



# 1 Deglacial and Holocene sea ice and climate dynamics <sup>in?</sup> at the

## 2 Western Antarctic Peninsula

Is it appropriate to have WAP in the title since the study site is at the northern tip of the Bransfield Strait with both Weddell and Bellingshausen Sea influences? perhaps NAP would be more accurate?

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## 25 **Abstract**

26 The reconstruction of past sea ice distribution in the Southern Ocean is crucial for an improved understanding of  
 27 ice-ocean-atmosphere feedbacks and the evaluation of Earth system and Antarctic ice sheet models. The Western  
 28 Antarctic Peninsula (WAP) is experiencing rapid warming and the associated decrease in sea ice cover contrasts  
 29 the trend of growing sea ice extent in eastern Antarctica. To reveal the long-term sea ice history at the WAP under

needs a time frame



30 changing climate conditions we examined a marine sediment core from the eastern basin of the Bransfield Strait  
 31 covering the last Deglacial and the Holocene. For sea ice reconstructions, we focused on the specific sea ice  
 32 biomarker lipid IPSO<sub>25</sub>, a highly branched isoprenoid (HBI), and sea ice diatoms, whereas a phytoplankton-derived  
 33 HBI triene (C<sub>25:3</sub>) and open ocean diatom assemblages reflect predominantly ice-free conditions. We further  
 34 reconstruct ocean temperatures using glycerol dialkyl glycerol tetraether (GDGTs) and diatom assemblages, and  
 35 compare our sea ice and temperature records with published marine sediment and ice core data. Our results  
 36 document a retreat of the WAP ice shelf at 13.9 ka BP (before present). Maximum sea ice cover is observed during  
 37 the Antarctic Cold Reversal, while a still extended but variable sea ice coverage characterized the core site during  
 38 the early Holocene. An overall decreasing sea ice trend throughout the Middle Holocene is accompanied by a  
 39 successive ocean warming and increasing phytoplankton productivity. The Late Holocene is characterized by  
 40 unstable (winter) sea ice conditions and a further sea ice decline until 0.5 ka BP.

cold, warmer or both open ocean group(s)?

Consider changing to 'coincides with', 'parallels',

41  
 42 **Key Words:** Western Antarctic Peninsula, Holocene, sea ice cover, IPSO<sub>25</sub>, highly branched isoprenoids, diatoms

## 43 1 Introduction Needs to be more precise/accurate - too many imprecise and ambiguous statements.

44 Sea ice significantly affects the global climate system through its impact on the atmosphere-ocean exchange of  
 45 heat and gas, physical and chemical properties of the water masses, ocean circulation, primary production and  
 46 biogeochemical cycles (Chisholm, 2000; Vancoppenolle et al., 2013). Sea ice cover limits evaporation, affects  
 47 precipitation and increases the reflection of solar radiation due to a high albedo (Allison et al., 1982; Butterworth  
 48 and Miller, 2016; Turner et al., 2017). When sea ice forms, cold and dense brines develop contributing to the  
 49 formation of intermediate and deep waters (Nicholls et al., 2009). Importantly, these dense water masses can prevent  
 50 warm currents from reaching the continental slope where they stimulate the basal melt of Antarctic ice shelves  
 51 with implications for the stability of ice sheets and eventually global sea level (Cook et al., 2016; Escutia et al.,  
 52 2019; Etourneau et al., 2019; Hellmer et al., 2012; Huss and Farinotti, 2014). During the spring season, sea ice  
 53 melting stimulates marine primary production by seeding algal cells, releasing the release of nutrients and by promoting  
 54 ocean stratification and a shallow mixed layer depth (Arrigo et al., 1997; Vernet et al., 2008). In addition, nutrient  
 55 supply can be locally enhanced by increasing wind-driven upwelling activity along the sea ice edge, thus triggering  
 56 phytoplankton blooms (Alexander and Niebauer, 1981). Enhanced carbon fixation through this nutrient-stimulated  
 57 biological pump hence leads to an increase of biological material transport and organic carbon export to the ocean  
 58 floor, thus contributing to lower surface pCO<sub>2</sub> (Kim et al., 2004; Schofield et al., 2018; Wefer et al., 1988).

should this be continental shelf or inner-shelf?

Repetition from line 50? consider replacing with 'boosts/augments/increases marine primary production' OR 'triggers phytoplankton blooms' if removed from line 55/56

assuming you mean increased nutrients related to sea ice rather than all/any nutrient-driven production? if so, please consider changing this to 'sea ice-stimulated'

consistency? sea ice or sea-ice

59 Since satellite based sea-ice data became available in 1979, fast and profound changes have been observed globally  
 60 due to anthropogenic global warming (IPCC, 2021). The Western Antarctic Peninsula (WAP), in particular, is

more recent references?

makes no sense unless you include what the changes relate too ... but lack of 21st Century warming in the AP (Turner et al. 2016 in Nature)



What about since 2003?

61 experiencing a **rapid** warming of the atmosphere (Vaughan et al., 2003) and the ocean (Cook et al., 2016). This is not clear whether 'this' refers to atmosphere or ocean warming.

62 accompanied by **rapidly** retreating glaciers and ice shelves (Cook et al., 2016; Rignot et al., 2019) and by

63 **significant** loss of sea ice cover in the adjacent seas (Parkinson and Cavalieri, 2012).

64 For an assessment of the region's past sensitivity to climate change, the deglacial and Holocene climate history of

65 the Antarctic Peninsula (AP) has been studied extensively. The Deglacial, the transition from the Last Glacial

66 Maximum (LGM, Clark et al., 2012) to the Holocene, is characterized by a **rapid** warming punctuated by a distinct

67 cold event, the so-called Antarctic Cold Reversal (ACR) from 14.7 ka to 13 ka BP (EPICA Community Members,

68 2004; Mulvaney et al., 2012; Pedro et al., 2016). This drastic cooling of both atmosphere and ocean temperatures

69 is recorded by stable isotopes in Antarctic ice cores and marine sediments (Blunier and Brook, 2001; Domack et

70 al., 2001; Jouzel et al., 1995; Morigi et al., 2003; Stenni et al., 2001). From the Deglacial towards the Middle

71 Holocene, the Antarctic Peninsula Ice Sheet retreated **rapidly** from the outer shelf to its modern configuration with

72 heavy melt water discharge (Bentley et al., 2014). Several <sup>comparisons between</sup> ~~syntheses of~~ Holocene climate <sup>records from</sup> ~~reflected in~~ marine and

73 lake sediment cores reveal that the timing of both hydrological and environmental changes is highly variable at <sup>across/through</sup>

74 the WAP (Allen et al., 2010; Ingólfsson et al., 2003; Minzoni et al., 2015; Roseby et al., 2022; Sjunneskog and

75 Taylor, 2002; Totten et al., 2022). The ice-core records from James Ross Island (JRI) at the northeastern tip of the

76 AP shows a pronounced warming between about 12 and 11 ka BP followed by a cooling trend until about 9 ka BP

77 and stable temperatures until 2.5 ka BP. <sup>After/From 2.5 ka BP in the Late Holocene, temperatures cooled until 0.6 ka BP</sup> Late Holocene cooling was reversed since 0.6 ka BP (Mulvaney et al.,

78 2012). An overall consensus is that ocean temperature <sup>WAP</sup> ~~in the WAP~~ <sup>was</sup> ~~was~~, in comparison to the Deglacial or the

79 Late Holocene, warmer during the Early and Middle Holocene <sup>Optimum</sup>. i.e. between 12 ka and 4 ka BP. In

80 contrast, the Late Holocene shows many different climate patterns around the AP, including a continuous

81 Neoglacial cooling (Etourneau et al., 2013) whereas other records resolve warmer and colder phases such as the

82 Medieval Climate Anomaly and/or the Little Ice Age (Bentley et al., 2009).

83 **Knowledge** <sup>Southern Ocean/Antarctic</sup> of past sea ice variability is crucial for modelling climate feedbacks <sup>for accurate/to improve</sup> ~~impacting the Antarctic ice sheet~~

84 ~~stability since the LGM~~ (Crosta et al., 2022). For periods beyond the satellite era, sea-ice **knowledge** is based on

85 <sup>proxy records</sup> ~~proxies from~~ marine sediments and ice cores (e.g. Bracegirdle et al., 2015, 2019; Crosta et al., 2022; Escutia et al.,

86 2019; Thomas et al., 2019). Models, <sup>At present/Currently</sup> ~~however~~, often fail to reproduce seasonal sea ice cycles <sup>to match reconstructions/observations</sup> for both glacial and

87 interglacial periods and often disagree with geological proxies (Roche et al., 2012). Ice-core based sea ice

88 reconstructions ~~for the LGM~~ are primarily based on the concentrations of sea salt sodium (WAIS Divide Project

89 Members, 2015). However, since sea-salt aerosols might be overprinted by the highly variable wind direction and

90 meteorological conditions in Antarctica <sup>sea salt records may not be consistent in reflecting...</sup> ~~and thus not reflect~~ regional sea ice conditions (Thomas et al., 2019).

91 Although marine sediments <sup>records usually/frequently/often</sup> ~~mostly~~ have a lower temporal resolution than ice cores, <sup>marine proxy reconstructions</sup> ~~they~~ can resolve regional

worth including that evidence/  
influence of ACR largely confined to  
high Southern latitudes

Seems like a contradiction here...  
line 73-74 'environmental  
changes are highly variable' then  
'An overall consensus...'

...should make it clear whether  
records come from lake, terrestrial  
ice or marine archives.

Also, snow petrol stomach  
oil deposits (McClymont  
et al. 2022 COP)

also true for satellite era  
trends

delete "for the LGM" unless  
the sea salt sodium only applied  
to LGM OR there is another  
reason why the LGM is important  
in this context.

what about after the LGM?  
what about MSA?



92 and/or large-scale changes in sea ice conditions as well as sea surface and subsurface ocean temperature, primary  
 93 productivity and marine ecology (Hillaire-Marcel and de Vernal, 2007). In addition to commonly used  
 94 geochemical, lithological and microfossil proxies (*e.g.* ice rafted debris (IRD), diatom assemblages, total organic  
 95 carbon), new approaches focus on specific organic biomarkers - highly branched isoprenoids (HBIs) - as reliable  
 96 proxies to distinguish between open marine and sea ice covered environments. The diunsaturated HBI IPSO<sub>25</sub> (Ice  
 97 Proxy for the Southern Ocean, C<sub>25:2</sub>, Belt et al., 2016; Massé et al., 2011) has already been applied for Antarctic  
 98 sea ice reconstructions (*e.g.* Barbara et al., 2013; Denis et al., 2010; Etourneau et al., 2013). Following the PIP<sub>25</sub>  
 99 approach for the Arctic (Müller et al., 2011), IPSO<sub>25</sub> has been combined with HBI trienes and/or sterols to  
 100 determine the phytoplankton-IPSO<sub>25</sub> sea ice index called PIPSO<sub>25</sub> (Vorrath et al., 2019), which has been  
 101 successfully evaluated with recent sea ice concentrations (Lamping et al., 2021), over periods of the industrial era  
 102 (Vorrath et al., 2020) and has been used in paleo sea ice studies from the Amundsen Sea (Lamping et al., 2020).  
 103 Hence, the combination of these new molecular proxies with the classical proxies offer a unique opportunity to  
 104 robustly reconstruct past sea ice at the WAP. Here, we present a marine sediment record covering the past 13.9 ka BP to reconstruct Deglacial and Holocene  
 105 environmental conditions at the WAP. Our study is based on a multiproxy approach focusing on the sea ice  
 106 biomarker IPSO<sub>25</sub>, a marine phytoplankton biomarker (HBI triene), and on glycerol dialkyl glycerol tetraether  
 107 lipids (GDGTs) for subsurface ocean temperatures (SOT). Additional information about the probability of winter  
 108 sea ice (WSI) and summer sea surface temperature (SSST) comes from diatom assemblages using transfer  
 109 functions. We discuss and compare our proxy results with other marine sediment and ice core records spanning  
 110 the Holocene providing further insight into environmental dynamics across the Deglacial and the Holocene.

## 112 2 Material and Methods

### 113 2.1 Study Area

114 The Bransfield Strait is located between the WAP and the South Shetland Islands (SSI) (Fig. 1a). Within this area,  
 115 a shallow shelf and deeper depressions characterize the Bransfield Basin with water depths exceeding 2000 m  
 116 (Fig. 1b). The shelf area was affected by intense ice sheet dynamics during the last glaciation (Canals and Amblas,  
 117 2016b; Ingólfsson et al., 2003) leaving ice sheet grounding lines and glacial troughs on the sea floor (Canals et al.,  
 118 2016; Canals and Amblas, 2016a). The Bransfield Basin is influenced by complex oceanic current systems, which  
 119 are not fully constrained because three different water masses enter the basin from the east and west (Moffat and  
 120 Meredith, 2018; Sangrà et al., 2011) and their mixing is not well understood. The cold (< 0°C) and relatively salty  
 121 Weddell Sea Water (WSW) enters from the east, flows alongshore the peninsula and fills the basin below 150 m



need to focus much more of the study region section on the sea ice and temperature setting including seasonal patterns.  
 What is the modern (winter/summer & mean annual?) concentrations of sea ice cover? SOT? etc.

watermasses are not currents  
 so these terms are not inter-  
 changeable.

Rephrase to make it clear that  
 fronts form between water  
 masses. Also not clear how WSW  
 produces surface front if it below  
 150 or 200m.

what is the mixed layer depth?

122 surface. The WSW is also observed at greater depths (200-600 m) north of the SSI and at Elephant Island due to  
 123 wind driven modulation (Meijers et al., 2016). In the western part of the Bransfield Strait, the WSW mixes with a  
 124 second water mass, the Bellingshausen Sea Water (BSW) (Collares et al., 2018), which is a branch of the Antarctic  
 125 Circumpolar Current (ACC). It conveys well-stratified, fresh and warmer ( $> 0^{\circ}\text{C}$ ) surface water (Sangrà et al.,  
 126 2011). A third water mass, originating from the Circumpolar Deep Water (CDW) is present between 200 m and  
 127 550 m (Sangrà et al., 2017). BSW and CDW flow eastward along the SSI, turn around and flow westward at the  
 128 northern tip of the islands (Sangrà et al., 2011). BSW forms the subsurface Bransfield front with the CDW at depth  
 129 and the surface Peninsula Front (PF) with the WSW, parallel to the Antarctic mainland (Sangrà et al., 2011, 2017).  
 130 The interplay of currents leads to a stratification of the water column of the upper 20 m in summer, with a steep  
 131 temperature gradient in the first 100 m below sea surface. This can be observed in CTD profiles in the Bransfield  
 132 Basin that show a dominance of WSW below 200 m (see Fig. 1c and Sangrà et al., 2011). The eddy system at the  
 133 Peninsula Front is assumed to play a key role for mixing and upwelling of the different surface and subsurface  
 134 water masses (Sangrà et al., 2011; Zhou et al., 2002), while several glaciers from the WAP influence coastal  
 135 surface water due to meltwater discharge and also transport dense bottom waters to the Bransfield Basin (Meredith  
 136 et al., 2018).  
 137 Primary production is mainly driven by mixing of water masses at the fronts (Gonçalves-Araujo et al., 2015),  
 138 mixed layer depth and upwelling (Sangrà et al., 2011), sea ice dynamics (Vernet et al., 2008) and iron availability  
 139 (Klunder et al., 2014). High concentrations of chlorophyll *a* and diatoms are distributed north of the Peninsula  
 140 Front and at the SSI, while lower production and communities of nanoplanktonic flagellates are found between  
 141 the Peninsula Front and the WAP (Gonçalves-Araujo et al., 2015). Further, changes in coastal primary production  
 142 are driven by upwelling, iron distribution and the retreat of sea ice cover in spring releasing nutrients and  
 143 stabilizing the water column (Vernet et al., 2008). A close link between marine primary production at the surface  
 144 and sediment composition at the ocean floor is reflected in high concentrations of total organic carbon (TOC),  
 145 pigments, sterols and diatoms (Cárdenas et al., 2019), and supported by studies confirming high fluxes of sinking  
 146 particles (Kim et al., 2004; Wefer et al., 1988). In the study area, particle flux is highly variable with seasonal  
 147 peaks occurring in late spring, which accounts for 85% of the total flux (Ducklow et al., 2008). Lithologically, the  
 148 sediments consist mainly of terrestrial silt and clay with varying amounts of diatom mud and ooze, and sand (Cádiz  
 149 Hernández, 2019; Lamy, 2016; Wu et al., 2019).

## 2.2 Sediment samples and age model

151 Piston core PS97/072-1 ( $62^{\circ} 0.39' \text{ S}$ ,  $56^{\circ} 3.86' \text{ W}$ , 1993 m water depth, 1583 cm in length) was recovered in the  
 152 eastern Bransfield Strait Basin during R/V *Polarstern* cruise PS97 (Lamy, 2016) (Fig. 1). The core is dominated

Confusing since this depth range is the same as WSW (Line 122)

make it clear where the westward flow of these waters occurs (along the AP?)

consistency - Front not front

Consider using PenF to distinguish from Polar Front.

How does 100m deep temperature gradient relate to the 20m stratified layer?

BSW is the only water mass described as 'sub-surface', no surface waters described.

only the Peninsula Front is mentioned above.

\*<https://paleowave.net/paleowave/2012/02/08/planktic-not-planktonic>

consider rephrasing as... the nutrient release and surface water stratification generated by melting sea ice in the austral spring.

more recent references?



153 by silt with thin layers of sand, clay, and traces of volcanic ash. Single pebbles are present below 630 cm. Since  
 154 we found disturbed sediments below 1015 cm depth, we only considered samples from above this level for our  
 155 analyses. After an XRF scan the core sampling was done at the Alfred Wegener Institute (AWI) where the samples  
 156 were stored frozen in glass vials (for biomarker analysis) and at 4° C in plastic bags (for micropaleontology).

157 The age model of core PS97/072-1 is based on  $^{14}\text{C}$  radiocarbon dating of eight calcite samples with the mini carbon  
 158 dating system (MICADAS) at AWI (Mollenhauer et al., 2021). From the conventional  $^{14}\text{C}$  age we subtracted a  
 159 reservoir age based on modelling by Butzin et al. (2017) and also subtracted an estimated ventilation age of 1200

add reference for choice of ventilation age here and/or supp table.

160 years (see table supplement section 1) before we calibrated the ages with the calibration curve SHCal20 (Hogg et  
 161 al., 2020) to calendar years before present (cal BP) with Calib 7.1 (Stuiver et al., 2018). To estimate the top age of  
 162 the core, TOC and biogenic opal data of the piston core were matched with data from a multicore from the same  
 163 sampling site that has been previously dated via  $^{210}\text{Pb}$  (Vorrath et al., 2020) (supplement section 2). We applied

164 the Bayesian age modelling tool *hummingage*, a freely available tool developed at AWI, that has been successfully  
 165 applied in previous studies (Ronge et al., 2021).

Does 'hummingage' not give errors or date ranges as part of its output?

What about the age of sediments below the deepest radiocarbon date at 868cm? linear extrapolation or Bayesian modelling to 1015cm?

### 166 **2.3 Organic geochemical analyses of piston core PS97/072-1**

Please add information on the number of samples and the min/max or mean+sd. sample intervals for each of the analyses described.

167 For the analyses of several organic components and biomarkers the sediments were freeze-dried and homogenized  
 168 in an agate mortar. Total carbon (C) and nitrogen (N) were measured with a CNS analyzer (Elementar Vario EL  
 169 III, error of standards and duplicates < 5%). TOC was measured on 0.1 g of acidified samples (500 µl HCl) and  
 170 determined in a carbon-sulphur determinator (CS-800, ELTRA, standard error < 0.6%). To identify the source of  
 171 TOC, measurements of stable carbon isotopes of bulk organic matter were done at Universität Hamburg (UHH),  
 172 Germany, and at Washington State University (WSU), USA. At UHH, the samples were acidified three times with  
 173 100 µl 1 N HCl and dried on a hotplate. High-temperature combustion was done in an Elementar CHNOS Vario  
 174 isotope elemental analyser at 950° C and the analysis was conducted with an Elementar IsoPrime 100 isotope ratio  
 175 mass spectrometer. We calibrated the pure tank  $\text{CO}_2$  with the International Atomic Energy Agency reference  
 176 standards IAEA-CH6 and IAEA-CH7. These and two other standards (IVA Sediment and Sucrose) acted as  
 177 internal standards in the measurement. The error of continuous standard duplicates was < 0.2‰ and < 0.06‰ for  
 178 sample duplicates. At WSU, 100 mg of freeze-dried sediment samples were used. An elemental analyzer coupled  
 179 with an Isoprime isotope ratio mass spectrometer (IRMS) was used, with a precision of 0.1‰. The running  
 180 standard was a protein hydrolysate calibrated against NIST standards. Isotope ratios are expressed in units per mil  
 181 (‰).  $\delta^{13}\text{C}$  values are expressed in ‰ against Vienna Pee Dee Belemnite (VPDB).

182 Biogenic opal was estimated following the alkaline extraction procedure described by Mortlock and Froelich  
 183 (1989), but using 0.5M NaOH as a digestion solution (Müller and Schneider, 1993). Extraction and analysis by



molybdate-blue spectrophotometry were conducted at the University of Concepción, Chile. Values are expressed as biogenic opal by multiplying the Si (%) by 2.4 (Mortlock and Froelich, 1989). We did not correct for the release of extractable Si from coexisting clay minerals, and thus biogenic opal values could be slightly overestimated (Schlüter and Rickert, 1998). Instrumental precision was  $\pm 0.5\%$ ; error of duplicates  $\leq 3\%$ . Details on the methodology used can be found in Cárdenas et al. (2019).

The extraction, purification and identification of HBIs followed the analytical protocol published e.g. in Belt et al. (Belt et al., 2014) and Vorrath et al. (2019). 7-hexylnonadecane (7-HND) and  $C_{46}$  served as internal standards. Lipids were extracted using ultra sonication and a mixture of  $CH_2Cl_2$ :MeOH (v/v 2:1; 6ml). HBIs and GDGTs were separated by means of open column chromatography using  $SiO_2$  as the stationary phase and hexane, and  $CH_2Cl_2$ :MeOH (v/v 1:1) as eluents. HBIs were analyzed by means of an Agilent 7890B gas chromatograph (30 m DB IMS column, 0.25 mm diameter, 0.250  $\mu m$  film thickness) coupled to an Agilent 5977B mass spectrometer (MSD, 70 eV constant ionization potential, ion source temperature 230° C). The initial oven temperature of 60° C was held for 3 min, ramped to 325° C within 23 min, and was held at 325° C for 16 min. HBIs were identified *via* comparison of their retention times and mass spectra with published mass spectra (Belt et al., 2000) and quantified using the ratio of peak areas of individual HBIs ( $m/z$  346;  $m/z$  348) and the 7-HND ( $m/z$  266) standard and consideration of instrumental response factors. The error of duplicates was  $<1.4\%$  for  $IPSO_{25}$ ,  $<2.6\%$  for HBI trienes. The phytoplankton- $IPSO_{25}$  index ( $PIPSO_{25}$ ) was calculated after Vorrath et al. (2019) as:

$$PIPSO_{25} = \frac{IPSO_{25}}{IPSO_{25} + (c \times \text{phytoplankton marker})} \quad (1)$$

The HBI z-triene <sup>is</sup> ~~was~~ considered as a phytoplankton biomarker and, since <sup>in these samples at</sup> the concentrations were on the same level as  $IPSO_{25}$ , the c-factor was set to 1 (Vorrath et al., 2019). To confirm the sea-ice origin of  $IPSO_{25}$ , the stable carbon isotope composition of  $IPSO_{25}$  was examined in 8 samples (with minimum 50 ng carbon) via GC-irm-MS at the GFZ Potsdam, Germany. The GC (7890N Agilent) equipped with Ultra1 column (50 m x 0.2 mm diameter, 0.33  $\mu m$  film thickness) was connected to a DeltaVPlus isotope ratio mass spectrometer through a modified GC-Isolink interface. Each sample was separated chromatographically with a temperature program that started with an oven temperature of 80° C, which was held for 3 min, ramped to 250° C with 3° C per min and then ramped to 320° C with 5° C per min and finally reached temperature of 325° C with a ramp of 1° C per min and held for 15 min. The organic substances of the GC effluent stream were oxidized to  $CO_2$  in the combustion furnace held at 940° C on a CuO/Ni/Pt catalyst. Samples were measured in duplicate and the standard deviation was  $\leq 0.5\%$ . The quality of the isotope measurements was checked <sup>regularly</sup> ~~by~~ by measuring different *n*-alkane standards with known isotopic composition (provided by Campro Scientific, Germany and Arndt Schimmelmann, Indiana University, USA).

please provide some quantification/approximation of what you mean by 'regularly'





GDGTs were re-dissolved in 120 µl hexane:isopropanol (v/v 99:1) and filtered through polytetrafluoroethylene  
 filters (0.45 µm in diameter) and analyzed using high performance liquid chromatography (HPLC, Agilent 1200  
 series HPLC system) coupled to a single quadrupole mass spectrometer (MS, Agilent 6120 MSD) via an  
 atmospheric pressure chemical ionization (APCI) interface. The individual GDGTs were separated at 30° C on a  
 Prevail Cyano column (150 mm x 2.1 mm, 3µm). After injection of the sample (20 µl) it passed a 5 min isocratic  
 elution with mobile phase A (hexane/2-propanol/chloroform; 98:1:1, flow rate 0.2 ml/min). The mobile phase B  
 (hexane/2-propanol/chloroform; 89:10:1) was increased to 100% in two steps: a linear increase to 10% over 20  
 min followed by an increase to 100% within 10 min. During the measurement, the column was cleaned after 7 min  
 via backflush (5 min, flow 0.6 ml/min) and re-equilibrated with solvent A (10 min, flow 0.2 ml/min). The  
 conditions of the APCI were a nebulizer pressure of 50 psi, vaporizer temperature and N<sub>2</sub> drying gas temperature  
 350°C, flow 5 l/min, capillary voltage 4 kV, and corona current 5 µA. The GDGTs were detected by selective ion  
 monitoring (SIM) of (M+H<sup>+</sup>) ions (dwell time 76 ms). In relation to the internal standard C<sub>46</sub> (*m/z* 744) the  
 molecular ions *m/z* of GDGTs-I (*m/z* 1300), GDGTs-II (*m/z* 1298), GDGTs-III (*m/z* 1296), and crenarchaeol (*m/z*  
 1292) were quantified. Also, the branched GDGTs-Ia (*m/z* 1022), GDGTs-IIa (*m/z* 1036), GDGTs-IIIa (*m/z* 1050)  
 were quantified. The hydroxylated GDGTs OH-GDGT-0 (*m/z* 1318), OH-GDGT-1 (*m/z* 1316), and OH-GDGT-2  
 (*m/z* 1314) were quantified in the scans of their related GDGTs (Fietz et al., 2013). The standard deviation was  
 0.01 units of TEX<sub>86</sub><sup>L</sup>.

Kalanetra et al. (2009) showed that GDGT-producing Thaumarchaeota are abundant in subsurface waters in both  
 Arctic and Antarctic regions. As Thaumarchaeota were found between 50 m and 200 m water depth in Antarctica  
 (Kim et al., 2012), temperatures based on GDGTs are suggested to reflect sub-surface waters (Etourneau et al.,  
 2013, 2019). Similarly, also RI-OH' based temperatures in Prydz Bay have been interpreted to reflect subsurface  
 water temperatures (Liu et al., 2020). We therefore consider our results to reflect subsurface ocean temperatures  
 (SOTs). We calculated TEX<sub>86</sub><sup>L</sup> after Kim et al. (2012) with the *m/z* 12963 (GDGT-3), *m/z* 1298 (GDGT-2), *m/z*  
 1300 (GDGT-1):

$$TEX_{86}^L = \log \left( \frac{[GDGT-2]}{[GDGT-1] + [GDGT-2] + [GDGT-3]} \right) \quad (2)$$

$$\text{and calibrated with } SOT = 50.8 * TEX_{86}^L + 36.1 \text{ (Kim et al., 2012).} \quad (3)$$

For the calculation of temperatures based on hydroxylated GDGTs we followed the approach of Lü et al. (2015)

$$RI - OH' = \frac{[OH-GDGT-1] + 2 * [OH-GDGT-2]}{[OH-GDGT-0] + [OH-GDGT-1] + [OH-GDGT-2]} \quad (4)$$

$$\text{and calibrated it with } SOT = (RI - OH' - 0.1) / 0.0382. \quad (5)$$





244 For the branched and isoprenoid tetraether (BIT) index for indicating terrestrial organic matter (Hopmans et al.,  
 245 2004) we used crenarchaeol (m/z 1292) and the branched GDGTs and calculated it as:

$$246 \quad BIT = \frac{[GDGT-Ia] + [GDGT-IIa] + [GDGT-IIIa]}{[Crenarchaeol] + [GDGT-Ia] + [GDGT-IIa] + [GDGT-IIIa]} \quad (6)$$

## 248 2.4 Diatom analyses

249 We selected a set of 76 samples for the analysis of diatom assemblages. Freeze-dried samples (20-120 mg) were  
 250 treated with hydrogen peroxide and sodium pyrophosphate to remove organic matter and clays, respectively,  
 251 washed several times with DI water until reaching neutral pH. The treated samples were then settled for six hours  
 252 in B-Ker2 settling chambers to promote an even distribution of settled particles (Scherer, 1994; Schrader and  
 253 Gersonde, 1978; Warnock and Scherer, 2015). Once the samples were dry, the quantitative slides were mounted  
 254 with Norland mounting medium (refraction index=1.56). Diatom valves per slide were counted across traverses  
 255 (at least 400 valves per slide) using an Axioscop 2 Plus and Olympus BX60 at a magnification of ×1000. The  
 256 counting procedure and definition of counting units followed those of Schrader and Gersonde (1978). We  
 257 performed two sets of counts, with and without *Chaetoceros* resting spores. Diatoms were identified to species or  
 258 species group level and, if applicable, to variety or form level following the taxonomy described by e.g., Gersonde  
 259 and Zielinski (2000), Armand and Zielinski (2001), Esper et al. (2010), Esper and Gersonde (2014a, 2014b).

260 Diatom analyses were done at the University of Concepción, Chile, and at Colgate University, USA, especially with regards to assigning morphological  
 varieties ie. T1 and T2 of *Thalassiosira antarctica*?

261 Diatom species were grouped into ecological assemblages reflecting i) seasonal sea-ice, ii) cold open ocean, iii)  
 262 warmer open ocean, and iv) benthic-epiphytic diatoms environments (Buffen et al., 2007; Cárdenas et al., 2019;  
 263 Esper et al., 2010). Additionally, a group of reworked diatoms was identified. A Spearman principal component  
 need to be clear which species  
 comprise each grouping (either here  
 or in the supplementary table). Were  
 all taxa assigned to an ecological  
 group or were some excluded?

264 analysis (PCA) was applied to the diatom assemblages to differentiate their temporal distribution. PCA will show which assemblage group best describes  
 the variability in the total assemblage not necessarily  
 their 'distribution or abundance.

265 For estimation of winter sea ice (WSI) concentrations we applied the transfer function MAT-D274/28/4an which  
 266 comprises 274 reference samples with 28 diatom taxa/taxa groups and considers an average of 4 analogues (Esper  
 267 and Gersonde, 2014a). The analogues refer to surface sediments from the Atlantic, Pacific and western Indian  
 268 sector of the Southern Ocean. The WSI renders sea-ice concentrations in a 1° by 1° grid for the September average

269 of the period 1981 to 2010 (Reynolds et al., 2002, 2007). The threshold of an open ocean to sea ice covered area  
 270 is set at 15% of sea ice concentration (Zwally et al., 2002) and the average sea ice edge is defined at 40% (Gersonde  
 271 et al., 2005; Gloersen et al., 1993). The qualitative estimation of sea ice concentration was derived from the  
 272 abundance pattern of diatom sea-ice indicators (Gersonde and Zielinski, 2000). The estimation of summer sea

273 surface temperature (SSST) came from the transfer function IKM-D336/29/3q comprising 336 reference samples  
 274 (Pacific, Atlantic and Indian Southern Ocean) with 29 diatom taxa and three factors (Esper and Gersonde, 2014b).

Please state how many reference  
 sites are in the WAP and whether  
 Transfer Function applied to total  
 counts or *Chaetoceros* free counts?

Is this used/presented in manuscript  
 or supplementary figures? If so please  
 explain how this differs from the  
 seasonal sea-ice assemblage  
 ecological group?



PCA? MAT? IKM?  
 275 The calculations were done with the software R (R Core Team, 2012) using the packages Vegan (Oksanen et al.,  
 276 2012) and Analogue (Simpson and Oksanen, 2012).

### 277 3 Results

278 Based on our age model the sediment core covers the last 13.9 ka BP with a mean sedimentation rate of 67 cm/ka  
 279 and a resolution ranging between 50 and 150 years per sample. We note a higher sedimentation rate of 95 cm/ka  
 280 between 5.5 ka and 3 ka BP and few short-term intervals of <sup>greatly?particularly?</sup>significantly lowered (19 cm/ka) and enhanced (190  
 281 cm/ka) sedimentation (Fig. 2).

This cannot refer to all sample intervals... (eg. Diatom samples: Some intervals of 20 to 52 cm = ~300 to 770 years!). Please amend or specify which analyses apply to.

282 Organic geochemical bulk parameters (TOC, biogenic opal), concentrations of HBI (IPSO<sub>25</sub>, C<sub>25:3</sub> HBI triene) and  
 283 diatom species assemblages of piston core PS97/072-1 are summarized in Figure 3 (additional data can be found  
 284 in the supplement section 3). TOC increases from very low values of 0.1 wt% at 13.7 ka BP to an average  
 285 concentration of ~0.8 wt% between 9.9 ka BP and the top of the core with recurring short-lived minima during the  
 286 Middle and Late Holocene (Fig. 3f). Some of these TOC minima may be associated with thin sandy layers of  
 287 volcanic ash. Biogenic opal shows a similar pattern with minimum values in the lower part of the record (3.2 wt%  
 288 at 13.0 ka BP) and increases throughout the Deglacial to Holocene with average values of 30 wt% and a maximum  
 289 of 54.4 wt% at 5.3 ka BP (Fig. 3e).

of ?? wt%  
 Including values in your description make it much easier to follow/understand so please do the same for your other parameters.

290 Between 13.9 ka and 13.4 ka BP, both IPSO<sub>25</sub> and HBI triene concentrations are close to or below the detection  
 291 limit. The IPSO<sub>25</sub> concentration ranges between 0.1 to 31.5 µg g<sup>-1</sup> TOC, while the concentration of the HBI triene  
 292 ranges between 0.1 and 6.6 µg g<sup>-1</sup> TOC (Fig. 3). IPSO<sub>25</sub> is absent before 13.5 ka BP and rises rapidly to maximum  
 293 values at 12.9 ka BP. Subsequently, concentrations decrease steadily until 8.5 ka BP and then remain on an average  
 294 level of ~4 µg g<sup>-1</sup> TOC with a slightly decreasing trend towards the present and smaller peaks at 6.0 and 3.0 ka  
 295 BP. The HBI triene is largely absent until 13.0 ka BP and shows high concentrations after 8.5 ka BP with large  
 296 fluctuations in the Middle Holocene and from 3.4 ka BP to the present.

Would be good if this was indicated on the plots using dotted line or similar.

297 The diatom composition has two contrasting groups indicating open ocean conditions (Fig. 3a) and seasonal sea  
 298 ice (Fig. 3c). Although the group reflecting seasonal sea ice is also present throughout the core (mostly >20%),  
 299 the highest contributions are seen before 12.8 ka BP, between 10.8 and 9.9 ka BP and around 3 ka BP. The  
 300 contribution of the open ocean assemblage is very low in the Deglacial and Early Holocene and rises to highest  
 301 values in the Middle Holocene and remains around 10% in the Late Holocene. A biplot of a principal component  
 302 analysis (PCA) shows the relationship of the ecological groups <sup>for three time intervals</sup>along the sediment core with clear dominance of  
 303 seasonal sea-ice before 13.3 ka BP and warmer open ocean conditions after 8.5 ka BP (supplement section 4).

are the two contrasting groups the cold and warmer open ocean assemblages? or the cold &/or warmer open ocean assemblage contrasting with seasonal sea ice?

which open ocean assemblage? cold or warmer?

304 Sea ice estimates based on diatom assemblages (WSI) and the IPSO<sub>25</sub> index as well as the content of IRD in  
 305 PS97/72-1 are summarized in figure 4 (a-c). Reconstructed winter sea ice concentrations (% WSI) derived from

not clear if this refers to the same as the 'open ocean assemblage' above (lines 297 and 300).

Not clear whether or not IRD content is included as tool for sea ice estimation or something else?

Only the 'open ocean' and 'sea ice' assemblages are plotted in Fig 3 and none in the supplementary file. Also, I think the 'open ocean assemblage' should be 'WARM/WARMER open ocean assemblage'

Not clear what 'also' refers to here? the HBI's or the open ocean group(s)?

Please explain how the IPSO index relates to sea-ice cover ie. does the scale of 0 to 1 equate to a scale from open ocean to full sea ice cover or similar?



Most of decrease is concentrated between 8 and 5 ka. Might be weak decrease earlier than 8 ka, but definitely not much evidence of decrease after 5 ka.

Need to have the sample points marked on the plots to evaluate whether sample resolution is influencing where these 'distinct fluctuations' occur

Are there any samples in the MAT results that either don't have 4 analogues or that only match with analogues of very different oceanographic setting(s)?

the MAT results from the

306 diatom assemblages range from 80% to 90% during the ACR and the Deglacial (13.9 ka – 11 ka BP) and exhibit

307 an overall decreasing trend over the Holocene with distinct fluctuations reaching minimum sea ice concentrations

308 of ca. 65% during the Middle and Late Holocene (Fig. 4a). PIPSO<sub>25</sub> values show a similar trend indicating higher

309 sea ice cover during the ACR, the Deglacial and the Early Holocene (PIP SO<sub>25</sub> > 0.8) and a successive decline to/of ?? µg g<sup>-1</sup>

310 throughout the Middle and Late Holocene with a distinct minimum at 0.5 ka BP (Fig. 4b). IRD (lithic particles

311 and pebbles > 5 µm) occurs frequently between 13.9 ka and 9 ka BP and is virtually absent in the younger part of

312 the sediment core (Fig. 4c).

313 Figure 5 provides ocean temperature reconstructions based on diatom assemblages (SSST) and GDGT-derived

314 RI-OH' and TEX<sub>86</sub><sup>L</sup> SOTs in core PS97/72-1 (Fig. 5 b-d). SSST estimates derived from diatom data generally

315 have minimum temperatures of -1.5°C to 0°C during the Deglacial and a warming trend (> 0°C) in the Middle and

316 Late Holocene with a distinct cold event at 3.1 ka BP (Fig. 5b). RI-OH'-derived SOTs reflect generally lower

317 temperatures between -1.9 to -1.2°C and a similar trend of rising temperatures until 4.2 ka BP followed by a subtle

318 cooling (Fig. 5c). TEX<sub>86</sub><sup>L</sup> data from GDGTs cover a temperature range of 0.7 to 3.8°C and display an opposite

319 trend to both SSST and RI-OH' SOT with decreasing temperatures since the Deglacial (Fig. 5d).

320

321 **4 Discussion**

322 **4.1 The late Deglacial (14 ka to 11.7 ka BP)**

323 In the oldest part of our sediment record covering the later part of the last Deglacial from 14 ka until 11.7 ka BP

324 we observe a remarkable environmental change indicated by significant shifts in the TOC, biomarker and diatom

325 records. Before 13.4 ka BP, the very low concentrations of biomarkers, TOC, and biogenic opal suggest that

326 primary production associated with open marine and also sea ice settings was diminished, while sea-ice related

327 diatom species show the highest contribution (Fig. 3) albeit very low concentrations (supplement section 5).

328 Highest PIPSO<sub>25</sub> and WSI values pointing towards maximum sea ice cover and lowest ocean temperatures

329 reflected in the RI-OH'-derived SOTs are well in line with peak ssNa concentrations and minimum δ<sup>18</sup>O values in

330 the EDML, WAIS and JRI ice core records referring to an extended sea ice cover and lowered atmospheric

331 temperatures (Fig. 4 and 5; EPICA Community Members, 2006; Fischer et al., 2007; Mulvaney et al., 2012; WAIS

332 Divide Project Members, 2015; WAIS Divide Project Members et al., 2013). The near absence of IPSO<sub>25</sub>, HBI

333 triene and open ocean diatom species between 13.9 ka and 13.5 ka BP evidences a very thick or permanent sea ice

334 cover. Similarly, Lamping et al. (2020) relate the absence of IPSO<sub>25</sub> and a phytoplankton biomarker in sediments

335 in the western Amundsen Sea to the re-advance of a floating ice shelf canopy during the ACR. The abrupt increase

You should perhaps mention that this is different to the abundance pattern in the seasonal sea ice assemblage presented above.

'significant' should be reserved for statistical results. Consider replacing with conspicuous, marked, noticeable, prominent...

would be better to include a plot of the diatom concentrations over time.

obvious/prominent/clear?

(fig 2d) (Fig 2f) (fig 2e)

limited?

of >?? %

In agreement with the biogenic opal and mod

warm, cold or both open ocean?



336 in IPSO<sub>25</sub> concentrations at 13.5 ka BP, may indicate the retreat of such an <sup>ice-canopy</sup> ice-shelf-cover from the core site  
 337 permitting sea ice growth during spring and a subsequent increase in primary production reflected in rapidly rising  
 338 HBI triene concentrations at 13 ka BP. A <sup>marked/prominent?</sup> significant decrease in sea ice associated diatoms between 13 ka and 12  
 339 ka BP (Fig. 3), however, is not mirrored by the still high WSI. We note that traces of biomarkers and diatoms  
 340 (supplement section 5) deposited in sediments older than 13.5 ka BP may reflect sub-ice shelf lateral advection  
 341 and reworking (Smith et al., 2019).

342 While the ACR lasts from 14.7 ka to 13 ka BP (Pedro et al., 2016), our sediment record shows that cold conditions  
 343 with extended sea ice cover and reduced ocean temperatures in the eastern Bransfield Strait lasted until ca. 11 ka  
 344 BP (Figs. 4 and 5). Further, the IRD content (including the presence of single large pebbles) points to the frequent  
 345 occurrence of icebergs during the Deglacial and the Early Holocene (Fig. 4c) related to the overall ice sheet

346 disintegration <sup>in/along</sup> at the WAP that occurred around 14 ka BP at the South Shetland Islands and at 13.2 ka BP at Palmer  
 347 Deep at the southern WAP (Domack et al., 2001; Domack, 2002; Jones et al., 2022; Milliken et al., 2009). A slight  
 348 decrease in PIPSO<sub>25</sub> values and rising RI-OH' SOTs characterize the late Deglacial between 13 ka and 11.7 ka  
 349 BP. This warming may relate to inter-hemispheric teleconnections through a global reorganization of atmospheric  
 350 and ocean circulation that is related to the bipolar seesaw pattern of opposite climate trends between the northern  
 351 and southern hemisphere (Anderson et al., 2009; Broecker, 1998; EPICA Community Members, 2006; Pedro et  
 352 al., 2016). With cooling of the northern hemisphere, a southward shift of the Intertropical Convergence Zone and  
 353 the southern hemisphere westerlies (Lamy et al., 2007) resulted in intensified wind stress in the Drake Passage  
 354 (Timmermann et al., 2007) and increased upwelling that may have driven the continued warming and sea ice  
 355 retreat in Antarctica towards the Holocene (Anderson et al., 2009).

## 356 4.2 Early Holocene warming from 11.7 ka to 8.2 ka BP

357 The Early Holocene from 11.7 ka to 8.2 ka BP is characterized by a progressively decreasing though highly  
 358 variable <sup>seasonal?</sup> winter and spring sea ice cover as shown by further declining WSI and PIPSO<sub>25</sub> values (Fig. 4a and b).  
 359 While biogenic opal and TOC contents exhibit increasing trends, concentrations of the HBI triene and open ocean  
 360 diatoms remain low and only a <sup>marked/prominent?</sup> significant increase after 9 ka BP suggests higher phytoplankton productivity (Fig.  
 361 3). Ocean warming is indicated by RI-OH'-based SOT, while TEX<sub>86</sub><sup>L</sup> SOT and diatom-derived SSST show  
 362 fluctuating temperatures without a clear trend (Fig. 5b, c and d).

363 While PIPSO<sub>25</sub> values display a rather gradual decrease in sea ice coverage, the WSI record suggests a highly  
 364 variable sea ice cover with few distinct sea ice minima between 11 ka and 10 ka BP and around 9 ka BP (Fig. 4a  
 365 and b). These sea ice minima may have resulted from punctuated warming events, e.g. at 10 ka BP, when SSST  
 366 shows a short temperature peak, which might have led to a delayed sea ice formation in autumn and winter (Fig.

Need to make it clear whether the cover is sea-ice, ice-shelf or could be either - ice canopy. If ice-shelf is a possibility you should propose viable source region(s)/glacier(s) and include any supporting evidence.

Can the presence of icebergs indicate whether ice canopy is more or less likely ice shelf or sea-ice?

Also Roseby et al., 2022 papers on Anvers Shelf cores with ramped pyrolysis C14 dates.

consider timing and mechanisms of this ocean warming with JRI/WAIS ice core temp records since you include them in Fig 5

WSI is variable but not convinced that 4a is showing a decrease between 11.7 and 8.2 ka. How do you reconcile declining WSI with the increased abundance of the sea ice assemblage shown in fig 3c?

Need to provide some explanation for the different SOT's and if water column structure accounts for difference with SSST's.



367 5b). Another WSI minimum at 9 ka BP coincides with a major (and final) peak in IRD deposition at the core site  
 368 (Fig. 4) evidencing iceberg discharge during episodes of peak ice-sheet loss at the WAP (Jones et al., 2022). As  
 369 sea ice melting may have been an important driver of the ocean stratification, we suggest warmer, stratified surface  
 370 waters with moderate production in summer, supported by increasing insolation in December (Fig. 5a).  
 371 Ameliorating climate conditions, ice-shelf retreat and the establishment of modern-like ocean conditions after 9  
 372 ka BP have also been proposed for the western Bransfield Strait (Heroy et al., 2008) and are well in line with the  
 373 rising contribution of open ocean diatoms and the phytoplankton-derived HBI triene at our core site (Fig. 3). Our  
 374 marine records of decreasing sea ice and rising ocean temperatures are consistent with the overall slight warming  
 375 trend recorded in the WAIS Divide ice core (Fig. 5h). Interestingly, neither this rise in RI-OH' derived SOTs nor  
 376 the highly variable TEX<sub>86</sub><sup>L</sup> temperatures correspond to the declining TEX<sub>86</sub> temperatures reported for ODP site  
 377 1098 at Palmer Deep (Shevenell et al., 2011) or the declining  $\delta D$  values recorded in the JRI ice core (Fig. 5;  
 378 Mulvaney et al., 2012). These regional differences may relate to changing ocean circulation patterns and associated  
 379 shifts in water mass distribution at the WAP and EAP during the Early Holocene. We further note that the partly  
 380 opposing trends in RI-OH' and TEX<sub>86</sub><sup>L</sup> temperatures at our core site could indicate that the respective GDGT  
 381 producing archaea thrive in different water depths or during different seasons.

please clarify whether this is local ice retreat or the WAP ice-loss referred to on line 368.

Open ocean diatoms only increase at the very end of this interval (after ~8.8 ka)

Would be useful to expand on this with any evidence of diverging temperature trends happening in other WAP regions or time intervals.  
 Also/or whether these different trends are present during the WAP warming of recent decades.

Excellent! Would be great to include more discussion along these lines for the other intervals & proxies.

### 383 4.3 Middle Holocene from 8.2 ka until 4.2 ka BP

384 The Middle Holocene from 8.2 ka to 4.2 ka BP is a period of significant sea ice retreat and minimum iceberg flux  
 385 at the core site indicated by decreasing WSI and PIPSO<sub>25</sub> values, virtually absent IRD, and an oceanic warming  
 386 reflected in SSST and RI-OH' SOT (Fig. 4 and 5). For the whole period, diatoms associated with warmer open  
 387 ocean conditions, peak HBI triene concentrations and maximum TOC and biogenic opal contents (Fig. 3) refer to  
 388 a high export production during the Middle Holocene (Abelmann et al., 2006; Smetacek et al., 2004). This can be  
 389 linked to a decrease of both winter and spring sea ice and potentially ice-free summers indicated by WSI and  
 390 PIPSO<sub>25</sub> minima (Fig. 4).  
 391 The retreat of the previously grounded AP ice-sheet between 10 ka and 5 ka BP finally opened the passage for  
 392 ACC surface waters to enter the Bransfield Strait from the west (Bentley et al., 2014; Ó Cofaigh et al., 2014). As  
 393 a result, we suggest that sea ice conditions at our core site were predominantly influenced by branches of the ACC  
 394 (the BSW and CDW) and inflow from the Weddell Sea was diminished due to the still grounded ice sheet at the  
 395 tip of the AP. The influence from the Weddell Sea was weak and opposite sea ice conditions were reconstructed  
 396 for the eastern AP where HBI biomarker and diatom assemblages record extended sea ice cover between 7 ka and  
 397 4.5 ka BP (Fig. 4e, Barbara et al., 2016a; Minzoni et al., 2015).

marked/prominent?

influence/activity?

not clear in Fig 3 that it is the warmer open ocean assemblage shown.

indicate/suggest?

need to clarify whether 'this' refers to the high export production, the warming or one of the proxies you mention.

Would be useful to have greater explanation on the evidence/rationale for 'potentially ice-free summers'.  
 How do the MAT results on WSI concentrations and ice-free summers compare with modern sea-ice seasonality?

Are these 'ACC surface waters' BSW or (modified?) CDW? If so better to state explicitly.

Please make it clear what the influence/impact on sea ice was (eg. longer/shorter sea ice season).

continued?

in/adjacent to/over the Bransfield Strait?

impacted/affected?

incursions of (warmer?) oceanic waters associated with

add reference for this.



398 Regarding ocean temperatures, we observe a sustained warming in RI-OH' SOT punctuated by a cooling at 5.5 ka  
 399 BP, while  $\text{TEX}_{86}^L$  temperatures depict a subtle cooling between 8.2 ka and 7 ka BP followed by a warm reversal  
 400 until 6 ka BP (Fig. 5). A Middle Holocene slight cooling trend has also been observed at core sites at Palmer Deep  
 401 at the WAP (Fig. 1 and 5, Etourneau et al., 2013; Shevenell et al., 2011) and contrasts a rapid warming observed  
 402 in JPC38 from the eastern AP between 8 ka and 6.5 ka BP (Fig. 1 and 5, Barbara et al., 2016; Etourneau et al.,  
 403 2019). Here, the near-coastal marine sediment core close to JRI (Fig. 5g, Barbara et al., 2016) records the transition  
 404 from cold and heavily sea ice covered conditions at 8.2 ka BP to a warmer environment with reduced ice cover  
 405 permitting phytoplankton growth between 6.5 ka and 4.2 ka BP (Barbara et al., 2016). Since stable temperatures  
 406 are inferred from the JRI ice core during the entire Middle Holocene (Mulvaney et al., 2012), we suggest that the  
 407 environmental changes recorded in JPC38 reflect ocean-driven rather than atmospheric processes.

Please add explanation(s) for the temperature patterns recorded at PS97/72-1 and if that reveals/suggests anything about ocean/atmospheric circulation in the NAP and/or WAP.

#### 409 4.4 Late Holocene and Neoglacial from 4.2 ka BP until today

410 The Late Holocene covering the past 4.2 ka BP is characterized by relatively stable environmental conditions at  
 411 our core site reflected in constant biogenic opal and TOC contents, low IPSO<sub>25</sub> and still variable but elevated HBI  
 412 triene concentrations. A gradual decline in PIPSO<sub>25</sub> values between 4.2 ka and 1.5 ka BP contrasts the highly  
 413 variable WSI concentrations (Fig. 4). Minimum PIPSO<sub>25</sub> values at 0.5 ka BP are related to the significantly reduced  
 414 IPSO<sub>25</sub> and HBI triene concentrations. Similar to our observation for the Deglacial, this pattern of low HBI triene  
 415 and minimum IPSO<sub>25</sub> concentrations may point to perennial cold conditions and the establishment of thick and/or  
 416 compacted sea ice limiting the productivity of both phytoplankton and also sea ice diatoms. This short period of  
 417 sea ice growth may be related to the Little Ice Age (LIA), when cooler conditions also triggered glacial readvance  
 418 at the Antarctic Peninsula 500 years ago (e.g. Simms et al., 2021). While a significant pulse in sea ice export from  
 419 the Arctic Ocean is proposed to have caused the LIA cooling (Miles et al., 2020), knowledge about LIA sea ice  
 420 conditions in the Southern Ocean is scarce and hence inconclusive (Parkinson, 1990). The Neoglacial cooling as  
 421 found in the JRI ice core, for example, is clearly reflected in an increased sea ice cover at Palmer Deep since 2 ka  
 422 BP (Fig. 4d, Etourneau et al., 2013) but this record, however, contains no clear evidence for a further expansion  
 423 of sea ice in response to the LIA. We conclude that a comprehensive assessment of how Antarctic sea ice  
 424 conditions changed during the LIA requires more studies of well-dated and ideally higher resolved sediment  
 425 records.

Please add a description of what the 'stable environmental conditions' are likely to be. eg. several months sea ice cover? fully open waters every summer?

You mention the similarities between the last 0.5 ka with the deglacial but don't account for the contradictory evidence in the other proxies. In particular TOC and biogenic opal, respectively: <0.3 & ~10wt% during the deglacial and >0.6 & ~30wt% during the last 0.5 ka. Also diatom conc. <25 mv/g during deglaciation but 50-120 mv/g during last 1ka.

What about other Bransfield Strait and NAP marine records? eg. Barbara et al., 2013; Barcena et al., 2006; Kyrmanidou et al., 2018; Milliken et al., 2009

426 Similar to our sea ice signals, also the Late Holocene ocean temperature reconstructions display different patterns.  
 427 While RI-OH' SOT remains relatively stable,  $\text{TEX}_{86}^L$  indicates a cooling around 3 ka BP followed by a warming  
 428 until 1.5 ka BP and another cooling towards the top of the core. Diatom-derived SSST also point to a cold period





429 around 3 ka BP immediately followed by a warming peak at ca. 2.5 ka BP. Evidently, temperature trends at the  
 430 WAP in the Late Holocene are highly variable between different areas (Allen et al., 2010; Barbara et al., 2016;  
 431 Bárcena et al., 1998; Bentley et al., 2009; Etourneau et al., 2013; Mulvaney et al., 2012; Shevenell et al., 2011)  
 432 and this may relate to the complex oceanographic and atmospheric settings. With regard to the diverging  
 433 temperature trends observed in sediment core PS97/72-1, we not that also inconsistencies between different  
 434 analytical approaches to reconstruct ocean temperatures need to be acknowledged and examined. As previously  
 435 stated, more information on the applicability and significance of GDGT-derived ocean temperatures in polar  
 436 latitudes is needed.

I realise this would probably be an entire study but it would be worthwhile to suggest some of the potential causes/explanations that may be relevant.

## 437 5 Conclusions

438 We reconstructed the sea ice and climate development at the northwest AP since the last Deglacial using the  
 439 sediment core PS97/072-1 from the Bransfield Strait. In our multiproxy study we focused on the sea ice biomarker  
 440 IPSO<sub>25</sub>, the HBI z-triene representing open marine environments, and GDGTs for ocean temperature  
 441 reconstructions. Diatom ecological groups characteristic of sea ice or open ocean conditions were used, as well as  
 442 diatom transfer functions to reconstruct winter sea ice and summer sea surface temperature. Additional information  
 443 was derived from sedimentological records such as IRD. Our results reveal the retreat of a floating ice shelf canopy  
 444 after the ACR and an overall sea ice retreat and ocean warming during the Holocene. The late Deglacial from 13.9  
 445 ka to 11.7 ka BP was a highly dynamic period: until 13.4 ka BP the sedimentation of organic proxies was  
 446 diminished due to a permanent ice cover during the ACR. The ACR terminated with a shift to warm conditions at  
 447 13 ka BP along with a retreat in spring sea ice. The Early Holocene from 11.7 ka to 8.2 ka BP was characterized  
 448 by warming, slightly decreasing spring sea ice cover and highly variable winter sea ice cover. In the Middle  
 449 Holocene from 8.2 ka to 4.2 ka BP, stable environmental conditions prevailed with elevated primary production  
 450 due to intervals of lower sea ice cover. In general, sea ice seasons were short and sea ice cover was significantly  
 451 reduced to a minimum around 5.5 ka BP, even though high seasonal amplitudes and short-term, centennial changes  
 452 in sea ice conditions occurred. During the Late Holocene, the core site experienced a variable WSI and a short-  
 453 term cooling at 3 ka BP. Phytoplankton biomarkers as well as sea ice proxies (IPSO<sub>25</sub>, PIPSO<sub>25</sub>, WSI) were lowest  
 454 during the period coincident with the Little Ice Age which we relate to the establishment of a multi-year sea ice  
 455 cover.

In the conclusions, too many of the climatic changes are described only in relative terms with very little context to calibrate the descriptions. eg. After the ACR, there is a 'retreat in spring sea ice cover', then during the early Holocene there is a further 'slight decrease in spring sea ice cover' then 'intervals of lower sea ice cover' and 'short sea ice seasons' until a 'minimum at 5.5 ka' with 'high seasonal amplitudes' and 'short-term changes in sea ice cover'. Please add information on WSI % or approximate weeks/months of sea ice cover or similar. If this isn't possible then quantify with the units from the PIPSO ratio or similar to convey a measure of the changes. Same for temperature and productivity.

warmer? cold? or both?

This should be either ice canopy or ice-shelf.

how does this relate to the post-ACR retreat of floating ice canopy/ice shelf (line 443)?

'intervals of lower sea ice cover' and 'high seasonal amplitude and short term changes in sea ice conditions' seem to contradict the description of the mid-Holocene 'stable environmental conditions'.

Does this mean that conditions during the LIA were similar to the the ACR? If so, how does that reconcile with high TOC and Biogenic opal?





457 **Data Availability**

458 All data mentioned in this paper will be available at the open access repository [www.pangaea.de](http://www.pangaea.de)  
459 (<https://doi.pangaea.de/10.1594/XXXXXXX>).

460

461 **Author contributions**

462 The study was conceived by MV and JM. Data collections and experimental investigations were done by MV  
463 together with LLJ (foraminifera, age model), SMS (age model, humming age), CBL (core description, sampling,  
464 diatoms, biogenic opal, age model), PC (diatoms), AL (age model, diatoms), OE (diatom transfer function), GM  
465 (GDGTs PS97/072-1,  $^{14}\text{C}$  dating), AVH ( $\delta^{13}\text{C}$  IPSO<sub>25</sub>), NL ( $\delta^{13}\text{C}$  TOC), Je, DE and CE provided temperature and  
466 salinity profiles near the study site. MV drafted the manuscript. All authors contributed to the interpretation and  
467 discussion of the data and the finalization of this manuscript.

468

469 **Competing interests**

470 None of the authors have a conflict of interest.

471

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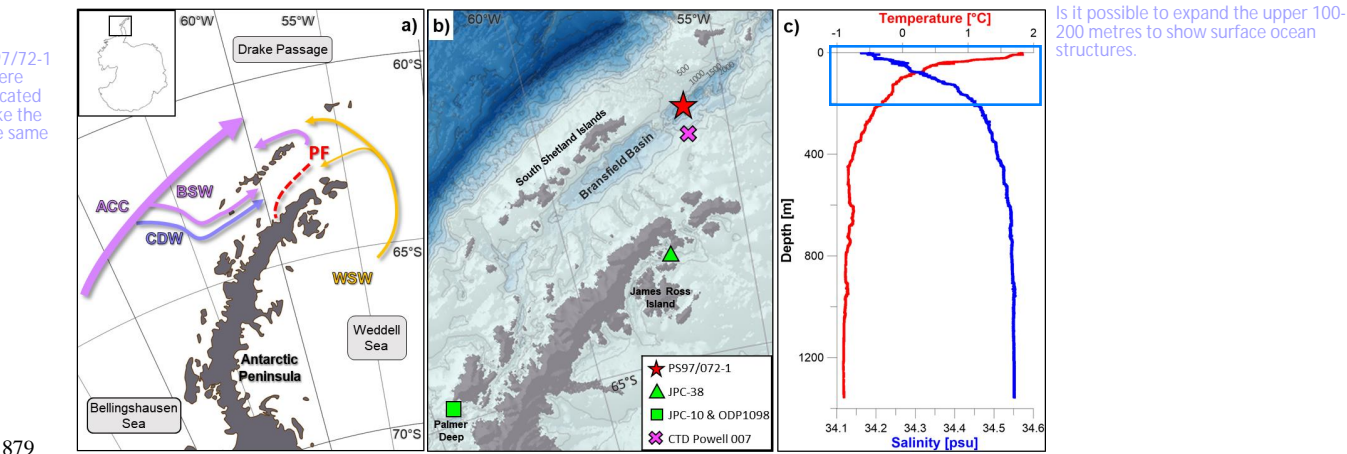
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878 **Figures** PF typically used as shortening for Polar Front. Suggest you change to Pen. F. or other term to avoid confusion.

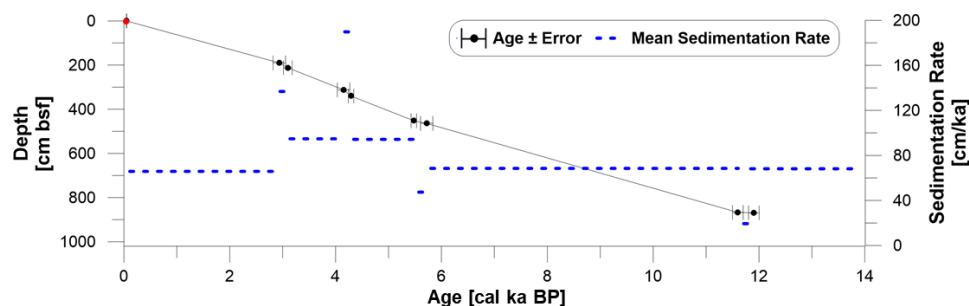
Please add the location of PS97/72-1 to the map 'a' so it is clear where oceanographic features are located in relation to core site OR make the map projections and areas the same in maps a & b.



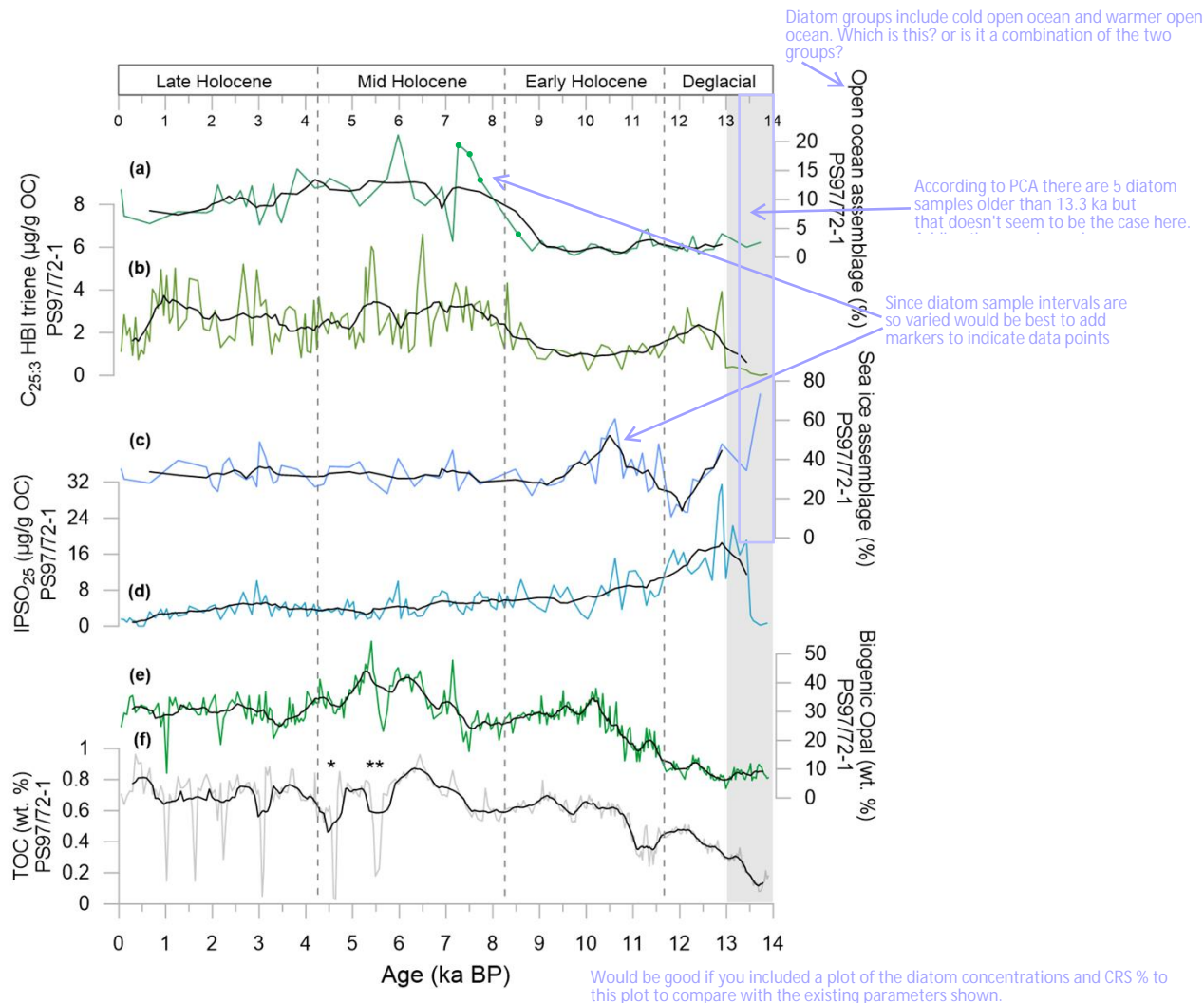
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880 **Figure 1:** a) Overview map with modern oceanography in the study area (Hofmann et al., 1996; Sangrà et al., 2011). ACC  
881 = Antarctic Circumpolar Current, BSW = Bellingshausen Sea Water, CDW = Circumpolar Deep Water, WSW =  
882 Weddell Sea Water, and PF = Peninsula Front. b) Bathymetric features in the Bransfield Strait with the location of  
883 sediment core PS97/072-1 (red star) and other sediment records discussed in the text (green), and the CTD station  
884 (purple cross) where c) the vertical profile of ocean temperature and salinity (cruise POWELL2020, CTD 007  
885 (62°09.075'S, 56°37.09'W) from 27.01.2020) shows a clear stratification of the upper 100 m of the water column. It  
886 indicates that surface waters are dominated by the BSW, while the basin is filled with WSW water. Maps were done  
887 with QGIS 3.0 (QGIS, 2018) and the bathymetry was taken from GEBCO\_14 from 2015.

See comment above.

Is it possible to expand the upper 100-200 metres to show surface ocean structures.



888  
 889 **Figure 2: Age-depth model for sediment core PS97/72-1 based on eight  $^{14}\text{C}$  dated calcite samples (black)**  
 890 **with error bars and mean sedimentation rates (cm/ka, dashed blue line). The core top age (red) was**  
 891 **estimated as 0.05 ka BP from matching with the  $^{210}\text{Pb}$ -dated multicore PS97/072-2 (Vorrath et al., 2019; see**  
 892 **supplement section 2).**  
 893  
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895

896 **Figure 3: Overview of organic geochemical parameters and main diatom assemblages determined in**  
897 **sediment core PS97/72-1 used to characterize the environmental setting over the past 14 ka BP. a) open**  
898 **ocean diatom assemblage, b) C<sub>25:3</sub> HBI triene, c) sea ice diatom assemblage, d) IPSO<sub>25</sub>, e) biogenic opal and**  
899 **f) TOC contents. Asterisks in f) mark layers of volcanic ash, where \*\* can be linked to a tephra layer in a**  
900 **sediment core from the Bransfield Strait at 5.5 ka BP (Heroy et al., 2008). Grey shaded interval refers to the**

Write in full?

901 **ACR.**

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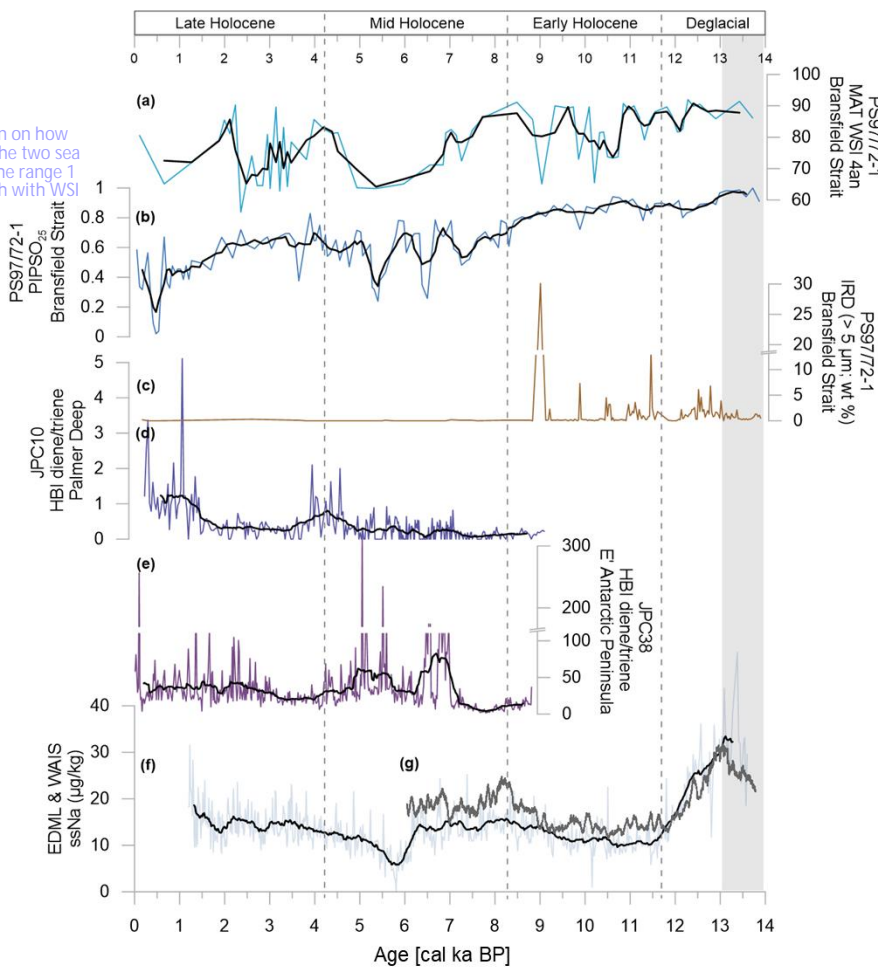
903



Please mark (line?) the modern concentration of winter sea ice for the site on plot a. Same with SOT and SST on plots b-d in Fig 5.

Please add comment/discussion on how the fluctuations and trends in the two sea ice proxies compare. ie. Does the range 1 to 0.4 on the PIPSO index match with WSI conc. of 90-60%?

Are there any MAT samples that either don't have 4 analogues or match only with analogues of a very different oceanographic context? If so it would be good if you marked these with a different symbol to the ones with full analogue matches from similar oceanographic setting(s).

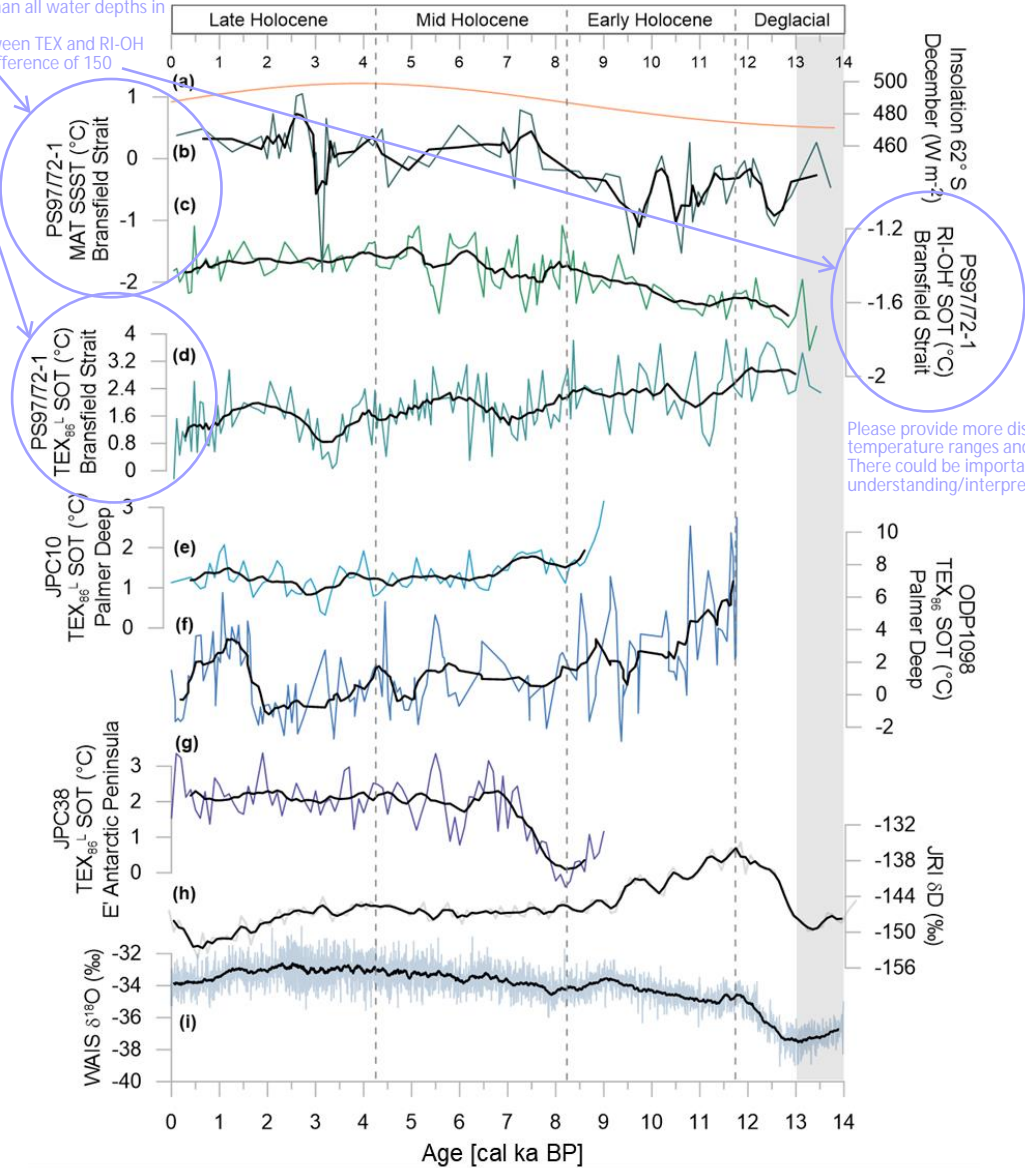


904  
905 **Figure 4: Sea ice related proxies in sediment core PS97/72-1 with a) the diatom based WSI, b) the sea ice**  
906 **index PIPSO<sub>25</sub>, and c) ice rafted debris (IRD). For comparison: the HBI diene/triene ratio of sediment core**  
907 **d) JPC10 from the Palmer Deep station (Etourneau et al., 2013) and e) JPC38 at the East Antarctic Peninsula**  
908 **(Barbara et al., 2016). ssNa records of f) the EDML ice core (Fischer et al., 2007) and g) the WAIS ice core**  
909 **(WAIS Divide Project Members, 2015). Grey shaded interval refers to the ACR.**



These temperature ranges do not match particularly well with the CTD profile in Fig 1c.  
> SSST here are about 2 deg lower than in the Jan CTD.  
> SOT from the RI-OH are colder than all water depths in the CTD  
> the temperature difference between TEX and RI-OH SOT's (>1.5 deg) suggests depth difference of 150 metres or more.

How do you reconcile the temperature range (2-3 deg) of TEX samples results with i) CTD profile, ii) the other temp records and iii) possible circulation/water mass changes?



Please provide more discussion/explanation on the different temperature ranges and trends in the two SOT proxies. There could be important clues to improve our understanding/interpretation of these proxies.

912  
913 **Figure 5: A comparison of a) December insolation (Laskar et al., 2004), b) diatom-based SSST, c) RI-OH'-**  
914 **derived SOT, d) TEX<sub>86</sub><sup>L</sup>-SOT of sediment core PS97/72-1, and temperature reconstructions e) TEX<sub>86</sub><sup>L</sup> from**  
915 **JPC10, Palmer Deep, d) TEX<sub>86</sub> from ODP1098, Palmer Deep, e) TEX<sub>86</sub><sup>L</sup> from JPC38, East Antarctic**  
916 **Peninsula, and ice core stable isotope records of h) JRI (Mulvaney et al., 2012) and i) WAIS Divide (WAIS**  
917 **Divide Project Members, 2013).**  
918