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Southern Ocean temperature records and ice-sheet models demonstrate rapid Antarctic ice sheet retreat under low atmospheric CO_2 during Marine Isotope Stage 31.

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Highlights

- We quantify precisely the magnitude of ocean warming using the **first ever** Southern Ocean molecular paleo-temperature reconstructions during the warm Marine Isotope Stage 31 (MIS31- 1.085 and 1.055 million years ago).
- Our data show a sustained surface Southern Ocean warming and a collapse of the Antarctic and sub Antarctic ocean fronts during that warm period and under low atmospheric CO_2 concentrations.
- We use sea surface temperature reconstructions to test the scenarios for the AIS retreat during MIS31 using coupled ice-sheet/ice-shelf model.
- We propose a two-step model for deglaciating West Antarctica which involves mild ocean warming (a new temperature threshold) which forces ice margin retreat followed by rapid ocean warming as the ice sheet retreats.
- Our work shows that the Paris Agreement target temperature of 1.5°C is sufficient to drive runaway retreat of the West Antarctic Ice Sheet. We derive this conclusion from the robust, ocean temperature proxy record and ice sheet simulation.

Abstract

Over the last 5 million years, the Earth's climate has oscillated between warm (interglacial) and cold (glacial) states. Some particularly warm interglacial periods (i.e. 'super-interglacials') occurred under low atmospheric CO_2 and may have featured extensive Antarctic ice sheet collapse. Here we focus on an extreme super-interglacial known as Marine Isotope Stage 31 (MIS31), between 1.085 and 1.055 million years ago and is the subject of intense discussion. We reconstructed the first Southern Ocean and Antarctic margin sea surface temperatures (SSTs) from organic biomarkers and used them to constrain numerical ice sheet-shelf simulations. Our SSTs indicate that the ocean was on average 5 °C (±1.2 °C) warmer in summer than today between 50 °S and the Antarctic ice margin. Our most conservative ice sheet simulation indicates a complete collapse of the West Antarctic Ice Sheet (WAIS) with additional deflation of the East Antarctic Ice Sheet. We suggest the WAIS retreated because of anomalously high Southern Hemisphere insolation coupled with the intrusion of Circumpolar Deep Water onto the continental shelf under poleward-intensified winds leading to a shorter sea ice season and ocean warming at the continental margin. In this scenario, the extreme warming we observed likely reflects the extensively modified oceanic and hydrologic system following ice sheet collapse. Our work highlights the sensitivity of the Antarctic ice sheets to minor oceanic perturbations that could also be at play for future changes.

Keywords

Super-interglacial, Pleistocene, Paleoceanography, Paleoclimate modeling, Antarctica, Southern Ocean, Organic Geochemistry

1. Introduction

Some geological records suggest that during the early Pleistocene Marine Isotope Stage 31 (MIS31 ~1.08 to 1.06 Ma), under preindustrial atmospheric CO_2 conditions (Lüthi et al. 2008; Hönisch et al., 2009), global mean sea level rose by ~20 m (Raymo et al., 2006). Recent climate simulations estimate that the global mean temperature increased by 1.2 ^{o}C during boreal summers in the Northern Hemisphere (Justino et al., 2019). The magnitude of the sea level high stand and geological evidence suggest a major retreat of the West Antarctic Ice Sheet (WAIS), part of the East Antarctic Ice Sheet (EAIS) and probably also the Greenland Ice Sheet (Hearty et al., 1999; Raymo et al., 2006; Dutton and Lambeck, 2012; Dutton et al., 2015) during this time. In sediment drill core records from around Antarctica, MIS31 is identified by the abrupt appearance of foraminiferal oozes and coccolith-bearing sediments in the Weddell Sea and Prydz Bay and of bioclastic limestone in the Ross Sea (Bohaty et al., 1998; Scherer et al., 2003; Villa and Wise, 1998; Villa et al., 2008; Villa et al., 2012). This indicates a significant southward migration of the Polar Front (PF), surface ocean warming and open-marine conditions in the Ross and Weddell Seas, with a reduction in sea ice extent relative to the present day (Becquey and Gersonde, 2002; Scherer et al., 2008; Winter et al., 2010; Martinez-Garcia et al., 2010; McKay et al., 2012). To date, elevated ocean surface temperatures have been inferred from paleoecological and sedimentological changes observed in sediment cores around the Antarctic margin (Bohaty et al., 1998; Scherer et al., 2003; 2008; Winter et al., 2010; Dunbar, 2012; Teitler et al., 2005). However, the scarcity of precise surface ocean temperature data has so far limited models from realistically simulating the ice sheet and Southern Ocean responses during MIS31 (Supplementary Information Section 1). Consequently, it has also not been possible to isolate the cause of ice melt and sea

level rise although an unusual orbital configuration and elevated polar summer insolation have been implicated (Laskar et al., 2004; Scherer et al., 2008).

Ice-sheet/ice-shelf and global climate simulations have highlighted the prominent role the ocean plays in controlling the position of the grounding lines of the WAIS and portions of the EAIS (Jacobs et al., 1992; Dinniman et al., 2016). Here we generate the first SST time series from molecular paleothermometric methods (alkenone unsaturation index $U^{k'}_{37}$ - Prahl et al., 1988; and the Long Chain Diol Index LDI - Rampen et al., 2012) from Southern Ocean and Antarctic margin deep-sea sediment cores. We then use these sea surface temperature reconstructions to test the scenarios for the AIS retreat during MIS31 using coupled ice-sheet/ice-shelf model (Golledge, et al., 2015).

2. Material and Methods

2.1. Site locations and modern oceanography

To document the evolution of the sea surface temperatures (SST) in the Atlantic sector of the Southern Ocean during MIS31, we have selected the ODP Site 1094, situated approximately 2° south of the present Antarctic Polar Front (Schneider-Mor et al., 2008 - Fig. 1). To investigate the SSTs in the peri-Antarctic region, we have selected the IODP Site 1361 (Wilkes Land - Escutia et al., 2011) and the ODP Site 1101 (Antarctic Peninsula – Barker et al., 1999). We supplement our latitudinal temperature dataset with the high resolution alkenone-derived SSTs record from Martinez-Garcia et al. (2010) for the ODP Site 1090 situated North of the modern subantarctic front (SAF) and the Mg/Ca measurements on planktic foraminifera from the Cape Roberts Project Site 1 (CRP-1) drilled beneath the northwestern Ross Ice Shelf (Dunbar 2012 - Fig. 1). Sample sites were selected on the basis of previously reported occurrence of MIS31 sediments in the literature.

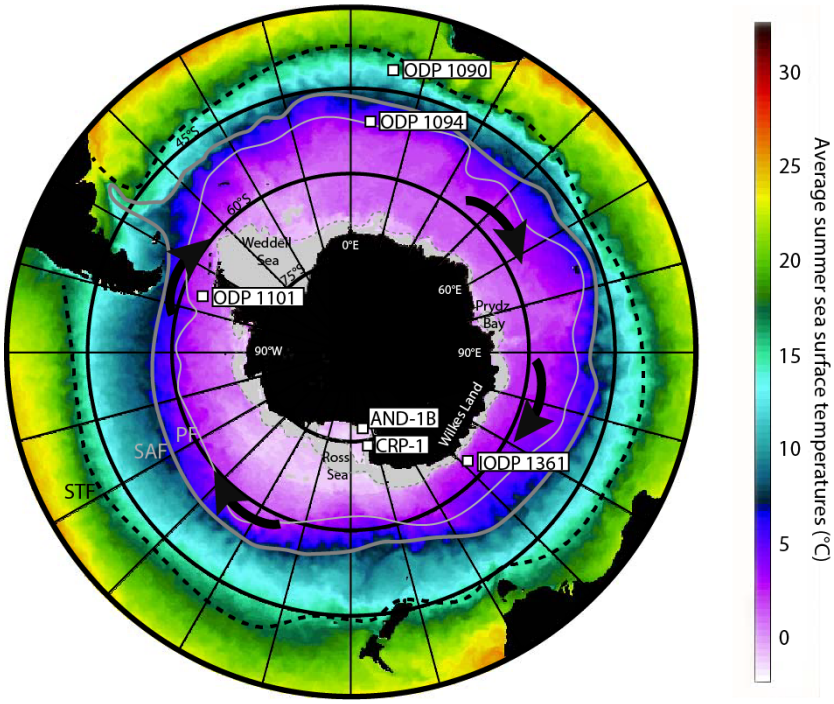

Figure 1: Location of the principal ocean frontal systems in the Southern Ocean (Orsi et al., 1995) and drill sites chosen for this study. The modern summer sea surface temperature map was modified after HYCOM Consortium for Data-Assimilative Ocean Modelling Southern Ocean Atlas (http://www7320.nrlssc.navy.mil/GLBhycom1-12/skill.html). The general path for the Antarctic Circumpolar Current (ACC) is shown with the black arrows (Orsi et al., 1995). STF: subtropical Front (black dashed)- SAF: Subantarctic Front (dark grey)- PF: Polar Front (light grey).

ODP Site 1090 (42°54.8'S, 8°53.9'E) was drilled on the southern flank of the Agulhas Ridge, at a water depth of 3699 m. It lies within the modern subantarctic waters between the subtropical (STF) and the subantarctic (SAF) fronts in the Atlantic sector of the Southern Ocean. The modern annual SST at this location is 10 °C (9 °C in July-September - 11 °C in January-March; Levitus, 1994). Pleistocene sediment consists of foraminiferal nannofossil ooze and foraminifer-bearing nannofossil ooze.

ODP Site 1094 (53°10.8'S, 5°7.8'E) is located south of the Polar Front, north of the Weddell Gyre (Antarctic Circumpolar Current (ACC) -boundary) and within the biogenic silica belt (Gersonde et al., 1999). The site is located close to Bouvet Island, in the ice-free Antarctic Zone, at 2807 m water depth. The investigated sediments

(1094D 11H to 13H) mainly consist of carbonate bearing diatom ooze, with varying amounts of foraminifera and nannofossils (between 20% and 45%). The mean annual SST in this area is 0.7 °C, with mean temperatures of 0.2 °C in winter and of 2.5 °C during austral summer (Levitus, 1994).

IODP Site U1361 (64.4°S, 143.88°E) is located on the Wilkes Land margin, on the eastern levee of the Jussieau sub-marine channel. The mean annual SST in this area is around 0 °C (Levitus, 1994). Sediments from 1361A-2H consist of silty clay and calcareous silty clay. Sediments from 1361A-3H are foraminifera-bearing silty clay and diatom-rich silty clay. The occurrence of calcareous nannofossils have been reported in section 5W. We focused on the interval comprised between 1361A-2H-6W and 1361A3H4W because in some discrete intervals planktonic foraminifers were abundant (from Sample 2H-CC contains >90%).

ODP Site 1101 (64.37°S, -70.26°E) is located on the continental rise, west of the Antarctic peninsula, 94 km offshore the shelf edge. The mean annual SST in this area is around 0 °C (Levitus, 1994). Between 1101A9H1W100 and 1101A10H3W117 the sediments consist of laminated clayey silts and foraminifera-bearing clayey silts containing well preserved calcareous nannofossils.

Cape Roberts Project Site 1 (CRP-1) (77.0°S, 163.8°E) was drilled on the continental shelf in the western Ross Sea near the Transantarctic Mountain belt. Previous studies (Scherer et al., 2008) have focused on the unusual early Pleistocene carbonate rich lithostratigraphic Unit 3.1 corresponding to the MIS31 interglacial. This unique interval contains laminated sediments with abundant diatoms, planktic foraminifera and thoracospherids (calcareous dinoflagellates - Scherer et al., 2008).

2.2. Age models

We developed a chronology for the IODP Site 1361 from new paleomagnetic measurements performed on u-channel samples at the University of Otago Paleomagnetic Research Facility (OPRF). Our LDI-derived SST data were then converted from depth to the time by linear interpolation between geomagnetic reversals. The alkenone-derived temperatures for the ODP Site 1094 and ODP Site 1101 were converted from depth to time using the chronologies from shipboard paleomagnetic data.

2.2.1. IODP Site 1361

U-channel samples were collected at the IODP Core Repository at Texas A&M University and shipped to the University of Otago's Paleomagnetic Research Facility (OPRF). U-channel samples were collected from the IODP Site 1361 archive half core to establish a more precise magnetostratigraphy. Archive half core sections were demagnetized in fields of 20 mT during the expedition to build an initial age model construction and core characterization. Magnetic moment measurements were made at 1 cm intervals using the 2G Enterprises pass-through superconducting rock magnetometer which is housed in a 150 nT magnetically shielded room. U-channels were demagnetized in alternating magnetic field increments of 5 mT between 20 mT and 50 mT, then at 60 mT, 70 mT and finally at 100 mT. Demagnetization data were analysed on orthogonal vector component plots using the PuffinPlot software (Lurcock and Wilson, 2012). Polarity was determined from principal component analyses (PCA, Kirschvink, 1980) of data between the 20 mT and c. 50 mT increments which we interpret is the primary depositional remanent magnetizations (DRM). PCAs were typically anchored to the origin with the exception of thin intervals which were offset from the origin where a high coercivity magnetisation prevents complete demagnetisation. Rock magnetic data of these intervals did not

reveal unusual changes in magnetic mineralogy therefore we suggest that strongly magnetised, high coercivity drop stones may be responsible for the offset origins.

Our new magnetostratigraphy does not change the correlation with the GPTS (Tauxe et al., 2012) but does result in more precise placement of reversal boundaries (Fig. 2). Our new magnetostratigraphy places the C1r.1r-C1r.1n (988 ka) boundary (at 17.77m ±0.03m, the C1r.1n-C1r.2r (1072 ka) at 19.02m ± 0.17m, the top and base of the Cobb Mt (1178–1215 ka, Channell, 2017) at 20.54m ±0.06m and 20.92m ±0.04m respectively (Fig. 2). The C1n-C1r.1r (781 ka) transition spans a core break between 8.1 m and 14.34 m (11.22 m mid-point) where the lower contact on core 2H is affected by coring disturbance and the C1r.2r-C2n (1778 ka) transition occurs at 32.33m ±0.025m (Tauxe et al., 2012).

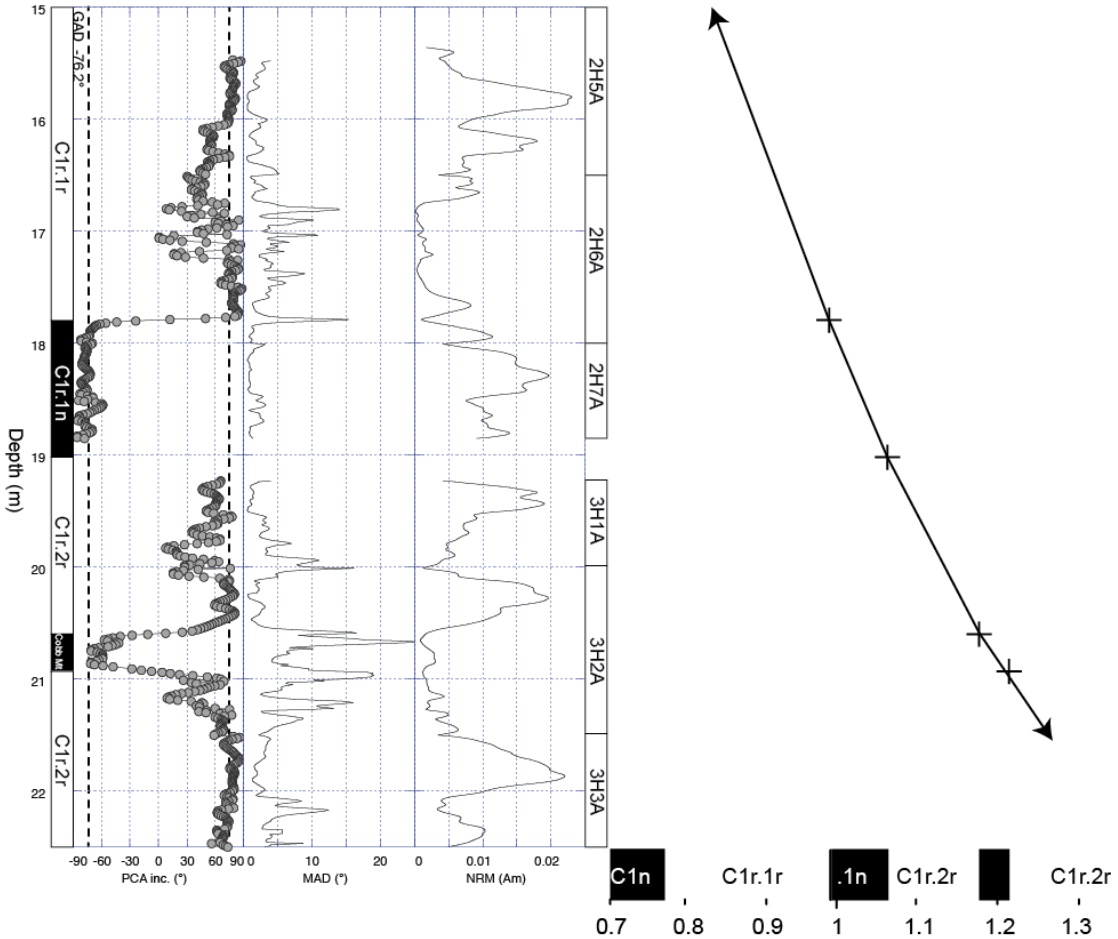

Figure 2: IODP 1361 age model developed in this study.

2.2.2. ODP Site 1094

The age model of ODP Site 1094A and 1094D was constructed from shipboard paleomagnetic data where the 20 mT demagnetisation step was taken as the characteristic remanent magnetisation as confirmed from complete laboratory demagnetisation of u-channel samples (Channell and Stoner, 2002). The C1n-C1r.1r boundary (781 ka) occurs at 99.61m, the C1r.1r-C1r.1n (988 ka) boundary at 123m ±2.22m, the C1r.1n-C1r.2r (1072 ka) at 129m ± 0.38m, and the base of the Cobb Mt (1178 ka) at 149.6 m.

2.2.3. ODP Site 1101

We adopted the age model (Acton et al., 2012) based on shipboard analyses and shore-based laboratory analyses from u-channel samples. The C1n-C1r.1r (781 ka) transition occurs at 55.05 m ±0.01 m, the C1r.1r-C1r.1n (988 ka) at 71.15 m ±0.19m, the C1r.1n-C1r.2r (1072 ka) at 76.12m ± 0.17m, and the top and base of the Cobb Mt (1178–1215 ka, Acton et al., 2012) at 81.61m ±0.65m and 82.53m ±0.01m respectively.

2.3. Temperature reconstructions

To reconstruct the latitudinal SST gradient during the MIS31 super-interglacial along a transect from Sub-Antarctic to Antarctic water masses of the Southern Ocean, we used two complementary organic geochemical proxies for SSTs: the alkenone unsaturation ratio (Prahl et al., 1988 - $U^{k'}_{37}$) and the Long Chain Diol Index (Rampen et al., 2012 - LDI). The use of alkenone thermometry was motivated by the unusual presence of surface-dwelling calcareous nannofossils in 1 Ma-old sediments at extreme southern latitudes around the Antarctic (Scherer et al., 2008; Villa et al., 2005; 2007; 2012). Specifically, representatives belonging to the alkenone (long-chain unsaturated ketones) producer Noelarhabdaceae family (*Emiliania*, *Gephyrocapsa*, *Reticulofenetra*, *Pseudoemiliania*, *Dictyococcites* sp. – Marlowe et al., 1990; Beltran et al., 2007; 2011) were reported at drilling sites located south of the Polar Front (Fig. 1).

Although alkenones were detected at Site 1094 (concentrations between 2 and 25 ng/g), the concentrations were below detection limits at the Antarctic margin IODP Site U1361. Thus, we chose to use the long chain Diol Index (LDI) to reconstruct the SSTs at the Antarctic margin. This newly developed proxy for paleotemperatures (see Lopes dos Santos et al., 2013; de Bar et al., 2018), is based on the relative abundances of lipids with chain length of 28 and 30 carbon atoms containing two alcohol groups (long chain alkyl 1,13- and 1,15-diols; C_{28}1,13-diol, C_{30}1,13-diol and C_{30}1,15-diol). In the marine environment, these long chain diols have been found in diatoms and marine algae (Rampen et al., 2012). Although the source of the long chain diols is still under discussion, these molecules have existed at least since the Mesozoic era and are ubiquitous in the world ocean from the equator down to very high latitudes (74°S). The LDI index correlates best with temperature for modern water depths of 0 to 30 m

during the warmest months of the year.

Alkenone unsaturation ratios were measured on samples from ODP Site 1094D-11H-1W to 1094D-13H-3W (between 115.9 and 137.8 m) at a resolution of one sample every 30 cm (temporal resolution of 2.2 kyrs) and from ODP Site 1101A-9H-2W to 1101A-10H-4W (between 68.7 and 80.4 m) at a resolution of one sample every 45 cm (temporal resolution of 7.3 kyrs). For IODP Site 1361, we measured the Long chain diol index (LDI) on 12 samples from 1361A-2H-6W to 1361A-3H-3W (between 17.7 and 21.7 m) at a resolution of one sample every 35 cm (temporal resolution of 23 kyrs). We found Long chain diols in 10 samples. For both methods, sediment samples (6 to 9 g) were freeze-dried, homogenized, and extracted using an automated solvent extractor a Dionex 300 Accelerated Solvent Extractor with dichloromethane and methanol (9:1, v/v) at 1500 psi and 100°C to obtain a total lipid extract.

Alkenones were isolated by silica gel chromatography using solvents of increasing polarity following the procedure of Sicre et al. (2001). The fractions containing alkenones were concentrated, transferred to clean glass-vials and evaporated under nitrogen. Gas chromatographic analyses were performed on a Hewlett Packard 6890 gas chromatograph with flame ionization detector using a fused silica capillary column (Rxi-1ms, 50 m long, 0.32 mm internal diameter, 0.25 μm film thickness). Helium was used as a carrier gas. We have converted the alkenone index into temperature using the Prahl et al. (1988) calibration (see Supplementary Information Section 1.3).

Long chain diols were isolated by silica gel chromatography following the procedure of Rampen et al. (2012). Compound identification of the long chain diols was conducted using an Agilent 6890N gas chromatograph (GC) equipped with an Agilent 5975B mass selective (MS) detector in selective ion monitoring mode operated at 100

eV (EI Source). They were separated using a fused silica capillary column (ZB-5MS, 30 m long, 0.32 mm internal diameter, 0.25 μm film thickness). The relative abundance of long chain diols was measured using single ion monitoring of m/z 299, 313, 327, and 411.

2.4. Ice sheet/shelves simulations

2.4.1. General Circulation Model (GCM) set-up and simulations

Our RegCM3 experiments were initialised from GENESIS3 GCM simulations previously run at UMass Amherst (DeConto, et al., 2012; DeConto and Kowalewski unpublished). The GCM was run for 30 years to equilibrate, and then for a further 11 years to provide input (initial / boundary conditions) for the RegCM3 regional climate model. To provide appropriate input for RegCM3, GENESIS3 was tuned for orbital forcing, CO_2 levels, and ice sheet configuration representative of MIS31. Ice sheet configurations for Neogene interglacials were provided from model results from the Pollard and DeConto (2009) ice sheet model (forced w/ GENESIS3 as well).

2.4.2. Regional Climate Model (RCM) set-up and simulations

The 11 years -period described above allows for a one-year spinup for the RCM, followed by a ten years model evaluation period. GCM derived boundary conditions were continuously updated every 6 hours of RCM simulation, with fields being interpolated by the 'RegCM3_Polar' package onto the new model grid. No proxy data were used during this phase of climate model set up, but RegCM3 simulations were validated against (1) GCM performance conducted with GENESIS3, and (2) RegCM3 forced with NCEP/NCAR reanalysis. Reanalysis driven runs are for the 1990-1999 period using observed SST forcing while the Genesis-driven runs employ SST forcing from a slab ocean model.

2.4.3 Ice Sheet Model (ISM) set-up and simulations

For our main experiments, we used climate model outputs from GCM experiments in which the ice sheet configuration was similar to present, since our ice sheet model experiments also start from a modern configuration. This avoids the circularity of simulation WAIS collapse by imposing a climatology that has been produced under conditions of a collapsed WAIS. We ran simulations with two different atmospheric greenhouse gas concentrations (280 and 400 ppm CO_2) to quantify the sensitivity of our results to different background climate states. The climatologies from the regional climate model RegCM3_Polar simulations (RegCM3 adapted to Antarctica; Fig. S3) were imposed at the beginning of each run and the ice sheet was allowed to freely evolve to the new boundary conditions. Ice sheet experiments used the Parallel Ice Sheet Model (PISM) version 0.6, whose dynamical core combines velocity solutions of the shallow shelf and shallow ice approximations for the entire domain (see Supplementary Information Section 2). We ran duplicate experiments both with, and without, the sub-ice shelf grounding line melt interpolation scheme turned on. Once afloat, we imposed two calving mechanisms – one based on horizontal strain rates (Levermann et al., 2012), and another in which a simple minimum thickness criterion (200m) must be satisfied. We ran experiments for two climate scenarios, reflecting atmospheric CO_2 concentrations of 280 and 400 ppm, and for each of these, ran duplicates with each of the two grounding-line parameterizations. All simulations were run for 10000 years and were initialized from a thermally and dynamically stable configuration used previously (Golledge, et al., 2015; 2017a; 2017b) which closely resembles the present-day ice sheet (Fig. S4 panel a). Ice sheet configurations after 5000 years for each of the experiments are shown in Figure S4.

3. Results

3.4. Early Pleistocene Southern Ocean sea surface temperatures

Our SST records were generated using the alkenone unsaturation proxy ($U^{k'}_{37}$ - Prahl et al., 1988) in the ice-free Antarctic zone (ODP Site 1094) and in the polar zone ODP Site 1101 and the Long Chain Diol Index (LDI – Rampen et al., 2012) at the IODP Site 1361 (Fig. 3). Two temperature proxies were used because although $U^{k'}_{37}$ is well-established, the calcareous nannofossils that produce alkenones (Marlowe et al., 1990; Beltran et al., 2007; 2011) are not always detectable in the coldest high latitude ocean regions. In contrast, the long chain diols used in the LDI index offer a good alternative as they are ubiquitous in the ocean, including at high latitudes (74°S) and LDI- and $U^{k'}_{37}$-derived SSTs record similar water depths and seasons (Rampen et al., 2012).

At ODP Site 1094, our record for the interval between 1.11 Ma and 0.95 Ma shows that the summer SSTs were on average 8.6 °C (between 5.3 °C and 10.8 °C), which is 3 °C warmer than today (Southern Hemisphere Summer months - Levitus, 1994; Fig. 3). This indicates sustained long-term elevated ocean surface temperatures in this sector of the Southern Ocean (Fig. 3A). At about 1.076 Ma, we identify a 10.8±1 °C temperature peak, i.e. 6 °C warmer than today, which we interpret as the MIS31 temperature optimum. Our data shows that during MIS31, SSTs closely track changes in Southern Hemisphere summer insolation with the warmest conditions reached during the highest Southern Hemisphere summer insolation maxima at 1.08 Ma (Laskar et al., 2004). We also observe large magnitude rapid temperature variations, e.g. 5.5 °C warming between 1082.69 ka and 1082.05 ka.

At ODP Site 1101 and IODP Site 1361, the age model and sampling resolution is

lower than at ODP Site 1094. However, at both sites, our data show that MIS31 corresponds to a summer SST increase of more than 5 °C compared to today (Levitus, 1994; Fig. 3A). The average SST through MIS31 at ODP Site 1101 is ~ 9.5 °C, peaking at about ~11 °C (1.069 Ma), which, is 7 °C warmer than the modern conditions. At IODP Site 1361, (the age model is less well constrained) the average surface water temperature during MIS31 was between 0 and 1°C, reaching a maximum value of 5 °C (6 °C warmer than today - Levitus, 1994). This result is in agreement with the Mg/Ca measurements on planktic foraminifera (Dunbar, 2012) and diatom based MIS31 temperature estimates (Scherer et al., 2008 and Scherer comm. pers.) from the Unit 3.1 at CRP-1B who suggests summer SSTs in the Ross Sea were warmer than today by 5-9°C.

Our paleotemperature reconstructions based on organic biomarkers, show that the surface ocean, south of 50 °S, warmed on average by 5 °C (± 1 to 2 °C) (Fig. 3 and 4), providing the first precise constraints on the maximum surface ocean temperature during MIS31.

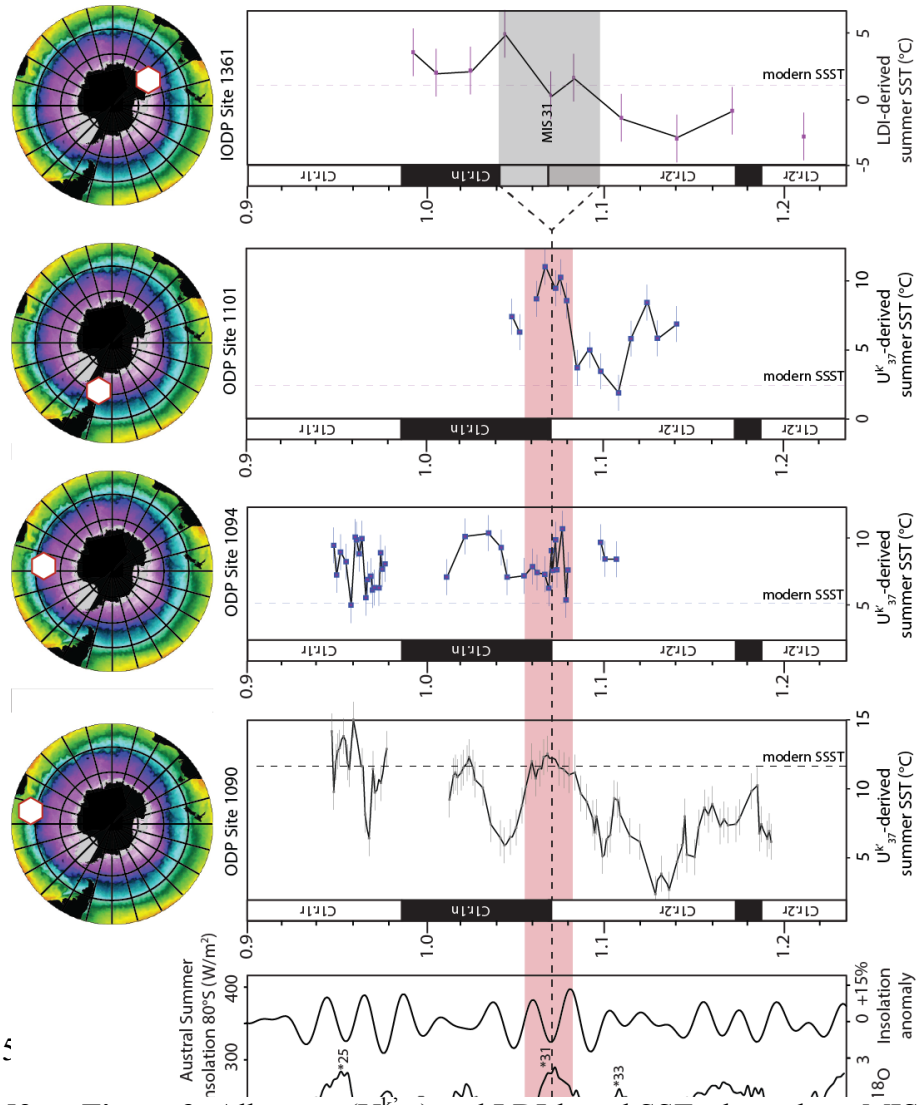

Figure 3: Alkenone ($U^{k'}_{37}$) and LDI-based SSTs throughout MIS31.
We show time series of deep-sea benthic $\delta^{18}O$ record (‰) (Lisiecki and Raymo, 2005), austral summer insolation at 80 °S and insolation anomaly relative to present (%increase relative to present; Laskar et al., 2004). Some of the MIS peaks are indicated with the symbol *. MIS31 interglacial duration is highlighted in red. The black dotted horizontal line shows the age of MIS31's peak (*31). Due to the uncertainties on the age model at Site 1361, MIS31 peak is highlighted in grey. Each map indicates the site locations. The error bars correspond to the error of each temperature proxy.
Summer southern hemisphere summer season which is here the average temperature for the months DJF

Our temperature record from ODP Site 1094 also indicate strong high-frequency variability within this warm period.

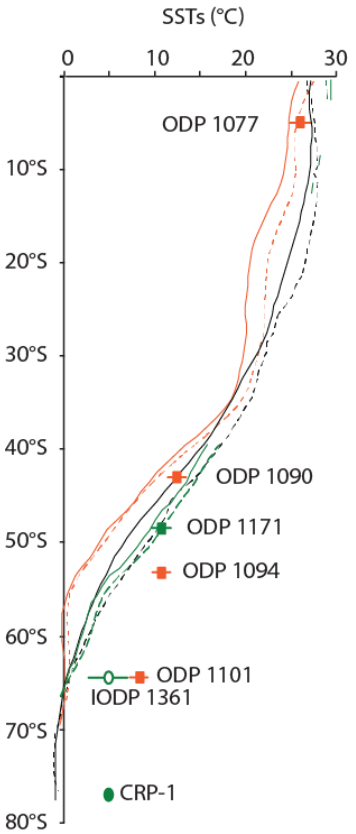

Figure 4: Comparison of the latitudinal sea surface temperature gradients during MIS31 warmth with modern conditions. MIS31 temperature data for Site ODP 1077 is from Schefus et al. (2004), for ODP Site 1171 from Beltran et al. (2016) and for CRP-1 from Dunbar (2012). The orange and green bars represent respectively the error bars for alkenone and LDI measurements.
Black dashed line: Modern Annual Southern Ocean SST - Green dashed line: Modern Annual Atlantic Ocean SST - Orange dashed line: Modern Annual Pacific Ocean SST - Green line: Modern Austral Summer Southern Ocean SST - Orange line: Modern Austral Summer Atlantic Ocean SST - Black line: Modern Austral Summer Pacific Ocean SST.

When comparing our high latitude records with the lower latitude alkenone-derived SSTs at the subantarctic ODP Site 1090 and the equatorial ODP Site 1077 (Schefus et al., 2004 - Fig. 4), we observe a significant polar amplification during MIS31, resulting in a reduced latitudinal SST gradient (Scherer et al., 2008; Martinez-Garcia et al., 2010; Dunbar, 2012; Beltran et al., 2016) of ~ -0.3 °C/ °latitude (Fig. 4).

3.2. Modelling the AIS during MIS 31

To investigate the degree of ice sheet retreat in relation to our SST records we conducted a suite of numerical simulations that use climate fields from a regional climate model (DeConto et al., 2012 - RCM, Supplementary Section 2). Our proxy data were critical to the simulations because they were the quality assurance that the simulations were capturing the correct magnitude of warming. We used the Parallel Ice Sheet-Model, which is an open source three-dimensional thermodynamic coupled ice sheet/ice-shelf model. Our RCM simulation represents a time slice reconstruction

rather than a transient prediction, but considers two atmospheric CO_2 concentration scenarios - 280ppm and 400ppm - so that the sensitivity of the ice sheet to atmospheric composition can be assessed (Supplementary Section 2.2). Using these two scenarios as inputs, we also then combined each one with either a conservative grounding line scheme (no SGm) or a more dynamic grounding line (SGm). Because our climate forcing is not transient, we were interested primarily in seeing whether significant ice sheet impacts could occur within relatively short time periods (centuries to a millennium) of warmth, since our proxy data indicate that such temperatures are not likely to have been sustained for the entire interglacial.

In all simulations, the WAIS melted within a few centuries (Fig. 5B, C, D, E and F) whereas the EAIS responded more slowly (Supplementary Section 2.3). The simulations revealed that regardless of the grounding line scenario and CO_2 forcing, no significant difference in the spatial extent of ice loss is observed after 5kyr (Fig. S4).

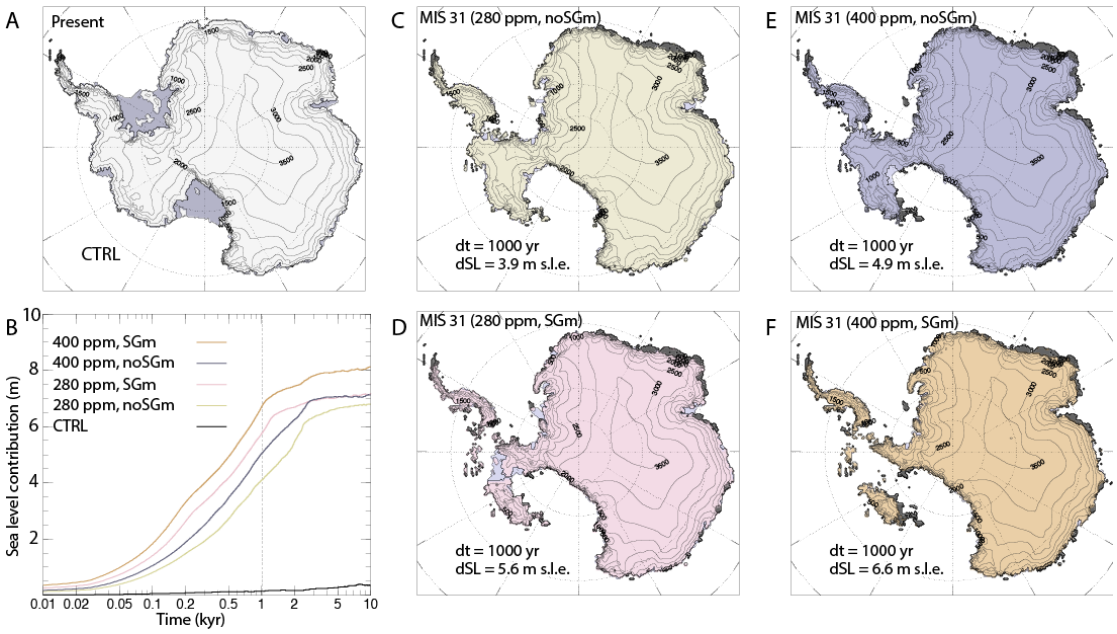

Figure 5: Snapshots of ice distribution for ice sheet experiment runs of 1000 years (dt). We present the results using a conservative groundling line behaviour (no SGm, C. and E.), a dynamic grounding line (SGm, D. and F.), each with two extreme CO_2 scenarios (280 ppm and 400 ppm). A. represents the modern ice distribution and B. is

the sea level contribution from ice melting through time. dSL is the sea level increase expressed in s.l.e (sea level equivalent).

Our results indicate that 2-m of sea-level rise from ice loss occur within a few centuries (Fig. 5B) and that persistent elevated SSTs would cause complete collapse of the WAIS, with additional thinning around the margins of the EAIS (i.e. the Wilkes Land basin, Weddell Sea, Aurora Basin).

4. Discussion and conclusions

Our temperature reconstructions indicate proximal ice margin summer SSTs of 4-5 °C in East and 9°C in West Antarctica (5-7 °C warmer than today) during MIS31, in agreement with Scherer et al (2008), Villa et al. (2008, 2012). Those temperatures are not compatible with the presence of a significant West Antarctic ice sheet (Joughin and Alley, 2011; Liu et al., 2015 - Fig. 5).

The causes of the MIS31 extreme warmth under pre-industrial atmospheric CO_2 conditions are still unclear. MIS31 interglacial coincides with a time of high obliquity and eccentricity, which led to strong high latitude austral summer insolation anomalies. How much of the ocean warming observed in our records can be attributed to the high summer insolation? Simulations of previous interglacials (Yin and Berger, 2012) showed that reduction of summer sea ice increases the sensitivity of southern high latitudes to solar radiation (Yin and Berger, 2012) resulting in warming of surface waters. According to Yin and Berger (2012), during interglacials with similar obliquity values as MIS31, increased insolation contributed to a maximum mean annual warming of only 0.5 °C of the high southern hemisphere latitudes. As a consequence, high insolation alone cannot account for the extreme ocean temperatures we observe around Antarctica during MIS31. We argue here that the exceptional warmth is the consequence of an increasingly stratified ocean due to freshwater input from ice loss (Golledge et al., 2014) rather than a trigger for ice loss.

Reduced surface ocean salinity is supported by the diatom assemblage composition at CRP-1 with increased abundance of *Chaetoceros* spp. correlated to the insolation maximum at 1.08 Ma (Scherer et al., 2008). The calcareous nannofossil assemblage composition at ODP Site 1094 also support this hypothesis with a clear increase of *Helicosphaera carteri* and *Syracosphaera* spp. (Flores and Sierro, 2007; Scherer et al., 2008).

We suggest ice margin retreat was caused by circulation changes in Antarctic coastal waters, as a result of changes in wind patterns (Gille et al., 2016). A poleward shift of the Southern Hemisphere Westerly Winds (SHWW) is seen in the decrease of aeolian dust in the Subantarctic Zone (Martinez-Garcia et al., 2010 - ODP Site 1090), raising the possibility that a wind-induced subsurface warming at the Antarctic margin took place during MIS31. In the modern ocean, cold Antarctic shelf waters are separated from warmer Circumpolar Deep Water (CDW) further offshore by the Antarctic Slope Front (ASF – Jacobs, 1991), a well-marked density front that is primarily maintained through coastal downwelling induced by the prevailing coastal easterly winds (Sverdrup, 1953). Ocean model simulations have demonstrated that a poleward shift of the SHWW reduces the extent and strength of the coastal easterlies and can produce strikethrough advective warming of subsurface waters at the Antarctic margin, via a weakening of the ASF that allows CDW to intrude onto the shelf (Schmidtko et al., 2014; Gille et al., 2016). Heat from the CDW would subsequently trigger basal melting of the ice shelves (Dinniman et al., 2016). Additional observational (Schmidtko et al., 2014; Dutrieux et al., 2014) and modelling evidence (Thoma et al., 2008; Dinniman et al., 2012) also implicates the SHWW in regulating poleward transport of warm CDW onto the Antarctic continental shelf in the modern-day ocean, and it therefore seems plausible that this onshore CDW transport was

enhanced by poleward-intensified winds during MIS31 (Martinez-Garcia et al., 2010). Precise dynamics underlying such wind-driven CDW inflow however remain an area of active research (Gille et al., 2016).

We propose that the exceptional surface ocean warming during the MIS31 super-interglacial occurred in two steps (Fig. 6). [Step 1] The unusual MIS31 orbital configuration resulted in an extended period (~ 2000 years) of warm and long-lasting summer seasons that caused mild warming of Southern Ocean surface waters and the reduction of sea ice development or survival. Concomitantly, the prevailing Southern Hemisphere Westerly Winds and the easterly coastal winds migrated to the south (Fig. 6). The poleward shifted winds warmed the coastal waters, thus bringing heat southwards thereby initiating basal melting of ice shelves and the retreat of the marine-based grounding lines. Our numerical ice sheet model demonstrates that a 0.5 °C ocean warming at the ice margin for c. 200 years is sufficient to cause ice retreat and that most of it can occur in less than 2000 years. [Step 2] The surface warming in the coastal regions (9.5 °C at ODP Site 1101 and 5 °C at IODP Site 1361) could only take place after complete loss of the WAIS (Fig. 6) and would likely be amplified by ocean stratification feedbacks.

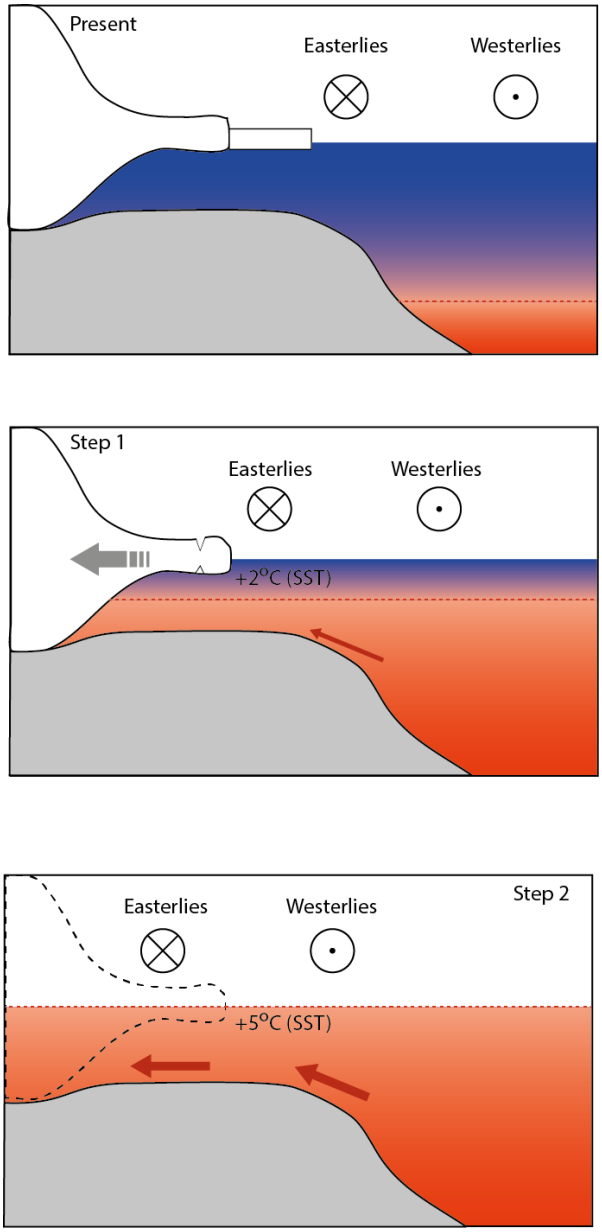

Figure 6: Schematic evolution of ice retreat during MIS31 in response to poleward wind shift and subsequent ocean warming. Step 1 shows the increased advection of warmer deep waters (red arrow, see Pedro et al., 2016), Step 2 represents the run over ice melting at the maximum of MIS31 warmth.

Our new data from the Antarctic sedimentary records demonstrate that a moderate surface ocean warming is sufficient to initiate a major West Antarctic Ice Sheet retreat. We also show that even after 10000 years of persistent elevated SSTs, marine basins of East Antarctica remain glaciated, with the majority of the EAIS sea-level contribution coming from coastal thinning around Dronning Maud Land. That the

marine basins of EAIS were not destabilized at MIS31 suggests that atmospheric warming (as simulated in the RCM) at that time was insufficient to bring about significant surface melting, thinning, and retreat from coastal pinning points (Golledge et al., 2017b).

Future research should focus on further documenting high latitude Southern Ocean SSTs using molecular thermometry to identify critical ocean temperatures that led to major Antarctic ice retreats in the past under varying atmospheric CO_2 and orbital conditions.

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Supplementary Information for:

Southern Ocean temperature records and ice-sheet models demonstrate rapid Antarctic ice sheet retreat under low atmospheric CO_2 during Marine Isotope Stage 31.

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Supplementary Information

Table of Contents

Section 1: Sea surface temperature proxies……. 3

1.1. Alkenone thermometry…….. 3

1.2. Long chain Diol Index thermometry for IODP Site U1361…….. 4

1.3. Alkenone Calibration……. 4

Section 2: Ice sheet/shelves simulations…… 5

2.1. Global Climate Modelling……. 5

2.2. Regional Climate Modelling…….. 6

2.3. Ice Sheet Modelling……... 7

References for Supplementary Information ..……. 11

Section 1: Sea surface temperature proxies

Climate simulations require unambiguous, continuous, and well-dated SST records to determine changes in latitudinal heat transport. Despite the emergence of new and well calibrated paleotemperature proxies based on diatom morphometry (e.g. Allen, 2014; Kloster et al., 2018; Shukla and Romero, 2018), the few available high latitude Southern Ocean SST reconstructions for MIS31 were either derived from 1) proxies based on calcareous nannofossil and planktonic foraminifera assemblage compositions that tend to be less reliable at high latitudes where the species diversity is low 2) diatom assemblages which composition reflect not only temperatures but also environmental parameters such as sea ice, nutrients and light etc… or 3) uncorrelated temperature proxies possibly reflecting temperatures of different water masses and/or seasons, which adds significant uncertainties to climate reconstructions. The lack of accurate and consistent SSTs for the Sub-Antarctic and Antarctic zones of the Southern Ocean during MIS31 greatly restricts our ability to precisely quantify the global latitudinal ocean heat transfer and subsequent sub-ice ocean heat flux, impacting our ability to simulate the AIS response to ocean temperature changes. To reconstruct the most continuous latitudinal record of the SSTs during the MIS31 super-interglacial along a transect from Sub-Antarctic to Antarctic water bodies in the Atlantic sector of the Southern Ocean, we used two complementary organic geochemical proxies for SSTs: the alkenone unsaturation ratio ($U^{k'}_{37}$, Prahl et al., 1988) and the Long Chain Diol Index (LDI, Rampen et al., 2012).

1.1. Alkenone thermometry

The C_{37} alkenone unsaturation index ($U^{k'}_{37}$; Eq. 1) is a robust proxy for SST, providing surface ocean temperature reconstructions with a precision of +/-1.2 °C (Sikes and Volkman, 1993; Muller et al., 1998; Conte et al., 2006).

$$U^{k'}_{37} = \frac{[C_{37:2}]}{[C_{37:2}]]+ [C_{37:3}]} \qquad \text{Eq. 1}$$

Global core-top studies demonstrate that $U^{k'}_{37}$ correlates best with mean annual temperatures at 0–10 m water depth range (Conte et al., 1993) and there is little evidence that the SST signal is affected by other environmental parameters such as productivity or salinity.

Previous studies (Sikes and Volkman, 1993; Conte et al., 2006;) showed that south of 40°S in the Pacific and Atlantic sectors of the Southern Ocean, alkenone production is limited to late spring-summer. As both ODP Sites 1090 and 1094 chosen in our study were drilled south of the subtropical front, the $U^{k'}_{37}$-derived SST reconstructions are likely to reflect the mixed-layer conditions during the warmest months of the year.

1.2. Long chain Diol Index thermometry for IODP Site U1361

The Long-chain diol index (LDI; Eq. 2) is defined as:

$$LDI = \frac{[C_{30}1,15\text{-diol}]}{[C_{28}1,13\text{-diol}]+[C_{30}1,13\text{-diol}]+ [C_{30}1,15\text{-diol}]} \qquad \text{Eq. 2}$$

Core top studies indicate that the LDI index is linearly correlated to SSTs (+/-2°C) in a range of temperatures from -3°C to 27°C (Eq. 3) with salinity and nutrient conditions having little to no influence (Rampen et al., 2002).

$$LDI = 0.033 \text{ x } SST + 0.095 \qquad \text{Eq. 3}$$

1.3. Alkenone Calibration

Most alkenone-derived SST reconstructions use the calibration of Prahl et al.(1988) based on the empirical linear relationship between $U^{k'}_{37}$ and SST obtained from laboratory cultures. However, several studies show that the $U^{k'}_{37}$ measured in particulate matter of surface waters disagree with this calibration (Prahl et al.,1988). These observations led to the development of alternate calibrations either global such

as the polynomial $U^{k'}_{37}$ -SST relationship of Conte et al. (2006) or region-specific (Muller et al., 1998; Sikes and Volkman, 1993).

To select the most accurate $U^{k'}_{37}$ calibration, we have extended the ODP Site 1094 sampling (Fig. S1) to cover the long term evolution of SSTs until today.

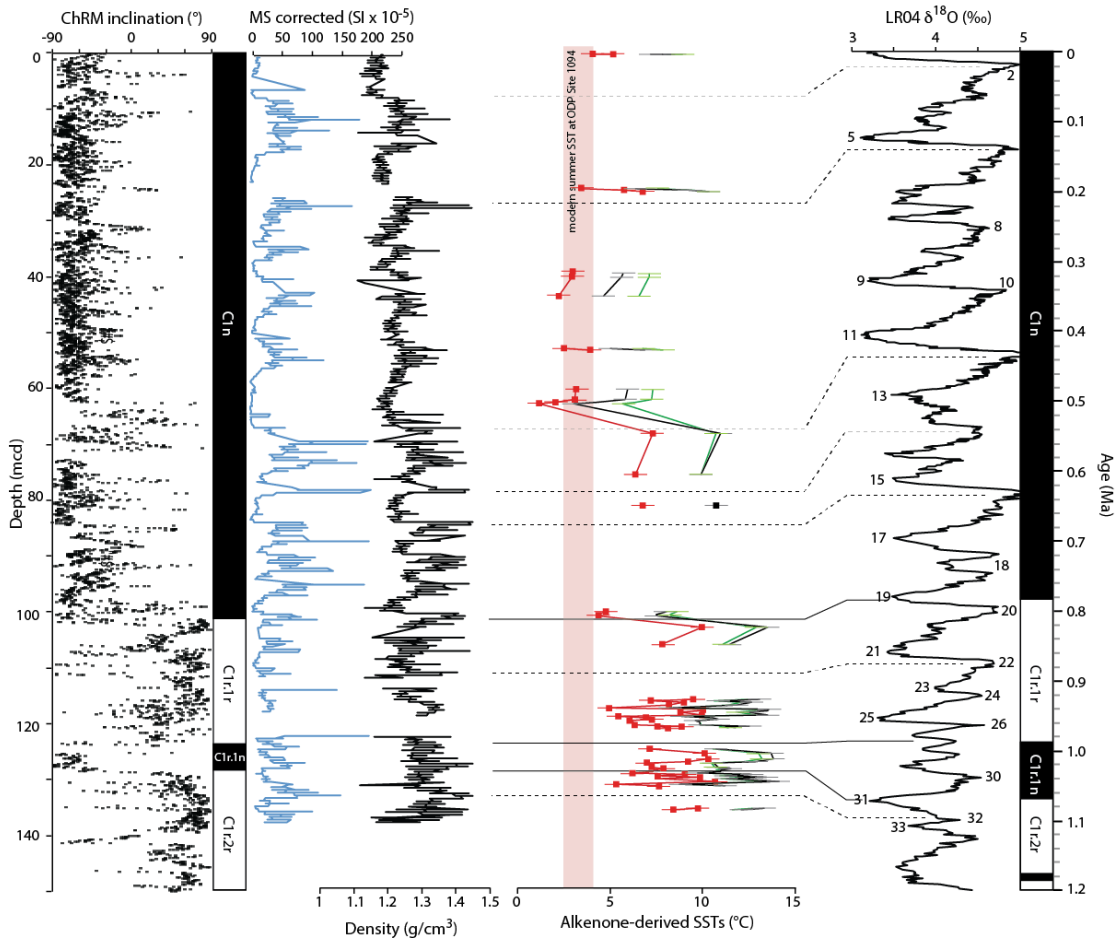

Fig. S1. *Long term evolution of the alkenone-derived sea surface temperatures at ODP Site 1094 using three different calibrations : Prahl et al. (1988) in red; Sikes and Volkman (1993) in green; Conte et al. (2006) in black. Our additional low resolution dataset shows that the Prahl et al.(1988) calibration leads to SSTs that are similar to the modern ones at the same location.*

When comparing the core top SSTs with the Levitus monthly modern SSTs it appears that the Prahl et al. (1988) calibration leads to values that are consistent with the modern February SSTs at the same location (Fig. S1). We have thus converted the alkenone index into temperature using the Prahl et al. (1988) calibration.

<u>Section 2: Ice sheet/shelves simulations</u>

2.1. Global Climate Modelling.

The relationship between air and ocean temperatures under different CO_2 conditions evolves through time (Rugenstein et al., 2016), such that in Antarctica, short-term (< 100 year) sea-surface temperature anomalies are considerably less than the equilibrium state (Fig. S2 below). When attempting to infer an atmospheric warming from our oceanic proxy data, therefore, we conservatively assume that the ocean / atmosphere warming has achieved balance (taking c. 1000 years), and thus mean annual sea temperature anomalies are likely to be c. 2.5 times the inferred mean annual air temperature anomaly (SST to air ratio of 0.4, Fig. S2).

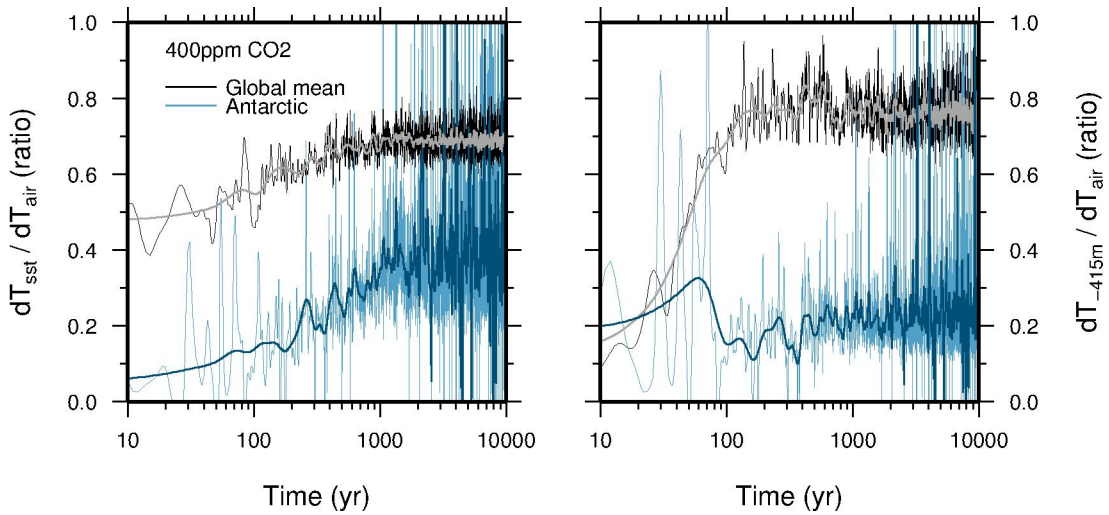

Fig. S2. *Evolution of the relationship between mean annual surface air temperature anomaly and that of the sea surface (left) and ocean at 415 m depth (right) for Antarctica, compared to the global mean (Rugenstein et al., 2016). An equilibrium SST to air ratio of 0.4 implies a 2.5 times greater air temperature anomaly than that felt by the sea surface.*

2.2. Regional Climate Modelling

Paleoclimate simulations were run using RegCM3_Polar. The regional climate model, RegCM3, was adapted to Antarctica (RegCM3_Polar) centered on the South Pole for Antarctic reconstructions. Boundary conditions for the regional model were taken from the library of existing GCM Pliocene simulations (GENESIS v. 3.0.; Pollard and DeConto, 2009). Simulations varied greenhouse gas, orbital, sea ice and SST

conditions; changes in environmental conditions were prescribed in both the RCM and GCM. The boundary forcing from GENESIS GCM was interpolated horizontally and vertically to the RegCM3_Polar grid and topography and was applied at 6-hour intervals for both 40km and 80km horizontal resolution regional climate simulations. Regional climate simulations were run using existing ice configuration and a MIS31 modeled ice sheet configuration imported from the Pollard and DeConto ice sheet / shelf model output (Pollard and DeConto, 2009).

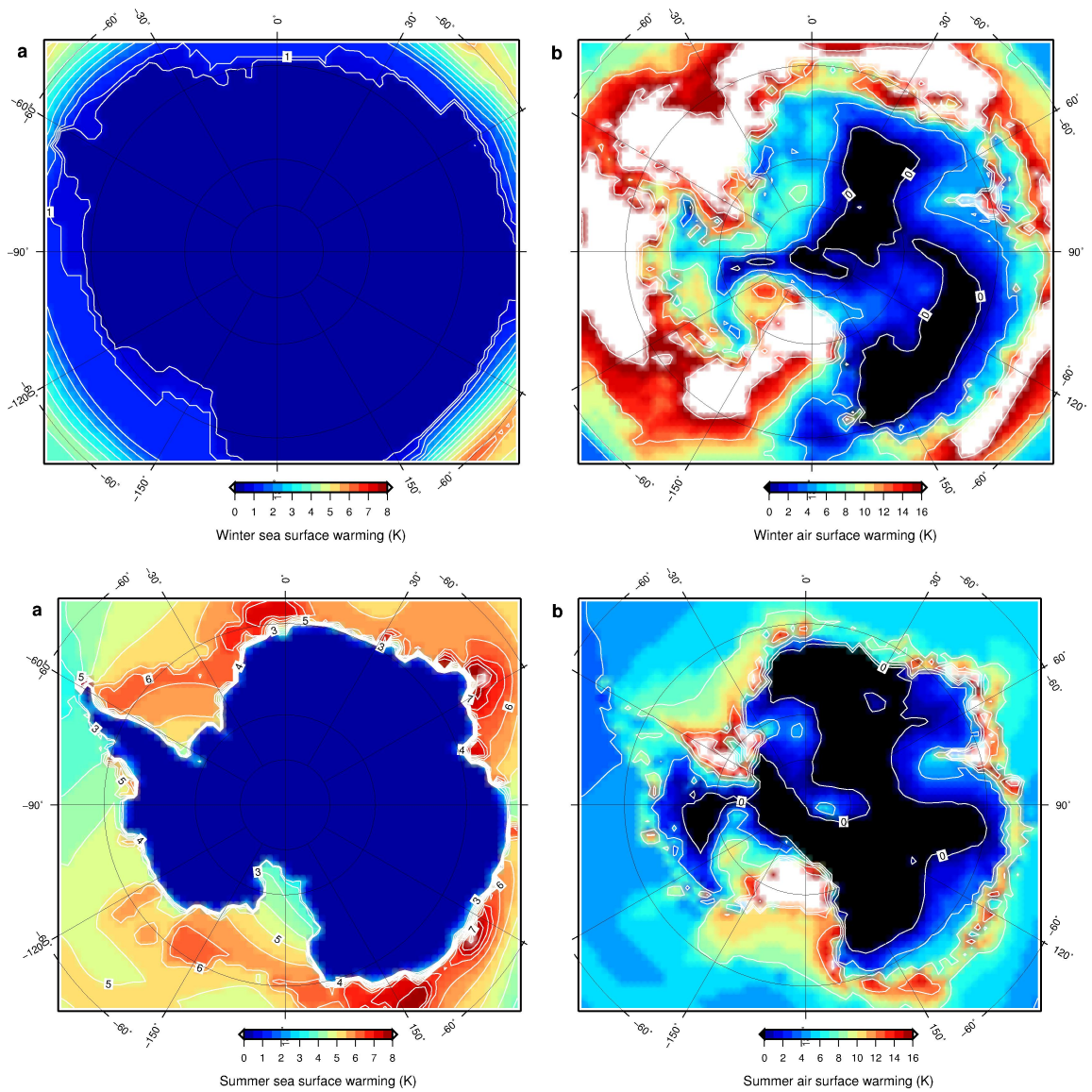

Fig. S3. *Seasonal climatologies for Antarctica during MIS31, as simulated by the Regional Climate Model, as anomalies from present-day.*

2.3. Ice Sheet Modelling

For our ice sheet experiments we use the climatologies from the RCM simulations (Fig. S3) as anomalies to present-day air temperature and precipitation fields (Lenaerts et al., 2012), in the same manner as employed previously (Golledge et al., 2017a). We use monthly climate fields in order to capture the high-latitude changes in seasonality brought about by the different orbital configuration of MIS31, compared to present. A positive-degree day model translates temperatures above freezing into surface melt, of which a proportion (60%) is retained in the snowpack through percolation refreezing processes. This value is difficult to constrain precisely (Kuipers Munneke et al., 2011) but here we use a constant value for both our control and our perturbation experiments, thus differences in the results are not likely to arise from errors in this parameterisation.

We use the Parallel Ice Sheet Model (PISM) version 0.6. Fast flow of grounded ice is facilitated by the plastic failure of subglacial sediments that arises as a consequence of an initially prescribed till friction angle, representing sediment strength, and its saturation state, which evolves dynamically in concert with the thermal properties of the basal layer of overlying ice. Thus where ice reaches sufficient thickness to allow basal melting to occur, meltwater begins to weaken the substrate until driving stresses exceed the evolving cohesiveness of the till. Failure of the substrate and the concomitant acceleration of overlying ice obeys a pseudo-plastic law (Schoof, 2006; Bueler and Brown, 2009), such that small increments of stress above the shear strength of the substrate lead to an increasing velocity response that ultimately thins the ice, reducing its driving stress and leading to deceleration of the ice. The cyclic

behaviour of ice streams that occurs as a consequence of this mechanism is described in more detail elsewhere (Van Pelt and Oerlemans, 20120. PISM employs a sub-grid grounding line scheme (Feldmann et al., 2014) in which sub-ice shelf melt may be interpolated across the grounded-floating ice junction, or not. The former approach tends to accelerate ice sheet retreat in marine basins, whereas the latter produces a slower response (Golledge et al., 2015). Equilibrium states tend to differ less between the two approaches than do retreat rates, however. For completeness, we run duplicate experiments both with, and without, the sub-ice shelf grounding line melt interpolation scheme turned on. Once afloat, we impose two calving mechanisms – one based on horizontal strain rates (Levermann et al., 2012), and another in which a simple minimum thickness criterion (200m) must be satisfied.

With the model setup described above, we undertook a series of simulations to explore the range of likely ice sheet responses to MIS31 climate states as simulated by the RCM. We ran experiments for two climate scenarios, reflecting atmospheric CO_2 concentrations of 280 and 400 ppm, and for each of these ran duplicates with each of the two grounding line parameterisations described above. We also ran a control experiment in which a present-day climate field was maintained throughout. All simulations were run for 10000 years and were initialised from a thermally and dynamically stable configuration used previously (Golledge et al., 2015; Golledge et al., 2017a; 2017b) which closely resembles the present-day ice sheet (Fig. S4 panel a). RCM-derived climatologies were imposed at the beginning of each run and the ice sheet was allowed to freely evolve to the new boundary conditions. Ice sheet configurations after 5000 years for each of the experiments are shown in Figure S4.

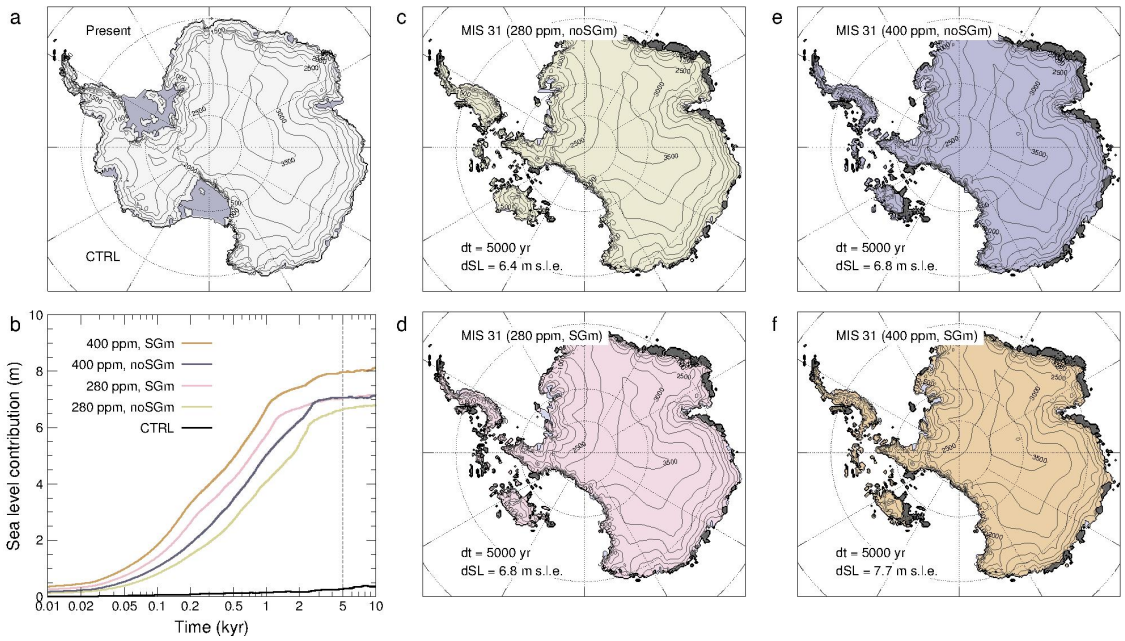

Fig. S4. *Ice sheet geometry and contribution to global sea level after 5000 years of simulated MIS31 climates, using either 280 ppm or 400 ppm atmospheric CO_2, and with one of two different grounding-line schemes (noSGm, SGm; see text for details). Note the rapid response of the ice sheet under all scenarios (panel b), corresponding to the collapse timescale of WAIS.*

Timeseries data illustrate that mass loss more-or-less stabilises after approximately 5000 years in most cases. In all MIS31 experiments, WAIS is lost relatively rapidly (first 1-2 millennia), with subsequent mass loss (from EAIS) taking place more slowly.

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