



## 1 Introduction

Physical paleoclimate models predict that with the current rate of ice volume loss (up to 97.8 Gt/yr, Pritchard et al. 2012) several sectors of the West Antarctic marine-based ice sheet will collapse within the coming few centuries (e.g., Joughin et al. 2014). Observations show that glaciers on East Antarctica are also vulnerable to basal melt through warming of the ocean waters when they are grounded below sea level (Greenbaum et al., 2015; Miles et al., 2016), making the East Antarctic ice sheet (EAIS) not as stable as previously thought (Mcmillan et al., 2014). Recent numerical modelling studies are more in line with the observed ice-sheet volume measurements, ~~because~~ <sup>as</sup> they incorporate positive feedbacks to global warming and more complicated physics into these models (Austermann et al., 2015; Deconto and Pollard, 2016; Fogwill et al., 2014; Golledge et al., 2017; Pollard et al., 2015), and show that sensitivity to global warming is particularly high where the ice sheet is grounded below sea level (Fretwell et al., 2013), such as the Wilkes Land basin.

Both on glacial-interglacial (Parrenin et al., 2013) and longer term Cenozoic timescales (Pagani et al., 2011; Zachos et al., 2008), Antarctic ice-volume changes have been mostly linked to changes in atmospheric CO<sub>2</sub> concentrations (pCO<sub>2</sub>, see e.g., Foster & Rohling 2013; Crampton et al. 2016), modulated by astronomical forced changes in solar insolation (e.g., Pälike et al., 2006; Liebrand et al. 2017). Foster & Rohling (2013) compiled past pCO<sub>2</sub> proxy data and associated sea level reconstructions for the last 40 million years (Myr). These data suggest that all ice on West Antarctica and Greenland <sup>may</sup> ~~could be~~ lost under current and near future atmospheric CO<sub>2</sub> conditions (400-450 ppmv) in equilibrium state. Projections of pCO<sub>2</sub> for the future emission scenarios of the latest IPCC Report (2014) show a range between 500 and 1000 ppmv for the year 2100, which could imply additional loss of East Antarctic ice-sheet volume. This range in atmospheric pCO<sub>2</sub> is similar to that reconstructed for the Oligocene epoch (e.g., Zhang et al. 2013), highlighting the importance to constrain near-field sea surface temperatures (SSTs) from the Oligocene Antarctic margin.

EAIS volume changes have been suggested for the Oligocene based on a number of deep-sea δ<sup>18</sup>O records (Liebrand et al., 2017; Pekar et al., 2006; Pekar and Christie-Blick, 2008), which reflect a combination of bottom-water temperature and ice volume. These records show long-term (1-3 Myr) trends: a shift towards lighter δ<sup>18</sup>O after the Oi-1 event and a steady increase towards 27 Ma, then a decrease to 24 Ma and a final increase leading to the Mi-1 event (Beddow et al., 2016; Cramer et al., 2009; Liebrand et al., 2016; Zachos, 2001). The high-resolution records show that these trends are punctuated by strong but transient glaciation events (Hauptvogel et al., 2017; Liebrand et al., 2017, 2016; Pälike et al., 2006). These glaciations are paced by periods of strong 110-kyr eccentricity fluctuations of up to 1% (Liebrand et al., 2017, 2016, 2011). Either these δ<sup>18</sup>O fluctuations are mostly resulting from the waxing and waning of the EAIS, in which case the ice sheet must have been highly dynamic, or they reflect large changes in deep-sea temperature, in which case large SST fluctuations in the region of deep-water formation must be expected. Considering the former, fluctuations between 50% and 125% of the present-day EAIS have been suggested (DeConto et al., 2008; Pekar et al., 2006; Pekar and Christie-Blick, 2008), but this amount of variability has not yet been entirely reproduced by numerical modeling studies (DeConto et al., 2008; Gasson et al., 2016; Pollard et al., 2015). Considering the latter, several studies have suggested that during the Oligocene the southern high latitudes were the prevalent source for cold deep-water formation (Katz et al. 2011; Goldner et al. 2014; Borelli

but also based on paleosea level studies e.g. Gallagher et al 2013



& Katz 2015). Hence, temperature records from the southern high-latitudes, particularly those capturing temperature  
85 changes on million-year as well as orbital timescales, may provide information on the relative contribution of deep-  
sea temperature variability in the  $\delta^{18}\text{O}$  records, and as such should reflect the sensitivity of Antarctic ice sheets to  
 $\text{CO}_2$  concentrations. Assuming that the deep-sea temperature trend captured in the Oligocene  $\delta^{18}\text{O}$  records is related  
to surface water temperature in the Southern Ocean similarly to today (Baines, 2009; Jacobs, 1991), ice-proximal  
SST would potentially gauge the Oligocene deep-sea temperature variability. Only few early Oligocene SST  
90 estimates are available for the Southern Ocean (Petersen and Schrag, 2015; Planck et al., 2014). Obstacles for  
reconstructing Oligocene SST in the Southern Ocean are the paucity of stratigraphically well-calibrated sedimentary  
archives, as well as suitable indicator fossils/compounds within these sediments that can be used to reconstruct SST.  
In this study we use the ratio between several glycerol dialkyl glycerol tetraethers (GDGTs), the  $\text{TEX}_{86}$  SST proxy.  
These resistant organic compounds are often the only fossil remains that preserve, because biogenic carbonate and  
95 silica dissolve under the corrosive bottom-water conditions of the high-latitude Southern Ocean.  
In 2010, the Integrated Ocean Drilling Program (IODP) drilled a sedimentary archive at the boundary of the  
continental rise and the abyssal plain offshore Wilkes Land that does contain a well-dated and complete Oligocene  
sequence: IODP Site U1356 (Fig. 1). We here reconstruct the first SST record from high Southern Ocean latitude  
based on  $\text{TEX}_{86}$  covering almost the entire Oligocene and compare it with the few existing early Oligocene SWT  
100 data from other high latitude Southern Ocean sites (ODP Sites 511 and 689) and with deep-water  $\delta^{18}\text{O}$  records from  
lower latitudes. Although the  $\text{TEX}_{86}$ -SST relation shows scatter in the low-temperature ( $<5^\circ\text{C}$ ) domain (Kim et al.,  
2010, 2008), more regional modern analogue calibration methods exist today to overcome some of the scatter  
(Tierney and Tingley, 2015, 2014). Still,  $\text{TEX}_{86}$  is known to overestimate temperatures at high latitudes due to  
multiple possible biases (Ho et al., 2014; Ho and Laepple, 2016; Schouten et al., 2013). However, there is general  
105 consensus that  $\text{TEX}_{86}$  is able to capture decadal and longer-term temperature trends (Ho and Laepple, 2016; Richey  
and Tierney, 2016), which is why our main focus lies on relative SST changes.  
The Wilkes Land region is one of the regions of East Antarctica sensitive to warming, because most of the bedrock  
lies below sea level today (Fretwell et al., 2013; Gollledge et al., 2017) and Site U1356 may therefore have recorded  
past dynamics of the East Antarctic ice sheet. However, during the Oligocene the ice sheet was likely not marine-  
110 based (Wilson et al., 2012). Still, detailed lithological logging of the section allows the distinction of glacial and  
interglacial deposits (Salabarnada et al., submitted this volume). This enables us to assess long-term evolution of  
SWT in proximity of the ice-sheet as well as the temperature differences between glacials and interglacials on  
orbital time scales, which have implications on the dynamics of the Oligocene Antarctic ice-sheet and its sensitivity  
to climate change.

## 115 2 Materials & Methods

### 2.1 Site description

Integrated Ocean Drilling Program (IODP) Expedition 318 Site U1356 was drilled about 300 kilometers off the  
Wilkes Land coast at the boundary between the continental rise and the abyssal plain at a water depth of 3992 m

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120 (Escutia et al. 2011, see Fig. 1). Today, this site is positioned south of the Antarctic Polar Front (PF) and is ~~thus~~  
under the influence of by Antarctic Bottom Waters (AABW), Lower Component Deep Water (LCDW), and  
Antarctic Surface Water (AASW).

### 2.2 Sedimentology

At present, IODP Site U1356 receives sediments transported from the shelf and the slope as well as *in situ* pelagic  
component. Detailed logging of the sediments recovered in Hole U1356A has revealed that the Oligocene  
125 sedimentary record (between 431.74 and 894.80 meters below sea floor) consists mostly of alternations of laminated  
and bioturbated sediments, mass transport deposits (MTDs), and carbonate beds (Salabarnada et al., submitted this  
volume) (Fig. 2). Samples from the mass-waste beds ~~seem to~~ contain the largest contribution of reworked older  
material transported from the continental shelf (Bijl et al., submitted this volume), while in the other lithologies, this  
component is much more reduced or absent ~~altogether~~.

130 Between 593.4 and 795.1 meters below sea floor, ~~clear~~ alternations between greenish, carbonate-poor laminated and  
grey bioturbated deposits with some carbonate-rich bioturbated deposit. These deposits <sup>are</sup> have been interpreted as  
contourite deposits recording glacial-interglacial environmental variability (Salabarnada et al., submitted this  
volume). Above 600 mbsf these alternations are <sup>not</sup> less clear, and the sediments mostly consist of MTDs (Fig. 2).  
However, between the MTDs greenish or grey laminated deposits and greenish or grey bioturbated deposits are  
135 preserved. Near the bottom of Unit III (around 433 mbsf) a different depositional setting is represented with  
alternations between pelagic clays and ripple cross-laminated sandstone beds (Escutia et al., 2011). Samples  
analyzed for TEX<sub>86</sub> <sup>were</sup> chosen from all the 7 different lithologies (Fig. 2). In particular the carbonate-poor laminated  
and (carbonate-rich) bioturbated deposits <sup>were</sup> have been sampled, so we can test whether the glacial-interglacial  
variability inferred from the lithology is reflected in our TEX<sub>86</sub> data.

### 140 2.3 Oceanographic setting

<sup>ok</sup> The details of the Oligocene Southern Ocean oceanography are not fully understood. Studies suggest most Southern  
Ocean surface and deep water masses were already in place by the Eocene-Oligocene Boundary ~~time~~ (Katz et al.,  
2011). Neodymium isotopes on opposite sides of Tasmania suggest that an eastward flowing deep-water current was  
present since 30 Ma (Scher et al., 2015). A westward flowing Antarctic Circumpolar Counter Current (ACCC) was  
145 already established during the late Eocene (49 Ma; Bijl et al. 2013) (Fig. 1). Opening of the Tasmanian gateway <sup>opening</sup>  
also allowed the proto-Leeuwin current flowing along southern Australia continue eastward (Carter et al., 2004; Stickley  
et al., 2004) (Fig. 1). Despite these reconstructions, numerical modeling studies showed that both Australia and  
South America were substantially closer to Antarctica (Fig. 1) than today, which ~~must have~~ limited throughflow of  
the Antarctic Circumpolar Current (ACC) during the Oligocene (Hill et al., 2013). Moreover, tectonic  
150 reconstructions and stratigraphy of formations on Tierra del Fuego suggest Drake Passage underwent temporal  
closure from 29 Ma onwards following <sup>open</sup> open conditions in the middle and late Eocene (Lagabrielle et al., 2009). If  
throughflow at Drake Passage was ~~indeed~~ limited in the Oligocene (Lagabrielle et al., 2009), the model study of Hill

*Tectonic reconstructions*



155 et al. (2013) suggests that the ACCC was more dominant than the ACC. Antarctica itself was positioned more eastward during the Oligocene relative to today (for ~~most~~ due to true polar wander; van Hinsbergen et al., 2015), leading to a relative northward position of U1356 during the Oligocene compared to today. Because of this, bottom-water formation did likely not occur at U1356A. Instead, bottom-water formation is expected in more southerly positioned shallow basins, and where glaciers extended onto the Antarctic shelf, such as the nearby Ross Sea (Sorlien et al., 2007).

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#### 2.4 Stratigraphic age model U1356

160 Oligocene sediments were recovered in the section from 894.68 mbsf (first occurrence (FO) *Malvinia escutiana*) to 432.64 mbsf (base of chron C6Cn.2n) at IODP Hole U1356A. The shipboard age model (Tauxe et al., 2012) was based on biostratigraphy with magnetostratigraphic tie points and chronostratigraphically calibrated to the Geologic Timescale of 2004 (Gradstein et al., 2004). We follow Bijl et al. (accepted), who recalibrated the existing age tie points to the Geologic Timescale of 2012 (GTS2012, Gradstein et al., 2012). The FO of *Malvinia escutiana* (894.68 mbsf; 33.5 Ma; Houben et al., 2011) and the last occurrence (LO) of *Reticulofenestra bisecta* (431.99 mbsf; 22.97 Ma) and the paleomagnetic tie points were used to convert the data to the time domain (see Fig. 4). For the Oligocene-Miocene Boundary, we also follow Bijl et al. (submitted) who infer a hiatus spanning ~22.5-17.0 Ma between Cores 44R and 45R (~421 mbsf). It is unknown whether hiatuses exist within the Oligocene record, but this is likely considering the presence of MTDs (Salabarnada et al., submitted this volume; Fig. S1). In addition, the poor core recovery in some intervals dictates caution in making detailed stratigraphic comparisons with other records.

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#### 2.5 Glycerol dialkyl glycerol tetraether extraction and analysis

175 A total of 129 samples from the Oligocene part of the sedimentary record (Table S1) have been processed for the analysis of glycerol dialkyl glycerol tetraethers (GDGTs) used for TEX<sub>86</sub>. Sample spacing varies due to variability in core recovery and GDGT preservation. Furthermore, sampling of contorted bedding was avoided. Sample processing involved manual powdering of freeze-dried sediments after which lipids were extracted through accelerated solvent extraction (ASE; with dichloromethane (DCM)/methanol (MeOH) mixture, 9:1 v/v, at 100°C and 7.6 x 10<sup>6</sup> Pa). The lipid extract was separated using Al<sub>2</sub>O<sub>3</sub> column chromatography and hexane/DCM (9:1, v/v), hexane/DCM (1:1, v/v) and DCM/MeOH (1:1, v/v) for separating apolar, ketone and polar fractions, respectively. Then, 99 ng of C<sub>46</sub> internal standard was added to the polar fraction, containing the GDGTs, for quantification purposes (cf. Huguet et al., 2006). The polar fraction of each sample was dried under N<sub>2</sub>, dissolved in hexane/isopropanol (99:1, v/v) and filtered through a 0.45-µm 4-mm-diameter polytetrafluorethylene filter. After that the dissolved polar fractions were injected and analyzed by high performance liquid chromatography/mass spectrometry (HPLC/MS) at Utrecht University. Most samples were analyzed following HPLC/MS settings in Schouten et al. (2007), while some samples (see Table S1) were analyzed by ultra-high performance liquid chromatography/mass spectrometry (UHPLC/MS) according to the method described by (Hopmans et al., 2016).  
185 Only a minor difference between TEX<sub>86</sub> index values generated by the different methods was recorded by Hopmans

were



U1356 might have been susceptible to Thaumarchaeotal community changes, which affect the GDGT-2/GDGT-3 ratio. As GDGT-3 is not included in the nominator of  $TEX_{86}^L$ , changes in the GDGT-2/GDGT-3 ratio will affect  $TEX_{86}^L$ -based SST reconstructions to a large extent. For this reason,  $TEX_{86}^L$ -based calibrations are not the focus of our study, but have been included together with all existing  $TEX_{86}^{(H)}$  calibrations as a supplementary figure (Figure S1).

It has been shown that highest GDGT fluxes are closely linked to highest organic matter, opal (diatom frustules) and lithogenic particle fluxes (Mollenhauer et al., 2015; Yamamoto et al., 2012). A lack of production of sinking particles that can incorporate GDGTs formed in deeper waters prohibits that surface-sediment  $TEX_{86}$  values are biased towards deep-water temperatures (Basse et al., 2014; Mollenhauer et al., 2015; Yamamoto et al., 2012). Still, particular environmental settings (e.g., upwelling regions, regions with oxygen-depleted deep waters, fresh-water surface waters) might favor the transport of a subsurface temperature signal to the sediments (Kim et al., 2012a, 2012b; Lopes dos Santos et al., 2010; Mollenhauer et al., 2015). Also for polar oceans it has been suggested that reconstructed temperatures reflect subsurface temperatures. <sup>512°C</sup> Today Thaumarchaeota are virtually absent in the upper 0-45 m of Antarctic low-salinity surface waters (Kalanetra et al., 2009). Surface water conditions over Site U1356 during the Oligocene were much like present-day regions just south of the Subtropical Front (STF) (see Fig. 1) (Bijl et al., submitted this volume). Thus, there is no reason to believe that  $TEX_{86}$  values are influenced by subsurface temperatures as a result of low-salinity surface waters due to sea ice melt. Therefore, subsurface temperature  $TEX_{86}$  calibrations are not discussed. However, they have been included in Figure S1.

Based on the above, we here use the linear SST calibration in Kim et al. (2010). Despite the inclusion of Arctic surface sediment samples with deviating  $TEX_{86}$ -SST relations, this calibration ( $SST = 81.5 * TEX_{86} - 26.6$  with a calibration error of  $\pm 5.2^\circ C$ ) has been shown to plot onto the annual mean sea surface temperatures of the World Ocean Atlas 2009 (WOA2009; Locarnini et al. 2010) for the  $TEX_{86}$  values obtained from surface samples in the Pacific sector of the Southern Ocean (Ho et al., 2014). However, this calibration is likely to be influenced by regional differences in water depth, oceanographic setting and archaeal communities (Kim et al., 2016; Tierney and Tingley, 2014; Trommer et al., 2009; Villanueva et al., 2015). To evaluate the regional variability of the  $TEX_{86}$ -SST relation, we have compared the linear  $TEX_{86}$  calibration of Kim et al. (2010) with the calibration model of Tierney & Tingley (2014; 2015). The latter calibration is based on a Bayesian spatially varying regression model (BAYSPAR), which infers a best estimate for intersection and slope of the calibration based on an assembly of  $20^\circ$  by  $20^\circ$  spatial grid boxes that statistically fit best with a prior estimate of average SST for our SST record. <sup>3?</sup> The prior <sup>?</sup> for site U1356 is obtained from recent clumped isotope measurements ( $\Delta_{47}$ ) on planktonic foraminifers from Maud Rise (ODP Site 689) (Petersen and Schrag, 2015), which show early Oligocene temperatures of  $12^\circ C$ .

To get an estimate for the long-term average SWT trends and confidence levels a Local polynomial regression model (LOESS) has been applied using R, which is based on the local regression model *cloess* of Cleveland et al. (1992). This method of estimating the long-term average trend is preferred over a running average, because it accounts for the variable sample resolution. For the parameter *span*, which controls the degree of smoothing a value was automatically selected through generalized cross-validation (R-package FANCOVA; Wang 2010).



### 3 Results

#### 260 3.1 Discarding potentially biased $TEX_{86}$ values

Of the 129 samples analyzed, 113 contained sufficient GDGTs to obtain a  $TEX_{86}$  value. However, only 69 of these 113  $TEX_{86}$  values could be used for SST reconstruction as discussed below.

265 Twenty-eight samples were taken within <sup>strata</sup> distorted beddings, likely the distal reaches of MTDs originating from the slope or outer shelf of the Wilkes Land Margin (lithological Units IV, VI, VIII and IX) (Escutia et al., 2011). Hence, samples from these beds may not reflect *in situ* material exclusively. In addition, dinoflagellate cyst assemblages show a high degree of reworked Eocene species below 880.08 mbsf (Houben et al., 2013). To avoid potential bias due to allochthonous input and reworking of older sediments, all samples from MTDs and from below 880.08 mbsf are also excluded from the SST reconstructions. In addition, the clast-bearing deposits of Unit IV and decimeter-thick granule-rich interbeds of Unit VIII (Fig. 2) are interpreted as ice-rafted debris (IRD) deposits (Escutia et al., 270 2011), and thus indicate the presence of ice bergs above the site during deposition of these intervals.

A contribution of terrestrial isoGDGTs can also bias the marine pelagic  $TEX_{86}$  signal, and can be verified by the BIT index (Hopmans et al., 2004; Weijers et al., 2006). In nine samples – none of which were derived from mass waste or IRD deposits – the BIT index value was  $>0.3$ , which indicates that the reconstructed  $TEX_{86}$  temperatures are likely affected by a contribution of soil-derived isoGDGTs (Weijers et al., 2006; Hopmans et al. 2004). For those 275 samples which were analyzed by UHPLC/MS, the composition of brGDGTs was analyzed to see if high BIT index values are the result of high marine brGDGT input, but none of these samples had  $\#rings_{tetra}$  above 0.7 (Sinninghe Damsté, 2016) meaning that a significant portion of the brGDGTs was likely soil-derived. All of these samples ~~were~~ <sup>therefore</sup> discarded.

Furthermore, the  $TEX_{86}$  signal may be influenced by a potential input of isoGDGTs from methanogenic archaea. 280 Since methanogenic Euryarchaeota are known to produce GDGT-0 (Koga et al., 1998), but not crenarchaeol such a contribution may be recognized by GDGT-0/crenarchaeol values  $>2$ , as well as values  $>0.3$  for the Methane Index (MI) (Blaga et al., 2009; Sinninghe Damsté et al., 2009; Weijers et al., 2011; Zhang et al., 2011). Thirteen samples have GDGT-0/crenarchaeol ratios  $>2$  and/or too high Methane Index values (Zhang et al., 2011). Eight of these also had too high BIT index values, and thus an additional five were discarded. No samples had GDGT-

285 2/crenarchaeol ratios  $>0.4$ , indicating no GDGT input from methanotrophs (Weijers et al., 2011). As a final exercise, the Ring Index ( $|\Delta RI|$ ) <sup>was</sup> ~~has been~~ calculated for our dataset to identify all other non-temperature related influences on  $TEX_{86}$  (Zhang et al., 2016). Using  $|\Delta RI| > 0.6$  as a cutoff, two more samples were discarded. All GDGT data and  $TEX_{86}$  values, including that of the discarded samples, are presented in the supplementary material (Figure S2, Table S1). Table S1 also shows the GDGT-2/GDGT-3 ratios, which show a large variation among the non-discarded samples. This justifies our choice for a  $TEX_{86}$ -based SST reconstruction over a  $TEX_{86}^L$ -based SST 290 reconstruction as large shifts in the GDGT-2/GDGT-3 ratio will affect the reconstructed SST trends, particularly below a value of 4 (Taylor et al., 2013).

#### 3.2 Relation between $TEX_{86}$ values and lithology



295 After ~~excluding~~ <sup>may</sup> samples with potentially biased  $TEX_{86}$  values, the remaining, ~~assumed~~ <sup>assumed</sup> *in situ* pelagic temperature  
signal ~~could~~ be interpreted. We note that the short-term variability in the record seems ~~to be~~ <sup>to be</sup> strongly linked to the  
lithology (see Fig. 2); Samples obtained from the greenish laminated, carbonate-poor (glacial) facies produce  
statistically significant ( $p$ -value $<0.002$  of  $t$ -test) lower  $TEX_{86}$  values than values obtained from the grey carbonate-  
rich bioturbated (interglacial) facies. Between 600 and 879 mbsf,  $TEX_{86}$  values are on average 0.51 and 0.56 for the  
300 glacial (laminated) and interglacial (bioturbated) lithologies, respectively. Also in Units III and IV the bioturbated  
beds are associated with relatively higher  $TEX_{86}$  values compared to the values obtained from the (cross-)laminated  
beds, although the difference between intervals in those Units is slightly less.

### 3.3 Oligocene long-term sea surface temperature trend

Based on the linear temperature calibration of Kim et al. (2010) (black curve in Fig. 3A), our  $TEX_{86}$  index values  
give an average SST of 16.3 ( $\pm 5.2^\circ\text{C}$  calibration error) for the Oligocene. Maximum and minimum temperatures are  
305  $25.1^\circ\text{C} \pm 5.2^\circ\text{C}$  and  $8.3^\circ\text{C} \pm 5.2^\circ\text{C}$ , respectively. For 90% of the samples reconstructed temperatures fall between  
 $10.5 \pm 5.2^\circ\text{C}$  and  $20.8 \pm 5.2^\circ\text{C}$  (Fig. 3A). Apart from the highest temperatures around 25.5 Ma, high SSTs are also  
reconstructed for the period around 30.5 Ma (up to  $22.6 \pm 5.2^\circ\text{C}$ ), whereas the interval after 23.5 Ma displays lower  
temperatures. SST variability increases significantly ( $p$ -value $<0.001$  in  $F$ -test) after 26.5 Ma (see Fig. 3B). Before  
26.5 Ma, the variation in the record has a double standard deviation ( $2\sigma$ ) of  $3.6^\circ\text{C}$ , while the  $2\sigma$  is  $6.8^\circ\text{C}$  after 26.5  
310 Ma. We note a strong ( $9.5^\circ\text{C}$ ) SST drop at the lower boundary of what is interpreted as chron C6Cn.2n (23.03 Ma)  
(see Fig. 2 and 4); at the stratigraphic position of maximum  $\delta^{18}\text{O}$  values related to Mi-1 in the deep-sea records  
(Beddow et al., 2016; Liebrand et al., 2011; Pálfi et al., 2006). Unfortunately, ~~due to core recovery issues, limited~~ <sup>lacks resolution</sup>  
high-resolution biostratigraphic control and the nature of the sediments, the age model is generally too crude to  
identify some of the other known transient temperature drops in our record ( $\sim 30$  Ma,  $\sim 24$  Ma) to Oligocene  
315 glaciation-related Oi-events (Fig. 4).

The BAYSPAR approach (Tierney & Tingley 2014; 2015) selects only those  $TEX_{86}$  values from the calibration set  
of Kim et al. (2010) that are relevant for a study site, thereby generating a more regional calibration. The SST curve  
for U1356 based on the BAYSPAR model shows the same trend as the SST record generated with the linear  
calibration, but is consistently offset by  $1.0 \pm 0.7^\circ\text{C}$ , and has a smaller calibration error ( $3.5^\circ\text{C}$ ) (red curve in Fig.  
320 3A). This offset and the smaller calibration error result primarily from the fact that BAYSPAR-calibration does not  
take the polar  $TEX_{86}$  core-top values into account. Instead, it bases its calibration mostly on the modern 30-50°  
northern and southern latitudinal bands (see map in Fig. 3). Nevertheless, the offset lies well within the  $\pm 5.2^\circ\text{C}$   
calibration error of the transfer function from Kim et al. (2010), as well as within the standard error of about  $3.5^\circ\text{C}$   
for the BAYSPAR calibration. In summary, SSTs from the BAYSPAR calibration are very similar to the SSTs  
325 derived with the linear calibration, in absolute values, long-term trends, as well as amplitude of variability.



## 4 Discussion

### 4.1 Oligocene Southern Ocean sea surface temperature estimates

Our  $\text{TEX}_{86}$ -derived Southern Ocean SWT record is the first that covers almost the entire Oligocene. Absolute temperature values are relatively high considering the high-latitude position of Site U1356. However, considering  
330 that  $\text{TEX}_{86}$ -based reconstructed SSTs obtained from interpreted glacial lithologies are generally lower than those  
obtained from interglacial lithologies (Fig. 4), strongly suggests that our  $\text{TEX}_{86}$  record is reflecting temperature. *variability*  
Several lines of evidence seem to support the relatively high Oligocene temperatures reconstructed for Site U1356. *isotopes*  
Dinoflagellate cyst assemblages from the same site (Bijl et al., submitted this volume) mostly contain taxa related to  
those found between the Polar (PF) and the Subtropical Front (STF), where mean annual sea surface temperature is  
335 between 8 and 16°C (Prebble et al. 2013), which is on the low end of our reconstructed SSTs. Furthermore, the  
abundance of *in situ* pollen of temperate vegetation in these sediments (Strother et al., 2017), which most likely  
derive from the Antarctic coastline, also suggests a relatively mild climate. Finally, the abundance of pelagic  
carbonaceous facies in some of the interglacial intervals of these high-latitude is interpreted to occur under the  
influence of northern-sourced surface waters at Site U1356 (Salabarnada et al., submitted this volume). *synchro*

340  $\text{TEX}_{86}$ -derived SSTs for IODP Site U1356 are generally higher than those reconstructed with other proxies in other  
high-latitude Southern-Ocean sites during the early Oligocene (between ~34 and ~32 Ma; Fig. 3A). However, the  
~12°C (standard error:  $\pm 1.1$ -3.5°C) based on clumped isotopes for Site 689 is derived from thermocline-dwelling  
foraminifera, whereas the temperature of the surface waters were likely higher than that at the thermocline (Petersen  
and Schrag, 2015). In addition, when the newest calibration for clumped isotope data is applied (Kelson et al., 2017)  
345 also higher temperature estimates, 12.8-14.5°C, are obtained. Temperatures between 6 to 10°C have been obtained  
from ODP Site 511 (see Fig. 1) based on  $\text{U}^{K}_{37}$  (Plancq et al., 2014) and  $\text{TEX}_{86}$  values (Liu et al., 2009), the latter  
re-calculated with the linear calibration of Kim et al. (2010) used here (Fig. 3A). The influence of the cold Antarctic-  
derived surface current prevailing at Site 511 (Bijl et al., 2011; Douglas et al., 2014) (Fig. 1) might be the reason of  
these colder estimates. *estimates*  
350 Meanwhile, similar to the Eocene, Site U1356 probably represents one of the warmest  
regions around Antarctica during the early Oligocene (Pross et al., 2012), situated at a relatively northerly latitude  
(Hinsbergen et al., 2015) and still under influence of relatively warm proto-Leeuwin current (PLC, Fig. 1) (Bijl et al.  
2011; submitted this volume). *with*

Biota-based temperature reconstructions at such high latitudes can be skewed towards summer, as has also been  
suggested for Site 689 and Site 511 (Petersen and Schrag, 2015; Plancq et al., 2014). An important reason for this  
355 could be the light limitation at high latitudes during winter (e.g., Spilling et al. 2015), which is unfavorable for the  
growth and bloom of phytoplankton, and organisms feeding on phytoplankton. Indeed, isoGDGTs likely require  
pelletting to sink effectively through the water column to the ocean floor (e.g., Schouten et al. 2013). As  
phytoplankton blooms mostly occur during Antarctic summer/autumn, as do their predators, the copepods (Schnack-  
Schiel, 2001), we expect the highest isoGDGT fluxes to the sediment during the summer in the Southern Ocean,  
360 despite their primary production during a different season (Church et al., 2003; Murray et al., 1998; Richey and  
Tierney, 2016; Rodrigo-Gámiz et al., 2015).





#### 4.2 Long-term Oligocene sea surface temperature variability

We aim to use our  $\text{TEX}_{86}$ -based SST record to distinguish between the temperature and the ice-sheet signal captured in the global  $\delta^{18}\text{O}$  trend. Due to the relatively low sample resolution and poor age model of our record in comparison  
365 to the  $\delta^{18}\text{O}$  records (Hauptvogel et al., 2017; Liebrand et al., 2017, 2016; Pälike et al., 2006), we will here focus on  
the long-term temperature trends. We can use the glacial-interglacial alternations in the lithology, which cover the  
period between 32 and 25 Ma, to differentiate between glacial and interglacial reconstructed SSTs (see Fig. 2).  
Separating glacial and interglacial signals allows us to interpret the long-term SST trend, as this removes a potential  
sampling bias towards more glacial or more interglacial deposits. To obtain both long-term glacial and interglacial  
370 SST trends, LOESS curves are plotted through SST estimates from the glacial and interglacial subsets. The LOESS  
curves through the glacial and interglacial data show similar trends (Fig. 4): SSTs increase from the earliest  
Oligocene towards 30.5 Ma, followed by a cooling trend until around 28-27 Ma, followed by warming towards a  
long-term optimum around 26 Ma, and then cooling towards the Oligocene-Miocene transition at 23 Ma. The  
recorded post-Oi-1 SST warming coincides with the disappearance of IRD (Escutia et al., 2011) and sea-ice related  
375 dinoflagellate cysts (Houben et al., 2013) in the same record (Fig. 4). The second temperature optimum between  
26.5 and 25 Ma is characterized by the influx of the dinocyst genus *Nematosphaeropsis* (Bijl et al., submitted this  
volume). This seems to indicate a strong influence of northern-sourced surface waters (PLC, Fig. 1) at Site U1356,  
as this species is currently associated with the Subtropical Front and mean annual temperatures above 11°C (Esper  
and Zonneveld, 2007; Marret and De Vernal, 1997; Prebble et al., 2013).

380 Site U1356 probably was not in the region of deep-water formation, given it probably was the warmest region  
around Antarctica (Pross et al., 2012) likely also in the Oligocene. Nevertheless, because Antarctic Circumpolar  
Countercurrent (ACCC) (Bijl et al., 2013) and bottom-water formation along the associated Antarctic Slope Front  
(ASF) were likely established near U1356 in the early Oligocene (Scher et al., 2015), we expect that bottom-water  
formation was under the influence of the same long-term (million-year) climatic trends recorded in the SST  
385 reconstruction of Site U1356. In turn, Southern Ocean surface-water temperatures were likely relayed to the deep-  
ocean. In fact, Southern Ocean sourced deep waters may have flowed all the way to the north Pacific during the  
Oligocene (Borelli & Katz 2015). This suggests that deep-sea benthic foraminiferal  $\delta^{18}\text{O}$  records have incorporated  
both a temperature and ice-volume signal present in the Antarctic-derived deep-waters. Indeed, the only bottom-  
water temperature record available for the Oligocene, which is based on Mg/Ca ratios of benthic foraminifera from  
390 Site 1218 (equatorial Pacific) (Lear et al. 2004; Fig. 4), shows a long-term warming between 27 and 25 Ma, similar  
to our SST record. Notably, Mg/Ca-based reconstructed bottom-water temperatures at Site 1218 are lower  
than reconstructed SSTs from Site U1356: 3.7°C on average (Lear et al. 2004; Fig. 4). Likely, this difference results  
from the fact that bottom water forms at higher latitudes in the subsurface in winter (Jacobs, 1991), while our  
 $\text{TEX}_{86}$ -based SST record likely reflects summer temperatures. The temperature rise before 30.5 Ma and the  
395 temperature decrease after 25 Ma in our  $\text{TEX}_{86}$ -based SST record cannot be recognized in the bottom-water  
temperature record of Site 1218. Uncertainties about Mg/Ca ratio of the seawater and the influence of a changing  
carbonate ion saturation state of the deep waters at Site 1218 (Elderfield et al., 2006; Lear et al., 2010), as well as a



lack of a high-resolution age model of our record and low sample resolution compromise a detailed comparison between the two records.

400 With the here reconstructed SST record, we can now evaluate what part of the long-term  $\delta^{18}\text{O}$  trend can be explained by high-latitude SST changes. We compare our SST record to the high-resolution Oligocene deep-sea  $\delta^{18}\text{O}$  record from the equatorial Pacific (Site 1218; Palike et al. 2006), which is the only high-resolution  $\delta^{18}\text{O}$  record that covers the entire Oligocene. To enable direct comparison, we have resampled the glacial (values above average  $\delta^{18}\text{O}$ ) and interglacial (values below average  $\delta^{18}\text{O}$ )  $\delta^{18}\text{O}$  trends at Site 1218 following the sample resolution of

405 UI356 (Fig. 4). Part of the long-term benthic  $\delta^{18}\text{O}$  trend from Site 1218 is reflected in our TEX<sub>86</sub>-based SST record (Fig. 4): both glacial and interglacial SST LOESS curves show a decreasing trend towards 27 Ma ~~co-occurring~~ with the long-term 0.6‰  $\delta^{18}\text{O}$  increase observed between 32 and 27 Ma. The subsequent decrease in  $\delta^{18}\text{O}$  is also matched by an increase in the long-term interglacial and glacial SST curves. This means that in this part of the record, long-term bottom-water temperature changes could account for part of the long-term  $\delta^{18}\text{O}$  trends. However, other parts of

410 the record show a mismatch between the long-term SST trend and the long-term  $\delta^{18}\text{O}$  trend. For example, our reconstructed SST record shows increasing temperatures between 33 and 30.5 Ma, while  $\delta^{18}\text{O}$  values decrease only between 33.5 and 32 Ma. Furthermore, reconstructed SSTs start decreasing after 25 Ma, while  $\delta^{18}\text{O}$  values start increasing only after 24 Ma. However, low sampling resolution and uncertainties in the age model might account for these mismatches.

415 The striking similarity between the SST and  $\delta^{18}\text{O}$  long-term trends in the interval between 30.5 and 25 Ma, despite reconstructed SSTs at Site UI356 being likely summer-biased, and deep-water formation likely occurring during wintertime and not exactly in the region of UI356, gives confidence that there is indeed a relation between Southern Ocean SSTs and benthic  $\delta^{18}\text{O}$  values. Although not yet quantifiable due to the above-mentioned uncertainties, this relation will have implications for  $\delta^{18}\text{O}$ -based ice-volume reconstructions. Therefore, we have conducted a thought

420 experiment by ignoring the uncertainties in the TEX<sub>86</sub>-based SST record resulting from low sampling resolution and poor age model, and by assuming the reconstructed SST trend has been relayed to the deep ocean. Then, the reconstructed 1°C cooling in the TEX<sub>86</sub>-based SST between 30.5 and 27 Ma could account for one third of the about 0.75‰  $\delta^{18}\text{O}$  long-term decrease between 32 and 27 Ma (1°C = 0.21–0.23‰; Ravelo & Hillaire-Marcel 2007).

425 Similarly, the late Oligocene warming of 1.6°C in our SST record between 27 and 25 Ma would account for much of the overall 0.50‰ decrease in  $\delta^{18}\text{O}$ . However, the further 0.02‰ decrease of benthic  $\delta^{18}\text{O}$  between 25 and 24 Ma should then be related to ice volume loss, since our SWTs decrease with 2°C at that time. Recent ice-volume calculations by Liebrand et al. (2017) suggest that the ice-sheet volume around 27 Ma is at least as large as today's East Antarctic ice sheet, assuming deep-sea temperatures cannot drop below current bottom-water temperatures

(Fig. 4). However, because their calculations are also based on a constant deep-sea temperature, they are overestimating the long-term growth and decline between 32 and 25 Ma (Liebrand et al., 2017). Hence, if the magnitude of deep-sea temperature long-term change is equal to our long-term reconstructed SST trends, current ice-volume estimates during periods of low benthic  $\delta^{18}\text{O}$  (see Fig. 4) are underestimating the size of the Antarctic ice sheet. After 24 Ma, and towards Mi-1, the long-term  $\delta^{18}\text{O}$  trend rises 0.6‰, while the reconstructed SST trends drop about 3°C. Although this cooling could fully account for the long-term  $\delta^{18}\text{O}$  trend between 24 and 23 Ma, we

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yes yes cannot really tell from Fig 4 record

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clarifies  
not  
sure

"though expert" is what you mean perhaps you hypothesize<sup>13</sup>?



435 acknowledge the existing physical evidence for a profound glaciation event during Mi-1 (Naish et al., 2001).  
Similarly, we fully recognize that ice volume variability took place during the Oligocene on the long term (Liebrand  
et al., 2017; Pekar et al., 2006; Pekar and Christie-Blick, 2008), but we suggest that ice-volume variability before 24  
Ma was less than previously assumed, as part of the variation has to be attributed to likely changes in deep-sea  
temperatures over these timescales.

What there is

controversy  
clarity

#### 440 4.3 Sea surface temperature variability at glacial-interglacial time scales

The offset between the glacial and interglacial LOESS curves is constant over time (Fig. 4). Irrespective of the  
chosen calibration (i.e. TEX<sub>86</sub> or BAYSPAR), SSTs are on average 1.8-3.2°C higher during interglacial intervals  
than during adjacent glacial times. This glacial-interglacial SST difference is smaller than the observed amplitude of

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the variability in our temperature record ( $2\sigma = 3.6^\circ\text{C}$  before 27 Ma), because it takes relatively warm glacials and  
cool interglacial SST values into account. Also considering that part of the  $2\sigma$  variability is due to the relatively  
large calibration error of the BAYSPAR calibration ( $\pm 3.5^\circ\text{C}$ ), the difference of 1.8-3.2°C may be a better  
representation of average glacial-interglacial SST variation than the  $2\sigma$ . If such glacial-interglacial SST variability  
is representative of the wider Southern Ocean region and relayed to the deep-sea, it should be considered when  
interpreting benthic foraminiferal  $\delta^{18}\text{O}$  records in terms of ice-volume variability. As such, a larger variability of  
 $\delta^{18}\text{O}$  than so far assumed (Liebrand et al., 2017; Hauptvogel et al., 2017) should be ascribed to deep-sea temperature  
rather than ice-volume changes. To be more specific, 40-70% of the 1‰ deep-sea  $\delta^{18}\text{O}$  variability over Oligocene  
glacial-interglacial cycles can be related to deep-sea temperature (Fig. 5). However, it is likely that not the entire  
amplitude of SST variability is relayed to the deep-sea. In fact, Mg/Ca-based reconstructed bottom-water  
temperatures from Site 1218 show much less glacial-interglacial variation (1.13°C, Fig. 5) (Lear et al., 2004). This  
suggests that the amplitude of the glacial-interglacial temperature variation of the surface ocean is strongly reduced  
at the nearby bottom-water formation sites. Likely, because bottom water has likely formed in the subsurface during  
winter like today (Baines, 2009; Jacobs, 1991). Still, our record provides additional evidence that polar SST  
experienced considerable variability, both on the short-term glacial-interglacial cycles as well as on the long-term.  
Our data furthermore suggest that glacial-interglacial SST variability increased after 27 Ma (Fig. 3B), which  
suggests a lower contribution of ice volume to the deep-sea  $\delta^{18}\text{O}$  signal during the late Oligocene. A major influence  
of deep-sea temperature on benthic  $\delta^{18}\text{O}$  could explain the level of symmetry in glacial-interglacial cycles in the  
Oligocene (Liebrand et al., 2017), as the temperature would vary in a sinusoidal fashion, whereas ice sheets would  
respond non-linearly to climate forcing. Note that this reasoning is still rather speculative as the sedimentary record  
of Site U1356 lacks the potential to obtain a resolution comparable to that of deep-sea  $\delta^{18}\text{O}$  records. However, ice-  
volume reconstructions from  $\delta^{18}\text{O}$  records on both long-term and short-term time scales should consider that an  
important component of the signal could potentially be ascribed to temperature variability.

variability

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out  
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speculative

#### 4.4 Implication for the Antarctic Circumpolar Current

intensification



Today the ACC and the associated Antarctic divergence isolate the continent from lower-latitude influence and are keeping sea-surface temperatures south of the divergence relatively cool (Orsi et al., 1995). The strength and development of the ACC during the Oligocene depends on the opening (and depth) of the Tasmanian Gateway and Drake Passage (Hill et al., 2013). If the Tasmanian Gateway and the Drake Passage were open to allow for some deep-water throughflow during the Oligocene, modeling of ocean currents suggests that a strong ACC could only develop once the ocean gateways are in line with the latitudinal position of the westerly winds (Hill et al., 2013). Dinoflagellate cysts (Stickley et al., 2004) and neodymium isotopes from east and west of Tasmania (Scher et al., 2015) indicate that significant eastward throughflow started south of Tasmania around 35.80 Ma. Instead, the TEX<sub>86</sub>-based SST record and dinoflagellate cyst record (Bijl et al., this issue) of Site U1356 suggest an influence of these warmer northerly waters also at higher latitudes. This suggests that Site U1356 is not fully isolated by a strong flowing ACC and that the eastward flowing current south of Tasmania might have been deflected towards Site U1356. Alternatively, warmer surface water may have reached Site U1356 through eddy-induced heat transport (Dufour et al., 2015; Thompson et al., 2014). It remains a matter of debate when Drake Passage opened sufficiently to allow for a significant ACC throughflow (Lawver & Gahagan 2003; Livermore et al. 2004; Scher & Martin 2006, 2008; Barker et al. 2007; Lagabrielle et al. 2009; Maldonado et al. 2014; Dalziel 2014). Depending on whether Drake Passage was closed or open, modeling suggests that the westward flowing ACC was stronger or weaker, respectively (Hill et al., 2013). Site U1356 would then be respectively more or less under the influence of the ACC. If Drake Passage was closed, a stronger ACC would be able to transport the warm waters recorded at Site U1356 further along the Antarctic coastal margin and thereby explain also the relatively warm temperatures at Site 689.

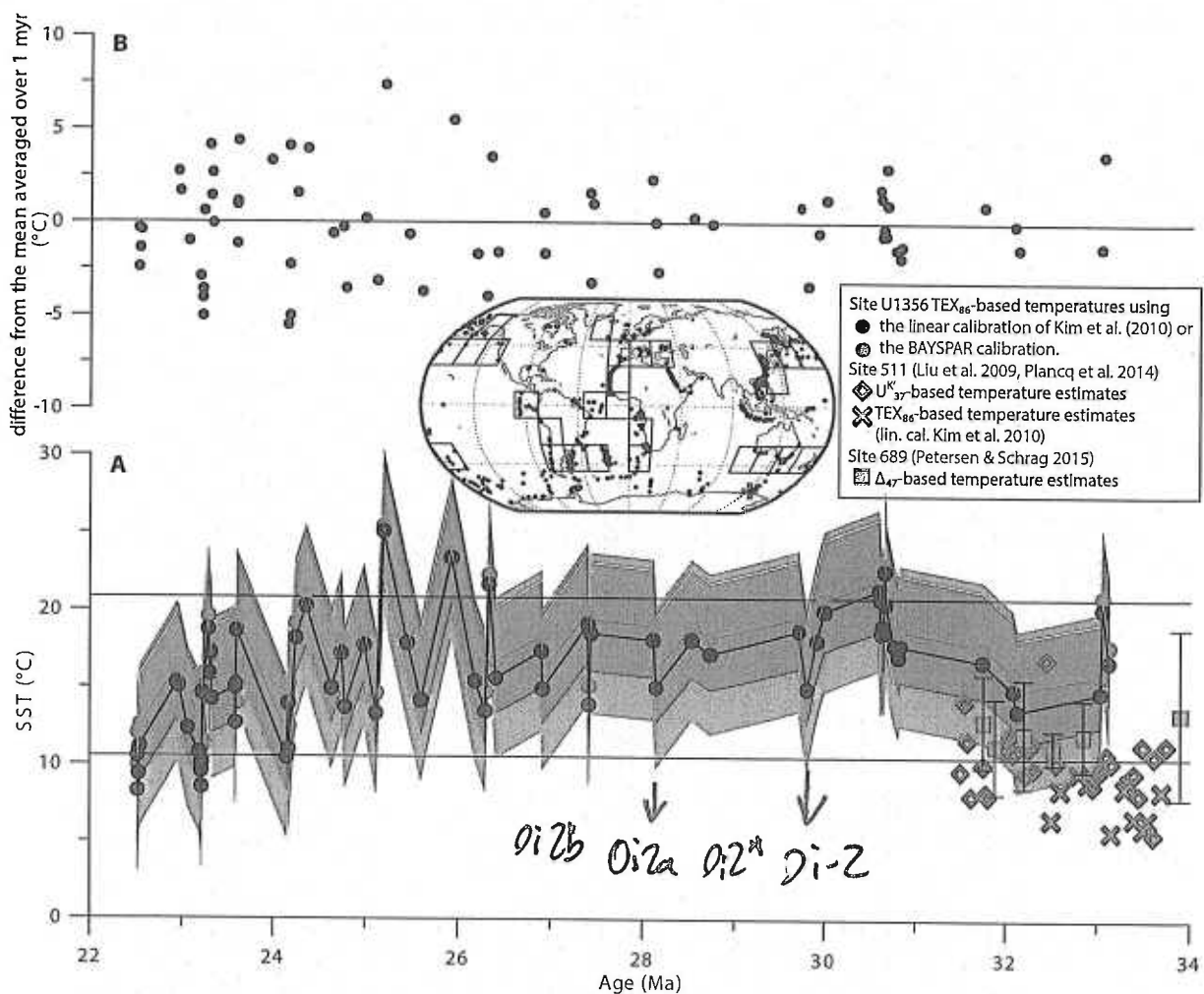
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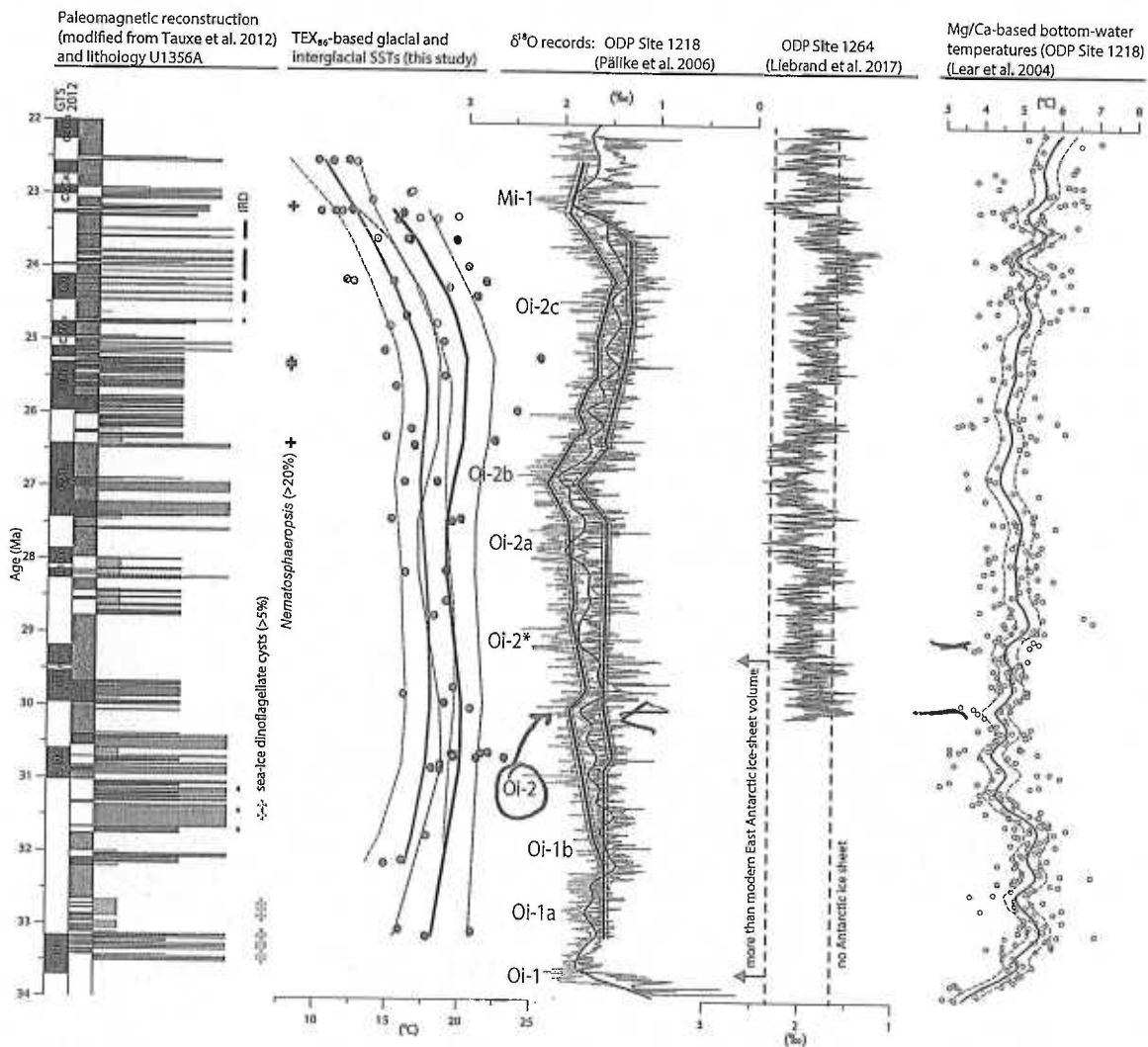
### 5 Conclusions

We reconstruct a summer SST of around 17°C on average for the Wilkes Land Margin during the Oligocene, albeit with much variability (up to a 6.8°C double standard deviation during the late Oligocene). The reconstructed temperatures are a few degrees higher than published high-latitude early Oligocene Southern Ocean estimates. Because alternations in the lithology reflect glacial-interglacial cycles, an estimated temperature difference of 2 to 3°C between glacials and interglacials could be determined. The long-term trends of both glacial and interglacial records show a temperature increase towards 30.5 Ma and a decrease after 25 Ma, generally following the long-term trends in the benthic  $\delta^{18}\text{O}$  record. Our results suggest that considerable SST variability prevailed during the Oligocene, which might have implications for the dynamics of marine-based continental ice sheets, if present, and the extent of the Antarctic ice sheet in general. Assuming that the reconstructed SST trends and glacial-interglacial variability have been relayed to the deep water at nearby bottom-water formation sites, our results imply that the long-term  $\delta^{18}\text{O}$  trend may for a considerable part be controlled by bottom-water temperature in addition to ice-volume fluctuations. This would mean that the Antarctic ice sheet was less sensitive to polar climate changes than previously assumed.

rewrite in light of edits above



**Figure 3:** The calibrations to  $TEX_{86}$  with corresponding standard errors (as shaded envelopes) (A) using the linear calibration of Kim et al. (2010) (in black) and the Bayesian spatially-varying regression model of Tierney and Tingley (2015) (in red). For the Bayesian calibration the used  $20 \times 20^\circ$  grid cells are indicated in the map (black dots represent surface sediment samples used for the calibration and the red cross is the location of Site U1356), the prior mean is derived from Petersen and Schrag (2015). Also plotted are temperature estimates from Site 511 and Site 689 (A). Differences from the LOESS mean of the SST reconstruction based on the calibration of Kim et al. (2010) are plotted above (B).



**Figure 4:** Lithology and paleomagnetic polarities as in Fig. 2, but plotted against age (Ma) updated to the GTS2012 timescale (Gradstein et al. 2012). Sections with IRD, and samples with sea-ice dinoflagellate cysts and *Nematosphaeropsis* dinocysts are indicated. Blue and red dots represent glacial and interglacial  $TEX_{86}$ -based SST estimates, respectively, based on the BAYSPAR calibration of Tierney and Tingley (2015). Thick red and blue lines are the LOESS curve averages of these interglacial and glacial SST curves with corresponding 95%-confidence interval (dotted lines). Benthic  $\delta^{18}O$  record of Site 1218 according to Pälike et al. (2006). Black LOESS curve average separates glacial (blue) and interglacial (red)  $\delta^{18}O$  values. Thick red and blue curves represent the long-term interglacial and glacial trends and 95% confidence intervals (dotted lines), respectively, based on resampling of the  $\delta^{18}O$  record at the resolution of the  $TEX_{86}$ -based SST record. Benthic  $\delta^{18}O$  record of Site 1264 with the  $\delta^{18}O$  equivalent of minimum Antarctic ice-volume estimates following Liebrand et al. (2017). These estimates are based on a constant bottom-water temperature of 2.5°C (modern-day value at Site U1364) and assume temperatures cannot drop lower. Oi-glaciation events according to Gradstein et al. (2012). Mg/Ca-based bottom-water temperatures of Site 1218 according to Lear et al. (2004) with LOESS curve average (thick black line) and 95%-confidence interval (dotted lines).