainty of the combined simulations (n=30000) when compared to the values obtained when using each DIC estimate in isolation. All carbonate system equilibrium constants are corrected for changes in Mg/Caₜw based on the [Mg]ₜw and [Ca]ₜw reconstructions of Brennan et al., (2013) and Horita et al., (2002) (Supp. Fig. 3) following Hain et al., (2015).
2.5 Estimating relative climate forcing

On timescales of less than a few million years, the close relationship between pH and atmospheric CO₂ forcing means that relative pH (ΔpH) can be used to determine the relative climate forcing from CO₂ change (∆F_{CO₂}; see Hain et al. (2018) for a full discussion). The estimates of δ¹¹B seawater, DIC, SSTs, SSSs and the δ¹¹B measurements (and the associated uncertainties) used in the calculation are the same as in Sections 2.3-2.4, however, in analysing ∆F_{CO₂} rather than absolute CO₂ forcing the uncertainty in the δ¹¹Bsw and secondary carbonate system parameter become less significant with the primary source of uncertainty originating from the δ¹¹B_{calcite} measurements (Hain et al., 2018).

∆F_{CO₂} is calculated from ∆CO₂ change using the equation:

\[ ∆F_{CO₂} = 5.32 \ln \left( \frac{C}{C₀} \right) + 0.39 \left( \ln \left( \frac{C}{C₀} \right) \right)² \] (4)

where C and C₀ are the calculated CO₂ values (Byrne and Goldblatt, 2014). Here C₀ corresponds to the oldest sample at 24.02 Ma and the climate forcing is calculated for the rest of the record relative to this point.

3. Results and Discussion

3.1 δ¹¹B and temperature changes across the Oligocene-Miocene transition

Our record from *G. praebulloides* at Site 926 shows high and relatively stable δ¹¹B values (17.1 +/- 0.4 ‰; hence lowest CO₂) prior to and during the OMT glaciation (Fig. 3). After 23 Ma, δ¹¹B decreases in a number of cycles reaching minimum values of 16.3 +/- 0.5 ‰ at 22.5 Ma (highest CO₂). The data from Site 872 extends the record from Site 926 between 21-22 Ma and while the samples from the two sites do not overlap in the time domain there appears to be good consistency with the data from Site 926, adding confidence to our treatment of vital effects for *G. praebulloides* at Site 926 (Fig. 3). When comparing the benthic foraminiferal δ¹⁸O record to our δ¹¹B data, there appears to be a decoupling between the two series in the lead up to the glaciation (Fig. 3). The δ¹¹B record during this interval shows little change, whereas the δ¹⁸O increases by ~ 0.6 ‰ between 23.2-23.1 Ma. During the deglaciation phase,
however, the $\delta^{11}$B rise broadly tracks the decrease in $\delta^{18}$O although the $\delta^{11}$B record shows a transient increase to pre OMT glaciación levels around 22.8 Ma that is less pronounced in the $\delta^{18}$O record. The $\delta^{11}$B data from Site 872 suggest that elevated CO$_2$ levels are only maintained until ~22.2 Ma, after which CO$_2$ returns to approximately pre-OMT event values. More data are needed to determine whether the $\delta^{11}$B change between 22.2 Ma to 22 Ma reflects a trend in CO$_2$ or whether orbital-scale variations have been under-sampled across this interval.

It has been widely hypothesised that a decrease in CO$_2$ prior to the OMT glaciación may have been one of the key triggers of the event (Mawbey and Lear, 2013; Paul et al., 2000; Zachos et al., 1997). Yet, we find no evidence, within the resolution of our data, for a $\delta^{11}$B increase (CO$_2$ decrease) across the benthic $\delta^{13}$C increase that has been suggested to signify organic carbon burial in the lead-up to the OMT glaciación (Paul et al., 2000; Zachos et al., 1997). That said, the relationship between CO$_2$ and positive benthic $\delta^{13}$C excursions is not always straightforward. For example, a $\delta^{13}$C increase during the warming into the Miocene Climate Optimum coincides with a well-documented CO$_2$ increase (Foster et al., 2012; Greenop et al., 2014) suggesting that organic carbon burial was not the dominant control on CO$_2$ during that interval. Consequently, while carbon burial may occur prior to the OMT, other factors may act to keep atmospheric CO$_2$ levels at approximately constant levels.

The Mg/Ca-derived surface ocean temperatures at Site 926 show no clear temperature decrease during the OMT glaciación event (Figure 3), consistent with estimates of thermocline temperatures and planktic $\delta^{18}$O estimates from the same site (Pearson et al., 1997; Stewart et al., 2017). Mg/Ca measured in thermocline dwelling *Dentoglobigerina venezuelana* at Site 926 shows no long-term change between 24.0 and 21.5 Ma, with temperature variations of less than 3°C across the interval, and no reduction in thermocline temperatures during the OMT glaciación (Stewart et al., 2017). In our new record, we see a counterintuitive multi-million year decrease in temperature of ~2°C between 24 and 22 Myrs and no clear relationship between temperature and $\delta^{18}$O$_{benthic}$. Temperatures decrease from ~28°C prior to the OMT, to values comparable to modern at 23 Ma (modern 26.7°C; Schlitzer, 2000). Several different factors could explain the lack of coherence between surface water
temperature and the other proxy records such as (i) non-thermal control on Mg/Ca
eq 365 (e.g. salinity; e.g. Hönisch et al. 2013), (ii) variable degree of post-depositional
eq 366 dissolution of higher-Mg phases (Brown and Elderfield, 1996), or (iii) local
influences on surface water temperature such as variability in the position of the ITCZ
or changes in latitudinal heat transport (Hyeong et al., 2014). The inferred
temperature offset between Site 926 and 872 may be real or attributed to the different
taxa used between sites. Further work is needed at multiple sites in order to better
understand the surface ocean temperature change associated with the OMT glaciation.
We should stress, however, that the temperature effect on the calculation of CO₂ from
δ¹¹B is relatively minor and we propagate a large uncertainty in SSTs (3°C; 2σ).

3.2 The relationship between δ¹¹B and δ¹⁸O_sw across the transition at ODP Site 926
Benthic δ¹⁸O is a compound record of local salinity, temperature and global
continental ice volume changes. Salinity changes in the deep-sea are typically
considered negligible and therefore if an independent reconstruction of temperature
can be made the ice volume component (δ¹⁸O_sw) of the δ¹⁸O record can be isolated. At
ODP Site 926, a δ¹⁸O_sw record was developed across the Oligocene-Miocene
transition using Mg/Ca temperature estimates from O. umbonatus (Mawbey and Lear,
2013). To evaluate the relationship between δ¹⁸O_sw and δ¹¹B across this interval we
have interpolated the δ¹⁸O_sw to our δ¹¹B age points and generated crossplots of the
time equivalent data. The crossplots are based on changes in δ¹¹B and relative δ¹⁸O_sw,
rather than CO₂ and ice volume, because the large uncertainties in δ¹¹B_sw and Mg/Ca_sw
make it difficult to analyse the relationship between the two parameters. This
treatment is appropriate because the seawater composition influences absolute values,
but has a negligible effect on relative changes. That said, the uncertainty of the δ¹¹B
and δ¹⁸O_sw records is still relatively large, and there are relatively few data points
defining each line, therefore these patterns should be treated as preliminary. While no
relationship exists between ice volume and δ¹¹B/CO₂ (R² = 0.06, p-value = 0.36)
across the whole dataset, when the δ¹⁸O_sw/δ¹¹B data points are split into peak glacial
conditions (low sea level; Fig. 4 blue data points) and pre/post δ¹⁸O excursion (Fig. 4;
red data points) the data fall along two distinct trends. The exceptions to this finding
are two δ¹¹B data points from within the OMT glaciation that coincide with the
maximum in eccentricity when $\delta^{18}O_{sw}$ values were similar to pre/post OMT event conditions.

Based on the central estimates of the data available, the two different trend lines are statistically significant at the 95% confidence level and thus could reflect the different sensitivity of the ice sheet to CO$_2$ forcing under different orbital forcing. It is possible that the cool summers associated with low eccentricity would enable the ice sheet to expand further for a given CO$_2$ forcing compared to high eccentricity conditions, shifting these points from the other trend lines. Alternatively, the observed relationships could be interpreted as evidence for there being two components to the cryosphere, which respond differently for a given CO$_2$ forcing. Statistical analysis of a long Oligo-Miocene benthic $\delta^{18}$O record from Walvis Ridge suggests that the OMT is characterised by more non-linear interactions compared to other intervals with similarly high amplitude $\delta^{18}$O change, possibly related to cryosphere changes (Liebrand et al., 2017). While we cannot identify the ice sheet that forms during the OMT glaciation, the Greenland ice sheet, the marine-based West Antarctic ice sheet and sections of East Antarctic ice sheet have all been shown to be highly sensitive to CO$_2$ and orbital forcing (DeConto et al., 2008; Gasson et al., 2016b; Pollard and DeConto, 2009). While these new $\delta^{11}$B data show some tentative evidence for both an orbital configuration and CO$_2$ control on ice sheet growth over the OMT, more data are clearly needed to further investigate these relationships.

3.3 $\Delta F_{CO2}$ associated with OMT deglaciation.

To assess the significance of CO$_2$ in driving the OMT deglaciation phase it is instructive to calculate the climate forcing change from the $\delta^{11}$B data. The uncertainty in $\delta^{11}$B$_{sw}$ and the secondary carbonate system parameter become less significant when considering the relative change in CO$_2$ forcing on climate ($\Delta F_{CO2}$) over short timescales (in this case over <1 million years), compared to when calculating absolute CO$_2$ (Hain et al. 2018). To further reduce uncertainty, we estimate the $\Delta F_{CO2}$ between two time windows, identified using the $\delta^{18}$O$_{benthic}$ records (Pälike et al., 2006a). A comparison is made between the peak glaciation (23.1-22.9 Ma) identified from the $\delta^{18}$O$_{benthic}$ record and a snapshot post event when $\delta^{18}$O$_{benthic}$ values have stabilised (22.7-22.2 Ma) following the post-OMT seafloor dissolution event (Mawbey and Lear, 2013). Based on this assessment we calculate that the rebound out of the OMT
glaciation was associated with a change in radiative forcing of 1.15 W/m² (2σ range
0.8-1.5 W/m²). However, we note that while comparing ΔF CO₂ between two time
windows reduces the calculated uncertainty, it may also underestimate the amplitude
of ΔF CO₂ as the CO₂ change associated with the maximum change in δ¹⁸O sw is not
captured.

Our new ΔF CO₂ estimate can then be compared to published estimates of Δδ¹⁸O sw to
investigate the sensitivity of ice to CO₂-forcing over the OMT. Combining several
estimates (Beddow et al., 2016; Mawbey and Lear, 2013; Mudelsee et al., 2014), the
change in δ¹⁸O sw associated with the ΔF CO₂ of ~1.15 W/m² can be estimated at -0.41
±0.19 ‰ (Fig. 5). Intriguingly, this estimate is consistent with the range in Δδ¹⁸O sw
modelled for a range of CO₂ change scenarios by Gasson et al. (2016b) (Fig. 5). In
this way, our data support predictions from new-generation ice sheet models of a
dynamic Antarctic ice sheet during the early Miocene that waxed and waned in
response to both orbital configuration and atmospheric CO₂. However, we note that
the changes in ice volume modelled by Gasson et al. (2016b) require extreme orbits in
favour of Antarctic deglaciation, and it is as yet unclear what effect our observed CO₂
change would cause in these models under variable or average orbital configurations.
Furthermore, the resolution of our data is not sufficient to determine whether the rate
and timing of CO₂ and ice volume change is strictly comparable to that used in the

3.4 CO₂ changes prior to the OMT glaciation
While more robustly determined relative change in ΔF CO₂ is clearly instructive,
absolute reconstructions of CO₂ are required to shed light on the role of atmospheric
CO₂ thresholds in the initiation of the OMT glaciation. Our new δ¹³C CO₂ data
suggest that CO₂ rises from a baseline value of ~265 ppm (2σ ± 166 ppm), to ~325
ppm (2σ ± 218 ppm) following the deglaciation (average CO₂ values are calculated
from the post- and peak- glaciation windows defined in Figure 5). While the
uncertainty on the CO₂ estimates is large, primarily as a result of large uncertainties
on δ¹³C sw and DIC estimates (Supp. Fig 5), our data show that, within 1σ uncertainty
(68% confidence interval; 200-345 ppm), CO₂ is below 400 ppm prior to, and during
the Oligocene-Miocene transition (Fig. 3). Previous estimates of CO₂ across the OMT
are sparse. Nonetheless, the absolute values of CO₂ reconstructed here agree well with
the published alkenone records of Pagani et al., (2005) and Zhang et al., (2013) (when
the data are plotted on the age model in Pagani et al., (2011) and updated to the
Geological Timescale 2012 (Gradstein et al., 2012), as well as leaf stomata CO₂
records of Kürschner et al., (2008) (Supp. Fig. 6). Based on the good agreement
between alkenone and boron-isotope based CO₂ records across the OMT, in figure 6
we have plotted records derived using both methodologies to evaluate the multi-
million year trends in CO₂ leading up to the OMT glaciation. The currently available
data for the late Oligocene are sparse, however it appears that the OMT glaciation
occurs following a multi-million year decrease in CO₂ and when the orbital forcing
was favourable for ice growth. According to our combined multi-proxy dataset, the
CO₂ decline begins at 29.5 Ma from values of ~1000 ppm to a minimum of ~265
ppm at 23.5 Ma (Fig 6).

A potential issue with the interpretation of a long-term late Oligocene CO₂ decrease is
that the CO₂ fall between 27 and 24 Ma is at odds with the ~1‰ secular decrease in
benthic δ¹⁸O across the same interval, interpreted as an interval of climate warming
and reduced ice volume (Mudelsee et al., 2014; Zachos et al., 2001a). One possibility
is that climate – as far as it is represented by benthic δ¹⁸O – and CO₂ were decoupled
during the late Oligocene (as has been proposed for the Miocene; Herbert et al.,
2016). A second possibility is that the relationship between Antarctic climate and
deep-water temperature is not straightforward (Lear et al., 2015). For instance, a
climate modelling study from the Mid-Miocene Climatic Transition suggests that the
emplacement of an Antarctic ice sheet caused short-term Southern Ocean sea surface
warming alongside deep-water cooling (Knorr and Lohmann, 2014). The
hypothesised initiation or strengthening of the Antarctic circumpolar current (ACC)
during the Late Oligocene (Hill et al., 2013; Ladant et al., 2014; Lyle et al., 2007;
Pfuhl and McCave, 2005) may also have resulted in large oceanographic changes,
with impacts on global temperatures and benthic foraminiferal δ¹⁸O, although the
timing of ACC development is uncertain. A third possibility is that the ice volume
accommodated on Antarctica was reduced during the Late Oligocene because of the
tectonic subsidence of West Antarctica below sea level (Fretwell et al., 2013; Gasson
et al., 2016b; Levy et al., 2016). Indeed, tectonic subsidence and a shift to smaller
marine based ice sheets on West Antarctica during the Late Oligocene has been
hypothesized to explain the long-term transition from highly symmetrical to saw-toothed δ18O glacial-interglacial cycles (Liebrand et al., 2017). Finally, it is possible that the current estimates of CO2 do not capture the full extent of the changes across this interval. More work is needed to better understand the relationship between ice volume and global climate changes of the Late Oligocene in order to give further context to the changes in CO2, ice volume and climate across the OMT glaciation.

4. Conclusions

The new CO2 data presented here, when combined with published Oligocene CO2 data, suggests that the timing of the OMT glaciation is controlled by a combination of declining CO2 below a critical threshold and a favorable orbital configuration for ice sheet expansion on Antarctica. This combination of factors has previously been used to explain the inception of sustained Antarctic glaciation across the Eocene-Oligocene transition, potentially pointing to a common behavior of the climate system as CO2 levels approach an ice sheet expansion threshold through the Cenozoic. Our best estimate of CO2 suggests that values were around ~265 ppm (2σ $^{\pm}$ 166 ppm) immediately prior to, and during the OMT glaciation and increased by ~65 ppm during the deglaciation phase. Further work is needed, however, to gain a deeper understanding of the background climate and CO2 conditions during the late Oligocene so that the relative contribution of the different ice sheets to the ice volume changes associated with the OMT glaciation can be better determined.

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Figures Captions

**Figure 1**: Climate and forcing over the Oligocene-Miocene transition. a) Cenozoic oxygen isotope composite (Zachos et al., 2008) (b) Oxygen isotope records from Site 926 (light blue) (Pälike et al., 2006a), Site U1334 (dark blue) (Beddow et al., 2016), Site 1264 (light green) (Liebrand et al., 2011) and Site 1218 (dark green) (Pälike et al., 2006b and references therein). (c) Eccentricity orbital forcing from Laskar et al., (2004). (d) Carbon isotope records from Site 926 (light blue) (Pälike et al., 2006a), Site U1334 (dark blue) (Beddow et al., 2016), Site 1264 (light green) (Liebrand et al., 2011) and Site 1218 (dark green) (Pälike et al., 2006b and references therein). (e) Obliquity orbital forcing from Laskar et al., (2004). (f) Previously published CO2 records from across the OMT glaciation. Alkenone reconstructions (light blue and purple) from Pagani et al., (2005) and (dark blue) from Zhang et al., (2013) plotted on the age model of Pagani et al., (2011) updated to Gradstein et al., (2012). Leaf stomata CO2 reconstruction (yellow diamond) from Kürschnner et al., (2008). The Oligocene-Miocene transition is highlighted in red.

**Figure 2**: Map of study sites and mean annual air-sea disequilibria with respect to pCO2. The black dots indicate the location of the sites used in this study. ODP Site 926 (3°43.148’N, 42°54.507’W) is at a water depth of 3598 m and the modern extent of disequilibria is ~ +22 ppm. ODP Site 872C (10°05.62’N, 162°52.002’E) is at a water depth of 1082 m and the modern extent of disequilibria is ~ 0 ppm. Data are from Takahashi et al., (2009). Plotted using ODV (Schlitzer, 2017).

**Figure 3**: New Oligocene-Miocene SST/CO2 estimates and published climate records. (a) δ18O record from Site 926 (Pälike et al., 2006a and references therein). (b)
Oligocene-Miocene transition $\delta^{11}$B from Site 926 (red) and Site 872 (blue) from this study and Greenop et al., (2017). The data are plotted on inverted axes and the error bars show the external reproducibility at 95% confidence. (c) Oligocene-Miocene transition Mg/Ca temperature estimates from Site 926 (red) and Site 872 (blue) from this study and Greenop et al., (2017). Temperature is calculated using the generic Mg/Ca temperature calibration of Anand et al., (2003). 3°C error bar reflects the 2σ temperature uncertainty that was propagated through the CO$_2$ calculation. (d) Eccentricity orbital forcing from Laskar et al., (2004). (e) Oligocene-Miocene transition CO$_2$ from Site 926 (red) and Site 872 (blue) calculated from $\delta^{11}$B data from this study and Greenop et al., (2017). Dark and light bands show CO$_2$ uncertainty at the 68% confidence interval and the 95% confidence interval respectively at Site 926 (red) and Site 872 (blue). Uncertainty was calculated using a Monte Carlo simulation (n=30000) including uncertainty in temperature, salinity, the DIC relationship, $\delta^{11}$B$_{sw}$ and the $\delta^{11}$B measurement. See text for details of the measurement and uncertainty. (f) Obliquity orbital forcing from Laskar et al. (2004). Orange shaded area highlights the Oligocene-Miocene transition.

Figure 4: The relationship between $\delta^{11}$B and $\delta^{18}$O$_{sw}$. (a) The $\delta^{11}$B record from Site 926 focused on 22.7-23.4 Ma from this study and Greenop et al., (2017). The pink circles highlight $\delta^{11}$B samples that fall within ‘peak glaciation conditions’ but show a better fit on the pre/post OMT glaciation event line (see text for details). Note the axis is reversed. (b) Relative $\delta^{18}$O$_{sw}$ change color-coded for peak glacial (blue) and pre/post glacial conditions (red) (Mawbey and Lear, 2013). Open circles are $\delta^{18}$O$_{sw}$ estimates within the ‘dissolution event’ and therefore bias towards negative values. The dashed black lines show the coincident timing of the two $\delta^{11}$B data points that sit on the pre-/post- OMT glaciation event line and the eccentricity paced high sea level events within the OMT glaciation. Note the inverted axis. (c) Time equivalent crossplot of $\delta^{11}$B (error bars external reproducibility at 95% confidence) and relative $\delta^{18}$O$_{sw}$ (error bars ±0.2‰). The peak glacial (blue) and pre/post OMT glaciation (red) data plot along two separate lines. Dotted lines are the 95% confidence intervals on the fit of the linear regressions. The pink data points fall within the glacial interval (circled in panel (a)) but plot on the pre/post glacial line (see text for details).
Figure 5: Oligocene-Miocene transition relative climate forcing. (a) $\delta^{18}$O record from Site 926 (Pälike et al., 2006a and references therein). (b) Relative climate forcing across the OMT calculated from $\delta^{11}$B data from this study and Greenop et al., (2017) (see text for details). Dark and light bands show the uncertainty on relative climate forcing at the 68% confidence interval and the 95% confidence interval respectively at Site 926 (red) and Site 872 (blue). All climate forcing is calculated relative to the data point at 24.02 Ma. The dashed box and purple shaded area highlights the two windows relative climate forcing is calculated from for the data in panel (c). In order to investigate the step-change in CO$_2$ associated with the deglaciation we have excluded any data within the deep-ocean dissolution event (Mawbey and Lear 2013) between 22.9 and 22.8 Ma where $\delta^{11}$B is highly variable. (c) Relative climate forcing (with 95% confidence interval; red box) for data from this study plotted with an estimate of OMT relative $\delta^{18}$O$_{sw}$ change (-0.41 ±0.19‰) (see text for details). The modelled CO$_2$ from Gasson et al., (2016a) converted to relative climate forcing is also plotted with the model output $\delta^{18}$O$_{sw}$ and shows good agreement with our data (orange circles).

Figure 6: Long-term Oligocene climate and CO$_2$. (a) $\delta^{18}$O record from Site 1218 (Pälike et al., 2006b and references therein). (b) Obliquity orbital forcing from Laskar et al., (2004) (c) $\delta^{11}$B-CO$_2$ from Site 926 (calculated from $\delta^{11}$B data from this study and Greenop et al., (2017)) in red and Site 872 (this study) in dark blue, alkenone-derived CO$_2$ from Zhang et al., (2013) in light blue and $\delta^{11}$B-CO$_2$ from Pearson et al., (2009) in orange. For $\delta^{11}$B-derived CO$_2$ records error bars represent 2σ uncertainty.
Figure 1

(a) δ¹⁸O (‰) vs. Age (Ma)

(b) δ¹³C (‰) vs. Age (Ma)

(c) Eccentricity

(d) Obliquity

(e) CO₂ (ppm)

(f) Comparison of sites U1334, 1264, 926, and 1218

Miocene and Oligocene periods indicated.
Figure 2

Site 926
Site 872
\( pCO_2 \) (ppm)
Figure 3

![Graph showing data points and regression lines](image)

**Age (Ma)**

- δ¹⁸Osw (‰)
- δ¹¹B (‰)

**Regression Lines**

- $y = 0.64x - 10.69$
- $R^2 = 0.82$
- $p$-value < 0.04

- $y = 0.24x - 4.26$
- $R^2 = 0.57$
- $p$-value < 0.02

**Additional Notes**

- $p$-values for significant relationships.
Figure 5

![Graph showing relative climate forcing and δ¹³C changes over time.]

1067 1068 1069 1070 1071 1072 1073 1074 1075 1076 1077 1078 1079 1080 1081 1082 1083 1084 1085 1086 1087 1088 1089 1090 1091 1092