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Fates of paleo Antarctic Bottom Water during the early Eocene — based on a Lagrangian analysis of IPSL-CM5A2 climate model simulations

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Key Points:

- A coupled climate simulation of the early Eocene shows a single intense overturning cell originating in the Southern Ocean.
- Lagrangian analysis of this paleo Antarctic Bottom Water reveals a major route to the surface mixed layer, with implications for carbon cycle.
- This upwelling mainly occurs in equatorial and tropical regions, in relation with the spatial pattern of the wind-driven Ekman pumping.

25 **Abstract**

26 Both deepwater formation and the obduction processes converting dense deepwater to lighter
27 surface water are the engine for the global meridional overturning circulation (MOC). Their spatio-
28 temporal variations effectively modify the ocean circulation and related carbon cycle, which
29 affects climate evolution throughout geological time. Using early-Eocene bathymetry and
30 enhanced atmospheric CO₂ concentration, the IPSL-CM5A2 climate model has simulated a well-
31 ventilated Southern Ocean associated with a strong anticlockwise MOC.

32 To trace the fates of these paleo Antarctic Bottom Water (paleo-AABW), we conducted
33 Lagrangian analyses using these IPSL-CM5A2 model results and tracking virtual particles released
34 at the lower limb of the MOC, defined as an initial section at 60°S below 1900m depth. Diagnostic
35 analysis of these particles trajectories reveals that most paleo-AABW circulates back to the
36 Southern Ocean through either the initial section (43%) or the section above (31%); the remaining
37 (>25%) crossing the base of the mixed layer mostly in tropical regions (up to half). The majority
38 of water parcels ending in the mixed layer experience negative density transformations, intensified
39 in the upper 500m and mostly occurring in tropical upwelling regions, with a spatial pattern
40 consistent with the wind-driven Ekman pumping, largely determined by the Eocene wind stress
41 and continental geometry.

42 In the same way as present-day North Atlantic Deep Water upwells in the Southern Ocean, our
43 results suggest that the strong tropical and equatorial upwelling during the Eocene provides an
44 efficient pathway from the abyss to the surface, but at much higher temperature, with potential
45 implications for the oceanic carbon cycle.

46

47 **Keywords:** the early Eocene, paleo Antarctic Bottom Water, Lagrangian analysis, deepwater
48 obduction, tropical upwelling, Ekman pumping, global carbon cycle

49 **1 Introduction**

50 The formation of dense deep and/or bottom water masses and their conversion into light
51 surface water act as an engine for the global meridional overturning circulation and characterizes
52 the interactions between the surface and abyssal (interior) ocean (e.g. Bullister et al., 2013;
53 Kuhlbrodt et al., 2007). Wind-driven upwelling in the Southern Ocean in present-day conditions
54 provides a pathway for North Atlantic Deep Water towards the thermocline and closes the Atlantic
55 Meridional Overturning Circulation (AMOC), forming the quasi-adiabatic pole-to-pole
56 overturning circulation (Anderson et al., 2009; Marshall & Speer, 2012; Wolfe & Cessi, 2011). Spatio-
57 temporal variations in deepwater formation and its obduction, such as intensity, location and
58 timescale, effectively modify the global ocean circulation that shapes the global climate on
59 multiple timescales (Burke & Robinson, 2012; Kuhlbrodt et al., 2007; Nof, 2000; Srokosz et al.,
60 2012; Toggweiler & Samuels, 1995). Ocean circulation carries a huge amount of heat and thus
61 largely alters global heat distribution (Ganachaud & Wunsch, 2000; Macdonald & Wunsch, 1996;
62 Trenberth & Caron, 2001). Meanwhile, long-term variations (on a scale of thousands to millions
63 of years) in deepwater dynamics modify the carbon storage capacity of the deep ocean and adjust
64 the carbon exchange between the deep-ocean carbon inventory and atmosphere, which plays an
65 important role in the evolution of the Earth's climate (Anderson et al., 2009; Broecker & Barker,
66 2007). For example, the variations in Southern Ocean upwelling, associated with related physico-
67 chemical and biogeochemical processes, have been found to regulate fluctuations of atmospheric
68 CO₂ and climate evolution across the late Pleistocene glacial cycles (Sigman & Boyle, 2000;
69 Watson & Naveira Garabato, 2006; Anderson et al., 2009; Burke & Robinson, 2012; Martínez-

70 Botí et al., 2015). Strengthened upwelling during deglaciations increases the exposure rate of the
71 deep-ocean, which can trigger a series of physico-chemical (e.g. solubility of seawater) and
72 biogeochemical responses (e.g. nutrient availability and export production) and enhance CO₂
73 release into the atmosphere. In contrast, weakened upwelling contributes to decreased carbon
74 release from the abyssal ocean, thereby facilitating the sequestration of atmospheric carbon
75 (Watson & Naveira Garabato, 2006; Anderson et al., 2009; Burke & Robinson. 2012; Martínez-
76 Botí et al., 2015). Understanding these changing interactions between the abyssal ocean's carbon
77 reservoir and surface ocean is thus of utmost importance for decoding climate evolution on various
78 timescales.

79 Paleoceanographic data has shown that interactions between the deep and surface ocean
80 have gone through multiple-scale changes over geological history in response to a variety of
81 forcings (e.g. Ferreira et al., 2018). Geological records indicate that the early Eocene was
82 characterized by an extremely warm climate (5–8°C warmer than the present day) and very active
83 global carbon cycle with enhanced exchanges among different components of the Earth system,
84 (Cramer et al., 2011; Dunkley Jones et al., 2013; Foster et al., 2017; Hollis et al., 2019; Pearson &
85 Palmer, 2000; Tripathi et al., 2003). Numerous CO₂-proxy data report high atmospheric CO₂
86 concentrations during the early Eocene (Breecker et al., 2010; Foster et al., 2017; Pagani et al.,
87 1999; Pearson & Palmer, 2000; Tripathi et al., 2003), which is undoubtedly a critical factor causing
88 this warm climate, if not the primary one. Although this enhanced atmospheric CO₂ level may
89 partially result from carbon input from the biosphere — such as in a response to irreversible
90 degradation of biomass as tropical ecosystems crossed a thermal threshold of 35°C for vegetation
91 (Huber, 2008) or as high-latitude permafrosts experienced oxidation of soil organic carbon
92 (DeConto et al., 2012; Kurtz et al., 2003) — the carbon cycle was definitely influenced by the

93 deep-ocean carbon reserves during the early Eocene. The deep ocean is a huge carbon reservoir,
94 and its re-organization during the early Eocene left an imprint in the global carbon cycle (e.g. John
95 et al., 2013, 2014). A plausible mechanism for this carbon perturbation remains elusive yet, leaving
96 our understanding of Eocene Earth system evolution incomplete.

97 Considering the crucial influence of the deep-ocean carbon reservoir on global carbon
98 cycle during the late Pleistocene glacial cycles (Anderson et al., 2009; Burke & Robinson, 2012),
99 carbon exchanges between the deep ocean and the surface may hold the key for understanding the
100 early Eocene climate system (John et al., 2013, 2014). Global acidification of the surface ocean
101 and widespread dissolution of deep-sea carbonates during the early Eocene also suggest a large-
102 scale reorganization of the oceanic carbon cycle system (John et al., 2013, 2014). Such large-scale
103 changes of carbon-related processes in the ocean may alter the carbon exchange between the
104 abyssal ocean carbon reservoir and the surface ocean, which likely link to a different early-Eocene
105 meridional ocean circulation. Within the framework of the Deep-Time Model Inter-comparison
106 Project (DeepMIP; Lunt et al., 2017), including 840 ppm atmospheric CO₂ (three times the pre-
107 industrial (PI) levels) and realistic Eocene bathymetry and other boundary conditions (e.g. no
108 Antarctic ice sheet), the IPSL-CM5A2 model has simulated a vigorous meridional overturning
109 circulation (MOC) originating in the Southern Ocean, as discussed in Zhang et al., (2020). This
110 MOC is sustained by deepwater formation occurring at high latitudes of the Southern Ocean, where
111 denser water sinks into the depths and fills the whole ocean basins. Given this very different
112 oceanic circulation and specificity of the early Eocene (regarding the climatic environment and the
113 global carbon cycle), it is of utmost interest to understand how these dense waters formed by deep
114 convection are transformed into light water returning to the surface thermocline, e.g. at what rate
115 they are transformed and where this process occurs.

116 One frequently used way to understand paleo-oceanic circulation for relatively recent
117 periods is to compare it with present-day conditions and to analyze circulation anomalies relative
118 to the PI condition. However, the fact that both bathymetry and MOC during the deep geological
119 time of Eocene (cf. Zhang et al., 2020) are so different from those of the present- day hinders this
120 direct comparison. With such a different MOC structure and in the absence of an Antarctic
121 Circumpolar Current, one can expect different thermodynamic processes and physical mechanisms
122 to close the global ocean circulation and associated transformation of deepwater into light surface
123 water during the early Eocene, which are still unknown yet. Through the diagnosis of virtual
124 particle trajectories, Lagrangian analysis provides an accessible way to investigate the movement
125 of dense deepwater and their density transformations, and to diagnose their interactions with the
126 surface thermocline in given three-dimensional, time-evolving velocity fields (e.g. Blanke et al.,
127 2002; Tamsitt et al., 2017; 2018). In this study, we applied such a Lagrangian analysis to
128 investigate the fate of the paleo-Antarctic Bottom Water (hereafter referred to as paleo-AABW,
129 for both bottom and deep waters in the absence of other deep water mass). Using the IPSL-CM5A2
130 climate model results for the early Eocene, this analysis allows to map the 3D circulation of the
131 paleo-AABW, and in particular how it upwells back towards the upper ocean. In order to better
132 understand the closure regime of the Eocene MOC in the model simulation, we specifically set out
133 to 1) map the general pathways of the paleo-AABW circulation and upwelling and 2) diagnose the
134 underlying dynamical and thermodynamic processes behind this upwelling and associated density
135 transformations. In addition, the results of this analysis can also extend our knowledge on the
136 closure of the MOC in a context that is quite different than the present day, especially in the
137 absence of a quasi-adiabatic pole-to-pole overturning circulation regime. We first introduce the
138 paleoclimate simulation that is used, then the details of the Lagrangian analysis (Section 2).

139 Section 3 and Section 4 present the Lagrangian results, and their discussion on and on implication
140 on Eocene global carbon cycle, respectively. Summary and conclusions are given in Section 5.

141 **2 Methods**

142 **2.1 The early-Eocene simulation and its evaluation**

143 The early-Eocene simulation used in this study has been discussed in Zhang et al. (2020)
144 with detailed information on boundary condition and on simulation setup, so we only provide a
145 brief summary of the simulation hereafter. The simulation was performed with the IPSL-CM5A2
146 climate model and set up with the boundary conditions reconstructed for the early Eocene (~55Ma),
147 following the DeepMIP protocol (Lunt et al., 2017). The IPSL-CM5A2 is a new version of IPSL-
148 CM5A: an Earth system climate model comprising the LMDz atmospheric model, and the NEMO
149 ocean model with a spatial resolution of $\sim 2^\circ$, coupled to the LIM3 dynamical sea ice model through
150 the OASIS coupler (Dufresne et al., 2013; Hourdin et al., 2013; Sepulchre et al., 2020). The NEMO
151 ocean model has 31 vertical levels with a finer resolution near the surface (10m) than in the abyss
152 (~ 500 m), and partial steps were used for the level directly above seafloor to better represent the
153 bathymetry. The previous version IPSL-CM5A has been extensively used in various contexts, such
154 as the Last Glacial Maximum, the mid-Holocene (Kageyama et al., 2013), and the Pliocene
155 (Contoux et al 2013; Tan et al 2017, 2020). The new version IPSL-CM5A2, recently developed
156 by Sepulchre et al. (2020) for faster computation, is dedicated to the analysis of deep time history
157 of the Earth system. For instance, this new version model has been used to simulated Pliocene
158 (Tan et al. 2018, 2020) and Eocene climate (Ladant et al., 2014; Zhang et al., 2020), in which plate
159 tectonic forces, such as changes in continents geography and seafloor bathymetry, are permitted.
160 The boundary conditions of this early-Eocene simulation include an atmospheric CO₂
161 concentration of three times that of PI levels (840 ppm), bathymetry and topography reconstructed

162 for the early Eocene (Herold et al., 2014), and prescribed tidally-induced mixing (Green & Huber,
163 2013). The simulation was initialized by following the DeepMIP protocol, i.e. the ocean was
164 initialized from zonally symmetric ocean temperature and a constant salinity of 34.7 psu (Lunt et
165 al., 2017). The simulation was run for 4000 years until the ocean reached thermodynamic
166 equilibrium, with the final trend of deep-ocean temperature changes lower than 0.05°C per century
167 (Zhang et al., 2020). Monthly-averaged velocities (and ocean properties), computed over the last
168 100-yr of the simulation to represent a climatological year, were used to conduct the Lagrangian
169 analyses of the present study. The mesoscale eddy-induced turbulence was parameterized as eddy-
170 induced bolus velocity (Gent & McWilliams, 1990), and these have been explicitly added to the
171 total velocity used in our Lagrangian analysis. This eddy-induced bolus velocity is generally small,
172 with a maximum magnitude of 2 cm/s over regions of low latitude.

173 The early-Eocene simulation used in this study has been systematically discussed in the
174 context of boundary conditions and detailed ocean circulation, especially in the light of comparison
175 with the present-day situation and also with different CO₂ levels in Zhang et al. (2020). Here we
176 only describe it briefly, regarding (well-concerned) sea surface temperature (SST) and ocean
177 circulation. The simulation reproduces a global mean SST of 28°C, with the annual mean SST
178 varying from 10–15°C in the southern-most Southern Ocean to 30–37°C near the Equator (Fig. 1).
179 The zonally-averaged annual mean SSTs were overall ~10°C warmer than in the present-day
180 simulation (PI-1x, performed with the IPSL-CM5A2, but with the present-day boundary
181 condition), with the largest differences of 12°C found in the Southern Ocean. Wide seasonal
182 variations further extended the highest summer temperature in southern high latitudes to 25°C (Fig.
183 1), reached in the southeast Atlantic-India basin around Australia (Fig. 2a of Zhang et al., 2020).
184 This overall warmth and enhanced Southern Ocean temperature bring them generally close to the

185 proxy reconstruction despite some exceptions. The mismatched points mainly include the Tex⁸⁶
186 biomarker records from the Southern Ocean at latitude 60°S that suggest the same temperature
187 (30°C) as the Equator (Fig. 1), which nevertheless has been very likely overestimated (Hollis et
188 al., 2019). Zhang et al. (2020) compared this simulated temperature site-by-site with the Eocene
189 temperature records from the most comprehensive temperature dataset in Hollis et al. (2019). The
190 model-data comparisons suggest that they are generally compatible, although certain differences
191 can be seen for some specific proxy data points. For instance, proxy-based temperature data and
192 simulated temperature show better consistency in the Atlantic-Indian basin than in the Pacific
193 where the model produced more homogenous temperatures, in contrast to large spatial variability
194 suggested by the proxy data (Fig.2a & S2A of Zhang et al., 2020). Zhang et al. (2020) carried out
195 extensive model-data comparisons using statistical methods, such as the root mean square
196 deviation (RMSD) and benchmarks following Kennedy-Asser et al. (2019) method. These
197 comparisons suggest that the performance of the model simulation is generally satisfactory,
198 because the data-model RMSD was of the same order of magnitude as the uncertainty of proxy-
199 based SST estimates, and the simulation was able to capture the global mean temperature and some
200 latitudinal gradient patterns from the benchmark analysis.

201 Models simulations for the Eocene usually could not be able to simulate the much reduced
202 meridional temperature gradient suggested by proxy data (the “equable climate problem”, Huber
203 & Caballero 2011, and references herein). Regarding this long-standing issue, the early Eocene
204 IPSL simulation produces a SST gradient from the Equator to the Southern Ocean of ~24°C in the
205 annual mean and 18°C in February (Fig. 1), leading to an only slightly reduced gradient compared
206 with the present-day 28°C. Comparison with other model results suggests that the meridional
207 temperature gradient in the IPSL-CM5A2 simulation is located in the upper end of the multi-model

208 range (e.g. Lunt et al., 2012; 2020; Hutchinson et al., 2018). Nevertheless, more recent studies
209 appear to suggest a mitigated model-data discrepancy (Cramwinckel et al., 2018; Lunt 2020).
210 Cramwinckel et al. (2018) recently updated the Eocene equatorial SSTs estimation with new proxy,
211 that reduce the temperature gradient to about 20–22°C, and Lunt et al. (2020) found similar values
212 in recent DeepMIP simulations. Reasons for the remaining mismatches, especially on a regional
213 scale, perhaps involve either model simulations, which may still underestimate some regional scale
214 variations, or proxy-based temperature reconstruction, because they can be scrambled by
215 vertically-varying processes in the water column (e.g. Ho & Laepple, 2016) and by seasonality
216 signals over high latitudes (e.g. Davies et al., 2019). Overall, despite sizeable uncertainties, the
217 Eocene simulation is nonetheless able to capture the basic temperature pattern suggested by most
218 proxy records and previous model studies.

219 This paragraph will briefly evaluate the simulated Eocene oceanic circulation in
220 comparison with other models' results and with proxy-based oceanic circulation reconstruction of
221 the Eocene. The simulation shows a well-ventilated Southern Ocean and a single anticlockwise
222 global overturning cell, with bottom/deepwater formation at high latitudes of the Southern Ocean
223 where dense paleo-AABW sinks into the depths and fills the whole ocean basins (Figure 2A,
224 redrawn from Zhang et al., 2020). This single MOC structure has been well investigated, in
225 relation with winter convection and deep-water formation, in comparison with the PI circulation,
226 and shown to be robust to different atmospheric CO₂ levels (Zhang et al., 2020). Although the
227 magnitude of the MOC can vary in some model results, this well-ventilated Southern Ocean and
228 anticlockwise Southern Ocean MOC pattern are common features of most model results for the
229 early Eocene or middle-to-late Eocene before the Oligocene transition (e.g. Baatsen et al., 2018;
230 Lunt et al., 2010). One exception is a simulation of the GFDL model with 38 Ma bathymetry and

231 boundary conditions, which produces a two-cell MOC with deep water formed in both the
232 Southern Ocean and the North Pacific (Hutchinson et al., 2018). The presence of deepwater
233 formation in the North Pacific in this GFDL simulation (in contrast with the IPSL-CM5A2
234 simulation used herein and several others) is probably due to its low topography over North
235 America that induces less rainfall over the West Pacific, leading to higher surface density and
236 facilitating deepwater formation (Zhang et al., 2020), as illustrated by sensitivity studies on
237 topography effects (Maffre et al., 2018; Schmittner et al., 2011). Apart from these modelling
238 studies, the MOC has been inferred from neodymium (Nd) isotope distributions, a proxy for ocean
239 circulation (Thomas et al., 2003; Thomas et al., 2014). The common features between the
240 simulated and proxy-reconstructed MOC are the well-ventilated Southern Ocean and the
241 predominant Southern Ocean origin of the bottom waters (Thomas et al., 2014). Differences
242 remain in the North Pacific where the ocean is less ventilated in the simulation than as suggested
243 by the proxy data, whereas the Nd-based results suggest that the water of North Pacific origin
244 never crossed the Equator. Overall, the well-ventilated Southern Ocean filling the most of the
245 global bottom waters is supported by previous modelling and proxy reconstructions. Further
246 evaluations of the simulation have been addressed in Zhang et al. (2020).

247 **2.2 Lagrangian experiments and analysis**

248 **2.2.1 Lagrangian analysis and Ariane tool**

249 Lagrangian analysis of oceanic model outputs provides an interesting and comprehensive
250 way to investigate the three-dimensional, time-evolving transport fields through a diagnosis of
251 trajectories of virtual particles (Bower et al., 2019; van Sebille et al., 2018; Tamsitt et al., 2018).
252 As a useful and complementary tool to Eulerian approaches, Lagrangian analysis can help answer
253 a range of theoretical and practical questions, as reviewed in Griffa et al. (2007) and more recently
254 in van Sebille et al. (2018) and. In particular, Lagrangian approach has been widely employed to

255 solve problems involving connectivity between oceanic regions such as the bottom and the surface
256 layers, and used to track the movement and pathways of various water masses (e.g. Blanke et al.,
257 1999; 2002; Döös et al., 2008; van Sebille et al., 2013; Tamsitt et al., 2017, 2018). Here, we use
258 the Lagrangian analysis to provide comprehensive information on the ocean circulation during the
259 early Eocene, when the bathymetry was tremendously different from the present day. For instance,
260 the much narrower North Atlantic limits deep convection there and the almost closed Drake
261 passage hinders ACC during the early Eocene, which makes the MOC structure very different
262 from the present day. In addition, divergences of flow within individual basins (Atlantic, Pacific
263 and Indian oceans) through tropical straits break the mass conservation of each basin and decrease
264 the value of Eulerian-velocity-inferred MOC streamfunction of individual basins for indicating the
265 overall circulation, which makes the separation of each basin's transport from the total global
266 transport impossible using only a typical Eulerian approach.

267 Our Lagrangian analysis was carried out using the Ariane tool that is based on an off-line
268 mass-preserving algorithm to compute particles trajectory (<http://www.univ-brest.fr/lpo/ariane>). A
269 detailed description of the method used to follow particles in an Ocean General Circulation Model
270 (OGCM) is given in the appendix of Blanke and Raynaud (1997). The Ariane tool provides a
271 “quantitative” mode that is specifically designed to effectively evaluate water mass transport, by
272 tracking trajectories of particles between an initial section and the other sections which close a
273 three-dimensional oceanic region. Sections can be defined in various ways: geographically
274 (vertical and horizontal cross-sections), or in more complex manners, as for example time-
275 dependent isolines of a chosen quantity (isopycnals, mixed layer). Particles are automatically
276 placed by Ariane software tool on each model grid cell of a specified initial section, by considering
277 only water masses flowing inward the closed domain in forward integration (outward the domain

278 in backward integration), and a given water transport is accordingly attached to those particles.
279 Each particle represents a fraction of the total transport through the initial section. Because the
280 transport across the section is not uniform, the water mass transports assigned to particles are
281 different. The algorithm of this transport assignment is that: each grid cell with a given transport
282 T_n is described with N_n particles, and this integer particle number N_n satisfies (Blanke et al., 1999):

$$283 \quad \frac{T_n}{N_n} \leq T_0 \quad (1)$$

284 where T_0 is the prescribed maximum transport associated with a single particle. This assignment
285 is made for each time step of the oceanic model data input of Ariane and during a period of time
286 defined for an experiment. The total number of particles is the sum of the N_n over all relevant grid
287 cells of the initial section. Because of the low resolution of the ocean model and the associated
288 rather linear dynamics, we have used a large value for T_0 (0.8 Sv) such that a single particle is
289 released in each grid cell of the initial section at each time step. The setting of this initial section
290 (i.e. velocity fields are sampled only on the initial section) implies that particles should be released
291 everywhere to sample the complete velocity fields of the whole ocean (Döös et al., 1995; Blanke
292 et al., 1999; Döös et al., 2008).

293 Subsequently, each particle is tracked over each time step and its trajectory is computed
294 according to the evolving velocity field, until it reaches any one of the specified receptor sections
295 (Blanke & Raynaud, 1997; Döös & Webb, 1994). Along its path, particle velocity, temperature
296 and salinity evolve over the time integration, according to the local Eulerian model fields (Blanke
297 & Raynaud, 1997; Döös & Webb, 1994). Each particle can be seen as a water parcel evolving over
298 time, and the terms particle and water parcel are interchangeably used in this sense. Ultimately,
299 statistical analyses of all these trajectories represent how the water masses are transported from
300 this initial section to another destination. These inter-section transports can be portrayed by means

301 of a Lagrangian streamfunction (calculation given below), by taking advantage of a mass-
302 preserving scheme (Blanke et al., 1999; Blanke & Raynaud, 1997; Döös, 1995).

303 **2.2.2 Lagrangian experiment setup**

304 In the present study, we employ the Ariane tool to trace the fate of the paleo-AABWs and
305 to analyze the dynamical processes contributing to the upwelling along their trajectories. Based on
306 the overall features of the ocean circulation, the initial section (the snapshot) was set in the deep
307 ocean at latitudes 60°S below 1900 m to sample the northward flowing bottom/deepwater export.
308 The Eulerian-velocity-inferred meridional streamfunction (Eulerian-MOC) in Figure 2A shows
309 the anticlockwise meridional circulation in the early-Eocene simulation. Surface water loses
310 buoyancy through ocean-atmosphere interaction over the high-latitude Southern Ocean and forms
311 denser water sinking into the depths, then flowing northward in the abyssal ocean and filling the
312 whole ocean basins (Zhang et al., 2020). Therefore, there is a strong net northward export of deep
313 water over the lower (deep) limb of the MOC that is supplied by the deepwater formation in the
314 Southern Ocean and the overall southward transport over the upper ocean. The Eulerian-MOC
315 maximum intensity (up to 40 Sv) is reached at 60°S at a depth ~1900 m, at which latitude and
316 depth the initial section was accordingly set. Zooming into the vertical section at 60°S, the majority
317 of gross northward transport occurs in the abyssal ocean (e.g. northward transport is greater than
318 10 Sv at depths of 1800–4500 m), whereas the southward transport mainly occurs at shallow depths
319 (e.g. 10 Sv for depths of 500–2500 m) (Fig. 2B & 2C). This asynchronous transport over depths
320 closes up the anticlockwise MOC, with a boundary between the upper southward and lower
321 northward transport of ~1900 m (longitudinally-averaged). Correspondingly, the initial section at
322 60°S was constrained to depths below 1900 m in the vertical direction. Therefore, this initial
323 section was chosen to sample the northward-flowing paleo-AABW mass over the lower limb of
324 the anticlockwise meridional circulation.

325 To trace the destinations of the paleo-AABW, 4797 virtual particles were released at the
326 initial section (at 60°S below the depth of 1900 m), with T_0 value of 0.8 Sv. Particles were released
327 at the middle of each month throughout the first year, according to the monthly evolving velocity
328 fields (only at grid cells where the flow was northward for the given month). The large number of
329 particles leads to a relatively small transport assigned to each particle, ensuring precise
330 representation of the water mass transport. These particles provided a total estimate of 81.0 Sv for
331 the gross northward-flowing bottom/deepwater transport over the lower limb of the MOC, which
332 reproduces the Eulerian description of the northward transport very well (Fig. 2C). This
333 Lagrangian-based transport estimate was almost twice that of the Eulerian-velocity-inferred MOC
334 value that is the net of two directions flow in the same section. This indicates that half of the
335 northward transports were compensated by their opposing southward transport and the other half
336 left the initial section and carried on towards the north.

337 These particles 3D trajectories were integrated for 4000 years with monthly updated
338 velocity fields until being intercepted by any one of the receptor sections. These receptor sections
339 include the initial section itself, the vertical section above this initial section (60Sab) and the
340 monthly evolving mixed-layer base (MLB), separating the surface homogeneous thermal layer
341 from the deeper stratified ocean below, as shown in Figure 2A. The mixed layer depth is defined
342 by a potential density difference smaller than 0.3 kg/m^3 with reference to the surface. It has been
343 found that the trajectories at eddies scale and small spatial scales were sensitive to the increasing
344 temporal resolution ranging from days to month, while at scales larger the trajectories were largely
345 independent of the temporal resolution increase from day to month (Iudicone et al., 2002; Qin et
346 al., 2014). Given its $\sim 2^\circ$ spatial resolution, the ocean model used in the IPSL-CM5A2 simulation
347 does not resolve mesoscale eddies (they are parameterized) and the monthly resolution used in the

348 present study is sufficient. Note that the time resolution refers here to the frequency at which the
349 velocity field is updated. Every year, the same climatological monthly velocities are repeated (i.e.
350 after December the velocity fields return to the January), with activated consecutively looped
351 output over time. The Lagrangian experiment was run for 4000 years to obtain the long-term
352 trajectories of particles. This long-term experiment allowed 99% particles to be intercepted at one
353 of the specific receptor sections. The terms of initial and final refer to the moments when particles
354 are released from the initial section and are received by one of the receptor sections, respectively.
355 Below, we analyze these particles statistically in terms of water-mass transport and their
356 corresponding evolution of water properties along their trajectories. In addition to this main
357 experiment, an auxiliary Lagrangian experiment was conducted in a backward mode, in order to
358 trace back the geographical origins, defined as the last point of contact with the surface mixed
359 layer, of these paleo-AABW.

360 **2.2.3 Lagrangian diagnostics**

361 Two types of diagnostic analysis of the Lagrangian experiment were carried out:
362 Lagrangian streamfunction and transport-weighted particle distribution.

363 Lagrangian streamfunction portrays pathways of water mass on vertical or horizontal
364 directions, by taking advantage of mass conservation (i.e. the corresponding velocity field is non-
365 divergent) (Döös, 1995; Blanke & Raynaud 1997; Blanke et al., 1999, 2006). The calculation of
366 Lagrangian streamfunction can be divided into three steps (Blanke et al., 1999, 2006; Döös et al.,
367 2008): 1) obtain a three-dimensional transport field that corresponds to the flow of the water mass
368 within the domain of integration of the trajectories. As the local three-dimensional non-divergence
369 of the flow is exactly preserved, each particle entering one model grid cell through one of its six
370 faces has to leave it, and the transport field satisfies

371
$$\partial_i T_x + \partial_j T_y + \partial_k T_z = 0 \quad (2)$$

372 where T_x , T_y , and T_z depict the transports in the three directions, and where i , j , and k refers to
 373 the grid index. 2) obtain trajectories of individual particles by computing the three-dimensional
 374 streamlines of the velocity, and achieve the transport fields on the model metrics by algebraically
 375 summing the individual particles on each junction of two cells (i.e. recording particles when
 376 trajectory crosses grid cell boundaries every time). 3) compute the streamfunctions by integrating
 377 these transport fields along the vertical or zonal direction, which is similar to the Eulerian
 378 streamfunction calculation. Accordingly, the meridional $\Psi_{j,k}^{LZ}$ and horizontal $\Psi_{i,j}^{LH}$ streamfunctions
 379 are defined as

380
$$\Psi_{j,k}^{LZ} - \Psi_{j,k-1}^{LZ} = - \sum_i \sum_n T_{i,j,k,n}^y \quad (3)$$

381
$$\Psi_{i,j}^{LH} - \Psi_{i-1,j}^{LH} = \sum_k \sum_n T_{i,j,k,n}^y \quad (4)$$

383 Contours of $\Psi_{j,k}^{LZ}$ and $\Psi_{i,j}^{LH}$ depict pathways of water mass movement projected onto the meridional
 384 (latitude-depth) and horizontal (longitude-latitude) directions respectively. Northward and
 385 eastward movements are counted positive, while southward and westward movements are counted
 386 negative. One advantage of using the Lagrangian streamfunction is that it can then be decomposed
 387 into partial streamfunctions, computed from trajectories of particles received by a relevant section
 388 (Blanke et al., 1999; Döös et al., 2008). The partial Lagrangian streamfunction is obtained by
 389 summing only the particle trajectories intersecting by a selected final section.

390 To obtain statistical patterns from an ensemble of particles, we use the transport-weighted
 391 particle distribution for any variable tagged along the particle trajectories, such as density, potential
 392 temperature and salinity. This transport distribution is roughly equivalent to the probability density
 393 function (as defined in Eq 1 of Tamsitt et al., 2018), but is weighted by the volume transport of
 394 each particle and without any standardization. It is numerically calculated by binning particles and

395 summing their transport in each bin of a given space. The transport distribution in χ space is
396 calculated as

$$397 \quad P(\chi) = \sum_{i=1}^N \xi_i T_i; \quad \xi_i = \begin{cases} 1, & \text{if } \chi - \frac{\delta}{2} < \chi_i \leq \chi + \frac{\delta}{2} \\ 0, & \text{otherwise} \end{cases} \quad (5)$$

398 where δ is the bin width, N is the total number of particles, and the χ_i is the property of the i^{th}
399 particle. In the following analysis, we replace χ with density, potential temperature, salinity, and
400 depth to analyze particles distribution at the different moments, such as when they are released
401 from and intercepted by a section. For instance, transport distributions of different fates when they
402 are released reveal their tendency to different fates, and the transport distribution differences
403 between the initial and final sections depict transformation/conversion of relevant quantities that
404 particles have gone through along their trajectories.

405 **2.3 Testing the Lagrangian analysis on the more constrained present-day conditions**

406 We applied the same Lagrangian analysis framework to present-day conditions (a more
407 constrained framework, and much better known circulation) and compared the results with our
408 current understanding to validate our analysis. For present-day conditions, the meridional
409 circulation is characterized by a two-cell structure with deep water formed both in the North
410 Atlantic and Southern oceans (e.g. Lumpkin & Speer, 2007). Accordingly, we carried out two
411 Lagrangian experiments by releasing particles from two initial sections, i.e. initial-N (at 56°N
412 below 750m) and initial-S (at 65°S below 1400m), at the lower limb of the MOC to trace the fates
413 of the North Atlantic Deep Water and Antarctic Bottom Water respectively (Fig. 3A). Particles
414 were released from the initial sections Equatorward and tracked until they were intercepted by one
415 of the receptor sections that include the initial sections, the sections above (above-S and above-N)
416 and the MLB, following the same approach as for the early Eocene (as described above).

417 Our investigation examines the North Atlantic Deep Water (NADW) and Antarctic Bottom
418 Water (AABW) in turn. For each, we first introduce the Lagrangian experiments and then present
419 their results for each deepwater mass, before comparing with our understanding of present-day
420 conditions. For the NADW, 2931 particles sample a total southward transport of 17.0 Sv of
421 deepwater at the initial North Atlantic section at 56°N, of which 7.7 Sv went back to the North
422 Atlantic region, either through the initial section itself or the section above (i.e. the above-N section
423 in Fig. 3A). The remaining 9.2 Sv of water (more than half) leaves the North Atlantic and
424 eventually crosses the MLB entering the surface thermal layer globally. The final locations where
425 these NADW enter the mixed layer are shown in Figure 3B: A large part of NADW enters the
426 thermocline in the Southern Ocean scattered over the Atlantic, Indian and Pacific sectors. The
427 transport distribution of paleo-AABW as a function of latitude (Fig. 3D, calculated from eq.5)
428 shows that the majority (6 Sv out of 9.2 Sv) came to the surface over the southern high latitudes,
429 through the well-known wind-driven upwelling in the Southern Ocean. Around 3 Sv of particles
430 reached the surface ocean over other regions, including tropical upwelling (mainly along eastern
431 boundaries) and the ventilation in the North Atlantic and few in the North Pacific. These
432 Lagrangian-experiment results on the fates of the NADW highlighted the role of the upwelling
433 branch of the MOC in the Southern Ocean, and fit our understanding of global circulations well.
434 Numerous studies, from multiple perspectives, have highlighted the role of Southern Ocean
435 upwelling, which is driven by strong westerlies across all longitudes, in closing the present-day
436 MOC (Blanke et al., 2002; Garzoli & Matano, 2011; Marshall & Speer, 2012; Wolfe & Cessi,
437 2014; Rhein et al., 2014). For instance, the inter-hemisphere pole-to-pole meridional circulation
438 has been illustrated from a physical dynamics perspective (e.g. Toggweiler & Samuels, 1998;
439 Wolfe & Cessi, 2011). Marshall and Speer (2012) further demonstrated that the Atlantic deep

440 water is brought to the surface via the Southern Ocean upwelling through inter-hemisphere
441 overturning circulation, based on observation data and numerical experiments. Using the
442 Lagrangian analysis, Blanke et al. (2002) showed that the Southern Ocean is an important
443 receptacle of deep water. More recently, Tamsitt et al. (2017) demonstrated that the northern-
444 sourced deep water (i.e. NADW) of the three ocean basins spiraling southeastward and upward
445 through the ACC, with enhanced upwelling at major topographic features. Rhein et al. (2014) used
446 Atlantic CFC (Chlorofluorocarbon) observations to examine the spreading velocity and pathways
447 of Labrador Seawater and overflow water from Denmark Strait, and found that Deep Western
448 Boundary Current is the fast pathway reaching the Southern Ocean.

449 For AABW, the total amount of 34.5 Sv northward-flowing water was captured in the
450 lower limb of the MOC, which is much larger than the net northward transport of ~ 10 Sv shown
451 by the Eulerian MOC streamfunction. This large amount of northward transport can be attributed
452 to deepwater formed locally that leads to compensation in the wide Southern Ocean basin (as in
453 MOC streamfunction). Of that amount of northward deepwater, 7.5 Sv returned to the initial
454 section, and the majority, up to 27 Sv, crossed the MLB and came out into the surface ocean. The
455 final locations where these particles cross the MLB are restricted to the Southern Ocean with the
456 most northern position of 40°S . This may be in line with an explanation that AABW is isolated
457 from surface ocean by the overlying NADW and it mixes with the southward NADW in the interior
458 (Marshall & Speer, 2012; Orsi et al., 1999). The southern component of this mixed upwelled water
459 mass goes almost immediately back into the deep ocean, forming the recirculation of water
460 between the Antarctic water masses and Circumpolar Deep Water (Gordon, 1971; Orsi et al., 1999;
461 Toggweiler et al., 2006; Johnson, 2008; Tamsitt et al., 2017). Therefore, the results of our
462 lagrangian analysis for the fate of NADW and AABW are in good agreement with the actual

463 understanding of the global circulation under present-day conditions, confirming the interest to
464 apply such Lagrangian analysis to the Eocene circulation.

465 **3 Results**

466 **3.1 Diagnostic analysis of the paleo-AABW source and of the initial state**

467 A diagnostic analysis of the source supplying the paleo-AABW mass at the lower limb of
468 the MOC can illustrate how the northward export of paleo-AABW is sustained, bringing valuable
469 insight into paleo-AABW dynamics. Further, probability distribution analysis of the initial state
470 of the water mass properties, such as temperature, salinity and density (at the initial section), can
471 assess the representativeness of virtual particles and record the starting point of these properties.

472 These paleo-AABW exports are largely supplied by the ocean subduction process
473 associated with winter ventilation over the high-latitude Southern Ocean. The auxiliary backward
474 experiment provided an estimate of 81.0 Sv for the total gross amount of water crossing the initial
475 section, of which 49.8 Sv originated from the surface ocean by crossing the MLB over the Southern
476 Ocean (Table 1). This total gross transport is exactly the same as the water transport from the initial
477 section in the main Lagrangian experiment (i.e. tracing the fates of paleo-AABW). The statistical
478 analysis on geographical locations of the last contact point reveals that the Atlantic-Indian sector
479 of the Southern Ocean are the main sources of paleo-AABW (Table 1), providing more than 80%.
480 The rest of the paleo-AABW parcels stem from the Pacific basin. This origin-tracking analysis
481 also suggests that the majority of deepwater formation is supplied by local processes, with deep
482 convection, deepwater formation and northward export occurring in the same basin. The only
483 exception is the Atlantic sector of the Southern Ocean, because the denser water of Atlantic origin
484 is partly exported eastward to the Indian Ocean sector due to the lack of bathymetry barriers.

485 Regarding the timescale, the occurrence of this ventilation is relatively fast, since 80% of AABW
486 originating from the mixed layer have completed their journey to 50°S within 20 years.

487 The Lagrangian experiment estimates paleo-AABW transport very well compared with the
488 Eulerian description (Fig. 2C), confirming the adequate particle sampling strategy. A large part of
489 this northward-flowing deep water starts its journey in the Atlantic-Indian sector of the Southern
490 Ocean (Table 1 & Fig. 2B). Within each basin, there is a clear west-east pattern with the majority
491 of deepwater parcels leaving the initial section from the west side of the basin (Fig. 2B), within
492 western boundary currents. This west-east pattern across the basin is consistent with the clockwise
493 subpolar gyre circulation in the horizontal plane, as shown by the Eulerian barotropic
494 streamfunction (Zhang et al., 2020).

495 There are no distinct differences in potential density (σ_0) of the northward-flowing
496 water masses among the three basins. The potential temperature-salinity (θ -S) diagram of water
497 mass shows that the potential density anomalies of the paleo-AABW are in the range of 26.6–27
498 kg/m^3 when leaving the initial section (Fig. 2D); these values are comparable to the typical value
499 of the interior ocean of 26.5 kg/m^3 (bulk volume of seawater) of the early-Eocene simulation. The
500 density differences among the three basins are relatively small (i.e. within 0.5 kg/m^3), due to the
501 compensating effect of changes in temperature and salinity. For instance, the Pacific paleo-AABW
502 with lower temperature (by 1.5°C) and lower salinity (by 0.2–0.3 psu) is less than 0.2 kg/m^3 lighter
503 than the paleo-AABW from the Atlantic sector. The Indian sector paleo-AABW shows no density
504 differences from the deepwater mass of the Atlantic basin, with an average of $\sim 26.7 \text{ kg/m}^3$.
505 Regarding PI conditions, the particles represent the southward flowing water mass for the NADW
506 and northward water mass for AABW. In present-day conditions, the density of NADW and

507 AABW is in the range of 27.5–27.7 kg/m³ in general, which is denser than the Eocene much
508 warmer deepwater mass.

509 **3.2 Fates of paleo-AABW and their pathway**

510 **3.2.1 Two fates of paleo-AABW**

511 Northward-flowing deepwater mass over the lower limb of the MOC shows two fates: 1)
512 entering into the surface ocean by crossing the MLB; 2) returning to the Southern Ocean across
513 60°S, either through the initial section (below 1900 m depth) or the section above (the 60Sab
514 section). One way to quantitatively elucidate the fate of paleo-AABW mass is to statistically
515 analyze the initial and final position of these water parcels. As shown in Table 1, the statistical
516 results indicate that more than a quarter of paleo-AABW parcels enter the surface ocean by
517 crossing the MLB, with an average travel time around 700 years (Fig. S2); nearly half return to
518 the initial section within 100 years; and the rest are intercepted by the 60Sab section with a travel
519 time of 300 years.

520 Another way to illustrate the fates of this paleo-AABW mass is to calculate the meridional
521 Lagrangian streamfunction from these water parcels all along their trajectory, which gives more
522 information on their temporal evolution along their trajectories in addition to the initial and final
523 stage. It is worth to notice that this meridional Lagrangian streamfunction represents only a fraction
524 (i.e. the water flow originating from paleo-AABW) of the Eulerian-velocity-inferred MOC
525 streamfunction, because water particles are released only from the lower limb of the MOC in our
526 Lagrangian experiment. Theoretically, taking a full-range of particles released from all grid cells
527 in the full-domain simulation into consideration will reproduce a Lagrangian streamfunction that
528 is exactly the same as the Eulerian-MOC streamfunction (Blanke et al., 1999; Blanke & Raynaud,
529 1997; Speich et al., 2001). The Lagrangian MOC streamfunction for all particles (Figure 4A,

530 calculated from eq. 3) suggests that around 40 Sv of water, indicated by the difference between
531 the bottom and the top of the section, was transported away from the initial section towards the
532 north. Along its journey northward, more and more water is transported upward, and either enters
533 the surface ocean after crossing the MLB or changes direction to the south to eventually reach the
534 60Sab section. Similarly, the total streamfunction (Fig. 4) shows that more than 15 Sv of paleo-
535 AABW enters the surface ocean and ~25 Sv of water goes back to the Southern Ocean through the
536 60Sab section. Along the latitudinal direction, there is about 20 Sv of paleo-AABW that crosses
537 the Equator, and around 15 Sv continues on to latitude 20°N. This meridional Lagrangian
538 streamfunction can be decomposed into partial streamfunctions for the particles that go back to the
539 Southern Ocean, and those intercepting the MLB separately (Fig. 4B, 4C, 4D).

540 Under PI conditions, the proportion of deepwater mass entering the surface mixed layer
541 (Table S2) is larger than that (25%) for the early-Eocene simulation, with 9 Sv out of the 17 Sv
542 for NADW and 27 Sv out of the 35 Sv for AABW. These large proportions are probably due to
543 the presence of the Antarctic Circumpolar Current, dragging the NADW through inter-hemisphere
544 circulations (e.g. Wolfe & Cessi, 2011) and may also induce meridional excursion for AABW,
545 because circumpolar circulation in the Southern Ocean is not strictly zonal (Volkov et al., 2010;
546 Tamsitt et al., 2017).

547 **3.2.2 What drives the different fates of paleo-AABW?**

548 Particles sampling the paleo-AABW show different fates depending on their initial depth
549 and their basin of origin. Figure 4E shows the transport distribution of particles as a function of
550 initial depth. The results suggest that water parcels originating in the deepest layers (4000–4500
551 m) are relatively more likely to be intercepted by the MLB, the lower part of the deep waters
552 (3500–4000 m) tend to be intercepted by the 60Sab section, whereas the upper part of the deep

553 waters (2000–3000 m) tend to return to the initial section. This depth-dependency was unexpected,
554 because, intuitively, the upper deep waters have a short distance to travel before reaching the MLB,
555 thus should have a higher chance of being intercepted by the MLB. There are two possible
556 explanations for this unexpected result: 1) the switch between northward-flowing and southward
557 flowing transport happens somewhere in the middle of the lower and upper deep water, making it
558 more likely for the paleo-AABW flowing in the lower part of the deep waters to return to the initial
559 section; 2) a “channel” (e.g. prevailing upwelling) may link the upper deep waters to the surface
560 ocean.

561 Among the different basins, the waters of Atlantic origin are more likely to enter the mixed
562 layer (7.6 out of 18.8 Sv) than that of the Indian (4.4 out of 20.7 Sv) and Pacific (4.4 out of 11.1
563 Sv) oceans (Table 1). Most paleo-AABW of Pacific and Indian Ocean origin return to the initial
564 section: 45% and 55% respectively (Table 1). In contrast to this spatial pattern, there are no distinct
565 patterns among these destinations in terms of initial salinity or temperature, as shown in the θ -S
566 diagram (Fig. 4F).

567 **3.3 Return of paleo-AABW to the surface ocean by crossing the MLB**

568 As mentioned above, a quarter of paleo-AABW parcels cross the mixed layer and
569 eventually return to the surface ocean, which is statistically represented by particles intercepting
570 the MLB (Fig. 4B). The trajectories of these deepwater parcels ending in the mixed layer illustrate
571 the pathways toward the surface ocean and the actual transformations of paleo-AABW into light
572 water. The amount of this deepwater crossing the MLB is up to half of the non-recirculating paleo-
573 AABW mass (20.7 Sv out of 46.1 Sv). This large proportion shows this upwelling towards the
574 mixed layer plays a crucial role in closing up the meridional overturning circulations. We will

575 investigate further in more details what are the main dynamical processes occurring along the
576 particle trajectory for this obduction of deep water.

577 **3.3.1 General routes and residence time of deep water towards the MLB**

578 From the initial section onward, general routes for the horizontal propagation of these
579 paleo-AABW ending in the mixed layer are shown by the barotropic Lagrangian streamfunction
580 (calculated from eq. 4, Fig. 5A). The results show varied water transport routes among different
581 basins. Over the Pacific sector, ~ 5 Sv of paleo-AABW leaves the Southern Ocean from the initial
582 section and eventually crosses the MLB and returns to the surface ocean. A small fraction of these
583 northward-flowing water masses branches off from the northward transport and joins the
584 southwest currents into the Indian basin before reaching the Equator. Around half of these
585 northward-flowing water masses come out into the surface mixed layer over the tropics (9.3 Sv).
586 The remaining small fraction continues its northward journey crossing the Equator and eventually
587 reaching the surface ocean over the northern high-latitude regions with deeper mixed layer base
588 (Fig. 4 of Zhang et al., 2020). In the Atlantic-Indian Ocean, paleo-AABW particles initially travel
589 toward the north, and most reach the tropical region. Over the equatorial region, a large fraction
590 of water parcels changes their direction toward the west and crosses the remaining Tethys Sea
591 (bounded by Africa-India in the south and Eurasia in the north). There are actually two fates for
592 these waters flowing westward: 1) around 5 Sv goes southward along the east coast of the paleo-
593 Indian continent; 2) 10 Sv water keeps travelling westward, crossing the Gibraltar Strait and being
594 transported southward afterward, forming the noticeable anticlockwise transport around the
595 African continent. Strong tropical upwelling over the Pacific basin is supplied by deep water
596 originating from multiple basins, including paleo-AABW of Pacific origin locally, paleo-AABW
597 of Atlantic origin transported from the east through the Panama Strait and paleo-AABW of Indian

598 origin from the southwest. About 5 Sv water transport toward the Atlantic basin crossing the
599 Panama Strait is also visible in these routes of deepwater crossing the MLB, as indicated by the
600 differences in the Lagrangian streamfunction (Fig. 5). Although this only partially represents the
601 net water transport across the strait, the direction seems compatible with studies on a more recent
602 opening of the strait (Sepulchre et al., 2014; Tan 2017; Nof & Van Gorder, 2003).

603 The final geographical location of where paleo-AABW parcels cross the MLB shows a
604 clear latitudinal pattern (Fig. 5B). The distribution of these water parcels in terms of volume flux
605 as a function of latitude highlights over three regions: Southern Ocean deep convection,
606 equatorial/tropical upwelling, and Northern Hemisphere high latitudes (NHH) (Fig. 5C). The
607 Southern Ocean convection is revealed by enhanced vertical diffusivity and deep mixed layer
608 depth. For the equatorial/tropical upwelling regions, water parcels enter the surface mixed layer
609 mainly along the Equator (Pacific and Atlantic), and over the eastern coastal regions in the Pacific
610 and Atlantic basins, where surface winds drive strong offshore Ekman transport. The averaged
611 time required to travel from the initial section to the MLB accordingly decreases from the north
612 toward the south, with 1400 years for the NHH region, 1200 years for the tropics, and 900 years
613 for the Southern Ocean. On average, 80% of the deepwater parcels crossing the MLB finish their
614 journey within 1000 years (Fig. S1).

615 **3.3.2 The role of tropical upwelling in deepwater obduction**

616 The important role of tropical upwelling in the deepwater obduction in the early-Eocene
617 simulation is highlighted by the large amount of upwelling water in the tropical regions. As shown
618 in Figure 6B & 6C, the amount of paleo-AABW upwelling in the tropical region is up to 9.3 Sv,
619 which consists of 45% of the total deepwater masses crossing the MLB. This upwelling is
620 associated with the upwelling branch of the Eulerian MOC streamfunction in this tropical region

621 (see Fig, 7C, between 300-1000m). The rest of the particles mainly reach the mixed layer through
622 vigorous ventilation in the Southern Ocean. This large fraction indicates that the contribution of
623 tropical upwelling to the deepwater obduction process is substantial in magnitude, which is
624 essential for closing the global ocean circulation.

625 This strong upwelling over tropical regions is generally consistent with the spatial pattern
626 of wind-driven Ekman pumping. With the easterlies blowing over the tropical surface, westward-
627 flowing water is dragged by the Coriolis force away from its original westward route on both sides
628 of the Equator. The opposite directions of the Coriolis force between the southern and northern
629 hemisphere result in divergences of surface water transport over the equatorial region and induce
630 the well-known Ekman pumping. In line with this, the Ekman pumping can explain the spatial
631 pattern of the particles final positions when entering the surface mixed layer. As shown in Figure
632 5B, water parcels enter the surface mixed layer mainly over the eastern coastal regions, where
633 surface waters are dragged offshore by wind, leading to divergence at the surface and inducing
634 coastal upwelling.

635 To further put this tropical upwelling in context, we compared it with PI conditions. First,
636 we need determine which deepwater mass (NADW or AABW) is more comparable to the paleo-
637 AABW in the early Eocene simulation. Intuitively, the AABW is the natural counterpart for the
638 early-Eocene paleo-AABW, because both share the same origin in the Southern Ocean. However,
639 the overlying NADW in the present conditions complicates and even hinders these direct
640 comparisons. In contrast, from the actual ocean circulation perspective, the NADW is a more
641 comparable alternative, because it is formed in high latitudes (albeit in opposite hemispheres) and
642 has direct access to the surface ocean. However, only 1.7 Sv (out of 9.2 Sv) of the southward-
643 flowing NADW comes out into the surface ocean in tropical regions and the majority returns to

644 the surface ocean through the Southern Ocean upwelling. Therefore, compared with PI conditions,
645 the proportion (9 Sv out of 21 Sv) of tropical upwelling of paleo-AABW into the surface ocean
646 during the early Eocene is much larger. This is in line with the stronger Eocene upwelling over the
647 tropics than the present-day (Fig. S6B). Indeed, because of the almost closed Drake passage, there
648 is no Antarctic Circumpolar Current in the Eocene, and the tropical gateways between the different
649 basins are not located in the westerlies latitude band, such that the paleo-AABW overturning cell
650 cannot function as a pole-to-pole circulation like the NADW today. Consequently, another
651 transformation route to surface waters needs to be found, and this is partly through equatorial and
652 tropical upwelling.

653 **3.3.3 Intense exchanges across the three basins**

654 The strong connection between the three basins can be seen from the well-mixed final
655 locations of particles, independently of their basin of origin at the initial section. As summarized
656 statistically in the right part of Table 1, only half the amount of deepwater parcels enter the mixed
657 layer in the same basin as they departed from, while the other half enters the surface ocean in other
658 basins. This initial-final analysis provides only a minimum estimate on exchanges between basins,
659 because some of water parcels that eventually end up in the same basin may travel to other basins
660 before reaching their final location. In addition, the amount of paleo-AABW water ending in the
661 Pacific mixed layer is more than twice as the amount of paleo-AABW initialized from this basin
662 (9.7 vs 4.4 Sv), demonstrating that deepwaters ending in the Pacific mixed layer are supplied from
663 multiple basins, i.e. the Pacific itself, the Atlantic and the Indian Ocean. Overall, these Lagrangian-
664 based statistics illustrate that the magnitude of inter-basin overturning transports is substantial.

665 **4 Discussion**

666 **4.1 Conversion of paleo-AABW into lighter surface water**

667 The principle of deepwater conversion is that deep water is transferred from the interior
668 ocean to the surface mixed layer by both diapycnal mixing (Munk, 1966) and obduction processes
669 (Blanke et al., 2002; Liu & Huang, 2012; Marshall & Speer, 2012). The mass flux escaping the
670 interior ocean in our Lagrangian experiment is documented by particles intercepted by the MLB.
671 During their journey, the physical characteristics of water parcels are modified by mixing with
672 surrounding water, in association with topographical interactions and mesoscale dynamics
673 (Rimaud et al., 2012). Tracing the evolution of properties of these water masses along the trajectory
674 can thus illustrate how these processes contribute to paleo-AABW conversion.

675 **4.1.1 Deepwater conversion accompanied by density transformation**

676 For those water parcels that eventually enter the surface ocean, most go through diapycnal
677 mixing processes with net negative density transformation during the journey, as shown by the
678 differences of their transport distribution of density anomaly between the initial and final state (Fig.
679 5D). At the initial section, paleo-AABW density falls in the range of $26.5\text{--}27\text{ kg/m}^3$, whereas their
680 density anomaly when they enter the mixed layer ranges from 17 to 27 kg/m^3 with one mode at
681 $21\text{--}22\text{ kg/m}^3$ and another around 26 kg/m^3 . Therefore, water mass density decreases conspicuously
682 from the initial to the final state, indicating that water parcels go through negative density
683 transformation along their journey. Actually, the net negative density transformation are the results
684 of two opposing processes: salinity and heat changes, as a result of mixing with the surrounding
685 waters. The majority of water parcels gain both salinity and heat at the same time along their
686 journey, but the latter dominates the density decrease, as shown by the probability distribution in
687 the salinity and temperature spaces (Fig. S3).

688 The following analysis on the full-range trajectory of water parcels allows us to pin down
689 where these processes happen exactly. Density changes along the water parcels' trajectory reveal
690 clear vertical and geographical patterns. In the vertical direction, the density transformation mostly
691 occurs in the upper ocean, whereas in the abyssal ocean, water parcels mostly travel along
692 isopycnals with conserved density. The scatterplot of density changes as a function of depth
693 (Figure 6A) shows considerable changes over the upper ocean (e.g. above 1000 m), but very weak
694 changes below the depth of 1500 m. For the upper ocean, the particles distribution as a function
695 of density change (Fig. 6B) is clearly skewed to the left, indicating overall net density loss. The
696 net negative density transformation is the balance of the density loss and density gain, with the
697 latter being much smaller in extent. Further investigation reveals that these positive values of
698 density change are due to the seasonal cycle effects, and examining the density changes at annual
699 frequency filters this out and shows negative density change along the trajectories (Fig. S4).
700 Comparisons of density changes among different vertical layers demonstrate a clear decrease in
701 this net density transformation with increasing depths. For instance, over the top 500 m there are
702 four times as many particles losing density larger than 0.025 kg/m^3 per month than over the 500-
703 1000 m depth range (Fig. 6B).

704 In line with this efficient density transformation in the upper ocean, the final depths where
705 water parcels reach the MLB are restricted to the upper 500 m. In contrast, the final depths of the
706 particles that do not experience significant density changes (defined as the final density ≥ 26.5
707 kg/m^3) vary from 100 m to 2800 m, and are geographically limited to within 50°S south. The
708 geographical distributions of these two types of water parcels are consistent with the findings of
709 Blanke et al. (2002), who found that deepwater masses enter the surface ocean in two ways: 1)
710 deepwater parcels accumulate upward displacement over time and eventually reach the surface

711 mixed ocean throughout the climatological year; 2) seasonal shallowing of the mixed layer can
712 absorb some water into the surface mixed ocean, leading to water transfer into the mixed layer. In
713 the horizontal direction, the large density change mainly occurs near the tropical upwelling regions
714 (Fig. 6C), primarily in response to wind-driven Ekman transport. Figure 6C presents the
715 geographical distribution of these water parcels whose density change is greater than 0.1 kg/m^3
716 per month. Although both density gain and loss related to seasonal cycle are found near the tropics
717 (Fig. 6C), the annual signal is dominated by a net density loss (Fig. S6B). This strong negative
718 density loss over the tropical regions matches the tropical upwelling pattern well, supporting its
719 role in density transformation. Some density transformations also occur over the high-latitude
720 regions, probably as results of strong air-sea flux and induced convective displacement/ventilation
721 of water parcels, but their contribution remains relatively small.

722 This efficient density transformation in the upper ocean is coupled to the effective net
723 upward transport at the similar vertical level. As illustrated by the depth changes as a function of
724 depth along the particle trajectories (Figure 7), the net upwelling mainly occurs in the upper ocean,
725 despite an overall small absolute change. In the upper ocean, the particles distribution as a function
726 of depth change (Fig. 7) is strongly left-skewed, indicating substantial net upwelling. In contrast,
727 despite the larger absolute vertical displacements (both upward and downward transport) in the
728 abyssal ocean, these two opposite displacements are of similar order of magnitude and therefore
729 cancel each other out, without causing net upwelling transport. These large absolute vertical
730 displacements in the abyssal ocean are probably due to seafloor perturbations, since the spatial
731 pattern matches the bathymetry very well (Fig. S5).

732 4.1.2 The tropical upwelling of paleo-AABW

733 The above analysis of particle trajectory demonstrates that important processes occur over
734 the tropical region, highlighting the importance of tropical upwelling in the obduction of paleo-
735 AABW. This section will focus on this tropical upwelling transport to investigate the mechanism
736 behind this prevailing upwelling and the potential analog to the present-day situation.

737 Apart from statistical summaries of the final transport over tropical regions, another way
738 to show vertical transport is to calculate actual vertical transport at each depth using the MOC
739 streamfunction. By definition, the MOC streamfunction differences for a given latitude-band (Ψ_{latn}
740 $- \Psi_{\text{latn}}$) represent the vertical transport of this region, with positive values indicating upward
741 transport and negative values reflecting downward transport. For instance, over the latitude band
742 of 10°S and 10°N, streamfunctions based on water parcels crossing the MLB and on all particles
743 both estimate a noticeable upward water mass transport at shallow to intermediate depths (e.g.
744 upper 1500 m), with good agreement between them (Fig. 7C). Compared with Eulerian-inferred
745 results, these particles reproduce Eulerian-inferred vertical transports well at intermediate depths
746 (i.e. 500–1500 m), and only a fraction of transport over shallow depths. In particular, water parcels
747 only represent a small part of upwelling transport over the top 500 m.

748 This efficient upwelling circulation over tropical regions during the early Eocene is favored
749 by the continental geometry, compared with PI conditions. First, the three basins (Pacific, Atlantic
750 and Indian) are well-connected over the tropical regions by the opened Panama Strait, the wide
751 Gibraltar Strait and the remains of the Tethys Sea. These well-connected ocean basins allow
752 circum-circulation across all longitudes, in a similar way as the ACC under the present-day
753 condition, despite with a weaker intensity, and at different latitude range. Second, the broad
754 extension of the Pacific basin along the east-west direction over the tropics, 20% larger than the

755 present-day over most of the tropical latitudes, provides more space for accumulating wind-driven
756 divergence at the surface (and shallow levels), and facilitates the strong upwelling on the east side
757 of the basin. For the Atlantic basin, although the overall basin size is smaller than today, the basin
758 width at 0–10°S (where numerous water parcels enter the thermocline) reaches up to 4000 km,
759 which is sufficient to permit upwelling over the eastern basin at this latitude band. The final
760 geographical locations where water parcels enter the mixed layer are constrained to upwelling
761 regions and can be inferred from the wind stress curl (Fig. 5B & Fig. S6A). There are no significant
762 differences in wind stress between the early Eocene and present-day conditions (Zhang et al., 2020).
763 Therefore, the continental geometry differences are probably responsible for this efficient
764 equatorial and tropical upwelling process.

765 This strong upwelling of deep water near the Equator serves as a return path for paleo-
766 AABW from the interior ocean to the surface mixed layer. In this sense, the tropical regions during
767 the early Eocene are analogous to the present-day Southern Ocean, because both of them transport
768 deep water upward and are essential for the closure of the overturning circulation. The role of
769 Southern Ocean upwelling that is driven by strong westerlies across all longitudes in closing the
770 present-day MOC has been highlighted by numerous studies (Anderson et al., 2009; Blanke et al.,
771 2002; Garzoli & Matano, 2011; Marshall & Speer, 2012; Wolfe & Cessi, 2014; Tamsitt et al., 2018). Both
772 observation data and numerical experiments confirm that the North Atlantic Deep Water is brought
773 into the surface by the Southern Ocean upwelling through inter-hemisphere overturning circulation
774 (Marshall & Speer, 2012). More directly, Lagrangian analysis and CFC tracer observation
775 demonstrate the deep water upwelling in the Southern Ocean through the ACC (Blanke et al. 2002;
776 Rhein et al., 2015; Tamsitt et al., 2017). In the absence of such a route to the surface for paleo-
777 AABW in the Eocene, equatorial and tropical upwelling provide the privileged return route to the

778 surface. As such, the strong tropical upwelling of deep water during the early Eocene can play an
779 analogous role to the present-day Southern Ocean upwelling.

780 **4.2 Implications of strong tropical upwelling on global carbon cycle**

781 The tropical upwelling during the early Eocene serves as an efficient return path for paleo-
782 AABW and thus links the interior ocean to the surface ocean and eventually to the atmosphere.
783 The potential efficiency of the upwelling process to affect the global carbon cycle has been well
784 illustrated by the Pleistocene glacial cycles (Sigman & Boyle, 2000; Anderson et al., 2009; Burke
785 & Robinson, 2012). Multiple processes operating among various reservoirs in the carbon system
786 may likely be involved in contributing to the full amplitude of CO₂ variability during the last
787 glacial-interglacial transition — up to 100 ppm as archived in the ice core (Barnola et al., 1987;
788 Sigman and Boyle, 2000). The continental reservoir of organic carbon decreased during the glacial
789 period mainly via two processes accompanied by the extension of glaciers at high latitudes: 1)
790 degraded vegetation cover reduced biomass carbon (~15ppm); 2) exposure of sediments on
791 continental shelves enhanced weathering. Nevertheless, terrestrial carbon is a small reservoir and
792 may only be a modest source for glacial-interglacial CO₂ variation, especially considering the
793 buffering effect of the ocean reservoir. For the ocean reservoir, dynamical carbon storage in the
794 ocean associated with physico-chemical and biogeochemical processes in response to changes in
795 oceanic chemical composition has been suggested to play an important role (François et al., 1997;
796 Sigman & Boyle, 2000). Regarding physico-chemical processes, changes in CO₂ storage of the
797 surface ocean due to cooler temperatures and high salinity may reduce atmospheric CO₂ by 23ppm.
798 Part of this reduction may be compensated by an associated weathering-induced alkalinity balance
799 (e.g 15-20ppm), i.e. temporary imbalances between input from weathering and output by the burial
800 of biogenic calcium carbonate in the ocean sediment. Biogeochemical processes refer to the

801 extraction of carbon from the surface ocean by biological production, allied with changes in the
802 marine calcium carbonate budget. The availability of ocean nutrients in the low-latitude ocean, the
803 key to this biogeochemical process, can regulate biological export production, which in turn
804 determines the sequestration of inorganic carbon in the deep ocean by photosynthesis and the
805 subsequent rain of organic carbon out of the upper ocean (Broecker, 1982; Sigman & Boyle, 2000).
806 For instance, it has been estimated that a 30% increase in oceanic nutrients can decrease
807 atmospheric CO₂ concentrations by 30–45 ppm (Sigman et al., 1998). More importantly, it has
808 been demonstrated that increased nutrient utilization in the high-latitude Southern Ocean can also
809 contribute to glacial CO₂ reduction (Knox & McElroy, 1984; Sarmiento & Toggweiler, 1984),
810 which requires increased isolation of deep-water masses from the atmosphere (Sigman & Boyle,
811 2000; Köhler et al., 2005; Watson & Naveira Garabato, 2006; Peacock et al., 2006). The Southern
812 Ocean plays a central role in regulating the glacial-interglacial variability of atmospheric CO₂
813 because deep-water masses outcrop in the Southern Ocean and exchange gases with the
814 atmosphere (François et al., 1997; Sigman & Boyle, 2000). More recently, using biogenic opal
815 data as an upwelling indicator, Anderson et al. (2009) proposed that a change in Southern Ocean
816 upwelling circulation (ventilation of the deep-water masses) substantially alters the partitioning of
817 CO₂ between the atmosphere and the deep sea. This potential link of changed ventilation in the
818 Southern Ocean to the fluctuations of atmospheric CO₂ across glacial-interglacial cycles during
819 the late Pleistocene has been further elaborated by recently-available data arising from the
820 development of novel methods (e.g. Burke & Robinson, 2012; Martínez-Botí et al., 2015). For
821 instance, radiocarbon records from deep-sea corals have suggested that enhanced ventilation in the
822 Southern Ocean since the last Glacial Maximum (21 ka) contributed to an increase in atmospheric
823 CO₂ concentration during the last deglaciation (Anderson et al., 2009; Burke & Robinson, 2012).

824 Recent boron isotope data, as a more direct tracer of oceanic CO₂ outgassing, further suggest that
825 strengthened upwelling caused a leakage of oceanic carbon over the Southern Ocean, driving the
826 increasing atmospheric CO₂ over the course of the last deglaciation (Martínez-Botí et al., 2015).
827 Such non-negligible changes, caused by the variation of the Southern Ocean upwelling over the
828 late Pleistocene glacial cycles, demonstrate that upwelling dynamics can significantly change the
829 global carbon distribution.

830 During the Eocene, the low-latitudinal geographical locations of the upwelling process and
831 associated high temperature can reinforce the influence of upwelling on the global carbon cycle
832 by inducing physico-chemical and biogeochemical processes. From the physico-chemical
833 perspective, the solubility of greenhouse gases is a strong inverse function of seawater temperature.
834 The higher temperatures over upwelling regions may lead to more complete degassing of upwelled
835 water and promote an effective solubility pump of greenhouse gases from the abyssal ocean into
836 the atmosphere, although the absolute gas content hold in water can be reduced by the overall
837 warm Eocene environment. Hypothetically, this intensified solubility of upwelling water can
838 transfer a tremendous amount of carbon from the abyssal ocean to the atmosphere, via efficient
839 exposure of carbon-rich deep water to the active surface mixed layer. From a biogeochemical
840 perspective, strong upwelling theoretically brings nutrients up to the upper ocean and enhances
841 marine primary productivity. However, biogenic silica accumulation rates, as an indicator of
842 biological productivity, in the equatorial zone of the Pacific during the Eocene have been found to
843 be lower than in the Neogene (Moore et al. 2004; 2008). Given the complex processes involved in
844 transforming biological production into preserved biogenic sediment (especially for such deep
845 geological time), several factors may explain this discrepancy between expected high primary
846 productivity and low sedimentation rates (Olivarez Lyle & Lyle, 2005, 2006). First, in addition to

847 primary production, complicated secondary processes influence the settling flux and final burial
848 of organic matter on the seafloor as sediment deposits, which can lead to substantial uncertainties
849 in using biogenic silica as a proxy for biological production for such an ancient time period (e.g.
850 Sigman et al., 1998; Sigman & Boyle, 2000). For instance, the extremely warm Eocene
851 environment may substantially enhance the basal metabolic rate of the oceanic biota in deeper
852 waters and accelerate nutrient recycling, which can reduce burial on the seafloor, thus reducing
853 the biogenic silica accumulation rate (Olivarez Lyle & Lyle, 2005, 2006). Meanwhile, increase in
854 the dissolved nutrient concentrations of ocean waters below the photic zone may support high
855 respiration rates and consume the primary production (as mentioned by Moore et al. 2008; Piper
856 & Calvert, 2009). Second, the preservation of sediment is also an important issue (Moore et al.,
857 2008), which can be affected by an erosion process or the secondary dissolution due to a totally
858 different physico-chemical environment (Pichon et al., 1992). In particular, the Eocene is at the
859 base of most sediment cores, and fewer sediment core data are available for the Eocene period than
860 for the later periods (e.g. Oligocene). Potentially, strongly connected tropical pathways may allow
861 enhanced bottom currents that further modify sediment deposition. These large uncertainties may
862 explain the discrepancies between the low silica accumulation rates and the expected enhanced
863 biological pump from the upwelling found in our study.

864 The potential sources of this abyssal ocean carbon are also subject to substantial uncertainty,
865 although they have been explored from different perspectives (Cui et al., 2011; Sexton et al., 2011).
866 The first hypothetical origin of this Eocene carbon input is oceanic dissolved carbon sequestered
867 in the abyssal ocean during previous geological times (Sexton et al., 2011), perhaps with a similar
868 process occurring during the younger intervals of Earth history as during the Pleistocene glacial
869 cycles (Anderson et al., 2009; Burke & Robinson, 2012). The second hypothesis attributes this

870 Eocene carbon input to the releases of carbon from buried sediments (Nicolo et al., 2007; Panchuk
871 et al., 2008). However, here, we could not pin down which of the above-mentioned sources would
872 be more plausible yet, and the exact magnitude of these processes remains to be determined from
873 more data on the total carbon reservoirs in the deep ocean.

874 Overall, despite uncertainties in biogeochemical processes and possible carbon sources,
875 our results, from the oceanic dynamics perspective, demonstrate the efficient communication
876 between the abyssal ocean and the surface mixed ocean, which could effectively influence the
877 related carbon exchange and perhaps regulate climate evolution during the Eocene.

878 **5. Conclusions**

879 Coupled model simulations of the Eocene suggest a very intense overturning originating
880 in the Southern Ocean, and no deepwater formation in the northern hemisphere. Here, we used the
881 Lagrangian analysis tool Ariane releasing thousands of particles from the lower limb of this MOC
882 at latitude 60°S to follow the paleo-AABW masses leaving the Southern Ocean. Their trajectories
883 were subsequently analyzed to trace the fates of this northward-flowing bottom and deep waters.
884 With a good representation of the northward-flowing deepwater transports at the initial section,
885 the Lagrangian analysis revealed two fates for the paleo-AABW. The majority of paleo-AABW
886 returns to the Southern Ocean either through the initial section (~43%) as part of recirculation or
887 through the section above the initial section (~31%), and >25% eventually enters the surface ocean
888 by crossing the base of the mixed layer. Nearly half of the latter water parcels (9 Sv) enters the
889 surface ocean in tropical regions (30°S–30°N). This is quite different from the present-day
890 situation, in which most of the NADW comes out into the surface ocean in the Southern Ocean,
891 and AABW comes out in the region of 40°S.

892 Most of these water parcels experience negative density transformation before reaching the
893 surface mixed layer. This density transformation mainly occurs in the upper 500 m of the tropical
894 upwelling regions, which is accompanied by an effective upward overturning. The spatial pattern
895 of this negative density transformation is consistent with wind-driven Ekman pumping that is
896 favored by the Eocene continental geometry, including the widely extended Pacific basin and the
897 presence of tropical gateways (permitting substantial inter-basin connection in the tropics).

898 Our results indicate that the strong tropical upwelling during the early Eocene —
899 resembling the present-day Southern Ocean upwelling for the NADW, but occurring at much
900 higher temperature — provided an efficient pathway from the deep ocean to the surface ocean,
901 which may promote carbon release from the deep ocean through physico-chemical and
902 biogeochemical processes. The exposure of the carbon-rich deep water in upwelling regions can
903 promote the solubility pump of greenhouse gases from the abyssal ocean into the atmosphere,
904 although the absolute gas content of seawater may be reduced in a warm Eocene environment.
905 Meanwhile, upwelling can modify the nutrients distribution, which may change biological export
906 production by triggering changes in biogeochemical processes. These physico-chemical and
907 biogeochemical responses related to upwelling potentially facilitate carbon release from the
908 abyssal ocean to the surface ocean (and eventually to the atmosphere), which perhaps in turn
909 affected other processes of the carbon cycle and climate evolution during the Eocene. These
910 speculations probably deserve a more detailed analysis with a full carbon cycle model for this
911 particularly warm period of the Earth climate.

912

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922

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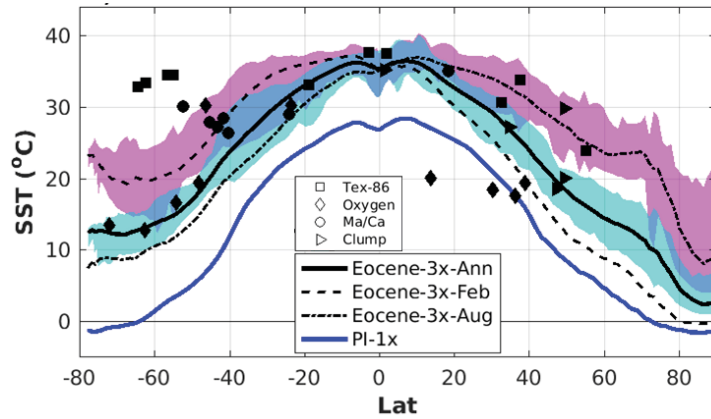
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1193 Figures

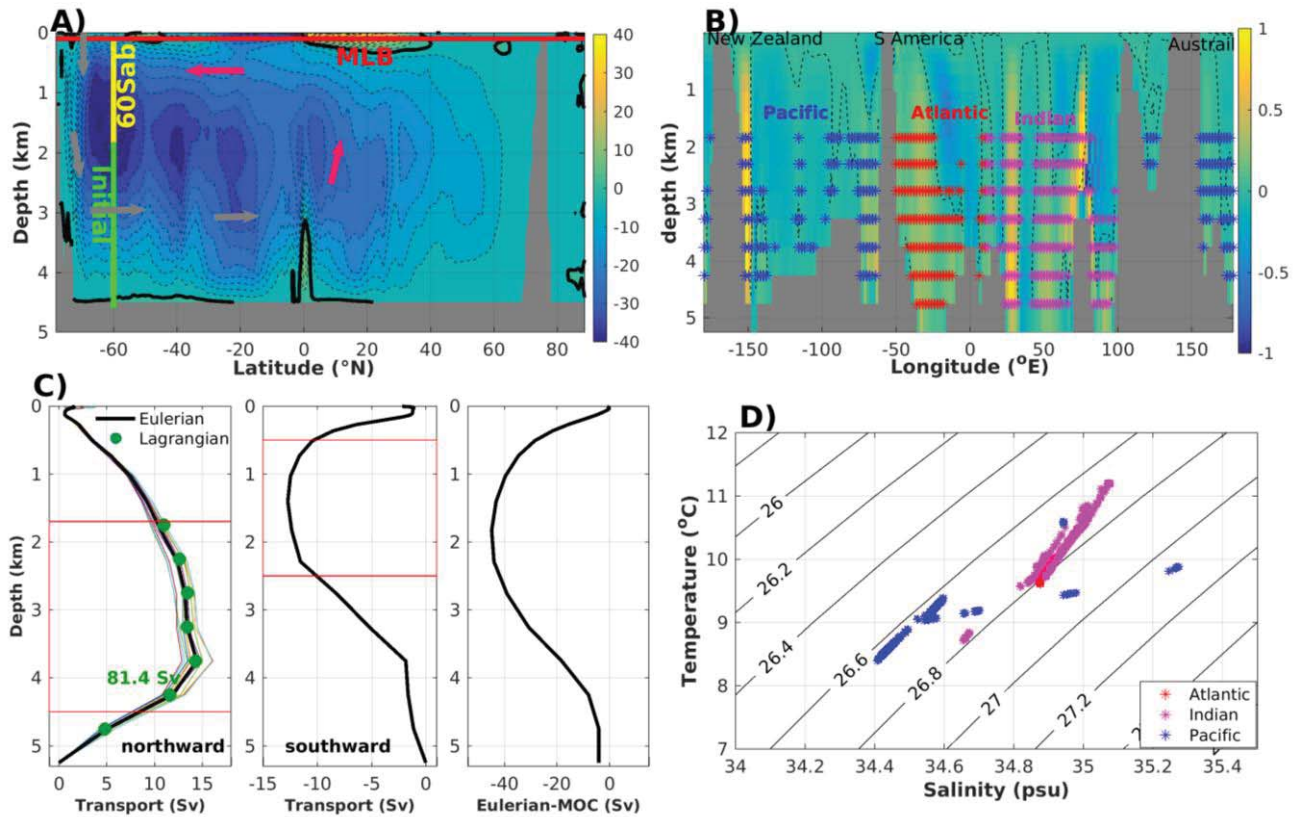
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1196 **Figure 1.** Simulated SST as a function of latitude and comparison with proxy-based temperature
1197 reconstructions. Annual mean (solid black), February (dashed) and August (dotted) are shown for the
1198 early Eocene; annual mean for the present-day is shown as the blue line. The different markers show
1199 reconstructed temperatures from different proxies. Shading indicates the simulated SST range over all
1200 longitudes.

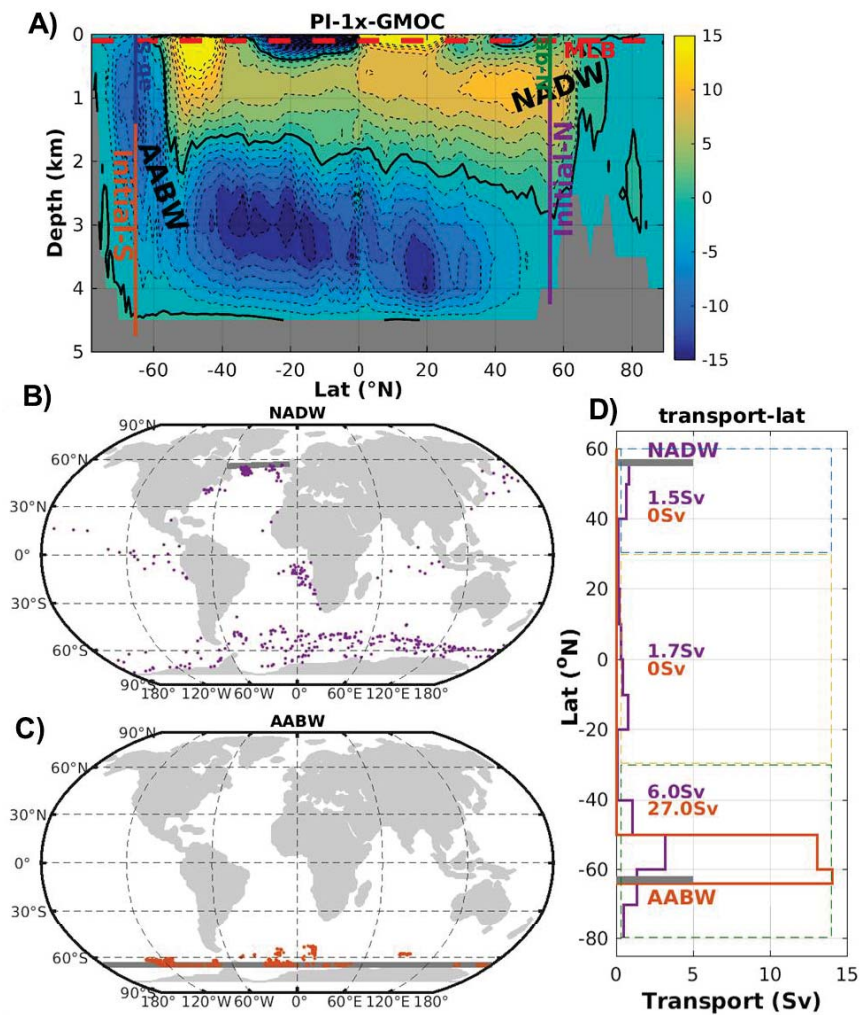
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1203 **Figure 2.** The Lagrangian experiment setup. (A) The specified sections used in the Lagrangian
 1204 experiment are indicated, including the initial, the 60Sab (the section above the initial section at
 1205 latitude 60°S) and the MLB (base of the mixed layer) sections. The Eulerian meridional
 1206 overturning circulation (MOC) streamfunction (Sv) of the early-Eocene simulation (adjusted from
 1207 Fig. 2a of Zhang et al., 2020) is also shown in the background with arrows to schematically indicate
 1208 the direction of circulation. (B) The initial position of particles is indicated with an asterisk and
 1209 their relationship to mass transport (with positive values for northward transport and negative
 1210 values for southward transport in Sv) at the vertical section of 60°S. (C) Particle-based water mass
 1211 transport estimates (Lagrangian approach) over the lower limb of the MOC, and relationship to
 1212 Eulerian-inferred transport (integrated along longitude for each layer), including northward and
 1213 southward transport and the Eulerian MOC stream function at the latitude of 60°S. (D) θ -S diagram
 1214 of particles (water parcels) at the initial state in the three ocean basins.

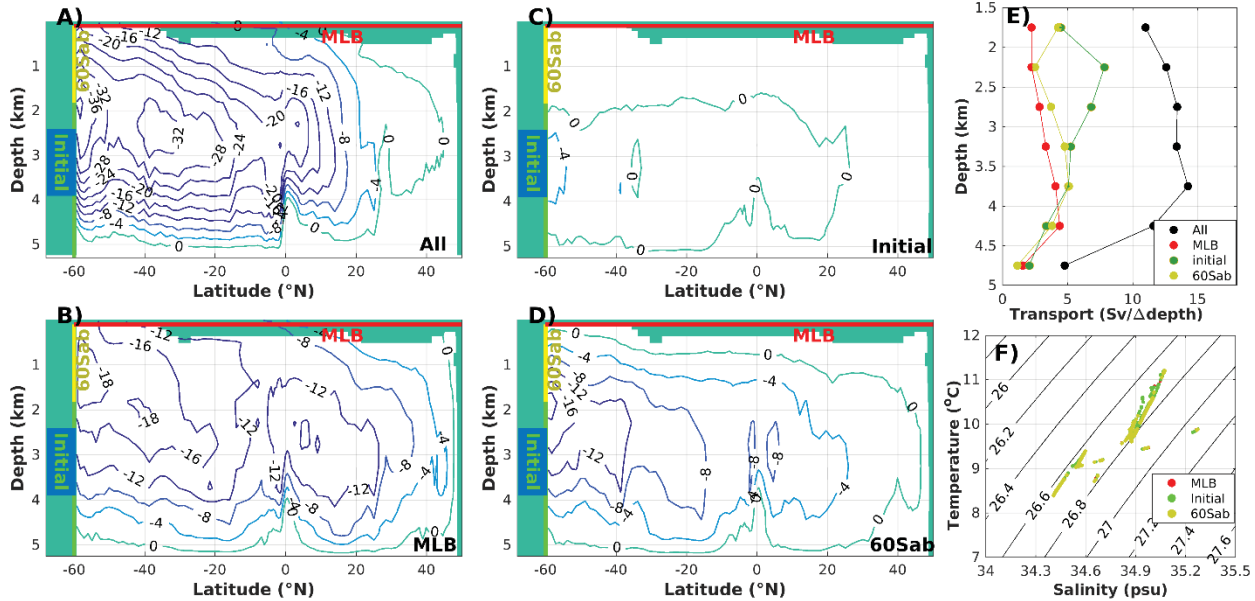
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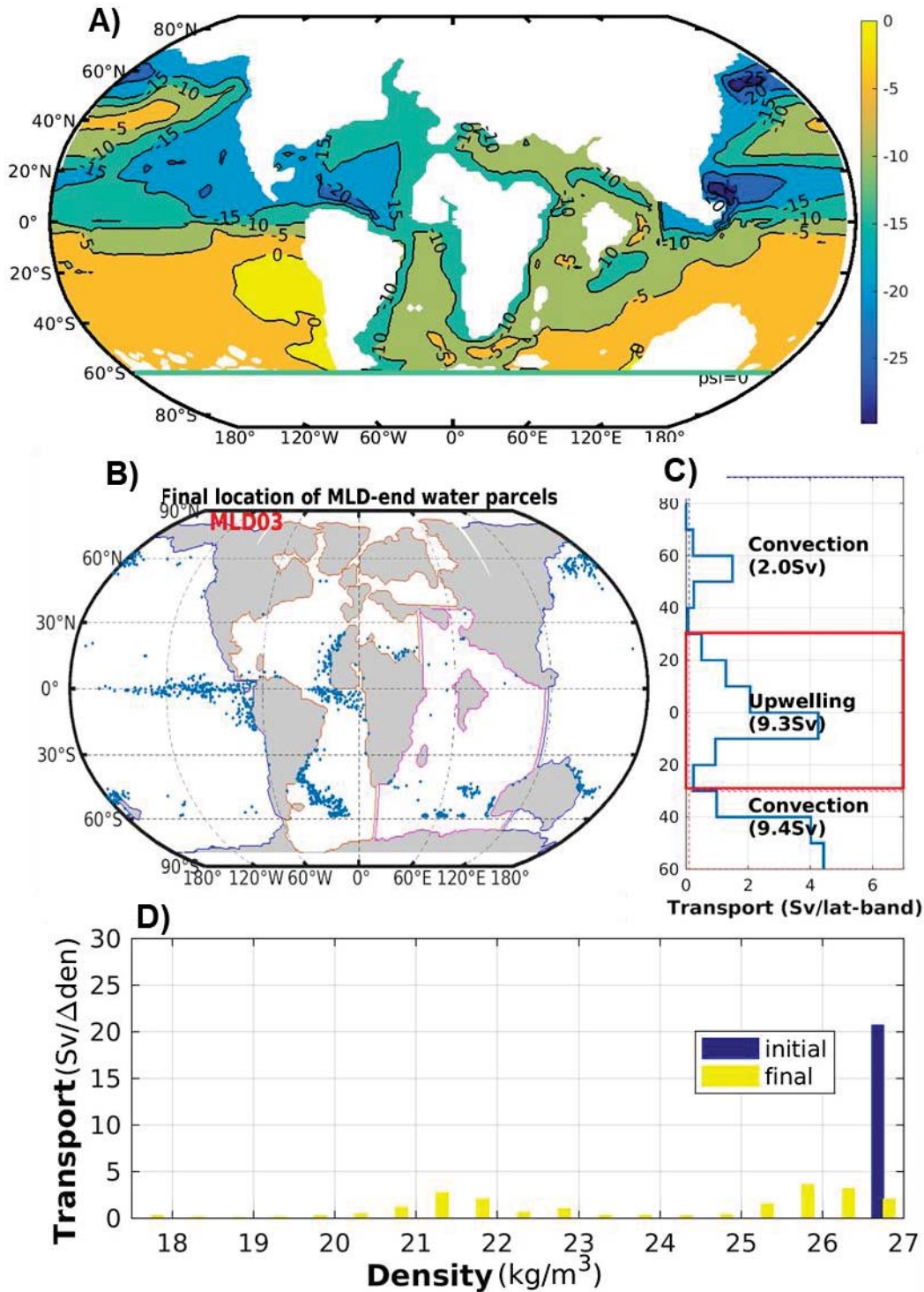
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1218 **Figure 3.** Validation of the Lagrangian method for the present-day (PI) circulation. A) Setup of
 1219 two Lagrangian experiments, with background showing the global MOC (CI: 2 Sv) adjusted
 1220 from Fig. 3b of Zhang et al., (2020). B) and C) show the final position of particles entering the
 1221 mixed-layer for North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) in
 1222 turn. D) Probability distribution as a function of latitude, shown as Sv per 10deg-lat-band. Initial
 1223 sections are marked by thick gray lines in B), C) and D).



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 1226 **Figure 4.** Fates of northward-flowing paleo Antarctic Bottom Water. Meridional Lagrangian
 1227 streamfunctions show the zonally integrated transport of paleo-AABW, A) for all particles, B) for
 1228 particles crossing the MLB; for particles back to the Southern Ocean through C) the initial deep
 1229 section and D) through the upper 60°Sal section (60Sab) respectively. E) Transport distribution of
 1230 paleo-AABW towards the different final sections as a function of their initial depth at 60°S. F) θ -
 1231 S diagram of water particles at the initial 60°S section, colored according to their fates.

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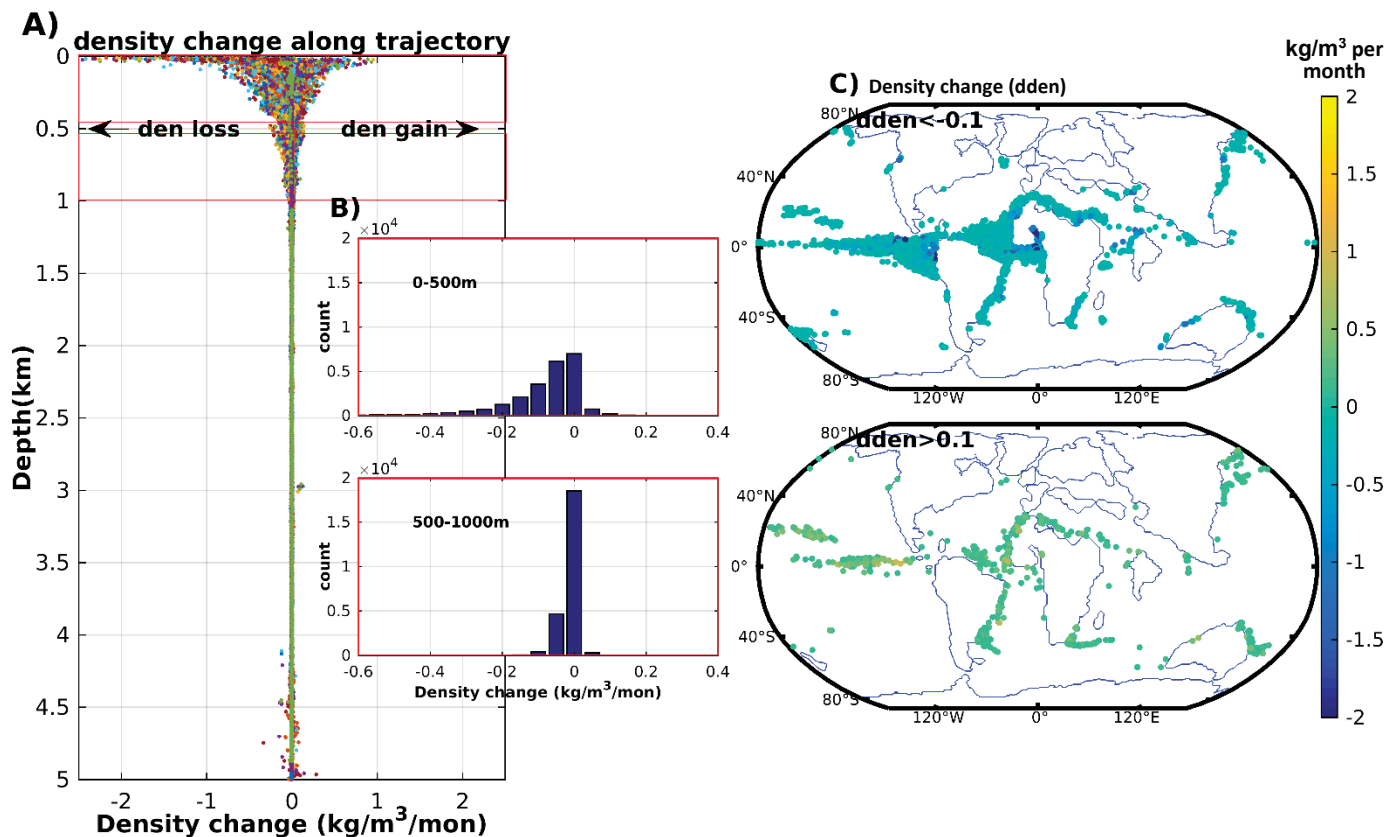
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1235 **Figure 5.** Horizontal pathway of paleo-AABW for particles entering the mixed-layer. **A)**
 1236 Horizontal Lagrangian streamfunction (in Sv), illustrating their pathways towards the mixed
 1237 layer. **B)** Final location of the paleo-AABW particles entering the mixed layer, with basin
 1238 contours in color. **C)** Upward transport distribution of paleo-AABW particles across the mixed
 1239 layer, in Sv per 10° latitude band. **D)** Transport distribution (Sv per density bin, particle number

1240 weighted by volume transport) of water parcels as a function of initial (blue) and final (yellow)
1241 potential density anomaly in 0.5 kg/m³ bins.

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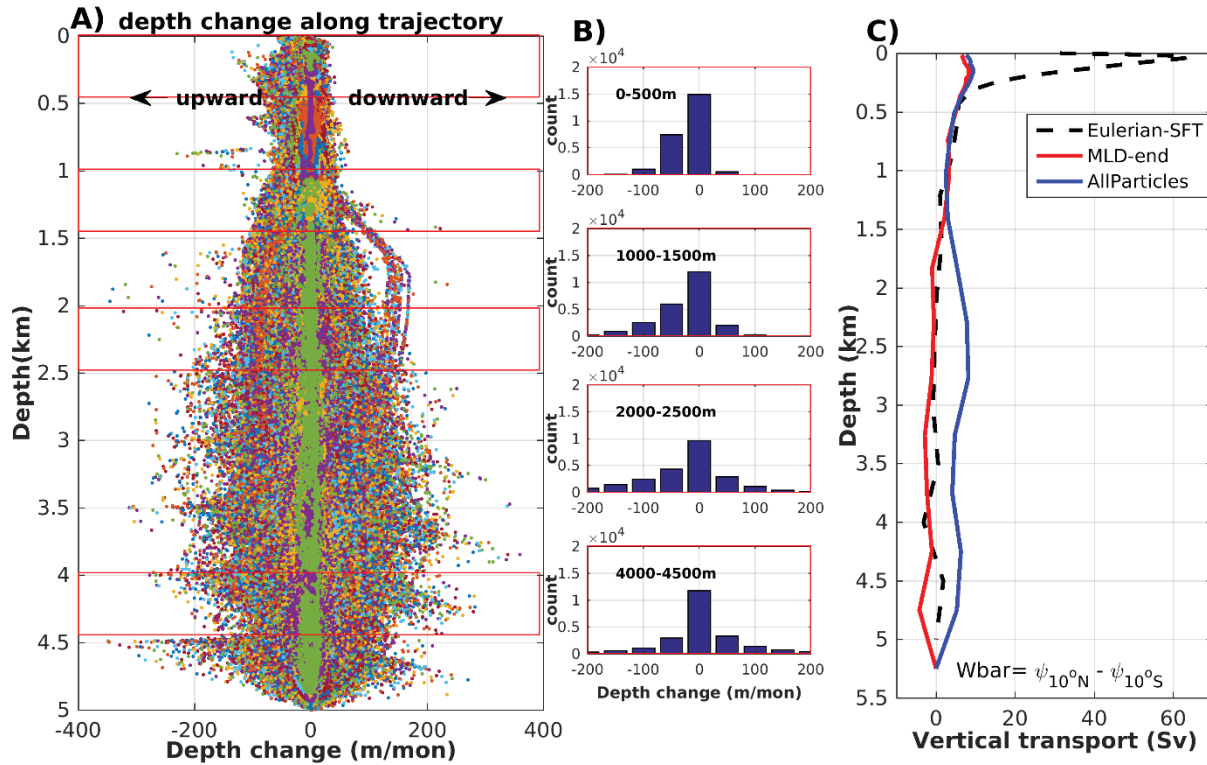
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1245 **Figure 6.** Density transformation along full-range trajectory of paleo-AABW parcels entering
1246 the mixed layer. A) Scatter plot of density change (kg/m³ per month) as a function of depth. B)
1247 Probability distribution as a function of density change at the two vertical layers as indicated in
1248 Fig 8A, with the per density change bin of 0.05 kg/m³. C) Geographical location of large density
1249 transformations, shown as the geographical scatterplots of water parcels whose density change is
1250 greater than 0.1 kg/m³, or less than -0.1 kg/m³.

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1254 **Figure 7.** Depth changes as a function of depth along paleo-AABW parcels trajectories. A)
 1255 Scatter plot of depth changes (in meters per month, negative: upward; positive: downward) as a
 1256 function of depth. B) Particle distribution as a function of depth change, shown at various
 1257 vertical layers as indicated by the boxes in Fig. 9A. C) Vertical transport over the tropical region
 1258 $10^{\circ}N-10^{\circ}S$, derived from the corresponding streamfunctions.

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1268 **Tables**

1269 **Table 1** Origins and fates (in Sv) of northward-flowing Antarctic Bottom Water (paleo-AABW),
 1270 and their distribution across the three basins. Initial: 60°S section below 1900 m; 60Sab, 60°S
 1271 section above 1900 m; MLB, the base of the mixed-layer. The left part (Origins columns) refers
 1272 to the auxiliary backward Lagrangian experiment. The right part (Fates columns) refers to the
 1273 forward Lagrangian experiment. The middle column (Initial gross) refers to both. The transport
 1274 of paleo-AABW upwelling in the mixed layer (MLB columns) is further decomposed among the
 1275 three basins (last 3 columns). These diagnostics illustrate the inter-basin exchanges, with the
 1276 grey shading highlighting the proportion that stays within the same basin.

1277

	Origins			Initial gross	Fates					
	Initial	60Sab	MLB_SO		Initial	60Sab	MLB			
							Global	Pacific	Atlantic	Indian
Global	28.17	3.04	49.84	81.04	34.90	25.35	20.73 (100%)	9.72 (44.4%)	8.49 (41.1%)	3.07 (15.0%)
Pacific	—	—	9.69	24.45	11.06	9.04	4.35 (100%)	2.47 (58.1%)	1.44 (32.6%)	0.44 (9.3%)
Atlantic	—	—	16.14	18.75	3.10	20.74	7.57 (100%)	2.98 (34.1%)	4.77 (54.5%)	1.05 (12.5%)
Indian	—	—	24.01	37.84	20.74	9.50	4.35 (100%)	3.72 (48.7%)	2.28 (30.3%)	1.58 (21.1%)

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Figure1.

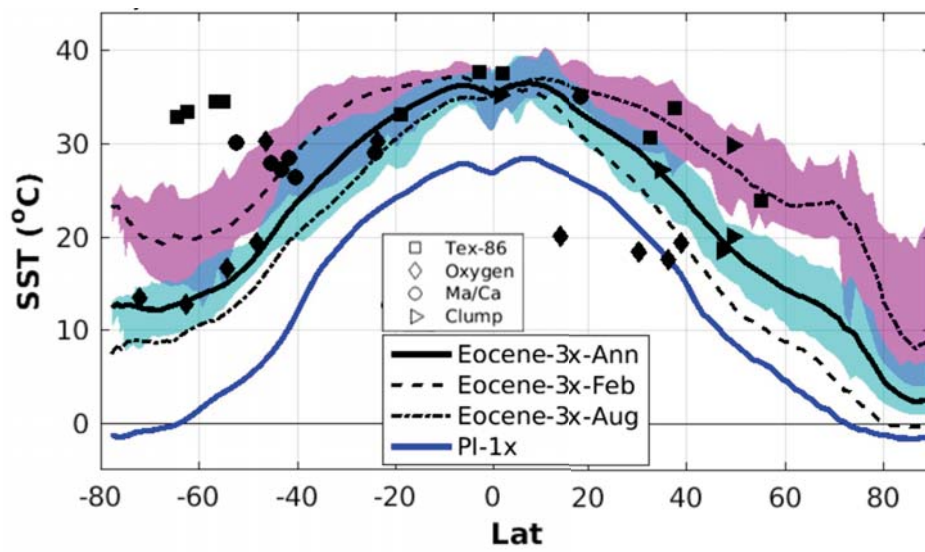


Figure2.

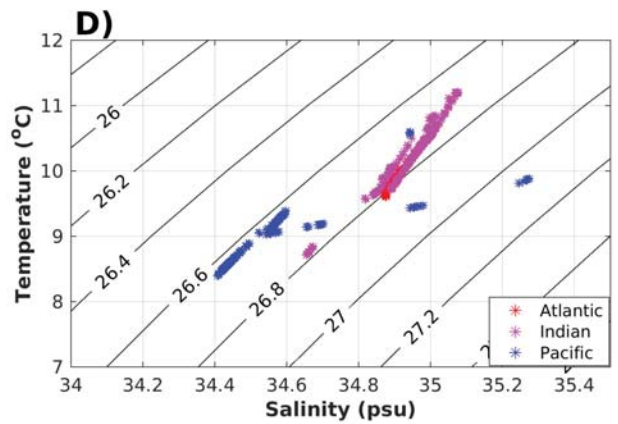
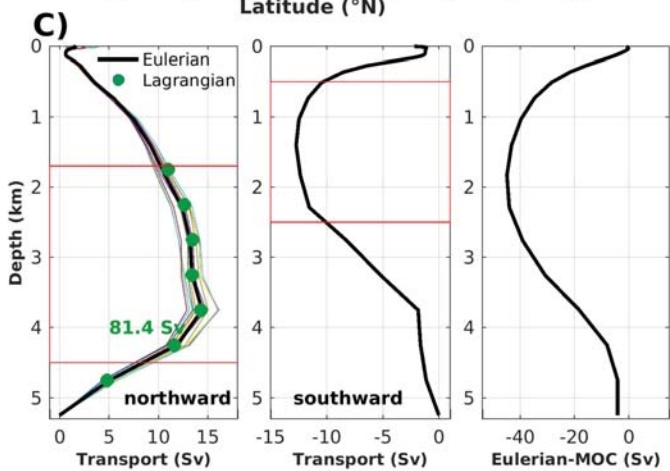
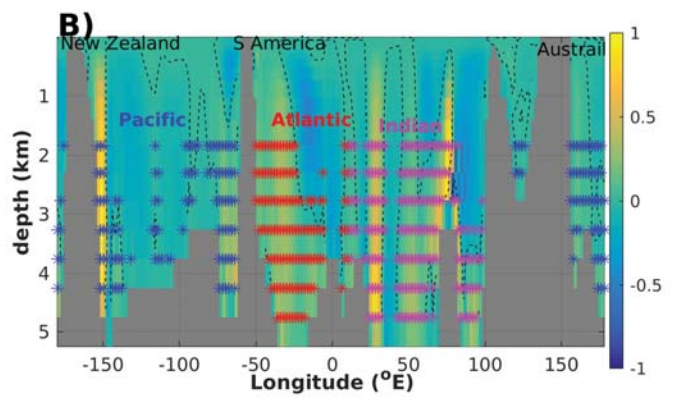
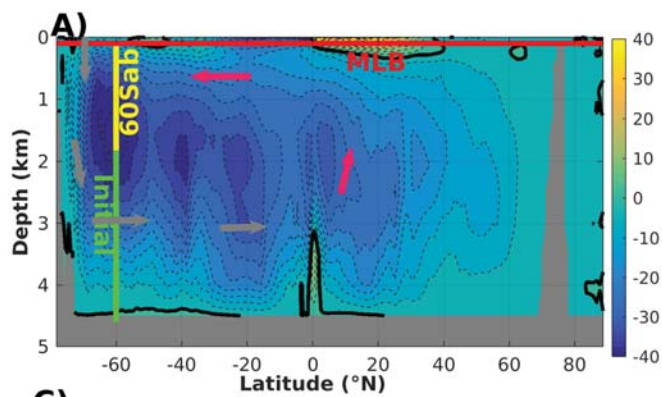


Figure3.

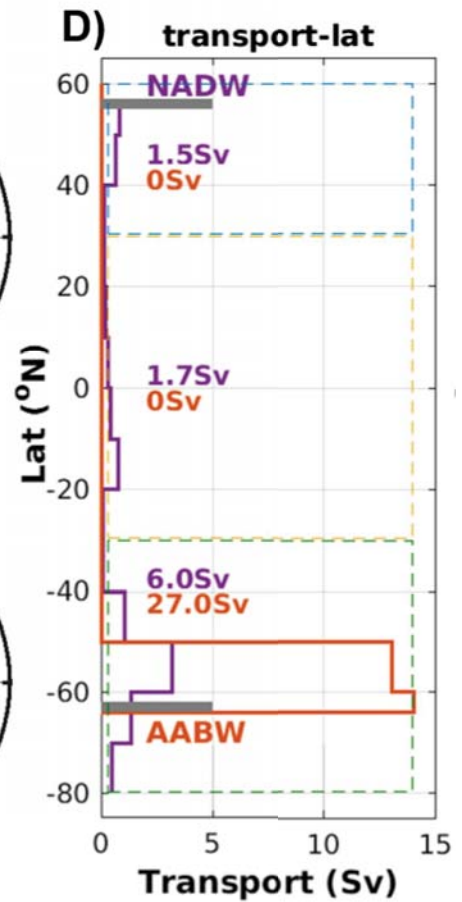
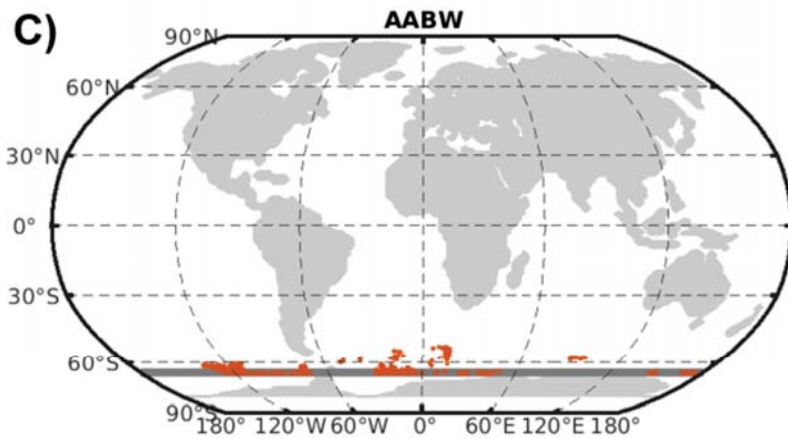
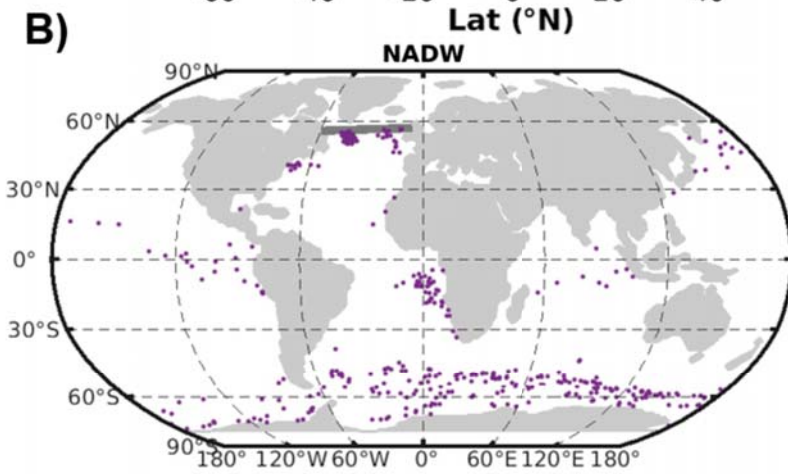
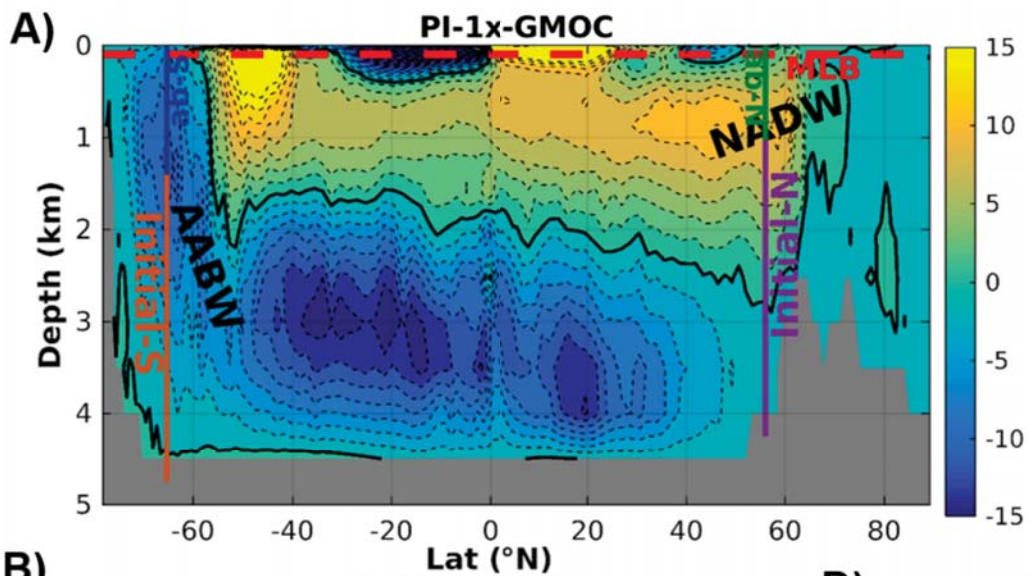


Figure4.

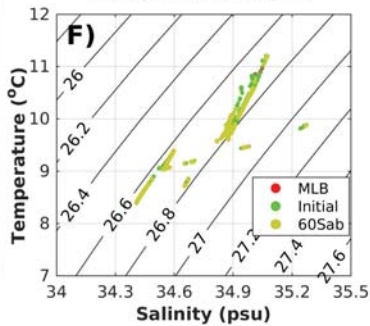
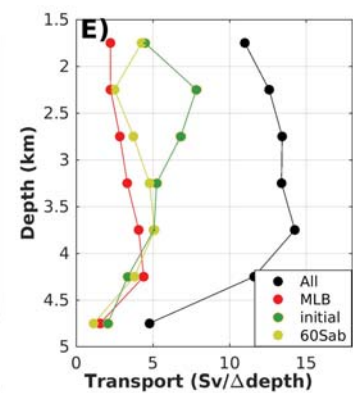
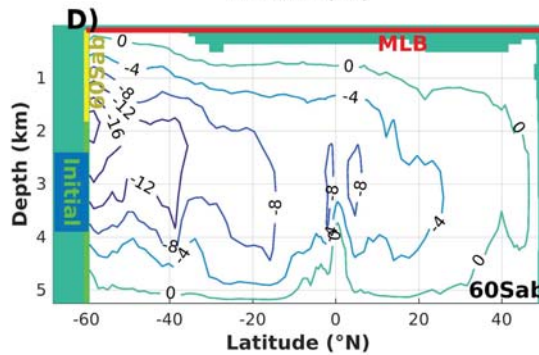
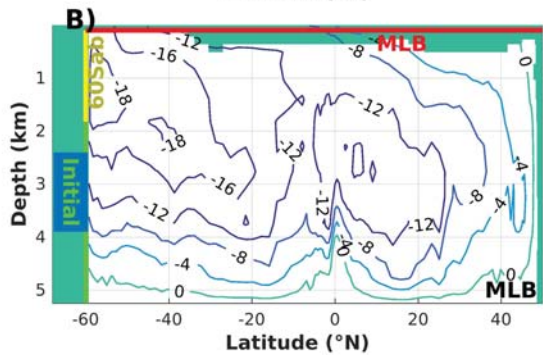
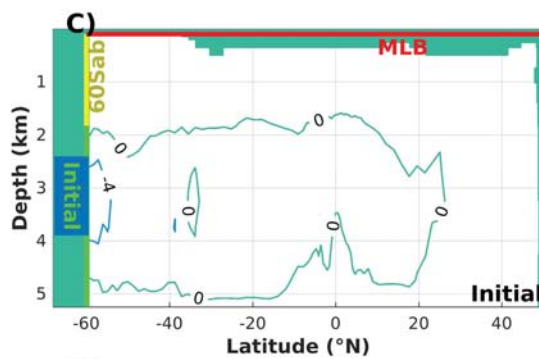
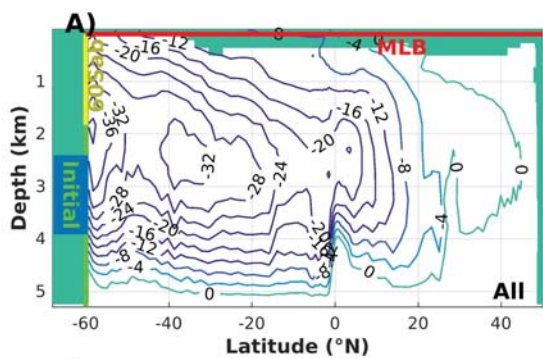


Figure 5.

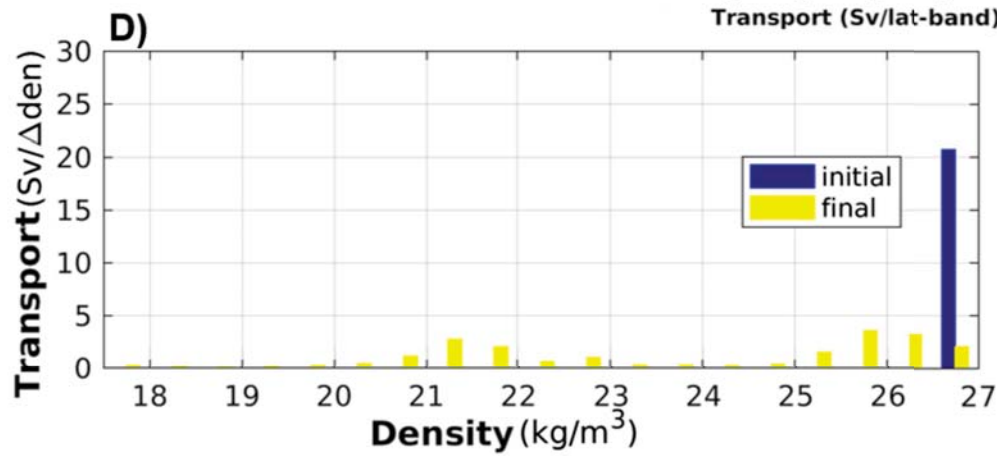
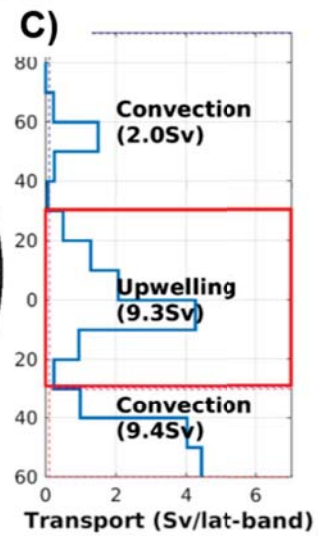
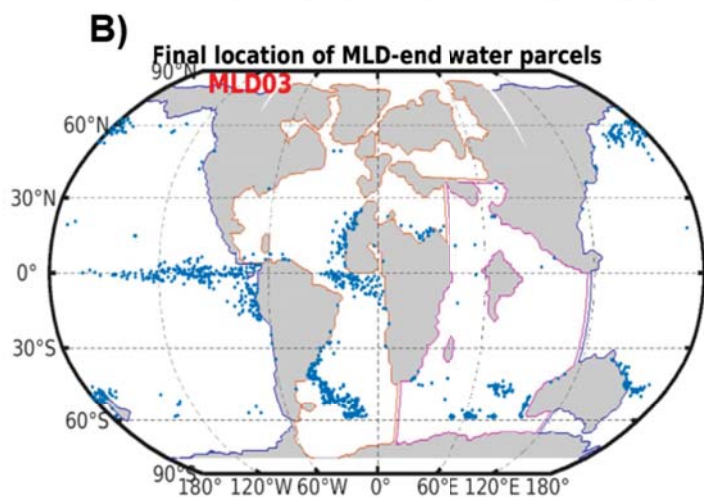
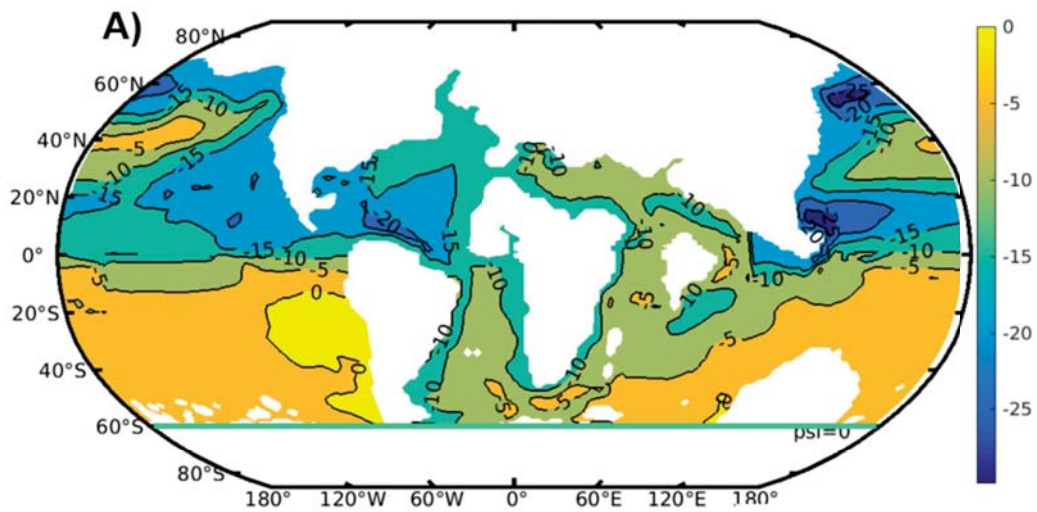


Figure6.

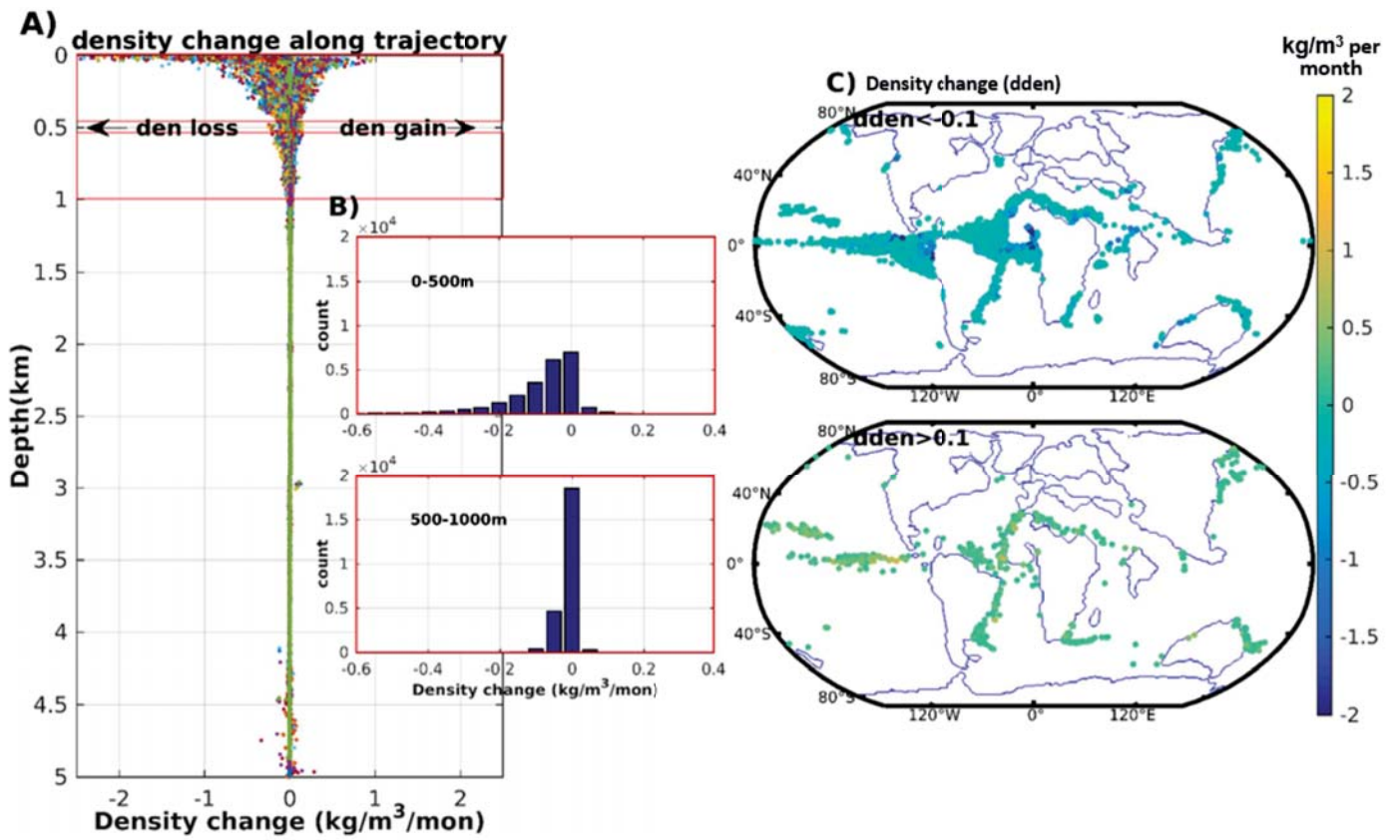


Figure 7.

