

Authors' response to Dan Lunt (Referee #1)

Interactive comment on “Early Eocene vigorous ocean overturning and its contribution to a warm Southern Ocean” by Yurui Zhang et al.

Dan Lunt (Referee)

I reviewed this paper for another journal. The CP editor informed me that this paper is unchanged from that version, so I am attaching here my original review (as such the line numbers refer to the previous version).

Please also note the supplement to this comment:<https://www.clim-past-discuss.net/cp-2019-163/cp-2019-163-RC1-supplement.pdf>

We are grateful to the reviewer for his insightful comments. We apologize about it but there has been some miscommunication between the editor, the reviewers and ourselves. Indeed, the submitted version of the paper to *Climate of the Past* was modified from the version the reviewer has previously reviewed, and most comments and suggestions were taken into account. Below we provide answers to the questions raised by the reviewer, i.e. we explained how and where these have been incorporated in the manuscript and clarified some points (we denoted the replies by blue color). The line numbers (in brackets) correspond to this revised version of the paper. These modified places are also marked by color in the annotated copy of manuscript.

In this paper, Zhang et al present results from IPSL-CM5A2 configured for the Eocene, and compare them with results of the modern. The focus is on ocean circulation, including regions of deep water formation, and the partitioning of heat transports into separate terms. Finally, sensitivity results to modified CO₂ are presented. The paper is very well written and clear in general, and in my opinion it is very appropriate for *Climate Dynamics*, with minor-moderate revisions.

General Comments

Line 100-127. The model description and experimental design needs considerably more detail. Given that Sepulchre et al is just “in prep”, we need many more details of the difference between IPSL-CM5A2 and IPSL-CM5A. For the experimental design, we need to know how soils, vegetation, sub-grid scale topography, and what frame of reference was used for the Herold reconstruction. The “bar” here is that the simulations should be approximately repeatable. The Lunt et al experimental design paper gives several options for many of the boundary conditions, so we need to know what choices have been made here.

This is a fair critic. First of all, since the initial submission, the paper by Sepulchre et al is now under review, and the preprint version is available from <https://www.geosci-model-dev-discuss.net/gmd-2019-332/>

Hence, we do not want to repeat too much of the material presented there. However, we have expanded largely the description of the model and of the different simulations, as our section 2.1 has been fully re-written. [line 90-166]

The model-data comparison needs some more work. In particular, the annual mean SSTs should be compared with the data in the main paper, and the seasonal in Supp Info, rather than the other way round.

We have moved the annual mean SST in the main text (Fig. 2A) and the summer SST in Supplementary information (Fig. S2A)

More explanation is needed of what the “uncertainty” in the proxies (Table 2) represents.

In response to this comment, we have added the uncertainty in the caption of Table 2, which is defined as follows: “The uncertainty range is defined as the 2σ deviations (2.2% and 97.8%) for $\delta^{18}\text{O}$, and as the range between 5% and 95% percentile SST estimates for TEX^{86} , Mg/Ca and clumped isotope data.”

Also, it is claimed that the model-data agreement is “overall consistent”, but this should be quantified more. For example, what RMS score would be obtained if it was assumed that the Eocene warmed uniformly, or with a $\cos(\text{latitude})$ response. This can give some quantification to the question “how good is good”. Also, the choice of $3x \text{CO}_2$ is rather arbitrary in the context of model-data comparison, so the GCM warming patterns could be scaled uniformly to best fit the data, and the RMS recalculated.

We are now presenting a more quantitative evaluation. First, we have justified better the fact that we mostly focus on the 55Ma-3x simulation, given that the agreement with the proxy-based SST reconstruction is better for this simulation than the other one (55Ma-1.5x simulation; see Table 2). Then, we have mostly followed the methodology proposed by Kennedy-Asser et al. (2019), that suggest 2 benchmarks to conduct such an evaluation. Benchmark 1 assumes a uniform mean temperature, and benchmark 2 is based on the assumption of the least squares linear fit through the proxy-based SST estimates with the cosine function of paleo-latitude from each-type-of-proxy sites. The results show the performance of the model simulation both RMSD metrics outperforms benchmark 1, but not benchmark 2 for some proxies. This means that the simulation outperforms the constant mean benchmark but not fully the latitudinal gradient benchmark, and thus it corresponds to the ‘moderate good performance’ following the wording used by Kennedy-Asser et al. (2019).

This is included in the text (lines 182-189), and Table 2, and in supplement materials (Lines 2-40, table S1 and figure 2B).

Specific Comments

Abstract - please mention the results of the CO_2 sensitivity here. Otherwise “different levels of atmospheric CO_2 ” on line 16 is hard to understand when reading for the first time.

The sentence ‘Simulations with different atmospheric CO_2 levels show that the ocean circulation and heat transport are relatively insensitive to CO_2 -doubling’ has been added at the end of the abstract. [line 17-18]

Abstract – please add something about the model-data comparison that you have carried out.

We have added the sentence of “When compared with proxy-based reconstructions, the simulations reasonably capture both the reconstructed amplitude and pattern of early Eocene sea surface temperature.” in the abstract. [line 4-5]

Line 19: Explain where (depth and latitude) that the 40Sv occurs.

The 40Sv MOC occurs at latitude of 60°S, as mentioned in the text. [line 8]

Line 32: Check whether 55 Ma is really the time period that DeepMIP focuses on; see e.g. Hollis et al (2019) and Lunt et al (2017).

We have modified this to 55Ma-50Ma, referring to the early-Eocene climate optimum (EECO) of DeepMIP framework, and added a sentence to clarify this. [Line 21, lines 24-26]

Line 40: I don't think it's correct that CO₂ can't explain the decreased meridional gradient at the EECO. The high CO₂ can lead to enhanced feedbacks at high latitudes.

The sentence has been changed to: ‘In the early Eocene, high levels of CO₂ in the atmosphere are undoubtedly a critical contributor to the extremely warm climate, but they do not fully explain the extreme warmth at high-latitudes and the reduced equator-to-pole temperature gradient.’ [line 29-32]

Figure S11: For the time series of temperature in the simulations, the evolution seems inconsistent with the statement that the 1.5x run is branched off from the 3x run after 1500 years. Please align the simulations correctly so that the 1.5x run starts after 1500 years.

The figure has been modified, as indeed, “the 1.5x run is branched off from the 3x run after 1500 years”.

As well as the time series of temperature (Figure S11), it is important to show the time series of some metric of overturning, e.g. maximum overturning, or averaged mixed-layer depth, so we can assess to what extent the ocean circulation is in equilibrium, and the inter-annual variability of circulation.

We have included in Figure S1 the time series of MOC and temperature (both the SST and volume averaged temperature).

Figure 4: maybe use a log scale rather than using different scales for each panel, or use the same scale for each panel.

We had tried the log scale and the same scale for all panels by following this comment, but the figure becomes messy. We would like to keep it as it is to improve readability.

Line 223: Not clear what “intermittent” means here – it implies temporal variability.

We mean deepwater is formed during some winters. The sentence has been modified to improve clarity. [Line 255]

Line 234: I would have expected the vertical gradient in density, rather than surface density, to be the key control.

The density at depth does not change much, meaning that the change in vertical density gradient is largely determined by the surface density, which can be seen as a proxy. The text has been modified to explain this better. [line 263-265]

Line 237: For the salinity and temperature values, please also give the percentage change that this induces in density (e.g. 80% and 20%)

Change in salinity induces a density change of 0.57 kg/m^3 that is partly balanced by temperature-induced change of 0.06 kg/m^3 . This is clarified in the text. [lines 266-268]

Line 243-254: I am not totally convinced by this mechanistic link to atmospheric circulation. Please either add some more quantitative analysis, or caveat this section with “possible” or “maybe”.

This section has been fully revised to take into account this comment. [lines 274-285]

Line 294: This additional simulation should be introduced in the methods section.

This additional simulation (55Ma-3x-noM2) is now presented in the method section. [line 157-158]

Line 335-336: Please reference/add a Figure for “visible on sea surface salinity”

Sea surface salinity is partly decided by the freshwater budget (P-E+R) (Table 3) with the atmosphere. This budget is largely affected by the precipitation (Fig S3) when other processes are similar. So we showed the driver of sea surface salinity, precipitation, as an indicator, and added references on this. [lines 366]

Line 356-366: This section may link more clearly to ocean circulation if you explore the wind stress curl rather than the wind stress.

We had calculated the wind stress curl, which however is not as clear as the wind stress. So we would like to keep the wind stress for clarity. [lines 385-394]

Line 498-499: This non-linearity in climate sensitivity is interesting and could be explored in a bit more detail. See e.g. discussion of why this may be in Farnsworth et al, in press, GRL.

The larger temperature response of the 55Ma simulation to CO₂ doubling is consistent with the results of Farnsworth et al. (2019), that we now reference explicitly in the text. [lines 519-520]

Line 552: It is not clear what you mean by “robust”.

We meant that the response is stable. Robust has been changed to stable. [line 565]

Line 560-570: Make clearer at the outset of this paragraph that it is a summary of the previous paradigm, not a summary of your results.

We have added “Numerous proxy based reconstructions have revealed that” in the first sentence to clarify the summary of previous studies, on which we started from. [lines 564-565]

Technical Comments

Line 37: “Ma ago” should be “Ma”.

Fixed. [line 28]

Line 44: “continental configuration” as well as “bathymetry”

Done [line 34]

Line 68: “latitude” should be “latitudes”

Done [line 56]

Line 82: remove “briefly”

Done [line 6-69]

Line 89: remove “IPSL in the following” and stick with the full name.

Done [line 75]

Equation 1: need to define E, W, x, z.

Done

Line 172: “clockwise” needs definition of which way we are “facing”.

The meridional transport was integrated from the west to east across the basin, implying we are facing the west. [line 208-209]

Line 221: “deep convection” instead of “convection”.

Done [line 253]

Line 242: net surface freshwater gain.

Fixed. [line 274]

Line 447: label the two terms in the equation, e.g. with curly brackets, OHT_{moc} and OHT_{gyre}.

We have labeled these two terms in the revised version. [line 471]

Line 530: HadCM3L.

Done [line 537]

Line 560: Warmest period in the Cenozoic.

Done [line 564-565]

Figure 2: Add labels for different lines.

Done

Figure 3: Add units for isopycnal contours in caption.

Done

Authors' response to Referee #2

Interactive comment on “Early Eocene vigorous ocean overturning and its contribution to a warm Southern Ocean” by Yurui Zhang et al.

Anonymous Referee #2

I also reviewed the paper for another journal and the line numbers below refer to that version.

We are grateful to the reviewer for these insightful comments. As explained to the other reviewer the submitted version of the paper to *Climate of the Past* was modified from the version the reviewer has previously reviewed, and most comments and suggestions were already taken into account. Below we provide one-by-one replies to answer the questions raised by the reviewer (we denoted the replies by blue color) and to explain how and where these have been incorporated in the manuscript, and we done and/or to clarify some points. The line numbers (in brackets) correspond to the resubmitted version of the paper.

The study by Zhang et al. presents new model simulations of the early Eocene based on paleo-geographic reconstructions from the DeepMIP initiative, using the AOGCM IPSL-CM5A2. A comparison to simulations with modern conditions is used to investigate the role of ocean circulation changes and meridional heat transport for high latitude warmth in the Southern Hemisphere. The authors find a strong abyssal overturning circulation in the Southern Hemisphere (SOMOC) that leads to enhanced poleward heat transport that maintains warm Southern Hemisphere high latitudes. Additionally, in contrast to earlier studies no deep water is formed in the Northern Hemisphere in the early Eocene simulations. These are novel points worth to publish and the authors provide an elaborate analysis of factors (e.g. tidal mixing scheme and CO₂ changes) controlling the vigorous SOMOC in their early Eocene simulations. Nevertheless, the current manuscript could be improved by testing their findings in the context of North Pacific deep-water formation as this might be the most important modulator of the presented results. In that sense, the absence of North Pacific deep-water formation might be inherently linked to the basic question ‘What does explain such a strong SOMOC?’ (line 275).

We have tried to elaborate a bit more in the revised version on why deep water formation occurs in the Southern Ocean but not in the Pacific, as explained in the following. [line 257-285]

Comments & Suggestions:

- To address this point, it might be helpful to include an additional experiment with a preindustrial CO₂ concentration in the atmosphere. If this scenario is unlikely to give North Pacific deep-water formation, an artificial modification of the continental run-off distribution in the Pacific might help. This is a good suggestion and indeed such an additional simulation (Eocene bathymetry and PI CO₂) would be useful. Yet, given the computational cost of running any of these simulations, it was not possible to run a new one for now (but we might in the future). However, the analysis of the Eocene 1.5-x simulation (which is not that different from what the reviewer is suggesting), and the comparison with the Eocene 1.5-x simulation suggests that the circulation is largely insensitive to the level of CO₂ during the Eocene (at least within this range), which makes us confident in the results presented in the manuscript.
- Line 234: In the Weddell Sea surface density is much larger than in the North Pacific (+0.51 kg/m³). Is it trivial from this value that North Pacific deep-water formation is absent? Do you expect a critical value?

What actually matters is the local change in stratification, of which the surface density is a good proxy in our case because the deep water density is much more similar in the two regions and between the different simulations (we have clarified this point in the revised version, see Lines 261-266). So, although it would be nice, we do not think there is such a thing as a threshold that allows (or not) for deep water formation in the Pacific. That said, such a statement should be checked more carefully, and could be a basis of a model intercomparison study based on the DeepMIP archive.

- It is recommended to include the CO₂ sensitivity into the section '3.3 Factors contributing to the vigorous SOMOC'

We respectfully disagree with the reviewer on that point. We feel that the storyline of the paper would get messy if we were to include the CO₂ sensitivity results earlier on, and we would lose the coherency that we believe section 6 has as it is.

- Please revise Table 2: the mean values and the relation to the uncertainty ranges is unclear, use units within the table for clarity, is the uncertainty range one or two sigma?

Uncertainty range had been added in Table 2. It read as "The uncertainty range is defined as the 2 σ deviations (2.2% and 97.8%) for $\delta^{18}\text{O}$, and as the range between 5% and 95% percentile SST estimates for TEX⁸⁶, Mg/Ca and clumped isotope data."

The units are now indicated in the table.

Summary: Although the study is already at a reasonable level, the authors are encouraged to address the potential influence of North Pacific deep water on the presented findings. I would be happy to have a look at a revised manuscript version.

Early Eocene vigorous ocean overturning and its contribution to a warm Southern Ocean

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1 **Abstract.** The early Eocene (~55 Ma) is the warmest period of the Cenozoic, and is most likely characterized by
2 extremely high atmospheric CO₂ concentrations. Here, we analyze simulations of the early Eocene performed with
3 the IPSL-CM5A2 earth system model, set up with paleogeographic reconstructions of this period from the DeepMIP
4 project and with different levels of atmospheric CO₂. When compared with proxy-based reconstructions, the
5 simulations reasonably capture both the reconstructed amplitude and pattern of early Eocene sea surface temperature.
6 A comparison with simulations of the modern conditions allows us to explore the changes in ocean circulation and
7 the resulting ocean meridional heat transport. At a CO₂ level of 840 ppm, the Early Eocene simulation is characterized
8 by a strong abyssal overturning circulation in the Southern Hemisphere (40 Sv at 60°S), fed by deep water formation
9 in the three sectors of the Southern Ocean. Deep convection in the Southern Ocean is favored by the closed Drake and
10 Tasmanian passages, which provide western boundaries for the build-up of strong subpolar gyres in the Weddell and
11 Ross seas, in the middle of which convection develops. The strong overturning circulation, associated with subpolar
12 gyres, sustains the poleward advection of saline subtropical water to the convective regions in the Southern Ocean,
13 thereby maintaining deep-water formation. This salt-advection feedback mechanism is akin to that responsible for the
14 present-day North Atlantic overturning circulation. The strong abyssal overturning circulation in the 55 Ma
15 simulations primarily results in an enhanced poleward ocean heat transport by 0.3–0.7 PW in the Southern Hemisphere
16 compared to modern conditions, reaching 1.7 PW southward at 20°S, and contributes to maintaining the Southern
17 Ocean and Antarctica warm in the Eocene. Simulations with different atmospheric CO₂ levels show that the ocean
18 circulation and heat transport are relatively insensitive to CO₂-doubling.

19 **Keywords:** Early Eocene, overturning circulation, deep water formation, oceanic heat transport.

20 1. Introduction

21 Proxy-based temperature reconstructions suggest that the early Eocene (55–50 Ma) was one of the warmest intervals
22 in the geological history and the warmest of the Cenozoic (Zachos et al., 2001; Cramer et al., 2011; Dunkley Jones et
23 al., 2013). More specifically, the EECO (Early Eocene Climatic Optimum) covers the interval between 53–51 Ma, but
24 shorter (less than tens of thousands of years) hyperthermal events, such as the PETM (Paleocene-Eocene Thermal
25 Maximum) about 55 Ma ago (Zachos et al., 2008), also spanned the early Eocene. The Southern Hemisphere was
26 particularly warm at that time, as shown by inferred surface ocean temperatures exceeding 20°C at high-latitudes (e.g.
27 Evans et al., 2018 and references therein), and by the absence of a perennial ice sheet over Antarctica until the Eocene-
28 Oligocene Boundary (~34 Ma when CO₂ abruptly declined below a certain threshold (Galeotti et al., 2016; Gasson
29 et al., 2014; Ladant et al., 2014). In the early Eocene, high levels of CO₂ in the atmosphere are undoubtedly a critical
30 contributor to the extremely warm climate, with the global temperature increasing by more than 5°C in less than
31 10 000 years (Zachos et al., 2001, 2008; Huber and Caballero, 2011; Anagnostou et al., 2016), but they do not fully
32 explain the extreme warmth at high-latitudes and the reduced equator-to-pole temperature gradient (Huber and
33 Caballero, 2011).

34 In addition to this much higher levels of CO₂ in the atmosphere, one of the main differences between the early Eocene
35 and our modern climate lies in the distinct bathymetric and continental configurations, likely resulting in very different
36 ocean circulation (Thomas et al., 2003; Voigt et al., 2013; Winguth et al., 2012; Zachos et al., 2001). In particular, the
37 opening or closing of major oceanic gateways (such as the Drake Passage or the Panama Seaway) during the Late
38 Paleogene and Neogene have been shown to exert a strong influence on the ocean circulation and its associated heat
39 transport (England et al., 2017; Ladant et al., 2018; Nong et al., 2000; Sijp and England, 2004; Toggweiler and
40 Bjornsson, 2000; Yang et al., 2014). Additionally, proxy-based reconstructions and results from Eocene model
41 simulations suggest that the Meridional Overturning Circulation (MOC) was also very distinct from our present day
42 MOC, with no evidence for deep water formation in the North Atlantic until the early Oligocene (Ferreira et al., 2018).
43 Instead, formation of deep water is suggested to happen in the North Pacific (Hutchinson et al., 2018; Winguth et al.,
44 2012) or only in the Southern Ocean (Sijp et al., 2014). Different ocean circulations resulting from different
45 bathymetries are expected to result in different Ocean Heat Transport (OHT). For instance, Sijp and England (2004)
46 found a 0.5 PW decrease in OHT in the Southern Hemisphere in response to the opening of the Drake Passage, but
47 other factors such as CO₂-induced radiative may also contribute to altering OHT (see Huber, 2012 for a review).

48 In our present-day climate, the ocean is an important actor of the Earth energy balance, as it contributes about one
49 third to the total redistribution of heat from the Equator to the Poles (e.g. Trenberth and Caron, 2001). Although
50 modifications of both the atmosphere and the ocean heat transport (OHT) tend to compensate (Trenberth and Caron,
51 2001), subtle changes in OHT could trigger large changes in atmospheric extratropical convection, modifying water
52 vapor greenhouse (Rose and Ferreira, 2013) and thereby affecting surface temperatures. The OHT itself results from
53 various contributions, and several attempts have been made in the literature to disentangle the relative roles of the
54 horizontal and overturning ocean circulations in the meridional OHT (Ganachaud and Wunsch, 2003). In the North
55 Atlantic, the strong Atlantic Meridional Overturning Circulation (AMOC) is fed by the formation of dense water by

56 convection **at high latitudes**, and the AMOC contributes up to 90% of the meridional OHT at 26.5°N (Msadek et al.,
57 2013), where the RAPID monitoring array is located (McCarthy et al., 2015). Based on hosing experiments performed
58 with a climate model (FAOM), Yang et al. (2013) found that the meridional OHT decreases rapidly in response to an
59 artificial shutdown of the AMOC, although Drijfhout and Hazeleger (2006) suggested that on longer (decadal)
60 timescale, the OHT might recover its initial level as the gyre contribution tends to compensate the decrease in OHT
61 associated with the AMOC shutdown. In contrast to the North Atlantic, Volkov et al. (2010) found that, in the Southern
62 Ocean, the OHT results roughly equally from the gyre and overturning contributions.

63 Given the importance of both horizontal and overturning ocean circulations for the meridional OHT, the different
64 MOC and horizontal gyres constrained by the Eocene bathymetry may result in different contributions to the
65 meridional OHT and may potentially contribute to the warm early Eocene climate, in particular in the much warmer
66 high-latitudes. Based on the analysis of early Eocene and present-day simulations performed with the **IPSL-CM5A2**
67 earth system model, the goal of this study is to better understand what sets the Eocene ocean circulation and the
68 importance of the ocean circulation for the poleward heat transport. The model and simulations used are described in
69 Section 2. In Sections 3 and 4, we examine the MOC and the horizontal circulation, respectively, for the present-day
70 and 55 Ma configurations, and these circulations are then linked to the OHT in Section 5. The sensitivity of the ocean
71 circulation and heat transport to the level of CO₂ in the atmosphere is discussed in section 6. A summary and
72 conclusions are given in Section 7.

73 **2. Numerical model data**

74 **2.1. Model setup and simulations**

75 The simulations used in this study are performed **with the IPSL-CM5A2 earth system model (Sepulchre et al., 2019).**
76 **The oceanic component of IPSL-CM5A2 is NEMOv3.6 (Madec and the NEMO team, 2016), which includes the LIM2**
77 **sea ice model (Fichefet and Maqueda, 1997) and the PISCES biogeochemical model (Aumont et al., 2015). The**
78 **atmospheric component is the LMDz model (Hourdin et al., 2013), which is coupled to the land surface model**
79 **ORCHIDEE (Krinner et al., 2005). Here, we use IPSL-CM5A2 in its standard resolution. NEMO is thus run at a**
80 **nominal resolution of 2°, increased down to 0.5° at the equator, with 31 levels that vary in thickness with depth. LMDZ**
81 **is run at a horizontal resolution of 3.75° longitude×1.875° latitude and 39 vertical levels. A full description and**
82 **evaluation of the IPSL-CM5A2 model can be found in Sepulchre et al. (2019).**

83 **Two sets of simulations are performed. The first set is composed of the reference preindustrial (PI) simulation (referred**
84 **to as PI-1x and described in Sepulchre et al., 2019) and another PI simulation in which the atmospheric CO₂**
85 **concentration is doubled (PI-2x). The second set consists in a baseline simulation of the Early Eocene (hereafter 55**
86 **Ma-3x), using a setup following the DeepMIP protocol described by Lunt et al. (2017), and two sensitivity experiments**
87 **to CO₂ (55 Ma-1.5x) and tidal mixing (55 Ma-noM2). In the following, we first briefly describe the baseline Early**
88 **Eocene simulation (as the DeepMIP guidelines from Lunt et al. (2017) give several options for implementing the Early**
89 **Eocene boundary conditions), and then precise the boundary condition differences of the two other simulations.**

90 2.1.1. General considerations

91 Most of the boundary conditions that were adapted for these IPSL-CM5A2 Early Eocene simulations are described in
92 Herold et al. (2014, hereafter H14). Following Lunt et al. (2017), the solar constant and orbital parameters are kept
93 to PI values, so are greenhouse gas concentrations with the exception of atmospheric CO₂. The latter is set to 840 ppm
94 (3x the PI value), so that the simulation is representative of the pre-PETM (following the terminology of Lunt et al.,
95 2017).

96 2.1.2. Oceanic boundary conditions

97 The ocean component (NEMO) is commonly run on a tripolar mesh grid (Madec and the NEMO team, 2016), which
98 avoids singularity points in the ocean domain. Because the implementation of the Eocene land-sea mask on the
99 ORCA2 grid would have shifted the singularity points into the ocean domain, we have constructed a new PALEORCA
100 grid, which is designed to run paleo-simulations with IPSL-CM5A2 (see more details in Sepulchre et al., 2019). The
101 bathymetry is obtained by masking out the H14 topography and remapping the resulting bathymetry onto the
102 PALEORCA grid using near-neighbor interpolation. This type of interpolation indeed ensures that small but crucial
103 features of the H14 dataset such as islands and seaways, which may strongly impact the modeled ocean circulation,
104 remain present in the interpolated bathymetry file. Handmade corrections have then been applied at some locations
105 (e.g. in the West African region) to retain oceanic straits that are sufficiently large to allow for exchanges. The Early
106 Eocene bathymetry is shown on Fig. 1.

107 Modern boundary conditions of NEMO also include forcings of the dissipation associated with internal wave energy
108 from the M2 and K1 tidal components (de Lavergne et al., 2019). The parameterization follows Simmons et al. (2004),
109 with refinements in the modern Indonesian Through Flow (ITF) region according to Koch-Larrouy et al. (2007). To
110 create an Early Eocene tidal dissipation forcing, we directly interpolate the H14 M2 tidal field (obtained from the tidal
111 model simulations of Green and Huber 2013) onto the NEMO grid using bilinear interpolation. In the absence of any
112 estimation of the K1 tidal component for the Early Eocene, we ignore this contribution. In addition, the
113 parameterization of Koch-Larrouy et al. (2007) is not used here because the ITF does not exist in the Early Eocene.
114 The geothermal heating distribution q is created from the 55 Ma global crustal age distribution of Müller et al. (2008),
115 on which the age-heatflow relationship of the Stein and Stein (1992) model is applied:

$$116 \quad q(t) = 510 t^{-1/2} \quad t \leq 55 \text{ Ma}$$

$$117 \quad q(t) = 48 + 96 e^{-0.0278 t} \quad t > 55 \text{ Ma}$$

118 In regions of subducted seafloor where age information is not available, we prescribe the minimal heatflow value
119 derived from known crustal age. The $1^\circ \times 1^\circ$ resulting field is then bilinearly interpolated on the NEMO grid. It must
120 be noted that the Stein and Stein parameterization becomes singular for young crustal ages, which yields unrealistically
121 large heatflow values. We thus set an upper limit of 400 mW.m⁻² on heatflow values, following Emile-Geay and
122 Madec (2009).

123 Salinity is initialized as globally constant to a value of 34.7 psu following Lunt et al. (2017). The initialization of the
124 model with the proposed DeepMIP temperature distribution (Lunt et al., 2017) led to severe instabilities of the model
125 during the spin-up phase. The initial temperature distribution has thus been modified to follow:

126 $T(^{\circ}\text{C}) = \frac{(1000-z)}{1000} 25 \cos(\varphi) + 10 \quad z \leq 1000 \text{ m}$

127 $T(^{\circ}\text{C}) = 10 \quad z > 1000 \text{ m}$

128 with φ the latitude and z the depth of the ocean. This new equation gives an initial globally constant temperature of
129 10°C below 1000 m and a zonally symmetric distribution above, reaching surface values of 35°C at the equator and
130 10°C at the poles. This profile gives a 5°C surface temperature reduction compared to the DeepMIP equation (Lunt et
131 al., 2017). No sea ice is prescribed at the beginning of the simulations.

132 The IPSL-CM5A2 model includes PISCES biogeochemical model. Biogeochemical cycles and marine biology are
133 directly forced by dynamical variables of the physical ocean and may affect the ocean physics via their influence on
134 chlorophyll production, which modulates light penetration in the ocean. However, because this feedback does not
135 significantly affect the ocean state significantly (Kageyama et al., 2013) and because the Early Eocene mean ocean
136 color is unknown, we have prescribed a constant chlorophyll value of 0.05 g.Chl/L for the computation of light
137 penetration in the ocean. As a consequence, marine biogeochemical cycles and biology do not alter the dynamics of
138 the ocean and as such, biogeochemical initial forcings have been kept to modern.

139 2.1.3. Continental boundary conditions

140 The atmospheric (LMDZ) and land surface (ORCHIDEE) models run on a low resolution grid but require input
141 forcings at higher resolution. The topographic field is created by masking out ocean points in the H14 reference file
142 and upscaling the 1°x1° masked H14 file to the required LMDZ input topographic resolution (1/6°), as LMDZ includes
143 a subgrid scale orographic drag parameterization requiring high-resolution surface orography (Lott and Miller, 1997;
144 Lott, 1999). A similar procedure is applied to the standard deviation of orography proposed by H14.

145 Following Lunt et al. (2017), the soil properties are prescribed as globally constant to the global mean of the PI
146 simulation. There is no lake module in this version of IPSL-CM5A2. The river routing proposed by H14 is passed to
147 ORCHIDEE at its original resolution of 1°x1°, which ensures an appropriate downscaling to the model resolution.
148 The vegetation cover is prescribed from the BIOME4 reconstruction of H14, using a lookup table (given in Table S2)
149 to convert the 10 megabiomes into ORCHIDEE Plant Functional Types (PFTs). Aerosol distributions are left identical
150 to PI values.

151 2.1.4. Sensitivity experiments and equilibrium

152 We perform two additional Early Eocene experiments. One has the same boundary conditions as the baseline Early
153 Eocene experiment (55 Ma-3x) but an atmospheric CO₂ concentration of 420 ppm (1.5x the PI levels, 55 Ma-1.5x).
154 The other differs from the baseline Early Eocene experiment by the absence of tidal dissipation forcing (55 Ma-3x-
155 noM2).

156 The 55 Ma-3x simulation is initialized from rest and run for 4000 years (Fig. S1). The 55 Ma-1.5x simulation is
157 branched from 55 Ma-3x at year 1500 and run for 4000 years. The 55 Ma-3x-noM2 is branched from 55 Ma-3x at
158 year 3000 and run for 2000 additional years. The two PI simulations are initialized from the Levitus climatology
159 (Boyer et al., 2005), and run for more than 2700 years (Table 1). At the end of all the simulations, the ocean has

160 reached a quasi-equilibrated state and trends in deep ocean temperatures over the final 1000 years of all simulations
161 are smaller than 0.05°C/century.

162 We use the monthly outputs of the last 100 years of each simulation to create a climatological year for each simulation.
163 In the following, we will mostly focus on the comparison between the baseline Early Eocene simulation (55 Ma-3x)
164 and the PI control simulation (PI-1x). The other simulations are analyzed in Section 3 to estimate the contribution of
165 tidal mixing to the oceanic overturning circulation, and in Section 6 to examine the sensitivity of the ocean conditions
166 to different levels of CO₂ in the atmosphere.

167 2.2. Evaluation of the simulated ocean temperature

168 This section examines the mean state and the seasonal variation of ocean temperature in the 55 Ma simulations.
169 Further, we evaluate the ability of IPSL-CM5A2 to reasonably simulate the early Eocene sea surface temperature
170 (SST).

171 The annual mean SST in the 55 Ma-3x simulation varies from 10–15°C in the Southern Ocean, to 37.2°C near the
172 Equator (Fig. 2A), with a global mean of 27.5°C. During summer (defined as July-August-September for the Northern
173 Hemisphere and January-February-March for the Southern Hemisphere), the simulated SST reaches ~20°C over most
174 of Southern Ocean (south of 60°S), and up to 38°C in parts of equatorial Indian and Atlantic oceans (Fig. 2A). In the
175 55 Ma-1.5x simulation, both SST and global mean temperature are ~5°C lower than in the 55 Ma-3x simulation (Table
176 1).

177 These simulated SSTs are further compared with proxy-based SST estimates for the early Eocene provided by a recent
178 data compilation performed within the DeepMIP framework (Hollis et al., 2019). The dataset includes 32 records in
179 total, from 4 proxy types (TEX⁸⁶, δ¹⁸O, Mg/Ca and Clumped Isotope data). The spatial pattern of the simulated SST
180 is overall consistent with the proxy based SST, although significant differences can be observed at some specific proxy
181 data sites (Fig. 2A and S2A).

182 In order to further compare the simulations with the proxy-based reconstructions, we also calculate the root-mean-
183 square deviations (RMSD) between the simulated SST and the reconstructions (Table 2). Although large, the RMSD
184 values are overall of the same order of magnitude as the uncertainties of proxy-based SST estimates, suggesting a
185 reasonable model-data consistency. More importantly, the RMSD values are smaller for the 55 Ma-3x simulation than
186 for the 55 Ma-1.5x simulation, suggesting that the 55 Ma-3x simulation captures better the signal of proxy-based SST
187 reconstructions. This is also consistent with proxy reconstructions suggesting that the atmospheric CO₂ concentrations
188 during the early Eocene were most likely three to four times the PI level (Foster et al., 2017). Additional details on
189 the model-proxy comparison can be found in Supplementary Material.

190 The zonal mean SST in the 55 Ma-3x simulation ranges from 30°C to 37°C in the tropics and decreases toward the
191 high latitudes (Fig. 2B). Within the 40°S–40°N latitudinal band, summer SST remains above 30°C, whereas around
192 60–70°S the annual mean SSTs are ~13°C with a seasonal amplitude of 10 to 15°C. These zonal mean SSTs are overall
193 ~10°C warmer than in the PI-1x simulation, with the largest warmings (as much 12°C) found in the Southern Ocean
194 (Fig. 2B). The warm SSTs found in the Southern Ocean in the 55 Ma simulations also extend at depth (Fig. 2C), with

195 a mean global temperature of 11.3°C in the 55 Ma-3x run, compared to 3.3°C in PI-1x (Table 1). The very warm
196 temperatures found at depth are compatible with several proxy-based temperature estimates (bottom temperatures at
197 1000-5000m between 10 and 15°C, e.g. Huber et al., 2000, their Fig. 4-1). More specifically, Mg/Ca-based
198 temperature estimates suggest that the bottom-water (below 1000 m) temperatures were around 15°C during the early
199 Eocene (Cramer et al. 2011), and Dunkley Jones et al. (2013) found a similar value based on $\delta^{18}\text{O-Mg/Ca}$ paleo-
200 thermometry for the PETM time-window.

201 3. The overturning circulation

202 Here we describe the simulated MOC in the different simulations, and investigate the links between the MOC and
203 deep-water formation.

204 3.1. Meridional overturning circulation

205 The global MOC is represented through the vertical streamfunction ψ computed from the zonally-integrated
206 meridional volume transport as:

$$207 \quad \psi(z) = \int_W^E \int_{-H}^0 v(x, z) dz dx \quad (1)$$

208 where v is the meridional velocity, z is the vertical coordinate, H is the ocean depth, x is the zonal coordinate
209 integrated from West (W) to East (E) boundary (meaning that we examine ψ while facing west).

210 In the 55 Ma-3x simulation, a single anticlockwise inter-hemispheric MOC cell fills the whole deep ocean, with a
211 maximum of 40 Sv at 1500 m depth and 60°S (Fig. 3). This strong MOC cell (referred to as SOMOC for Southern
212 Ocean MOC) is associated with the formation of (Eocene) Antarctic Bottom Water (AABW) in the Southern Ocean,
213 flowing northward below about 2000 m. The SOMOC is associated with an upwelling branch extending over the
214 whole Northern Hemisphere, with almost 26 Sv crossing the equator northward (22 Sv at 20°N and only ~5 Sv at
215 60°N). There is no deep-water formation in the Northern Hemisphere, neither in the North Atlantic nor in the North
216 Pacific.

217 In contrast to the 55 Ma-3x run, the MOC in the PI-1x simulation is composed of the traditional upper and lower cells.
218 The upper cell is clockwise and associated with the AMOC, with a maximum strength of 11–12 Sv reached at 800 m
219 depth around 50–60°N. This cell is fed by the formation of North Atlantic Deep Water (NADW) around 60°N. NADW
220 is transported southward all the way to the Southern Ocean at depth between 1000 and 3000 m, where it is brought
221 back to the surface through wind-induced upwelling, forming the quasi-adiabatic pole-to-pole overturning circulation
222 regime (Marshall and Speer, 2012; Wolfe and Cessi, 2014). The lower cell is anticlockwise, with a maximum strength
223 of 15–16 Sv at 3000 m depth. This anticlockwise circulation is fed by the Antarctic Bottom Water formed by surface
224 buoyancy loss related to ocean-sea-ice interaction (Abernathey et al., 2016), and consumed through mixing process
225 induced by the breaking of internal wave at the sea floor and geothermal heating (Nikurashin and Ferrari, 2013; de
226 Lavergne et al., 2016).

227 The simulated deep-water formation in the Southern Ocean in the 55 Ma simulation is compatible with currently
228 available proxy-based reconstructions of the MOC. Although proxy data constraining the ocean circulation are very
229 limited for the early Eocene, these data seem to point to a common deep-water source around the Austral Ocean during
230 that period (Abbott et al., 2016; Batenburg et al., 2018; Frank, 2002; Thomas et al., 2003, 2014). For instance,
231 Batenburg et al. (2018) reported the convergence of Nd-isotopic signature of sea water across the Atlantic basin from
232 59 Ma onward, and proposed that this convergence is result from an intensification of the intermediate and deep ocean
233 circulation in the Atlantic Ocean, with the dominant deep-water mass originating from the southern high latitudes. In
234 addition, the reconstructed concentration of deep-sea carbonate ion ($[\text{CO}_3^{2-}]$) during the Eocene shows a reversed
235 inter-basin gradient compared to the present-day, suggesting a reversed ocean circulation at depth compared to the
236 present-day circulation (Zeebe and Zachos, 2007). Therefore, these proxy data and the simulated ocean circulation are
237 compatible, at least on the direction of the deep circulation and the source of the deep water masses.

238 3.2. Convection and deep-water formation

239 The abyssal circulation described in Section 3.1 is fed by deep convection processes that mainly occur in winter. We
240 examine the simulated Mixed Layer Depth (MLD) at the end of the winter season (Fig. 4), which is an efficient
241 indicator of convection. In the 55 Ma-3x simulation, deep convection occurs only at high-latitude in the Southern
242 Hemisphere, with maximum MLD reaching up to 4000 m in the Weddell Sea, and large areas of MLDs deeper than
243 2000 m around Antarctica in the Ross and Amundsen seas (Fig. 4). In contrast, MLDs remain shallow in the Northern
244 Hemisphere, suggesting the absence of any deep convection there. In this hemisphere, the maximum MLDs are found
245 in the North Pacific (350 m) over the poleward western boundary current between 35–50°N, and in the North Atlantic
246 between 35–40°N (300 m), but the deepest MLD at high latitude are only 200 m over the northwest Pacific. Note that
247 the early Eocene North Atlantic basin is limited to a narrow region west of Greenland and poleward of 52°N. In the
248 absence of other sources of deep waters, waters sourced in the Southern Ocean around Antarctica fill the whole abyssal
249 ocean (Fig. 3).

250 In contrast, deep-water formation occurs both in the northern North Atlantic and in the Weddell Sea in the PI-1x
251 simulation (Fig. 4, right column). The deepest MLD (up to 1500 m) are found in the Nordic Seas between 70–75°N
252 and just south of Iceland around 60°N. MLD larger than 1200 m are also found in the eastern Weddell Sea around
253 65°S, indicative of the formation of AABW. This pattern is consistent with the MLD simulated by the CMIP5
254 ensemble for modern conditions (Heuzé et al., 2015).

255 In 55 Ma-3x, the permanent absence of deep convection in the North Pacific is at odds with a few previous model
256 studies and proxy-based reconstruction of the ocean circulation of the early Eocene that have suggested that deep
257 water could form in the North Pacific (Lunt et al., 2010; Abbott et al., 2016). The reasons why deep water forms in
258 the Southern Ocean in the 55 Ma-3x simulation and not in the North Pacific can be linked to the different ocean
259 thermo-dynamical properties of these two regions. Table 3 summarizes key ocean surface variables in the convective
260 regions of the Weddell Sea, where the deepest MLD are found, and of the North Pacific. These two regions roughly
261 correspond to closed Sea Surface Height (SSH) contours around deep convection regions, which are associated with

262 a minimum of SSH— the geostrophic ocean surface currents circulate cyclonically along the SSH contours, so that
263 the interior regions remain largely isolated from surrounding waters. The Weddell Sea is characterized by a strongly
264 reduced vertical stratification compared to the North Pacific, the surface signature of which is much larger surface
265 density (+0.51 kg/m³). This reduced stratification provides favorable conditions for the emergence of deep convection.
266 According to the equation of state of seawater, the larger surface density found in the Weddell Sea convection region
267 is due to higher salinity (+0.73 psu) contributing to a density increase of 0.57 kg/m³, partly balanced by warmer
268 temperature (+0.36°C) contributing to a 0.06 kg/m³ density decrease.

269 Two major causes may explain these large differences in salinity between the two regions: (i) the atmospheric
270 circulation and freshwater fluxes to the ocean and (ii) the ocean circulation and positive salt-advection feedback. The
271 surface freshwater budget for two extended regions over the Weddell Sea (78°S–61°S and 62°W–8°E) and the North
272 Pacific (48°N–67°N and 124°E–143°W) is shown in Table 3. Averaged precipitation and runoff in the North Pacific
273 box (respectively 1.48 and 0.37 m/yr) exceed those in the Weddell Sea box (1.01 and 0.22 m/yr) by roughly 50%,
274 whereas the evaporation rates are almost similar (0.75 vs 0.71 m/yr). Overall, the average net surface freshwater input
275 is 0.53 m/yr in the Weddell Sea compared to 1.11 m/yr in the North Pacific. Reduced freshwater input by precipitations
276 and continental runoff is related to different atmospheric circulation in the Northern and Southern Hemispheres. In
277 the Southern Ocean, winds largely follow the Antarctic orography (as shown by geopotential height at 850 hPa, Fig.
278 S3), and induce almost no precipitation by orographic uplift over Antarctica coastal regions and no runoff to the
279 Southern Ocean. In contrast, in the North Pacific, the westerlies are blocked by the paleo-Rocky Mountains, especially
280 in the region between 50°N and 70°N. The orographic uplift of moist air masses induces high precipitation (up to 2–
281 3 m/yr) and runoff into the North Pacific (as found in several other models, Carmichael et al., 2016), leading to low
282 sea surface salinity (below 30 psu) along the Pacific coast of North America (not shown) and consequently to increased
283 surface stratification. The upper branch of the MOC and the associated poleward advection of saline subtropical waters
284 constitutes the other contribution to the larger salinities found in the Southern Ocean relative to the North Pacific. This
285 process was sorted out as a positive salt-advection feedback by Ferreira et al. (2018)

286 In contrast, present-day circulation, characterized by deep-water formation in the North Atlantic, is maintained by
287 higher salinities in the North Atlantic than in the North Pacific, which are partly sustained by atmospheric fluxes and
288 the salt-advection feedback (Ferreira et al., 2018). A recent sensitivity study of the impact of topography on modern
289 ocean circulation reveals that the presence of the Rocky Mountains influences the global salinity pattern and the
290 regions where deep convection occurs, through the adjustments of the freshwater transfers between the Pacific and
291 the Atlantic Oceans (Maffre et al., 2018).

292 3.3. Factors contributing to the vigorous SOMOC

293 Different factors contribute to the intense SOMOC simulated by the 55 Ma-3x simulation (40 Sv) in comparison with
294 typical present-day MOC, the intensity of which reaches around 18 Sv and 20 Sv for the upper and lower cells
295 respectively (Lumpkin and Speer, 2007). Deep-water formation occurs in the three sectors of the Southern Ocean
296 (Pacific, Atlantic and Indian) but the zonal connections between the different basins hamper a clear quantification of

297 the contribution of each deep-water formation sector to the SOMOC. However, the contributions of the Weddell Sea
298 and the Pacific sector of the Southern Ocean can be estimated because the narrow width of Drake and Tasman passages
299 at 55 Ma (Fig. 1) creates latitudinal continental boundaries on western and eastern sides of these regions. The Weddell
300 Sea, which exhibits the largest MLD in 55 Ma-3x, and the Pacific sector contribute roughly equally to the SOMOC
301 intensity (~ 19 Sv). The South Indian sector contribution is more difficult to assess directly because of the large open-
302 ocean zonal connection with the Atlantic sector.

303 The shallow Drake Passage at 55 Ma provides a western boundary for the development of a subpolar gyre in the
304 Weddell Sea (see Section 4 for more details). This clockwise gyre produces a favorable environment to trigger deep-
305 water formation (known as preconditioning) through isopycnals doming in the center of the gyre, thereby bringing
306 weakly stratified waters of the ocean interior close to the surface (Marshall and Schott, 1999). Clockwise subpolar
307 gyres, and associated deep-water formation by winter convection, are also present in the Pacific sector of the Southern
308 Ocean, in the Ross and Amundsen seas. Previous numerical investigations of the effects of a closed Drake Passage on
309 ocean dynamics have revealed that the closure of the Drake Passage tends to promote the existence of subpolar gyres
310 in the Southern Ocean and vigorous deep-water formation (Nong et al., 2000; Sijp and England, 2004; Ladant et al.,
311 2018). Additionally, it has been recently suggested that the effects of the closure and opening of the Drake Passage
312 and the Panama Seaway may not be independent (Yang et al., 2014; England et al., 2017; Ladant et al., 2018). For
313 instance, Yang et al. (2014) found that closing the Drake Passage tends to suppress the AMOC and to promote the
314 emergence of a strong SOMOC when the Panama gateway is open, whereas the AMOC may remain intense when the
315 Panama gateway is closed. It is thus very likely that, in our 55 Ma simulations, the very shallow Drake Passage,
316 possibly in combination with the opened Panama gateway, contributes to the strong SOMOC.

317 In addition to the influence of the different gateways, tidally-induced mixing, which represents the enhanced vertical
318 diffusivity resulting from the breaking internal waves generated by the interaction of tidal currents with rough bottom
319 topography (St. Laurent et al., 2002), is another factor that contributes to the strong SOMOC found in the 55 Ma-3x
320 simulation. A twin experiment of the 55 Ma-3x simulation, in which no tidal-induced mixing is prescribed (55 Ma-
321 3x-noM2), simulates a SOMOC with a similar structure but an intensity that is 7 Sv weaker (33 Sv, compared to 40
322 Sv in the reference 55 Ma-3x simulation, Fig. S4). It should be noted that the additional simulation has only been run
323 for 2000 years, so that a small part of the difference between the two runs could arise from different equilibrium states.
324 Yet, the difference between these two simulations is consistent with the recent results of Weber and Thomas (2017),
325 who find a 10 Sv MOC enhancement in an early Eocene simulation with the ECHAM5/MPIOM model that explicitly
326 simulates tides compared with a simulation that does not. This suggests that our parameterization for tidally-induced
327 mixing based on the M_2 dissipation fields of Green and Huber (2013) reasonably represents the effect of early Eocene
328 tides on global ocean circulation. The strengthening of the MOC induced by tidal mixing can be directly related to the
329 driving role of diapycnal mixing on the overturning circulation. Numerous studies have demonstrated that both the
330 magnitude and the vertical distribution of the diapycnal mixing largely affects the strength of the MOC (Bryan, 1987;
331 Manabe and Stouffer, 1999). As already pointed out by Green and Huber (2013), the Eocene MOC may be much more
332 sensitive to the intensity of the abyssal mixing than the present day AMOC, that is largely isolated from the ocean

333 floor by the presence of AABW and sustained quasi-adiabatically by the wind-driven upwelling in the Antarctic
334 Circumpolar Current (ACC; Marshall and Speer, 2012). Diapycnal mixing is the only process that can warm up the
335 dense waters formed in the Eocene Southern Ocean, and these dense waters are directly exposed to the tidally-induced
336 mixing at the bottom, such that the abyssal dissipation becomes the main controlling factor of the SOMOC intensity.
337 In this respect, the large tidal dissipation rate suggested by Green and Huber (2013) in the Pacific is particularly
338 important.

339 Another often-mentioned factor affecting ocean circulation is the climate state, in response to the atmospheric CO₂.
340 Based on highly simplified models, theoretical studies have suggested that ocean ventilation tends to increase under a
341 warmer climate state due to a higher sensitivity of seawater density to temperature (e.g. de Boer et al., 2007). However,
342 the deep-water formation is a very regional phenomenon and, thus, this idealized relationship might be complicated
343 by other regional-scale factors. Indeed, we do not see any systematic change in SOMOC between two simulations
344 with different CO₂ levels (1.5x and 3x) (see Section 6 for further discussion).

345 4. Horizontal circulation and winds

346 In the present day North Atlantic, vertical and horizontal circulations are intimately connected, especially in the
347 subpolar gyre (Marshall and Schott, 1999). For instance, the warm North Atlantic Current flowing northward and the
348 cold East Greenland Current flowing southward are found roughly at the same depth at 60°N, such that the MOC in
349 z-coordinates does not capture the associated water mass transformation at high latitude, whereas the MOC in density
350 coordinates does (Zhang, 2010). It is only when the cold branch deepens in the Labrador Sea and becomes the Deep
351 Western Boundary Current off Cape Hatteras that the overturning streamfunction in z-coordinates provides a good
352 estimate of the water mass transformation. From this perspective, the horizontal subpolar gyre is integral part of the
353 North Atlantic thermohaline circulation. By analogy, a similar intricacy between horizontal gyres and the thermohaline
354 circulation can be expected to exist in the early Eocene Southern Ocean, where the near-closure of the Drake and
355 Tasmanian passages allows for the emergence of intense subpolar gyres, which can precondition and feed the
356 formation of deep water sustaining a strong SOMOC. We thus examine in the following the horizontal ocean
357 circulation during the Eocene.

358 4.1. Gyre circulations

359 In the 55 Ma-3x run, several horizontal gyres are well developed in both hemispheres, as shown by the barotropic
360 streamfunction (Fig. 5). In particular, the quasi-closed Drake and Tasmanian passages support the western boundary
361 currents necessary for the buildup of intense subpolar gyres in each sector of the Southern Ocean. The intensity of
362 each gyre is respectively 40 Sv in the Weddell Sea, 35 Sv in the Indian sector, and 28 Sv in the Ross and Amundsen
363 Seas in the Pacific sector, and winter convection and deep water formation occurs in the center of these gyres, as it
364 is the case in the modern Labrador Sea (Marshall and Schott, 1999). The formation of deep water in the subpolar gyres
365 is also promoted by the advection of saline subtropical waters from the subtropical gyres southward flowing branch
366 (visible for instance on sea surface salinity), as shown by Ferreira et al. (2018). Compared to the PI conditions, the

367 subtropical gyres are strongly perturbed by the numerous open gateways connecting the different basins. In the
368 Southern Hemisphere, due to the large opening between Australia and Asia, the main subtropical gyre extends over
369 both the paleo-Indian and Pacific oceans, with a western boundary current leaning partly on the northern coast of
370 Australia (up to 52 Sv), Madagascar (60 Sv) and Africa. This ‘super’-subtropical gyre is partly fed by a 31 Sv eastward
371 flow south of the Cape of Good Hope originating from the South Atlantic subtropical gyre. Fig. 5 also reveals the
372 existence of a strong anti-clockwise subtropical gyre south of a clockwise subpolar gyre in the North Pacific, with
373 maximum intensity of ~42 and 13 Sv respectively. In contrast, the gyres in the tectonically-restricted North Atlantic
374 basin (Fig. 1) are weak, with maximum intensity of 22 Sv and 2 Sv for the subtropical and subpolar cells respectively.
375 The numerous gateways in the tropical band clearly complicate the traditional gyre pattern in each basin, and increase
376 global connectivity between ocean basins.

377 In the PI-1x simulation, the maximum intensity of the North Atlantic subtropical and subpolar gyres are respectively
378 37 and 19 Sv, which is comparable with the intensity of the gyres found in the Southern Hemisphere in the 55 Ma
379 simulations. The most salient feature of the PI-1x simulation is the existence of the Antarctic Circumpolar Current
380 (ACC), with an eastward transport of 108 Sv through the Drake Passage, which totally disrupts the subtropical-
381 subpolar gyres circulation in the different basins of the Southern Hemisphere. In many respects, the South Atlantic
382 and the Weddell Sea during the Eocene are thus an analogue of the present-day North Atlantic in terms of intertwining
383 of subtropical, subpolar and overturning circulation.

384 **4.2. Wind stress**

385 Modern ocean circulation, in particular in the surface layers, is largely driven by the surface winds (e.g. Munk, 1950),
386 and it is thus interesting to examine the difference between the PI and 55 Ma horizontal circulation in light of changes
387 in the wind pattern.

388 Overall, the patterns of the wind stress at the ocean surface in the 55 Ma-3x and PI-1x simulations are similar (Fig.
389 6), although the magnitude differs between the simulations. The sign of this wind stress difference depends on the
390 hemisphere considered. For instance, wind stress is about 30% weaker in 55 Ma-3x than in PI-1x in the Southern
391 Hemisphere (by 0.05 N m^{-2} at 45°S) whereas it is slightly stronger in the Northern Hemisphere. In the Southern
392 Hemisphere, the difference is particularly striking in the South Atlantic and Indian basins and is largely due to the
393 blocking position of Australia in the early Eocene. The paleogeographic context also explains the increased symmetry
394 in zonal wind stress fields between the Eocene hemispheres relative to the PI.

395 The more symmetrical pattern of wind stress at the ocean surface in the 55 Ma-3x simulation (compared to PI-1x) is
396 also observed in the zonal wind fields from the surface to 500 hPa in the atmosphere (Fig. 6D). The zonal wind strength
397 is largely determined by the meridional temperature gradient in the atmosphere through the thermal wind relation
398 (Holton and Staley, 1973). Indeed, the meridional temperature gradient in the 55 Ma-3x run (compared to PI-1x) is
399 much reduced in the Southern Hemisphere south of 40°S in a large part of the air column, (roughly from the surface
400 to at least 500 hPa) in good agreement with the weaker westerly winds found in the 55 Ma-3x run (Fig. 6D). Moreover,

401 the positions of Australia, Africa, and South America are located much more southward during the Eocene than the
 402 present-day, result in a blocking effect on zonal winds, reducing the wind stress at 500 hPa by a maximum of 18% at
 403 40°S. In the Northern Hemisphere, the meridional temperature gradient in the 55 Ma-3x simulation is reduced at the
 404 surface only north of 60°N and similar at 500 hPa relative to the PI-1x, whereas the zonal winds are slightly stronger
 405 in 55 Ma-3x simulation throughout the air column compared to the PI-1x. At the surface, the maximum zonal wind
 406 stress is 19% stronger and shifts poleward by 2°, mainly due to land-sea distribution.

407 Changes in ocean gyre circulation between the 55 Ma and PI configurations are mostly due to the large changes in the
 408 ocean basin geometry and gateways rather to the moderate changes in the strength and patterns of the wind stress (and
 409 its curl). A full understanding of what sets the gyres intensity in the 55 Ma simulations (as well as in the PI simulations)
 410 would require further investigations, which are beyond the scope of the present paper.

411 **5. Oceanic heat transport and its decomposition**

412 Modern ocean circulation plays a key role in the regulation of climate through its contribution to the redistribution of
 413 heat from the Equator to the Poles. In the tropics, the ocean transports roughly 50% of the 3 PW carried northward by
 414 the ocean–atmosphere system, but less than 10% of the total at high latitude (Trenberth and Caron, 2001). The relative
 415 contributions of the horizontal and overturning ocean circulations to the meridional heat transport also vary greatly
 416 over different latitudes and between oceanic basins (Ganachaud and Wunsch, 2003). In light of the different ocean
 417 circulations found in the Eocene and PI conditions, we further investigate in the following how efficient the ocean
 418 was at transporting heat across latitudes in the early Eocene.

419 **5.1. Oceanic heat transport**

420 The total meridional OHT at a given latitude y is defined as the sum of advective and diffusive contributions:

$$421 \quad OHT_{total} = \rho_0 C_p \int_W^E \int_{-H}^0 \left(v\theta + K_H \frac{d\theta}{dy} \right) dz dx \quad (2)$$

422 where ρ_0 is the seawater density, C_p is the specific heat capacity of sea water, v is the meridional velocity, θ is the
 423 potential temperature, K_H is the horizontal diffusivity coefficient, and H is the ocean depth. Here the computation is
 424 performed from the model outputs at each model time step. In the early Eocene simulations, there is a significant
 425 increase in OHT in the Southern Hemisphere compared to the present-day simulations, but a decrease in the Northern
 426 Hemisphere (Fig. 7). As a result, the simulated OHT in the 55 Ma-3x experiment is remarkably asymmetric between
 427 hemispheres. The mean OHT difference between the 55 Ma-3x and PI-1x simulations is of the order of 0.2 PW (1 PW
 428 = 10^{15} Watt), peaking at 0.5 PW around 35°S. The 55 Ma-3x OHT reaches a maximum of 1.7 PW at 15–20°S, that is
 429 ~0.3 PW larger than in the present-day simulations at the same latitude but also ~0.5 PW larger than the maximum PI
 430 value of the Northern Hemisphere. This enhanced OHT in the Southern Hemisphere contributes to maintaining the
 431 Southern Hemisphere particularly warm in the early Eocene, especially south of 50°S, as can be seen on the SST
 432 distribution (Fig. 2).

433 Previous studies have examined the role of OHT during the Eocene with a particular focus on the response of the OHT
434 to the opening of Southern Ocean gateways in either realistic Eocene paleogeography or more idealized modern
435 configurations. The OHT simulated by our 55 Ma-3x experiment lies within the range of values found in the literature.
436 For instance, using the NCAR CCSM ocean model with surface heat flux boundary conditions mimicking an energy-
437 balanced atmospheric model, Nong et al. (2000) found that closing the Drake Passage in a modern configuration
438 results in a stronger SOMOC (24 vs 12 Sv), associated with an increased poleward OHT in the Southern Hemisphere
439 (+0.2 PW, from 1 to 1.2 PW) and a decreased OHT in the Northern Hemisphere. Similarly, using the UVic
440 intermediate complexity Earth System Climate Model, Sijp and England (2004) found a strongly enhanced heat
441 transport (from 1.6 to 2.4 PW) in the Southern Hemisphere in response to the closure of Drake Passage in a modern
442 configuration. Using the fully coupled NCAR model with a realistic Eocene configuration in which the Drake Passage
443 has a similar configuration to the one we use but the closed Tasmanian gateway is closed, Huber et al. (2004)
444 performed a set of simulations that exhibited rather weak poleward OHT in the Southern Hemisphere (with a
445 maximum of 0.9 PW at $\sim 10^{\circ}\text{S}$), likely because of the absence of a strong SOMOC in their simulations. When closing
446 the Drake passage in an otherwise modern configuration with the GFDL model, Yang et al. (2014) found that the
447 change in OHT was much larger when the Panama Seaway was open, with a strong increase of the OHT in the
448 Southern Hemisphere. More recently, Baatsen et al. (2018) used the higher resolution CESM earth system model to
449 simulate the later Eocene (38 Ma) climate, with levels of CO_2 and CH_4 in the atmosphere prescribed to 2 and 4 times
450 the PI levels. In these two simulations, Baatsen et al. (2018) found a maximum of ~ 1.5 PW OHT at 20°S in the
451 Southern Hemisphere, associated with a 14–16 Sv SOMOC (compared to the 40 Sv and 1.7 PW at the same latitudes
452 in our 55 Ma-3x simulation).

453 Although these previous studies are all based a variety of models with different complexity and resolution, the results
454 consistently suggest that the OHT is largely coupled to the structure and strength of the MOC cells. For instance, the
455 difference in MOC intensity simulated by coupled and uncoupled models could be primarily caused by the positive
456 salt-advection feedback and the self-stabilizing thermal feedback (Sijp and England, 2004). It is noteworthy that the
457 strengthening influence of tidal-induced mixing on the MOC (+7 Sv) is associated with a rather weak increase in
458 OHT, lower than 0.03 PW on average, in agreement with Weber and Thomas (2017). Such a nonlinear relationship
459 between OHT and MOC has also been suggested by Boccaletti (2005), who argued that the shallow circulation
460 locally can be as important as the deep overturning for determining the OHT.

461 **5.2. Decomposition of the meridional ocean heat transport**

462 In order to understand the differences in OHT among our simulations (Fig. 7A), we further split up the OHT into an
463 advective contribution (OHT_{adv}) and a diffusive contribution, corresponding respectively to the first and second term
464 on the RHS of Eq. 2 (Fig. 7B and 7C). This decomposition reveals that, in both the 55 Ma-3x and the PI-1x simulations,
465 the advective part dominates the OHT at all latitudes, except at 40°S/N in PI-1x where the presence of large
466 temperature gradient (Fig. 2B) results in larger diffusive heat transports. Figs. 7B and 7C also reveal that different
467 OHT between PI-1x and 55 Ma-3x are mainly due to differences in the advective components, and the differences in
468 diffusive OHT being rather small.

469 The advective component, OHT_{adv} , can be decomposed further into an overturning (OHT_{MOC}) and a gyre (OHT_{gyre})
 470 component, following for instance Bryan (1982) or Volkov et al. (2010):

$$471 \quad OHT_{adv} = \underbrace{\rho_0 C_p \iint \bar{v} \bar{\theta} dx dz}_{OHT_{MOC}} + \underbrace{\rho_0 C_p \iint v' \theta' dx dz}_{OHT_{gyre}} \quad (3)$$

472 where \bar{v} , $\bar{\theta}$ represent the zonal averages of the velocity and temperature, respectively, and v' , and θ' the deviations
 473 from these zonal means. The first term of the RHS of Eq. 3 corresponds to the overturning component (OHT_{MOC}) and
 474 the second term corresponds to the horizontal transport associated with the large-scale gyre circulation (OHT_{gyre}).
 475 Note that due to limitations on the availability of model outputs, the different terms of Eq. 3 presented on Figs. 7D
 476 and 7E are computed from monthly means. This explains why the sum of the two terms does not completely equal
 477 OHT_{adv} shown on Fig. 7C, since the latter is computed at each model time step during the simulations. The differences
 478 between the two computations can be seen on Fig. 7C.

479 The decomposition reveals that the enhanced Southern Hemisphere OHT_{adv} in 55 Ma-3x (compared to PI) is overall
 480 due to differences in OHT_{MOC} (Fig. 7D, 7E). The contribution from gyre circulation varies with latitude, with a
 481 compensation effect between OHT_{gyre} and OHT_{MOC} in the low-latitudes and an enhancement in mid-to-high latitudes.
 482 Consequently, the overall larger OHT_{adv} in the tropics of 55 Ma-3x simulation is primarily contributed from the strong
 483 Eocene OHT_{MOC} (up to 2 PW), which is ~ 0.3 PW larger than in the PI simulations. In the mid-latitudes of the Southern
 484 Hemisphere, where the OHT_{adv} is overall smaller than in the tropics, the OHT_{MOC} in 55 Ma-3x is almost 1 PW stronger
 485 than in PI-1x. By contrast, south of $60^\circ S$, enhanced OHT_{adv} in 55 Ma-3x (compared to PI) results from the combination
 486 of stronger OHT_{gyre} and OHT_{MOC} . It is worth noticing that OHT_{gyre} at $\sim 40^\circ S$ in the PI-1x simulation is very likely
 487 underestimated in our decomposition computed from monthly mean data, because higher-frequency processes
 488 (resulting for instance from atmospheric synoptic variability) could contribute significantly to OHT_{gyre} in regions
 489 where the mean meridional currents are weak (Volkov et al., 2010).

490 It is obvious from Figs. 3 and 7 that the vigorous SOMOC simulated in the 55 Ma-3x experiment drives a strong net
 491 OHT toward the South Pole. This strong SOMOC is associated with poleward transport of warm waters at shallow
 492 depths where zonal oceanic temperature gradients are larger and with a returning equatorward transport of colder
 493 water at depth where ocean temperature tends to be more homogenous (Fig. 2C). Remarkably, although the ACC is
 494 absent from the Eocene simulation because of different configurations of the Drake and Tasmanian passages that
 495 constitute latitudinal barriers (Munday et al., 2015), the contribution of the gyre circulation and diffusive process to
 496 poleward heat transport (OHT_{gyre}) is smaller than in the preindustrial simulations. This small OHT_{gyre} in the Eocene
 497 Ma simulation is unexpected, given previous hypotheses on the climatic effects of the ACC (e.g. Nong et al., 2000;
 498 Toggweiler and Bjornsson, 2000; Sijp and England, 2004). The ACC has indeed been suggested to be a barrier for
 499 poleward heat transport, so that the onset of the ACC could be a potential driver for the Antarctica cooling Eocene-
 500 Oligocene Transition around 34 Ma. However, these studies may not have captured the full complexity of the links
 501 between the ACC and OHT in the Southern Ocean. The analysis of both in-situ observations (Watts et al., 2016) and

502 the CESM1.0 model (Yang et al., 2015) have indeed revealed that the ACC is composed of meridional excursions of
503 the mean geostrophic horizontal shear flow, energetic eddies and large diffusive heat transport, which balances out
504 equatorward OHT due to Ekman transport and leads to a net poleward OHT in the Southern Ocean (Volkov et al.,
505 2010).

506 **6. Sensitivity of the ocean response to a doubling of the levels of atmospheric CO₂**

507 Our analysis has so far focused on the comparison between the 55 Ma-3x and PI-1x simulations, as the former is
508 performed with atmospheric CO₂ levels considered to be representative of the early Eocene (Foster et al., 2017; Lunt
509 et al., 2017). The analysis of two additional simulations (55 Ma-1.5x and PI-2x) allows us to investigate the robustness
510 of the ocean circulation in this range of atmospheric CO₂ concentrations. These simulations also help us to quantify
511 the sensitivity of the oceanic conditions to a doubling of the CO₂ levels in the atmosphere under the early Eocene and
512 modern setting.

513 The mean ocean temperatures are very sensitive to the atmospheric CO₂ concentration in the Eocene configuration
514 (Table 1). The global mean ocean temperature in the Eocene simulations increases by 4.9°C in response to CO₂
515 doubling from 1.5x to 3x, which is much larger than the 1°C increase in PI simulations from 1x to 2x. The global mean
516 SST in the Eocene also shows a much larger increase as response to a CO₂ doubling (+4.7°C) than in the preindustrial
517 (+2.6°C), especially at high latitudes and in the regions of deep-convection of the Southern Ocean (Fig. 8), which is
518 similar to the changes in air temperature at 2 m (+5.6°C vs +3.5°C respectively, this difference is known as the climate
519 sensitivity). **Such contrasted values of climate sensitivity between 55 Ma and PI are in good agreement with the recent
520 results of Farnsworth et al. (2019) obtained with the HadCM3BL model.** In the absence of perennial sea-ice in 55 Ma
521 simulations, the winter SST in deep-convection regions is largely influenced by the air-sea interactions and thus
522 directly related to air temperature, which rarely decreases below 10°C (resp. 5°C) in 55 Ma-3x (resp. 1.5x), such that
523 the deep waters filling the whole ocean interior vary accordingly with temperature. This is not the case in the present-
524 day configuration where deep-water formation is tied to marginal ice zones, such that dense waters formed by brine
525 rejection have initial temperatures close to freezing. Intuitively, we could have expected a larger sensitivity of ocean
526 temperature to atmospheric CO₂ in the PI runs because of the ice-albedo feedback. Yet, the effect of this feedback
527 appears to be limited to the high latitudes (Fig. 8) and only plays a marginal role in the changes in global mean SST
528 or temperatures of the deep water masses (Table 1).

529 The ocean circulation exhibits only a minor response to a CO₂ doubling in both the Eocene and preindustrial
530 configurations, although this response locally varies (Table 4). In the 55 Ma simulations, the doubling of CO₂ level
531 marginally enhances the maximum abyssal SOMOC (by 0.3 Sv out of 40 Sv in total), while the intensity of the shallow
532 MOC cell and of the barotropic streamfunction are slightly reduced. In the PI simulations, the doubling of atmospheric
533 CO₂ levels has the opposite effect on the AABW, whose maximum intensity reduces by 1.2 Sv (out of ~15 Sv), whilst
534 the AMOC slightly increases by 0.3 Sv (out of ~11 Sv). This small increase in the steady-state AMOC is quite
535 interesting and needs to be contrasted with the transient response of the AMOC intensity to global warming (Gent,

536 2018; Jansen et al., 2018). Indeed, CMIP-type climate models consistently project strong decline of the AMOC
537 strength when forced with a range of increasing greenhouse gas emission scenarios (Schmittner et al., 2005; Cheng et
538 al., 2013). Yet, when run for longer integrations until full equilibrium, models suggest that the AMOC tends to recover
539 and that the AMOC is ultimately not very sensitive to the CO₂ levels (Jansen et al., 2018; Thomas and Fedorov, 2019),
540 as is the case in our PI simulations. Once equilibrium is reached, the most significant effect of a CO₂ doubling in our
541 PI experiments is a sharp increase in the ACC transport (+22 Sv out of 108 Sv in PI-1x), in response to stronger
542 westerlies in the Southern Ocean. In contrast, the barotropic circulation remains almost the same in the Pacific, but
543 varies in the North Atlantic with a 5 Sv stronger (weaker) subpolar (subtropical) gyre in the PI-2x simulation compared
544 to PI-1x.

545 A number of paleoclimate studies have investigated the influence of CO₂ levels on ocean circulation in coupled
546 models, with contrasted responses depending on the period considered and the model used. For instance, the deep
547 overturning circulation in the HadCM3L climate model shows an overall high sensitivity to CO₂ concentrations in the
548 PETM simulations (Lunt et al., 2010), whilst the GENIE model only exhibits a small response in the Cretaceous
549 simulations of Monteiro et al. (2012). Winguth et al. (2010) further suggest that, in a given simulation of the Paleocene-
550 Eocene period, the different MOC cells (e.g. in the Northern and Southern hemispheres) could respond differently to
551 a change of CO₂ levels. These various responses can be attributed to different factors. First, it is clear that the resolution
552 and overall complexity of the model may partly control the sensitivity of the MOC to CO₂ levels, as in models of
553 modern climate (e.g. Bryan et al., 2006). Second, a variety of time scales are intertwined in the adjustment of ocean
554 circulation to external perturbations, from decades for the dynamical adjustment to millennia for the thermo-dynamical
555 response of bottom waters through vertical advective-diffusive balance (e.g. Donnadieu et al., 2016). The transient
556 response of the ocean can therefore differ, or even be in the opposite direction, from the final equilibrium response.

557 The OHT response to a doubling of CO₂ in the Eocene simulations is also small, with a slight decrease, in contrast to
558 the OHT increase seen in the PI simulations (although the magnitude of the change is smaller; Fig. 9A). The OHT in
559 the 55 Ma-3x simulation is about 0.15 PW smaller than in 55 Ma-1.5x simulation over most latitudes. Both overturning
560 and horizontal components contribute to this overall smaller OHT in the 55 Ma-3x simulation (Fig. 9D, 9E). A weak
561 OHT in the tropics is due to weaker MOC at the same latitudes, while a weak OHT at high latitudes can be attributed
562 to the smaller amplitude of horizontal gyres. In the PI simulations, the OHT in the PI-2x simulation is ~0.1 PW larger
563 than in PI-1x at high latitudes. This larger OHT in PI-2x is mostly due to the gyre component, which is in good
564 agreement with the ~5 Sv stronger North Atlantic subpolar gyre and almost no change in the AMOC for instance.

565 Our results therefore support a stable, yet rather small, response of global ocean circulation and heat transport to the
566 doubling of atmospheric CO₂ levels. Nevertheless, we have only investigated a limited range of CO₂ levels (from 1.5x
567 to 3x) and cannot exclude that the sensitivity of ocean circulation to CO₂ concentrations may change at more extreme
568 CO₂ levels, as this sensitivity may be highly non-linear (Lunt et al., 2010). Given that the early Eocene atmosphere
569 CO₂ levels are relatively poorly constrained by proxy reconstructions, additional experiments are underway to explore
570 the model response to higher CO₂ concentrations in the early Eocene configuration.

571 **7. Conclusions**

572 Numerous proxy based reconstructions have revealed that the early Eocene (~55 Ma) was most likely the warmest
573 period in the Cenozoic. During that period, paleogeographic restrictions of certain modern basins and gateways, such
574 as the North Atlantic and the Drake and Tasmanian passages, suggest fundamental differences with the modern large-
575 scale ocean circulation. It has been proposed that the distinct mode of ocean circulation operating during the early
576 Eocene may have contributed to the significant polar warmth recorded by observational evidence. There is however
577 no consensus on the modes of early Eocene ocean circulation or on the relative influence of the overturning and
578 horizontal circulation on the poleward heat transport. Here we revisit this question by analyzing the ocean circulation
579 and its contribution to the meridional OHT using simulations of the early Eocene performed with the IPSL-CM5A2
580 earth system model set up with the recent paleogeographic reconstruction of this Eocene time slice distributed as part
581 of the DeepMIP project. Our main results are summarized hereafter.

582 A strong abyssal overturning circulation (up to 40Sv) is found in the 55 Ma simulation, with deep water formed only
583 in the Southern Ocean (mainly in the Weddell Sea), whereas there is no deep water formation in the Northern
584 Hemisphere, in contrast to some previous work on the early Eocene (e.g. Winguth et al., 2012). This situation is
585 favored by orographically-induced freshwater fluxes (precipitation and runoff) and maintained by a salt-advection
586 feedback. Indeed, the atmospheric circulation around Antarctica induces relatively low precipitation rates in the
587 Southern Ocean, resulting in higher salinity, and hence larger surface density than in the North Pacific, where the
588 orographic precipitations and runoff induced by the presence of the paleo-Rockies tend to reduce surface salinity and
589 inhibit deep water formation.

590 The paleogeography (and paleobathymetry) and tidally-induced mixing are the main drivers of the strong SOMOC
591 during the early Eocene. The (nearly-)closed Drake and Tasmanian passages are of fundamental importance for
592 sustaining the SOMOC, via their effect on horizontal ocean circulation. More specifically, with the (nearly-)closed
593 Drake and Tasmanian Passages serving as a western boundary, clockwise subpolar gyres are well-developed (~40 Sv)
594 in the Weddell and Ross Seas, thereby favoring the emergence of deep convection and deep-water formation through
595 isopycnal doming and salt-advection feedback. Tidal-induced mixing also contributes to 7 Sv (out of 40 Sv) to this
596 SOMOC, but with only a limited impact on the heat transport.

597 The vigorous SOMOC simulated for the Eocene is associated with a larger poleward OHT (by a maximum of 0.5 PW
598 relative to PI) in the Southern Hemisphere that largely contributes to maintaining a warm Southern Ocean and
599 Antarctica. Perturbation experiments have been conducted in present-day coupled models to evaluate the impact of an
600 AMOC shutdown. In their model, Vellinga and Wood (2008) found that a 10 Sv reduction in AMOC, associated with
601 a change of its structure, leads to a 1.7°C cooling of the Northern Hemisphere, with a local stronger cooling by 5°C
602 in the northern North Atlantic. This gives credit to the importance of the 40 Sv SOMOC for maintaining the Southern
603 Ocean warm in the 55 Ma simulations. However, other factors than a strong SOMOC and associated OHT could
604 contribute to the warm Southern Ocean. Rose and Ferreira (2013) have shown that changes in OHT can induce changes
605 in global mean temperature and meridional temperature gradient through convective adjustment of the extratropical

606 troposphere and increased greenhouse effect. According to their results, the magnitude of those changes could be up
607 to 1°C and 2.6°C for every 0.5 PW enhancement in OHT. Further investigations would be required to examine if this
608 mechanism is also at play in our simulation and whether it contributes significantly to Southern Ocean warmth.

609 A further decomposition of the OHT reveals that the different overturning circulations between the early Eocene and
610 the preindustrial explain most of the increase of the OHT in the Southern Hemisphere. The contribution of gyre
611 circulation to the OHT is only secondary, and varies with latitude, with a compensating effect between the MOC and
612 gyre circulations in the low- and mid-latitudes, whereas the two contributions add up in the high latitudes. More
613 importantly, the latitudinal distribution of the gyre contribution to the OHT only marginally varies between the early
614 Eocene and preindustrial simulations, despite the absence of the ACC in the early Eocene experiment. Given that the
615 55 Ma paleo-bathymetry does not allow the existence of a strong ACC, this result questions strongly the idea that the
616 ACC could be a strong barrier for OHT in our modern climate, and suggests that the meridional excursions of the
617 ACC might indeed play an important role in the gyre-related OHT (Volkov et al., 2010; Watts et al., 2016; Yang et
618 al., 2015).

619
620 **Acknowledgements.** This research has received partial funding from the French National Research Agency (ANR)
621 under the ‘Programme d’Investissements d’Avenir’ ISblue (ANR-17-EURE-0015) and LabexMER (ANR-10-LABX-
622 19) for the COPS project. Additional funding has been received from Ifremer and ‘Université Bretagne Loire’ to
623 support the postdoc of Yurui Zhang. Jean-Baptiste Ladant received funding from the ANR project ANOX-SEA during
624 the early phases of this work. We thank the GENCI TGCC at CEA for providing HPC computational resources. We
625 are grateful to Pierre Sepulchre for the initial design of the numerical model. We thank Arnaud Caubel, Anne Cozic,
626 Agnès Ducharne, Josefina Ghattas, François Lott, Olivier Marti, Jean-Yves Peterschmitt and Alistair Sellar for their
627 help in implementing the Early Eocene boundary conditions and in resolving technical issues.

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Tables

Table 1 Summary of the simulation setup and key diagnostics in the different simulations used in this study. All the values presented are averaged over the last 100 years of each simulation.

Simulation	Setup			Ocean surface			Full depth ocean			Atmos.
	CO ₂ (ppmv)	Bathymetry	Duration (yr)	SST (°C)	SSS (psu)	Sigma (kg/m ³)	T (°C)	S (psu)	Sigma (kg/m ³)	T at 2m (°C)
55 Ma-3x	840	55Ma	4000	27.51	34.04	21.33	11.30	34.68	26.24	25.12
55 Ma-3x-noM2	840	55Ma	2000	27.47	34.16	21.44	10.85	34.68	26.34	24.98
55 Ma-1.5x	420	55Ma	4000	22.82	34.36	22.94	6.38	34.68	26.94	19.56
PI-2x	560	PI	2910	20.14	34.40	23.63	4.32	34.61	27.27	16.88
PI-1x	280	PI	2790	17.51	34.43	24.32	3.34	34.61	27.40	13.33

Table 2 Root-mean-square-deviation (RMSD) of simulated annual mean SST and proxy-based SST estimates (in °C). RMSD metrics are defined in Supplementary material. The proxy-based SST estimates are from the DeepMIP dataset for the early Eocene (Hollis et al. 2019), and the number of data points and the uncertainty for each proxy type is also indicated. The uncertainty range is defined as the 2 σ deviations for $\delta^{18}\text{O}$, and as the range between 5% and 95% percentile SST estimates for TEX⁸⁶, Mg/Ca and Clumped isotope data.

Type of proxy	TEX ⁸⁶	$\delta^{18}\text{O}$	Mg/Ca	Clum. isotope
Number of data points	10	10	7	5
Uncertainty range of proxy-data (°C)	15.1	3.4	6.7	5.1
RMSD (°C)	55Ma-3x	13.7	7.5	6.4
	55Ma-1.5x	18.2	7.5	10.6

Table 3 Key ocean surface parameters in the North Pacific and Weddell Sea in 55Ma-3x simulation. The Weddell Sea and the North Pacific convection regions are defined by the deepest mixed layer depth at high-latitudes and the extended regions are defined as boxes over the Weddell Sea (78°S-61°S and 62°W-8°E) and the North Pacific (48°N-67°N and 124°E-143°W) roughly corresponding to closed contours of sea surface height. SST, SSS, sigma are winter average (January-February-March for the Northern Hemisphere and July-August-September for the Southern Hemisphere), MLD is given for the end of winter (March for the Northern Hemisphere and September for the Southern Hemisphere), while precipitation, runoff and evaporation are annual means.

	Convection region				Extended region						
	SST (°C)	SSS (psu)	sigma (kg/m ³)	MLD (m)	Area (km ²)	Precipitation (P) (m/yr)	Runoff (R) (m/yr)	Evaporation (E) (m/yr)	P+R-E (m/yr)	SST (°C)	SSS (psu)
North Pacific	9.21	34.06	26.22	111	9.23e6	1.48	0.37	0.74	1.12	10.33	32.89
Weddell Sea	9.57	34.79	26.73	3129	4.70e6	1.01	0.22	0.69	0.54	10.13	34.77

Table 4 Intensity of the gyres (SPG_{SA}: South Atlantic subpolar gyre for 55 Ma; STG_{NA}: North Atlantic subtropical gyre; SPG_{NA}: North Atlantic subpolar gyre; STG_{NP}: North Pacific subtropical gyre; SPG_{NP}: North Pacific subpolar gyre; ACC: Antarctic circumpolar current for PI) and overturning cells (SOMOC: Southern Ocean MOC for 55Ma; AABW: Antarctic bottom water for PI; NADW: North Atlantic deep water for PI).

Simulation	Overturning (Sv)		Gyre intensity (Sv)						
	SOMOC/AABW	NADW	SPG _{SA} /ACC	SPG _{SI}	SPG _{SP}	STG _{NA}	SPG _{NA}	STG _{NP}	SPG _{NP}
55 Ma-3x	-40	—	40	35	28	22	-2	42	-13
55 Ma-1.5x	-40	—	38	46	35	29	0	53	-8
PI-2x	-15	11.6	130	—	—	32	-24	48	-19
PI-1x	-16	11.3	108	—	—	37	-19	48	-20

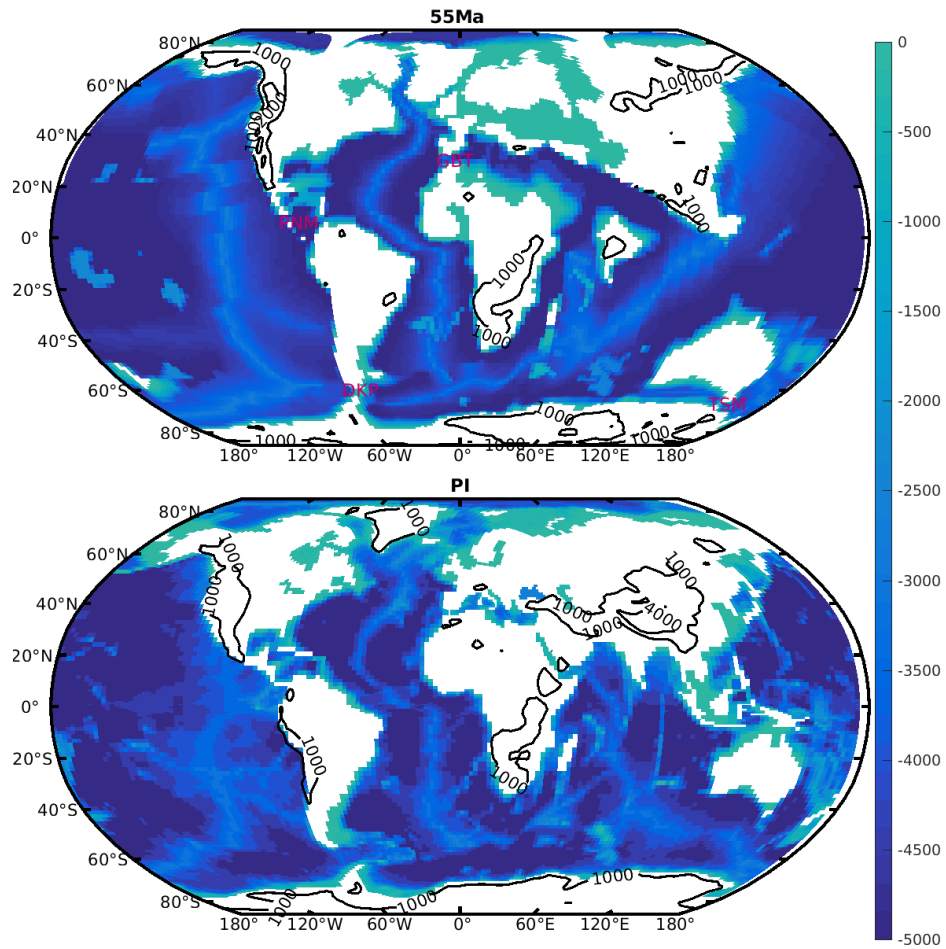


Figure 1. Bathymetry/topography (m) boundary conditions used in the 55 Ma (based on Herold et al. 2014) and PI simulations. The black contours indicate the 1000 and the 4000m altitude. DKP indicates the Drake Passage, TSM the Tasmanian Passage, PNM the Panama passage and GBT Gibraltar.

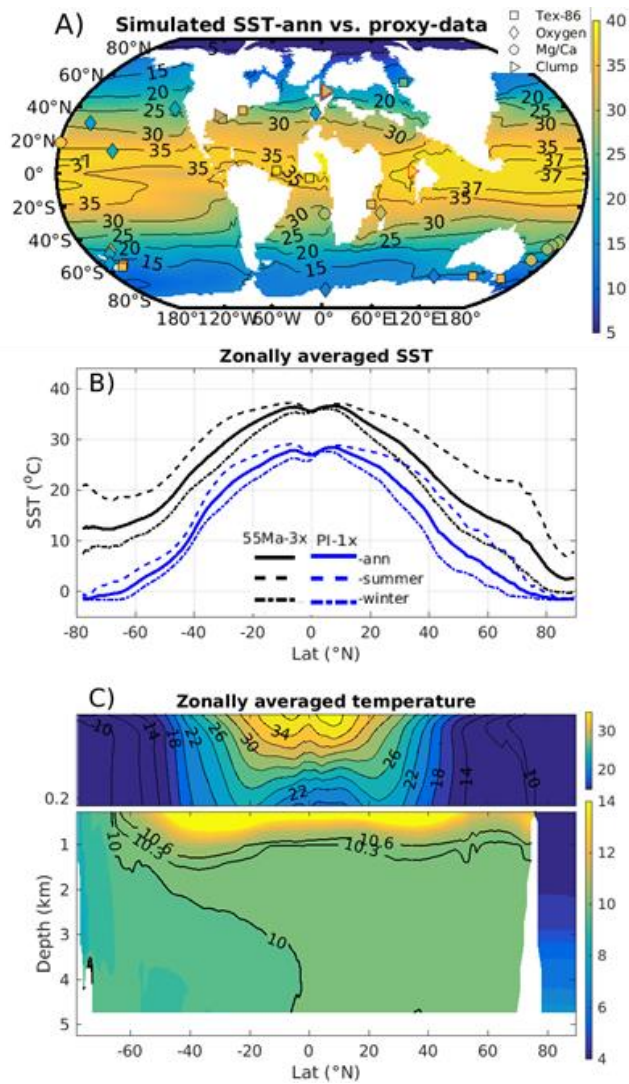


Figure 2. (A) Annual mean SST (in °C) in the 55 Ma-3x simulation, and point-to-point comparisons with proxy-based SST estimates from the DeepMIP dataset for the early Eocene (Hollis et al. 2019). The different symbols represent different proxies. (B) Simulated zonally-averaged annual mean SST (in °C, solid lines) in the 55 Ma-3x (black) and PI-1x (blue) simulations. The dashed and dotted lines are indicating the means for Summer and Winter, respectively. (C) Zonally-averaged ocean temperature in the 55 Ma-3x simulation, with a zoom in the upper 200 m (contour interval: 2°C and 0.3°C for the top and bottom panels, respectively).

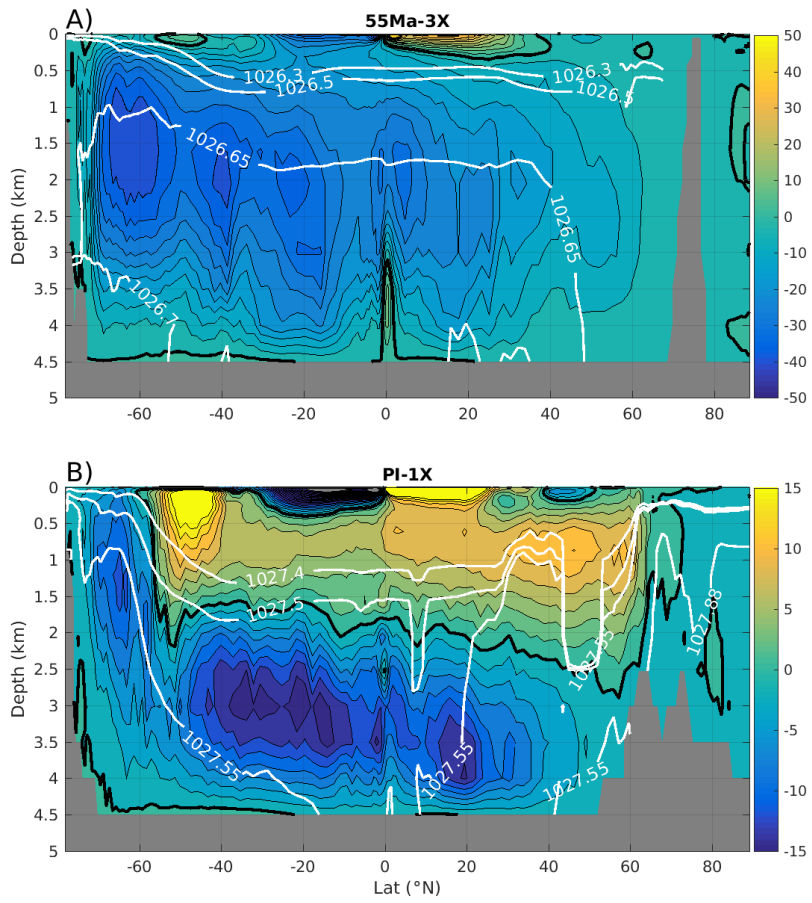


Figure 3. Streamfunction of the meridional overturning circulation (in Sv) in the 55 Ma-3x (A, contour interval: 4 Sv) and PI-1x (B, contour interval: 2 Sv) simulations. Note the different colorbars of the two subplots. The black thick lines indicate the zero-contour, with positive values indicating clockwise circulation, and negative values anti-clockwise circulation. White lines show selected zonally-averaged isopycnal contours (potential density in kg/m^3).

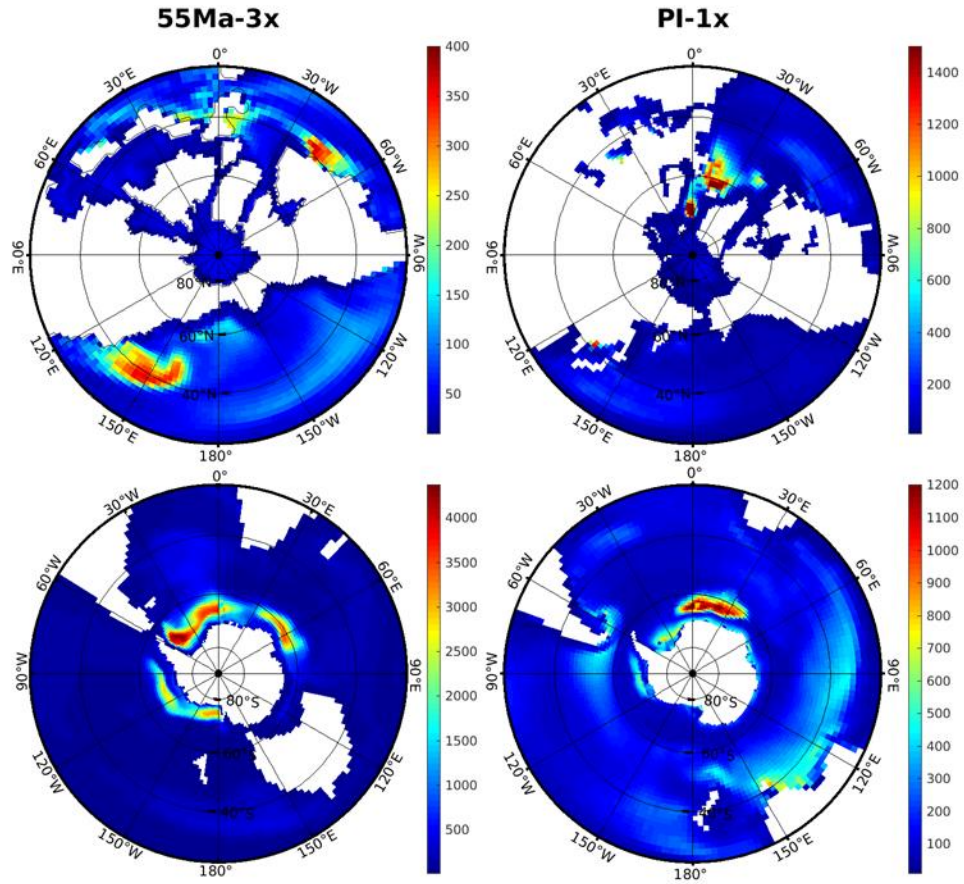


Figure 4. Winter (i.e. March in the Northern Hemisphere – top panels; and September in the Southern Hemisphere – bottom panels) mixed layer depth (in m) in the 55 Ma-3x (left) and PI-1x (right) simulations. Note the very different colorbars among plots. Mixed layer depth is defined by the potential density difference of 0.3 kg/m^3 with reference to the surface.

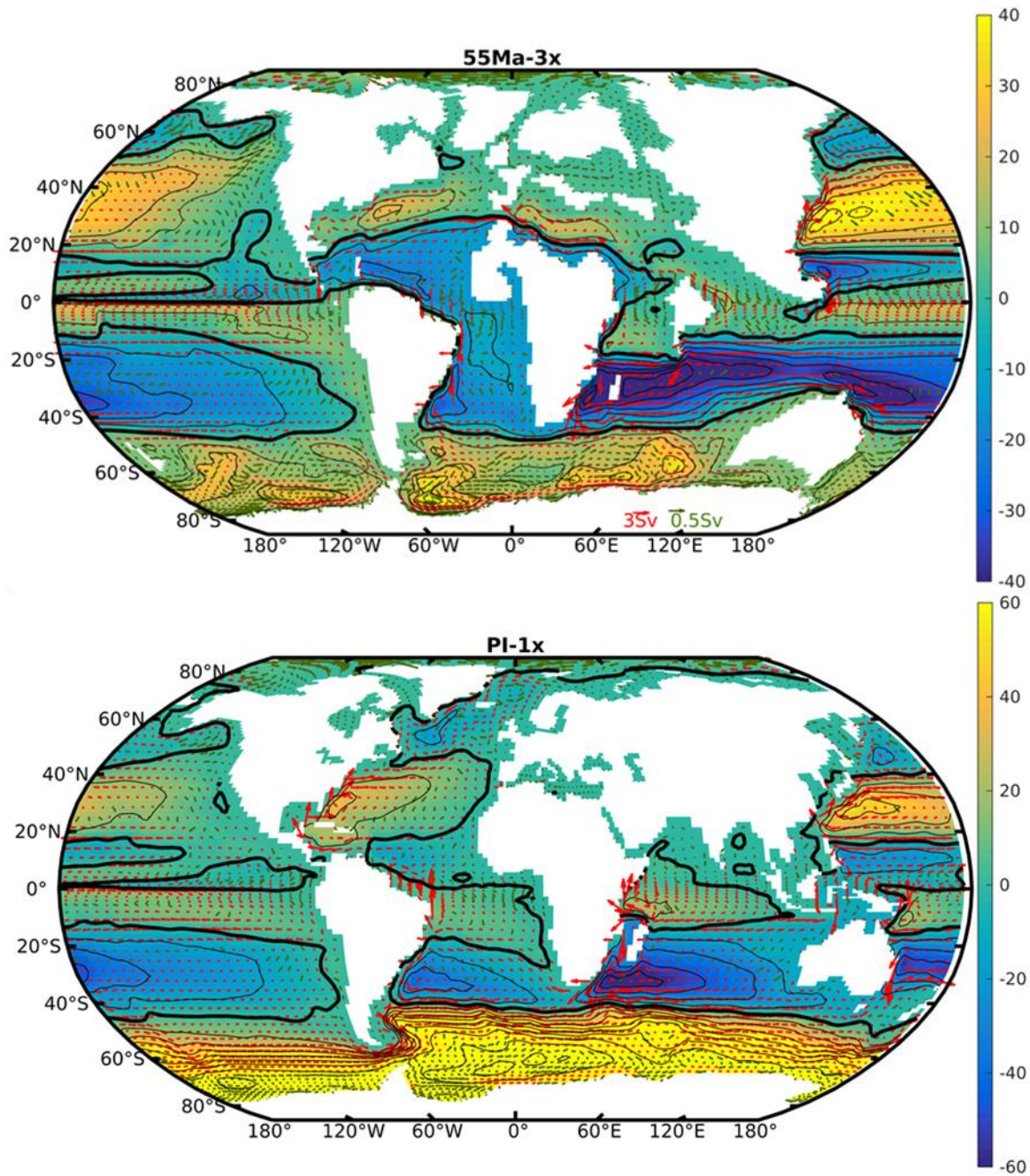


Figure 5. Barotropic streamfunction (in Sv) in the 55 Ma-3x (top, contour interval: 10 Sv) and PI-1x (bottom, contour interval: 15 Sv) simulations, integrated northward from Antarctica. Note the different colorbars of the two subplots. The black thick lines indicate the zero-contour. The mean transport (in Sv) integrated over the top 300m is indicated with vectors (note that only one every two point is plotted to increase the readability). Two scales are used to represent transport larger (in red) or lower (in green) than 0.5 Sv. The transports through the key gateways in the 55 Ma-3x simulation are respectively: Drake Passage 3.1 Sv, Tasmania 1.3 Sv, Panama 4.7 Sv, Gibraltar -14.3 Sv (positive transports are eastward, negative westward). The Drake Passage throughflow, corresponding to ACC, in the PI-1X simulation is ~108 Sv.

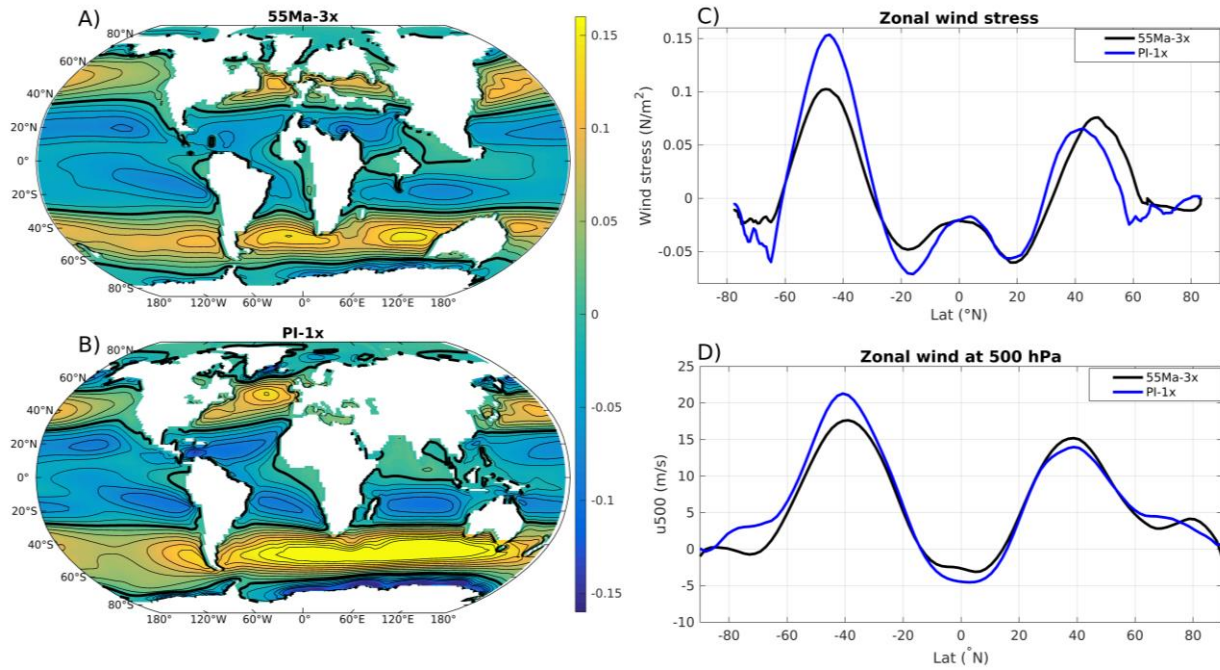


Figure 6. Zonal wind stress (in N/m^2) in the 55 Ma-3x (A) and the PI-1x simulation (B). (C) Zonally averaged zonal wind stress (in N/m^2) as a function of latitude in the 55 Ma-3x and PI-1x simulations. (D) Zonally averaged zonal wind (in m/s) at 500 hPa in the atmosphere as a function of latitude in the 55 Ma-3x and PI-1x simulations.

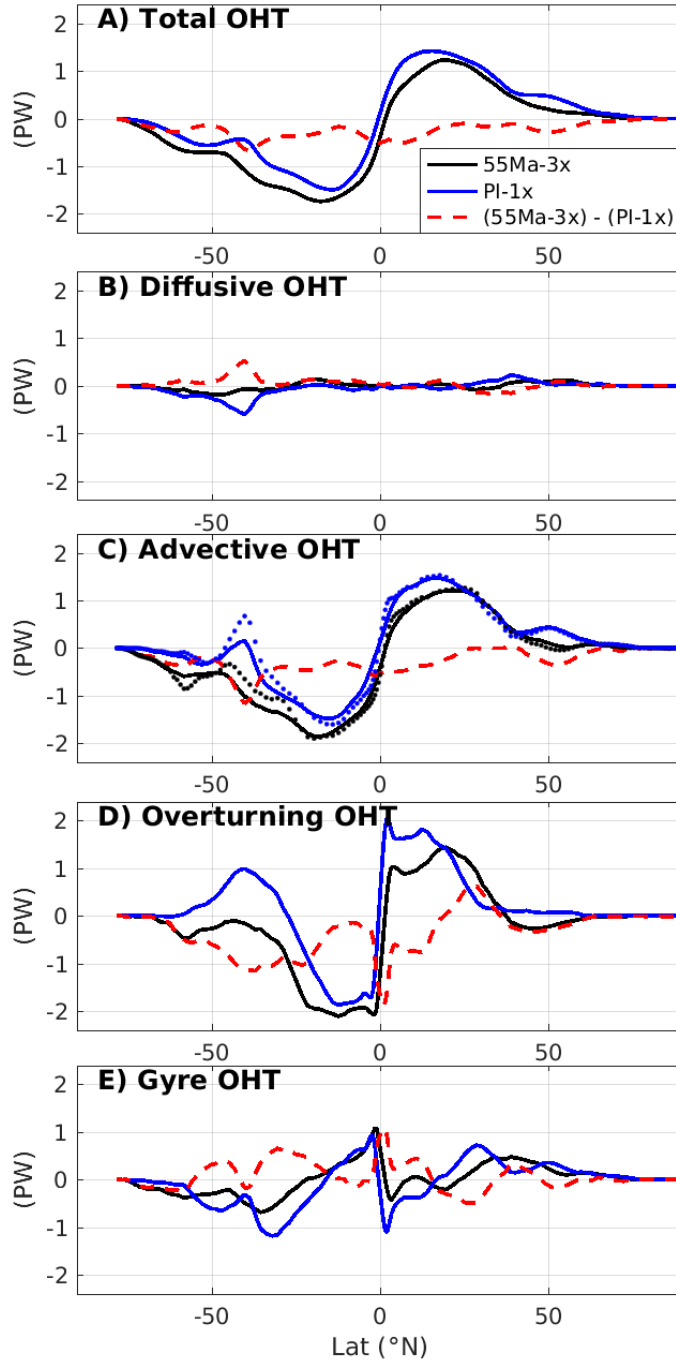


Figure 7. Meridional oceanic heat transport (in PW) as a function of latitude (positive contribution is northward) and its decomposition according to Eqs. 2 and 3. Results from the 55 Ma-3x and the PI-1x runs are shown in black and blue, respectively, and the difference between the two is in red. On panel C, the dotted lines indicate the sum of OHT_{MOC} and OHT_{gyre} estimated from monthly means, while the solid lines are computed at the model time step.

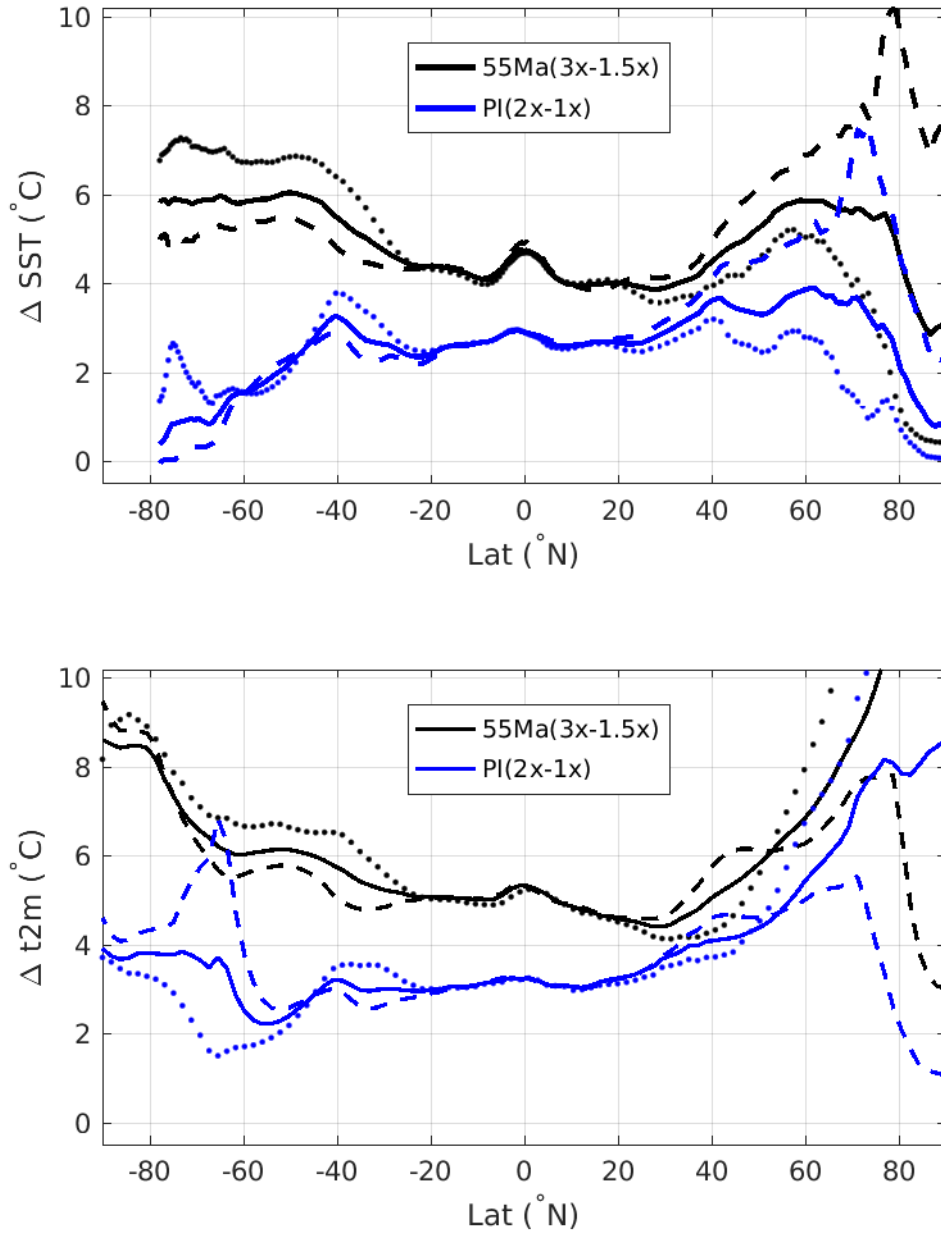


Figure 8. Difference of the zonally-averaged SST (top, in $^{\circ}\text{C}$) and air temperature at 2m (bottom, in $^{\circ}\text{C}$) as a function of latitude between the 55 Ma-3x and the 55 Ma-1.5x runs in black, and the PI-2x and PI-1x in blue. The solid line indicates the annual mean, the dashed line the mean over July-August-September, and the dotted line the mean over January-February-March.

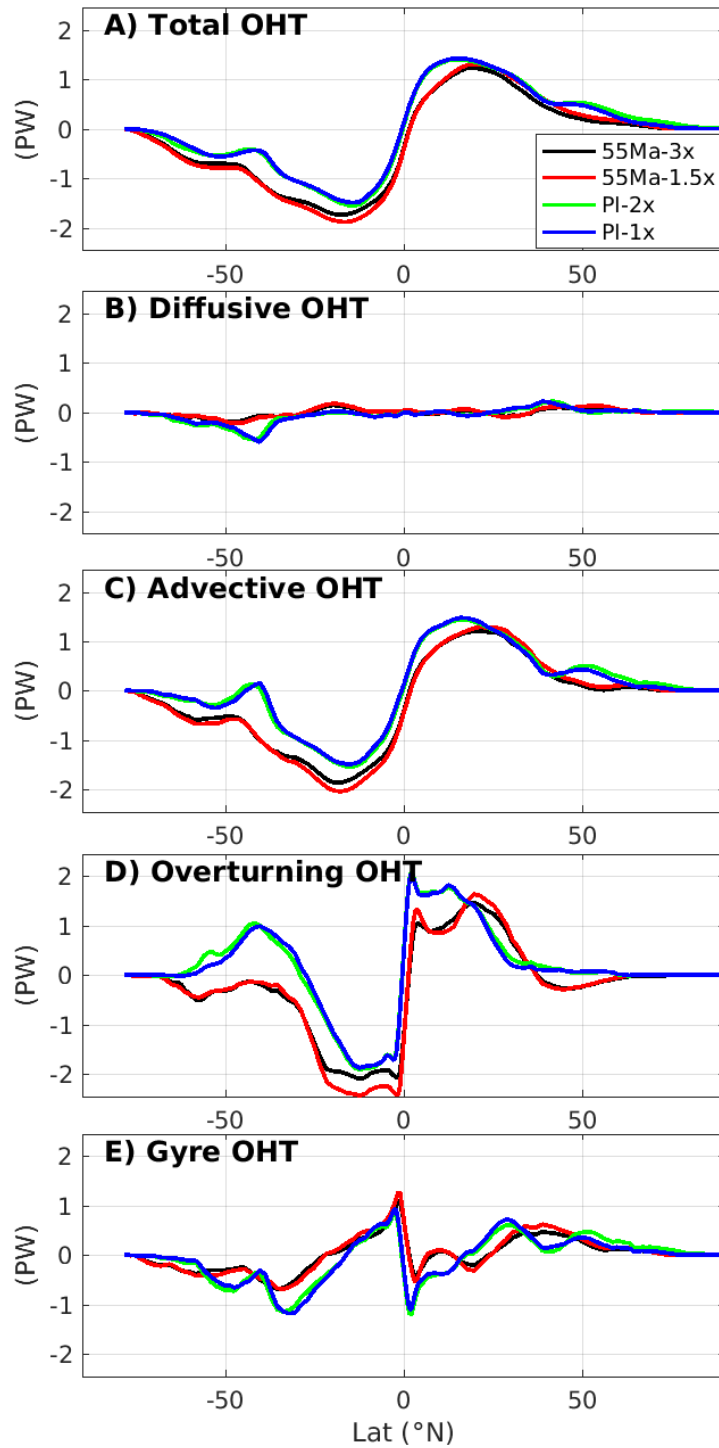


Figure 9. Meridional ocean heat transport (in PW) as a function of latitude (positive contribution is northward) and its decomposition according to Eqs. 2 and 3 in the 4 simulations.