

1 **Response of atmospheric pCO₂ to a strong AMOC**
2 **weakening under climate change**

3 **A. A. Boot¹, A. S. von der Heydt^{1,2}, and H. A. Dijkstra^{1,2}**

4 ¹Institute for Marine and Atmospheric research Utrecht, Department of Physics, Utrecht University,
5 Utrecht, the Netherlands

6 ²Center for Complex Systems Studies, Utrecht University, Utrecht, the Netherlands

7 **Key Points:**

- 8 • First results on the carbon cycle response to AMOC weakening in a CMIP6 Earth
9 System Model are presented.
10 • Strong weakening of the AMOC does not result in a large response of atmospheric
11 pCO₂ under climate change.
12 • The spatial patterns of the carbon cycle response to an AMOC weakening are not
13 dependent on cumulative CO₂ emissions.

Corresponding author: Amber Boot, a.a.boot@uu.nl

Abstract

The Earth System is warming due to anthropogenic greenhouse gas emissions which increases the risk of passing a tipping point in the Earth System, such as a collapse of the Atlantic Meridional Overturning Circulation (AMOC). An AMOC weakening can have large climate impacts which influences the marine and terrestrial carbon cycle and hence atmospheric $p\text{CO}_2$. However, the sign and mechanism of this response are subject to uncertainty. Here, we use a state-of-the-art Earth System Model, the Community Earth System Model v2 (CESM2), to study the atmospheric $p\text{CO}_2$ response to an AMOC weakening under low (SSP1-2.6) and high (SSP5-8.5) emission scenarios. A freshwater flux anomaly in the North Atlantic strongly weakens the AMOC, and we simulate a weak positive $p\text{CO}_2$ response of 0.45 and 1.3 ppm increase per AMOC decrease in Sv for SSP1-2.6 and SSP5-8.5, respectively. For SSP1-2.6 this response is driven by both the oceanic and terrestrial carbon cycles, whereas in SSP5-8.5 it is solely the ocean that drives the response. However, the spatial patterns of both the climate and carbon cycle response are similar in both emission scenarios over the course of the simulation period (2015-2100), showing that the response pattern is not dependent on cumulative CO_2 emissions up to 2100. Though the global atmospheric $p\text{CO}_2$ response might be small, locally large changes in both the carbon cycle and the climate system occur due to the AMOC weakening, which can have large detrimental effects on ecosystems and society.

Plain Language Summary

The Atlantic Meridional Overturning Circulation (AMOC) modulates global climate by transporting heat from the Southern to the Northern Hemisphere. The AMOC is considered to be a tipping element with a possible future collapse under climate change. An AMOC weakening can have large climate impacts which influences the marine and terrestrial carbon cycle and hence the atmospheric $p\text{CO}_2$. Here, we use a state-of-the-art Earth System Model to study the atmospheric $p\text{CO}_2$ response to an AMOC weakening under low and high emission scenarios. We use simulations where we artificially weaken the AMOC, which results in a weak positive response of 0.45 and 1.3 ppm $p\text{CO}_2$ increase per decrease in Sv for low and high emissions, respectively. For low emissions this response is driven by both the oceanic and terrestrial carbon cycle processes, whereas in the high emission scenario it is solely the ocean that drives the response. Spatial patterns, both the climate and carbon cycle response, are similar in both emission scenarios over the course of the simulation period (2015-2100). The global atmospheric $p\text{CO}_2$ response is small, but locally large changes in both the carbon cycle and the climate system can occur due to the AMOC weakening.

1 Introduction

Anthropogenic emissions of greenhouse gases cause the Earth System to change and warm up. As temperatures increase, we are at risk of crossing tipping points with possibly large detrimental effects on our climate, biodiversity and human communities (Lenton et al., 2008; McKay et al., 2022). One of these tipping points can occur in the Atlantic Meridional Overturning Circulation (AMOC) (Lenton et al., 2008). Currently, the AMOC is in a so-called on-state where it transports heat from the Southern Hemisphere to the Northern Hemisphere and thereby modulates global and especially European climate (Buckley & Marshall, 2016). In models, the AMOC can be strongly weakened and in this so-called collapsed state (or off-state), the northward heat transport is disrupted with large global climatic effects (Orihuela-Pinto et al., 2022).

Proxy-based evidence suggest that AMOC collapses have occurred frequently during the Pleistocene where they are a main source of millennial variability (e.g. the Dansgaard-Oeschger cycles; Rahmstorf, 2002; Lynch-Stieglitz, 2017). The disrupted heat transport causes warming of surface air temperature (SAT) and sea surface temperature (SST) in

64 the Southern Hemisphere, while the Northern Hemisphere cools (also called the ‘bipo-
65 lar seesaw’; Vellinga & Wood, 2002; Caesar et al., 2018), with local SAT changes up to
66 10°C (Cuffey & Clow, 1997; Rahmstorf, 2002). In models, the bipolar seesaw results in
67 an increased northern hemispheric sea-ice extent and changes in atmospheric dynamics
68 (Vellinga & Wood, 2002; Orihuela-Pinto et al., 2022). The changes in atmospheric dy-
69 namics are, for example, seen in wind fields with strengthened trade winds and strength-
70 ened Pacific Walker Circulation (Orihuela-Pinto et al., 2022), and a southward shift of
71 the Intertropical Convergence Zone (ITCZ) (Zhang & Delworth, 2005; Jackson et al., 2015).
72 The tipping threshold for the AMOC is estimated to be around 4 °C of warming rela-
73 tive to pre-industrial climate (McKay et al., 2022).

74 In addition to the climate system, also the carbon cycle is affected by an AMOC
75 collapse. In the ocean, the change in ocean circulation affects the advection of impor-
76 tant tracers such as Dissolved Inorganic Carbon (DIC) and nutrients (Zickfeld et al., 2008).
77 An AMOC collapse can also change upwelling rates and surface stratification, processes
78 that are important for driving Net Primary Production (NPP) and carbon sequestra-
79 tion in the deep ocean. Terrestrial primary productivity is affected by the changing tem-
80 perature and precipitation patterns. Locally, this can lead to both a reduction or an in-
81 creased uptake of CO₂ (e.g. Köhler et al., 2005). Several studies have looked into a po-
82 tential feedback between AMOC dynamics and atmospheric pCO₂, which is controlled
83 by the exchange of the atmosphere with the ocean and land carbon stocks. These stud-
84 ies (e.g. Marchal et al., 1998; Köhler et al., 2005; Schmittner & Galbraith, 2008), mostly
85 focused on Pleistocene and pre-industrial conditions, show a wide range of possible re-
86 sponses. There is no clear consensus on the responses of the terrestrial and ocean car-
87 bon stock to an AMOC weakening, or to the net effect on atmospheric pCO₂, which can
88 be attributed to different climatic boundary conditions, timescales assessed, and model
89 detail used (Gottschalk et al., 2019). In CMIP6 models, the AMOC gradually weakens
90 up to 2100 and, independent of the used emission scenario (Weijer et al., 2020), no AMOC
91 tipping is found. However, these models are thought to be biased towards a too stable
92 AMOC (e.g. Cheng et al., 2018; Weijer et al., 2019), and a recent observation based study
93 has indicated that the AMOC may tip between 2025 and 2095 (Ditlevsen & Ditlevsen,
94 2023).

95 The carbon cycle is also affected by climate change. In the ocean, the effect on the
96 solubility pump is relatively straight forward: increased warming, and increased CO₂ con-
97 centrations, reduce ocean pH and the solubility of CO₂, which reduces the uptake ca-
98 pacity of the ocean (Sarmiento et al., 1998). The biological pump in Coupled Model In-
99 tercomparison Project 6 (CMIP6; Eyring et al., 2016) models is much more uncertain
100 though (Henson et al., 2022; Wilson et al., 2022), especially given that the spread in NPP
101 and Export Production (EP) has increased from CMIP5 to CMIP6 (Kwiatkowski et al.,
102 2020; Tagliabue et al., 2021). The terrestrial biosphere is affected for example through
103 increased primary production related to CO₂ fertilization (Zhu et al., 2022), but also in-
104 creased respiration due to permafrost melt (Burke et al., 2020).

105 Studies looking at the combined effect of strong AMOC weakening and anthropogenic
106 climate change on the future carbon cycle are limited. A projected AMOC weakening
107 affects both the solubility and the biological carbon pumps (Liu et al., 2023), and gen-
108 erally leads to reduced uptake of (anthropogenic) carbon in the ocean (Obata, 2007; Zick-
109 feld et al., 2008; Liu et al., 2023), which can be partially compensated for by the terres-
110 trial biosphere (Zickfeld et al., 2008). However, the net effect has been found to be small
111 due to competing effects (Swingedouw et al., 2007; Zickfeld et al., 2008). Though global
112 effects might be weak, local effects can be quite strong. For example, a weakening of the
113 AMOC can also result in a local reduction in primary productivity (Whitt & Jansen,
114 2020), changes in the plankton stock (Schmittner, 2005) and plankton composition (Boot
115 et al., 2023a), which all can lead to reduced CO₂ uptake of the ocean (e.g. Yamamoto

116 et al., 2018; Boot et al., 2023a). These local changes related to an AMOC weakening are
117 strongest in the Atlantic Ocean (Katavouta & Williams, 2021).

118 The novel aspect of this paper is that we consider the effect of AMOC weakening
119 on the carbon cycle under climate change in a state-of-the-art global climate model, the
120 Community Earth System Model v2 (CESM2; Danabasoglu et al., 2020), as explained
121 in section 2. We use a strong freshwater forcing in the North Atlantic to artificially weaken
122 the AMOC and consider two different emission scenarios, Shared Socioeconomic Path-
123 ways (SSPs), with low (SSP1-2.6) and high (SSP5-8.5) emissions (O’Neill et al., 2020).
124 In the results of section 3 and the subsequent analysis, we focus on the mechanisms how
125 a forced AMOC weakening affects atmospheric $p\text{CO}_2$ under climate change.

126 2 Method

127 In the CESM2 (Danabasoglu et al., 2020), the atmosphere is represented by the
128 CAM6 model, the land by the CLM5 model (Lawrence et al., 2019), sea ice by the CICE
129 model, ocean circulation by POP2 (Smith et al., 2010), and ocean biogeochemistry by
130 MARBL (Long et al., 2021). The ocean models POP2 and MARBL are both run on a
131 displaced Greenland pole grid at a nominal horizontal resolution of 1° , with 60 non-equidistant
132 vertical levels. The ocean biogeochemical module MARBL is based on a NPZD-model,
133 where four nutrients (N, P, Fe, and Si) together with light co-limit the production of three
134 phytoplankton groups (diatoms, diazotrophs and small phytoplankton) which are grazed
135 upon by one zooplankton group. The terrestrial carbon cycle is represented with CLM5.
136 This module represents several surface processes such as biogeochemistry, ecology, hu-
137 man influences, biogeophysics and the hydrological cycle. As we use the default CESM2
138 version, there is no dynamic vegetation. For a complete overview of the CESM2 model
139 and submodules we refer the reader to Danabasoglu et al. (2020) (CESM2), Long et al.
140 (2021) (MARBL), and Lawrence et al. (2019) (CLM5).

141 We performed emission forced CESM2 simulations with two different emission sce-
142 narios, the low emission scenario SSP1-2.6 (126) and the high emission scenario SSP5-
143 8.5 (585). For each emission scenario, a control (CTL) and a hosing (HOS) simulation
144 were carried out. The CTL simulations were only forced with the greenhouse gas emis-
145 sions, while the HOS simulations were forced with greenhouse gas emissions and an ad-
146 ditional, artificial freshwater flux in the North Atlantic. This freshwater forcing is located
147 in the North Atlantic Ocean over the latitudes 50°N - 70°N (Fig. S1), and is kept con-
148 stant at a rate of 0.5 Sv over the entire simulation period. We will refer to the simula-
149 tions by their simulation type (CTL or HOS) and the respective emission scenario (126
150 or 585), e.g. as CTL-126 and HOS-585. All simulations are run from year 2015 to year
151 2100 and are initialized by values of the NCAR CMIP6 emission driven historical sim-
152 ulation (Danabasoglu, 2019). The used model output is based on monthly means, and
153 line plots are smoothed with a 5 year running mean. When looking at the difference be-
154 tween the HOS and CTL simulations, we subtract the CTL simulations from the HOS
155 simulations.

156 3 Results

157 3.1 Climate response

158 In CTL-126, an increase in atmospheric CO_2 concentration from 400 ppm to 467 ppm
159 in the 2050s is found, after which the concentration decreases to 432 ppm in 2100 (Fig.
160 1c). This is accompanied by an increase in global mean surface temperature (GMST)
161 of 1°C (Fig. 1b), and an AMOC decrease from 17 Sv in 2015 to 9 Sv in 2100 (Fig. 1a).
162 The weakening of the AMOC results in a cooling of the North Atlantic Ocean, while the
163 rest of the Earth warms with largest temperature increases found near the poles (Fig.
164 2a, b) as a response to the increase in greenhouse gas concentrations. In the water cy-

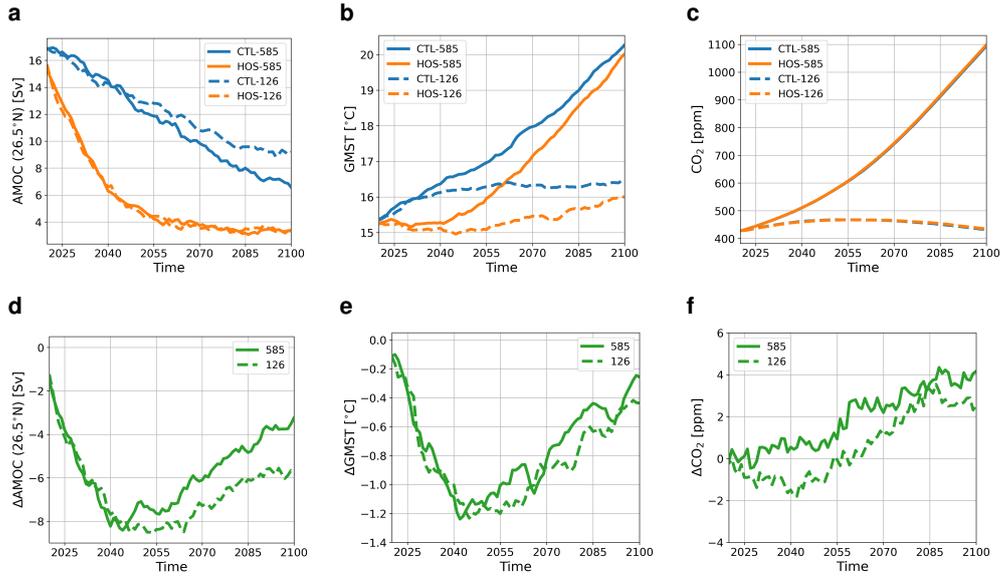


Figure 1. (a) AMOC strength at 26.5°N in Sv. (b) GMST in °C. (c) Atmospheric CO₂ concentration in ppm. In (a-c) blue lines represent the control (CTL) simulations, and orange lines the HOS simulations. (d-f) as in (a-c) but for the difference between the HOS simulations and the control simulations. In all subplots dashed lines represent SSP1-2.6 (126) and solid lines SSP5-8.5 (585).

165 cle we see a southward shift of the Pacific InterTropical Convergence Zone (ITCZ) of a
 166 few degrees (Fig. S2a, b). Furthermore, wind fields in the Northern Hemisphere show
 167 a small weakening, whereas in the Southern Hemisphere the winds intensify (Fig. S3a,
 168 b).

169 In CTL-585, the emissions increase the atmospheric CO₂ concentration from 400 ppm
 170 to 1094 ppm in 2100 (Fig. 1c) which results in a GMST warming of 5 °C (Fig. 1b). The
 171 AMOC weakens from 17 Sv to 7 Sv (Fig. 1a), which leads to a region without warm-
 172 ing in the North Atlantic, whereas we see strong warming everywhere else (Fig. 2d, e).
 173 There is a strong southward shift of the ITCZ in the Pacific and a moderate shift in the
 174 Atlantic Ocean (Fig. S2d, e). The changes in the wind field show similar patterns as CTL-
 175 126 but with a larger amplitude (Fig. S3d, e).

176 The net effect of the AMOC weakening (i.e. HOS minus CTL) is shown in Fig. 1def.
 177 In the year 2100, atmospheric CO₂ concentrations are 2.6 ppm and 4.2 ppm higher in
 178 HOS-126 and HOS-585 compared to their respective CTL simulations. In both HOS sim-
 179 ulations the AMOC quickly weakens from 17 Sv in 2015 to 6 Sv in 2045 after which the
 180 AMOC weakening starts to level off until the AMOC is weaker than 4 Sv in 2100 (Fig.
 181 1d). Due to the AMOC weakening we observe a relative cooling of (locally) more than
 182 3 °C in the Northern Hemisphere and warming in the Southern Hemisphere (Fig. 2c,
 183 f) (i.e. the bipolar seesaw). The cooling in the Northern Hemisphere results into an in-
 184 crease in sea-ice cover of the Arctic Ocean (Fig. S4), which for HOS-126 persists through-
 185 out the entire simulation period. The AMOC weakening also results into a stronger south-
 186 ward shift of the ITCZ in both the Pacific and Atlantic Ocean (Fig. S2c, f), and winds
 187 are relatively intensified in the Northern Hemisphere and weakened in the Southern Hemi-
 188 sphere (Fig. S3c, f), with a stronger response in SSP5-8.5.

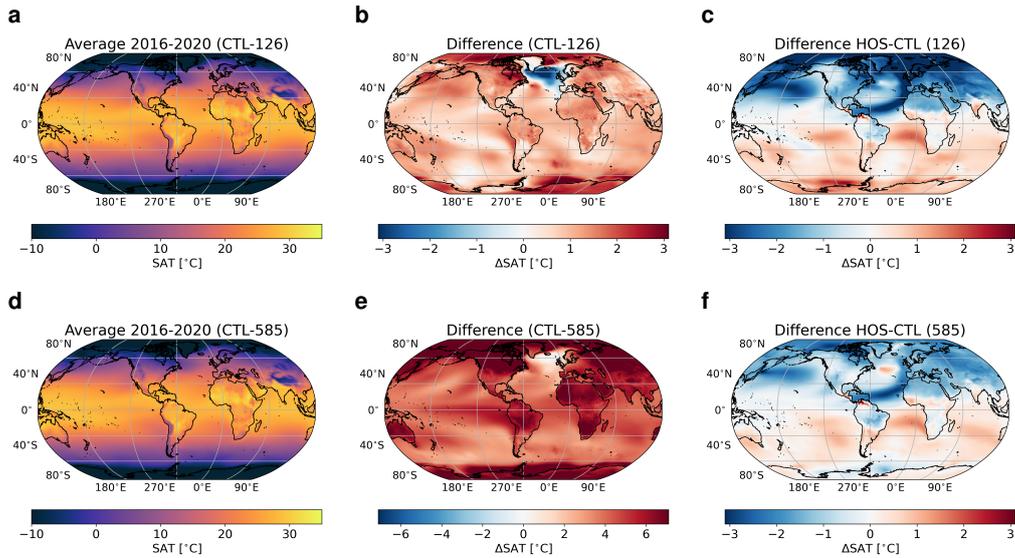


Figure 2. Results for Surface Air Temperature (SAT) in $^{\circ}\text{C}$. The top row (a-c) is for SSP1-2.6, and the bottom row (d-f) for SSP5-8.5. The left column (a, d) represents the average over 2016-2020 in the control simulations. The middle row (b, e) represents the difference between the average of 2096-2100 and 2016-2020 for the control simulations. The right row (c, f) represents the difference between the HOS and CTL simulations averaged over 2096-2100. Note the different scaling between b and e.

189

3.2 Marine carbon cycle response

190

191

192

193

194

195

196

197

198

199

200

201

In CTL-126 we see that, integrated over the entire simulation period, there are regions in the ocean with net carbon uptake, and net carbon outgassing (Fig. 3a). The Southern Ocean between 45°S and 60°S , and the equatorial Pacific Ocean, are regions of carbon release from the ocean to the atmosphere. The region of strongest outgassing in the Pacific is located in the upwelling regions on the eastern side of the basin. Carbon uptake generally occurs in the rest of the ocean with the strongest uptake located in the Sea of Japan and the high latitude North Atlantic Ocean. Looking at the development over time (Fig. 4a, b) we see a negative trend over almost the entire ocean, meaning regions which take up carbon in the beginning of the simulation have lower uptake at the end, and regions which emit carbon in 2015 emit more carbon at the end of the simulation. Some regions, e.g. in the Southern Ocean, shift from a carbon uptake region to a region of outgassing.

202

203

204

205

206

207

208

209

210

211

212

In CTL-585, also integrated over the simulation period, only the eastern equatorial Pacific shows strong outgassing (Fig. 3d). In the other equatorial basins, there are also some small patches that show net outgassing, but the rest of the ocean shows net carbon uptake. Except for the high latitude North Atlantic Ocean and some small other regions, we see a positive trend (Fig. 4d, e), meaning that regions that take up carbon in the beginning, take up more carbon at the end of the simulation, and regions which show outgassing in the beginning show either reduced outgassing or go from being a region of outgassing to a region of CO_2 uptake. A remarkable region is the high latitude North Atlantic Ocean where the flux from the atmosphere into the ocean strongly decreases while atmospheric pCO_2 almost triples. Integrated over time, the spatial pattern of regions that see increased or decreased exchange with the atmosphere is very sim-

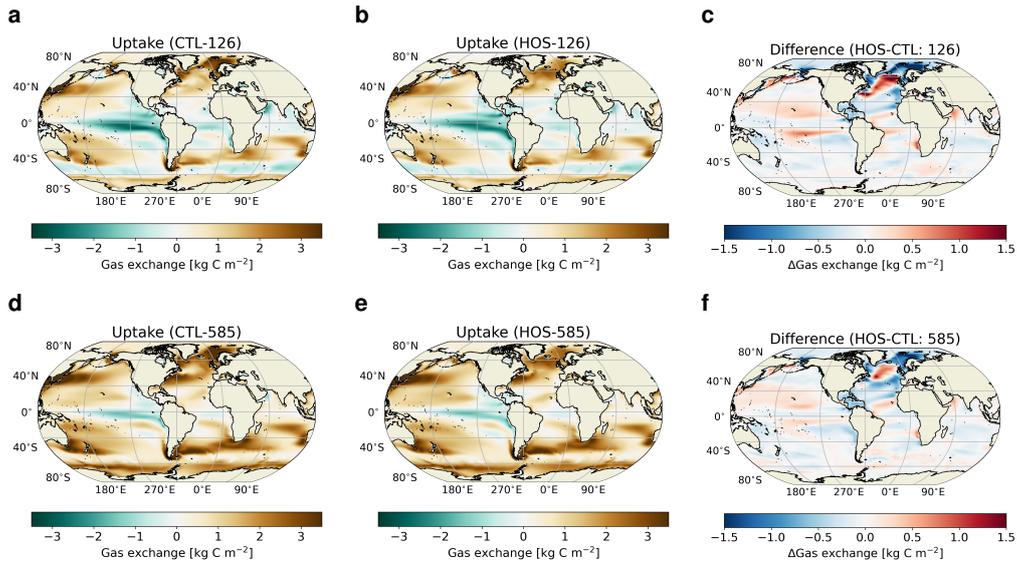


Figure 3. Results for the oceanic CO₂ uptake integrated over the entire simulation period in kg C m⁻². The top row (a-c) represents SSP1-2.6 and the bottom row (d-f) represents SSP5-8.5. The left column (a, d) represents the uptake in the control simulations, the middle column (b, e) the uptake in the HOS simulations, and the right column (c, f) the difference between the HOS and CTL simulations. In a, b, d, and e positive values (brown colors) represent net uptake, and negative values (blue colors) represent net outgassing.

213 ilar for SSP1-2.6 as for SSP5-8.5 (Fig. 3c, f). In total, the ocean takes up 7.4 PgC less
 214 due to the AMOC weakening in SSP1-2.6 and 15.6 PgC less in SSP5-8.5 (Fig. 5a, d).

215 Even though the climate system changes a lot due to the AMOC weakening, the
 216 CO₂ uptake of the ocean does not change a lot because of compensating effects. To ob-
 217 tain a better understanding of the mechanisms behind the reduced uptake, we have di-
 218 vided the ocean into 5 basins: the Arctic (north of 66°N), the Southern (south of 35°S),
 219 the Atlantic, Pacific and Indian Ocean (Fig. 5b, e). In the response (i.e. HOS-CTL), for
 220 both emission scenarios, all basins show the same sign, i.e. more uptake or less uptake
 221 due to the AMOC weakening.

222 In both emission scenarios the Arctic Ocean shows a decreased uptake (-6.0 PgC
 223 in SSP1-2.6 and -4.4 PgC in SSP5-8.5), which can be explained by looking at the sea-
 224 ice cover (Fig. S4). The cooling in the Northern Hemisphere following the AMOC weak-
 225 ening in the HOS simulations, increases the sea-ice cover. The increase in sea-ice cover
 226 has two effects on the uptake of CO₂: (1) it reduces the ocean area available for exchange
 227 with the atmosphere; and (2) it increases light limitation and thereby reduces net pri-
 228 mary production (NPP; Fig. S6) and the carbon export to the subsurface ocean. In SSP5-
 229 8.5 most of the sea ice still disappears due to the strong warming, but in SSP1-2.6 most
 230 of the sea ice persists throughout the simulation period, which explains why the Arctic
 231 Ocean in SSP1-2.6 responds stronger compared to SSP5-8.5. We also find this effect in
 232 the sea-ice covered regions in the North Atlantic (e.g. the Labrador Sea).

233 The Pacific Ocean takes up more carbon in the HOS than in the CTL simulations
 234 (+4.9 PgC in SSP1-2.6 and +1.7 PgC in SSP5-8.5). To analyze what is happening in
 235 the Pacific, we considered three different regions: (1) the North Pacific (20°N-66°N), the
 236 Equatorial Pacific (20°N-10°S), and the South Pacific (10°S-35°S). In the North Pacific,

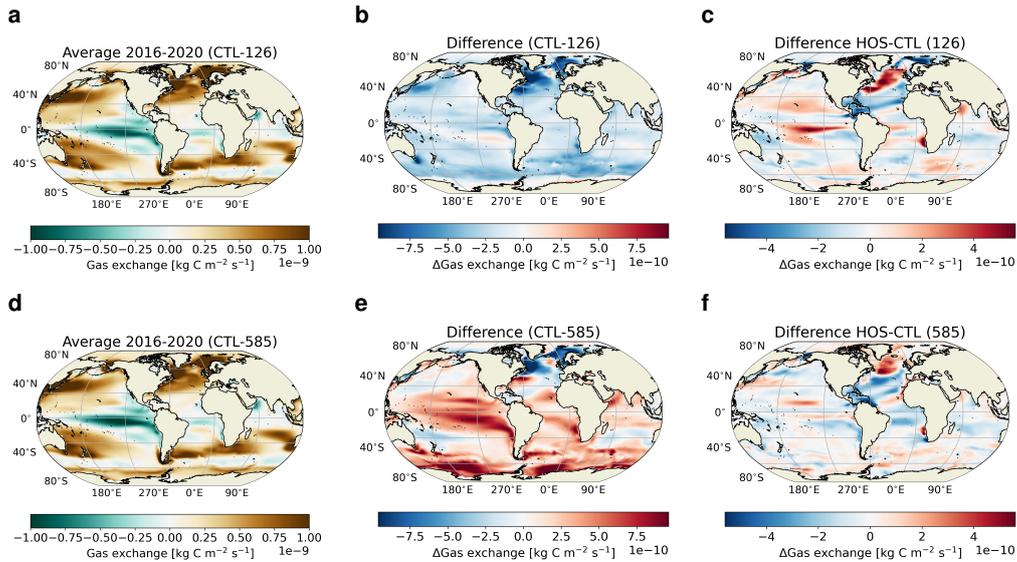


Figure 4. Results for oceanic CO_2 uptake in $\text{kg C m}^{-2} \text{s}^{-1}$. Panels represent the same as in Fig. 2. Positive values (brown colors) in a and d represent uptake by the ocean and negative values (blue colors) represent outgassing.

237 the relative cooling of the surface ocean (Fig. S7) results in an increase of solubility of
 238 CO_2 driving increased uptake (Fig. 3e, f). A similar, but opposite, response is seen in
 239 the South Pacific. Here the surface ocean becomes relatively warmer inhibiting the up-
 240 take of CO_2 . The equatorial Pacific is characterized by a band with reduced uptake and
 241 one with increased uptake. This can be related to the stronger southward shift of the
 242 ITCZ in the Pacific in HOS compared to the CTL (Fig. S2). Due to this shift, the di-
 243 lutive fluxes related to net precipitation shift southward, causing relative increases of salin-
 244 ity in the northern section due to reduced precipitation, and relative decreases due to
 245 increased precipitation in the southern section (Fig. S8). This, in turn, also affects the
 246 stratification in these regions with a weakening in the north and a strengthening in the
 247 south (Fig. S9). These changes affect the solubility of CO_2 in the equatorial regions caus-
 248 ing decreased uptake in the northern section and increased uptake in the southern sec-
 249 tion.

250 We find the largest difference in carbon uptake (-2.0 PgC in SSP1-2.6 and -9.3 PgC
 251 in SSP5-8.5) in the Atlantic. The regions with sea ice show similar behavior as the Arct-
 252 ic Ocean with decreased uptake related to a larger sea-ice cover in the HOS simulations.
 253 In the ice-free subpolar region, an increase in uptake is observed which is associated to
 254 decreases in sea surface salinity (SSS; Fig. S8) due to the applied freshwater forcing in
 255 this region which promotes the uptake of CO_2 . In the subtropical region we generally
 256 see a decrease in uptake. To explain this we consider several variables, i.e. SST (Fig. S7),
 257 SSS (Fig. S8), DIC (Fig. S12), Alk (Fig. S13) and NPP (Fig. S6), which all show a rela-
 258 tive decrease in this region. The net effect of the changes in these variables is a reduc-
 259 tion in pH (Fig. S16) and reduced uptake capacity of the ocean. In the Canary Upwelling
 260 System and along the North Equatorial Current we do see an increase in NPP (Fig. S6),
 261 due to increased nutrient concentrations (Fig. S11) related to increased upwelling of nu-
 262 trients (Fig. S10 and S15). In the region of the North Equatorial Current this leads to
 263 increased uptake of the ocean, and only in SSP5-8.5 also in the Canary Upwelling Sys-
 264 tem. Outside the North Atlantic, large responses are seen in the equatorial region and
 265 the Benguela Upwelling System which are characterized by reduced upwelling (Fig. S10),

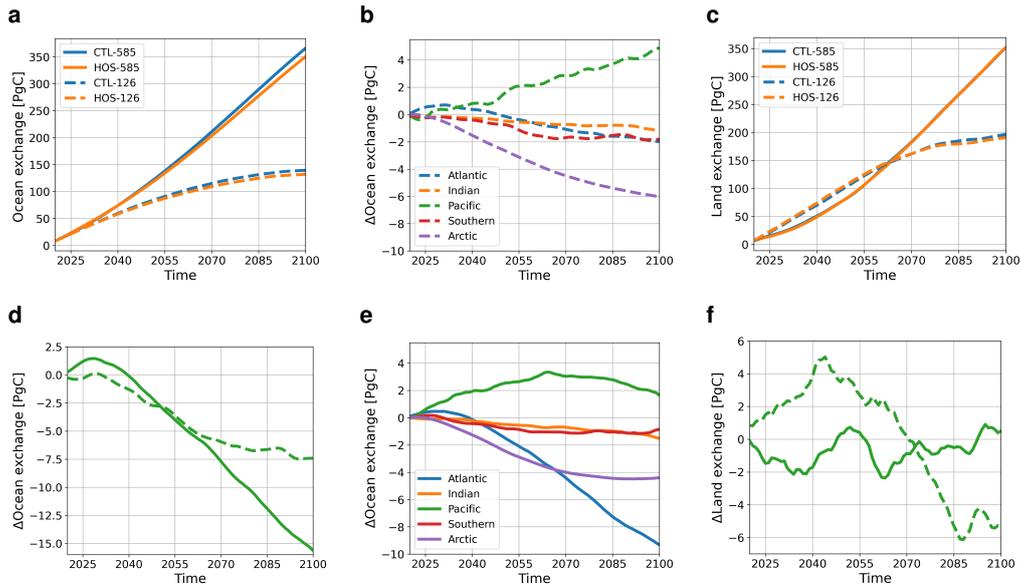


Figure 5. (a) Cumulative uptake of CO_2 in the ocean from 2016 onward in PgC. (b) Difference in the cumulative oceanic CO_2 uptake between the HOS and CTL simulations in SSP1-2.6 for different ocean basins. (c) As (a) but for the land. (d) The difference in the cumulative oceanic CO_2 uptake between the HOS and CTL simulations. (e) As in (b) but for SSP5-8.5. (f) As in (d) but for the land. In a and c blue lines represent the control simulations, and the orange lines the HOS simulations. In all subplots dashed lines represent SSP1-2.6 and solid lines SSP5-8.5. Negative values in b, d-f represent reduced uptake in the HOS simulations compared to the CTL simulations.

266 promoting additional uptake of CO_2 in the ocean. In the Atlantic Ocean, we find that
 267 DIC (Fig. 6) and nutrient (Fig. 7) concentrations decrease in the surface ocean due to
 268 the weakening of the AMOC and increase in the deep ocean. The reduction in DIC clearly
 269 shows the reduced uptake capacity of the ocean, and the reduction in PO_4 also explains
 270 the decrease in NPP (Fig. S6) observed in the Atlantic basin.

271 The Indian Ocean has a relatively weak response and is very similar for both emis-
 272 sion scenarios with a small decrease in uptake (-1.2 PgC in SSP1-2.6 and -1.5 PgC in
 273 SSP5-8.5). This is related to the relatively warmer SSTs in the HOS simulations (Fig.
 274 S7). The Southern Ocean also has a small decrease in uptake, with a larger decrease in
 275 SSP1-2.6 (-1.8 PgC compared to -0.9 PgC in SSP5-8.5). This larger decrease can be ex-
 276 plained by the fact that the sea-ice cover is larger in SSP1-2.6 compared to SSP5-8.5 (Fig.
 277 S5).

278 3.3 Terrestrial carbon cycle response

279 In CTL-126, the terrestrial biosphere, integrated over the entire simulation period,
 280 shows a net uptake of CO_2 in most regions (Fig. 8a). The Net Biosphere Production (NBP)
 281 maxima are located on the equator for the tropical rainforests, the boreal forests in the
 282 high latitude Northern Hemisphere, and the eastern United States and China. The few
 283 locations that show net emission of CO_2 are very local and present in the high latitude
 284 Northern Hemisphere, the Tibetan Plateau, South East Asia and South America. If we
 285 look at the development over time (Fig. 9a, b) we see that the tropical rainforests have

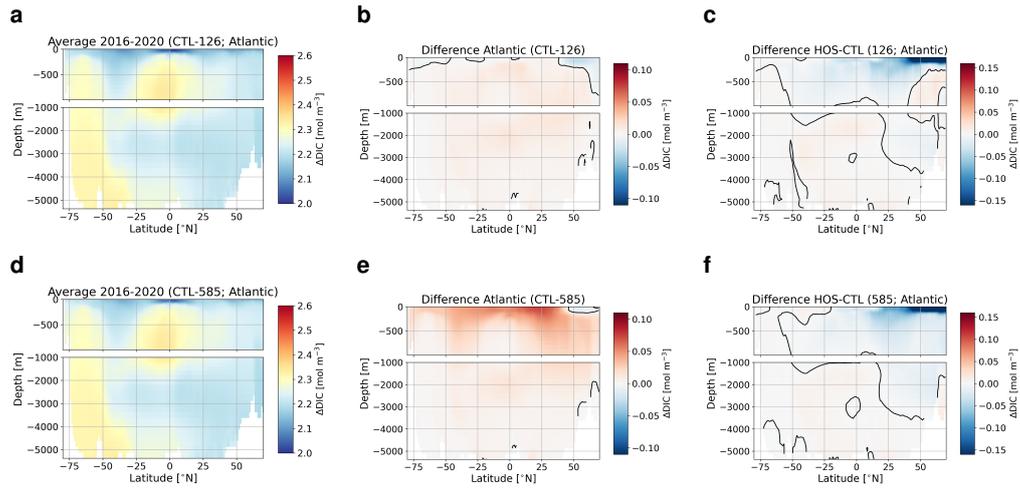


Figure 6. Results for zonally averaged DIC concentrations in the Atlantic basin in mol m^{-3} . Panels represent the same as in Fig. 2. Black contour lines in b, c, e and f represent the 0 mol m^{-3} contour. Note the different scaling of the surface ocean (top 1000 m) compared to the deep ocean.

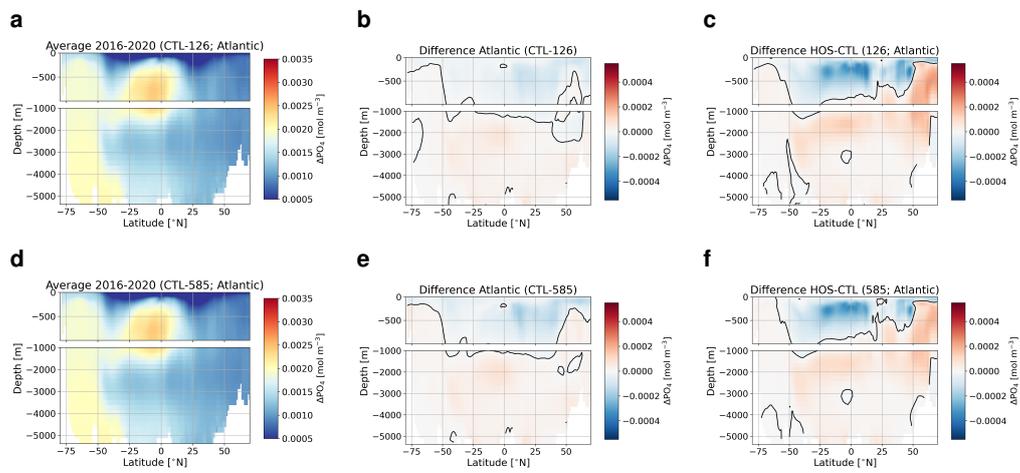


Figure 7. Results for zonally averaged PO_4 concentrations in the Atlantic basin in mol m^{-3} . Panels represent the same as in Fig. 2. Black contour lines in b, c, e and f represent the 0 mol m^{-3} contour. Note the different scaling of the surface ocean (top 1000 m) compared to the deep ocean.

286
287

a lower NBP at the end of the simulation. There are some regions that have a higher NBP in 2100, e.g. the boreal forests in Scandinavia.

288 The response in CTL-585 is very similar to CTL-126 with respect to the spatial
 289 pattern, except in central Africa (Fig. 8d). However, the amplitude of the response is
 290 much larger due to the CO₂ fertilization effect. Especially the tropical rainforests, but
 291 also the boreal forests, show more carbon uptake compared to CTL-126. The same is
 292 also true for regions that emit carbon, i.e., the region in the high latitude Northern Hemi-
 293 sphere that emits carbon is larger, and the amount of carbon emitted is also higher. The
 294 main difference with respect to CTL-126 is a region in the Congo basin which emits CO₂
 295 in CTL-585 whereas in CTL-126 it is a region of relatively strong uptake, which is possi-
 296 bly related to increased wildfire activity in this region in SSP5-8.5 (Fig. S17). When
 297 we look at the development over time (Fig. 9d, e) we find a completely different pattern
 298 in CTL-585 compared with CTL-126. The tropical rainforests show an increase in NBP
 299 related to the CO₂ fertilization effect whereas northern Siberia shows a decrease related
 300 to increased respiration due to permafrost melt (Fig. S19 and S20).

301 Integrated globally the terrestrial biosphere takes up 5.3 PgC less in SSP1-2.6 and
 302 0.5 PgC more in SSP5-8.5 (Fig. 5) in the HOS simulations compared to the CTL simu-
 303 lations. However, looking at spatial patterns of the cumulative uptake, we see a very
 304 similar response to the AMOC weakening (HOS-CTL) for both emission scenarios (Fig.
 305 8c, f). In both emission scenarios we find that the increased southward shift in the ITCZ
 306 in the HOS simulations lead to decreased NBP in central America, and increased NBP
 307 in Southern America. A similar shift can be seen in Africa, but with a smaller latitudi-
 308 nal shift and amplitude. The shift and amplitude are slightly stronger in SSP1-2.6. The
 309 boreal forests become relatively lower in NBP in the HOS simulations with a larger am-
 310 plitude in SSP1-2.6. This is because in SSP1-2.6, the forests have lower Gross Primary
 311 Production (GPP; Fig. S18) over the course of the century which can be related to the
 312 relative cooling in the Northern Hemisphere seen in the HOS simulations (Fig. S8). This
 313 relative cooling is stronger in SSP1-2.6, related to the increased sea-ice cover and there-
 314 fore higher albedo in the Arctic. Another effect of the Northern Hemispheric cooling is
 315 an increase in NBP in the permafrost regions in Siberia and North America in the HOS
 316 simulations. The cooling reduces permafrost melt (Fig. S19) and therefore reduces soil
 317 respiration (Fig. S20), with a larger amplitude in Siberia for SSP5-8.5.

318 3.4 Total response

319 In total we see an increase of atmospheric CO₂ concentration of 2.6 and 4.2 ppm
 320 in 2100 in SSP1-2.6 and SSP5-8.5 due to the AMOC weakening (HOS-CTL). In SSP1-
 321 2.6 this response is caused partly due to reduced uptake of the ocean and partly due to
 322 reduced uptake of the land. In SSP5-8.5 it is completely driven by the ocean as the glob-
 323 ally integrated uptake over the land is approximately the same in CTL-585 as in HOS-
 324 585. Eventually the AMOC strength in 2100 has decreased by 5.8 and 3.2 Sv in the HOS
 325 simulations compared to the CTL simulations. Under the assumption of linearity, this
 326 results in a positive feedback strength of 0.44 ppm Sv⁻¹ and 1.3 ppm Sv⁻¹ for SSP1-
 327 2.6 and SSP5-8.5 respectively. This can be considered a positive feedback since increased
 328 CO₂ concentrations in future climates are generally associated with a weakening of the
 329 AMOC (e.g. Weijer et al., 2020). This AMOC-pCO₂ feedback is small on the global scale,
 330 due to competing effects but locally large changes in carbon uptake can occur.

331 Fig. 10 gives an overview of the most important climate changes and how the ma-
 332 rine and terrestrial respond to these changes. In Fig. 10c, d the difference between SSP1-
 333 2.6 and 5-8.5 is highlighted. In the terrestrial biosphere the prime effect of the AMOC
 334 weakening is the southward shift of the GPP maxima in the tropical rainforests (Fig. S18).
 335 Though this could potentially have beneficial effects for the southern regions, it could
 336 have detrimental effects for the northern regions (e.g. the Sahel region) and could for
 337 example increase the latitudinal extent of the Sahara desert. This shift, caused by a shift
 338 in precipitation (Fig. S2), also has effects for the probability of wildfires (Fig. S17), which
 339 can increase in regions with reduced precipitation. We cannot conclude whether the AMOC

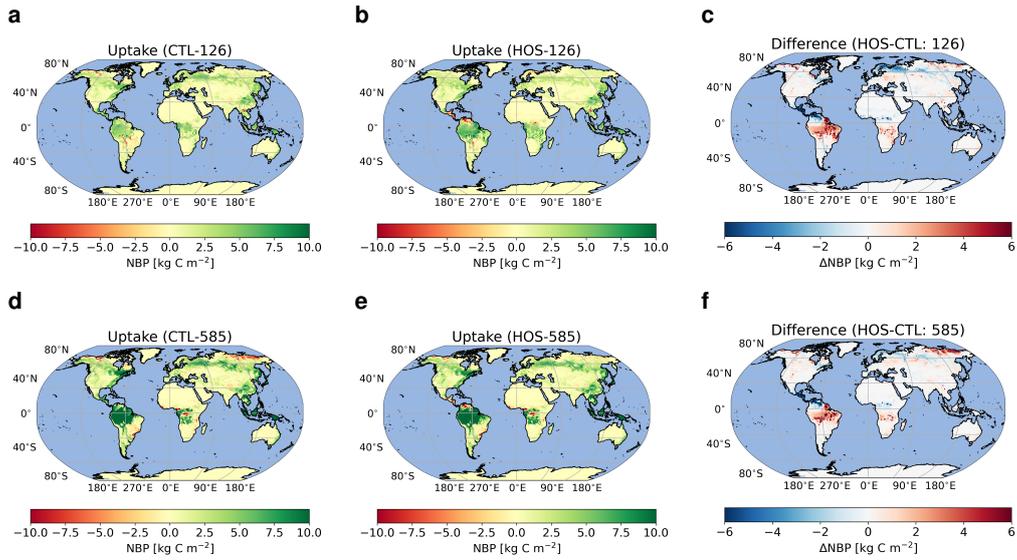


Figure 8. Results for the CO_2 exchange with the land integrated over the entire simulation period in kg C m^{-2} . The top row (a-c) represents SSP1-2.6 and the bottom row (d-f) represents SSP5-8.5. The left column (a, d) represents the uptake in the control simulations, the middle column (b, e) the uptake in the HOS simulations, and the right column (c, f) the difference between the HOS and CTL simulations. In a, b, d, and e green colors represent net CO_2 uptake by the land, and red colors represent net emissions into the atmosphere.

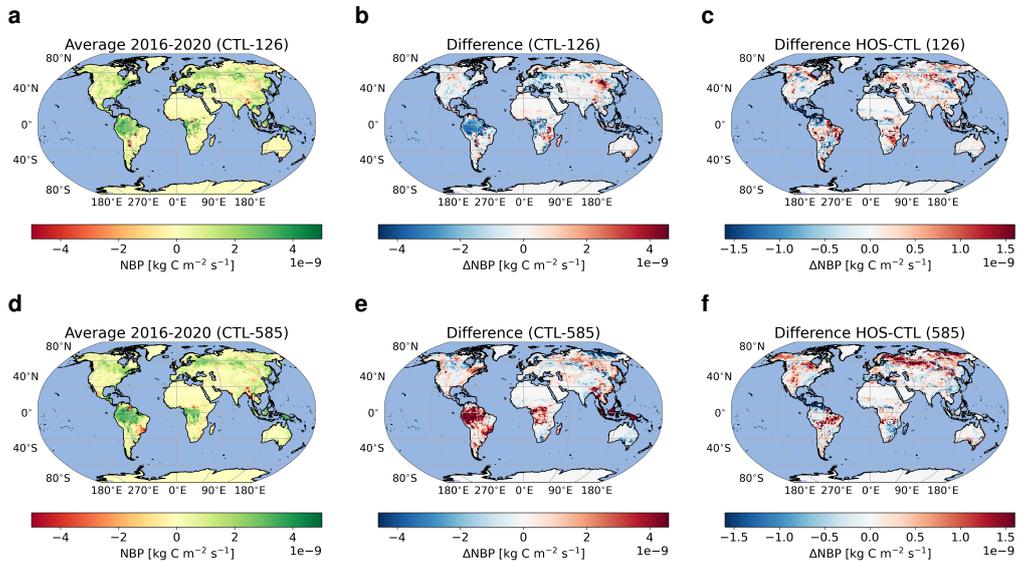


Figure 9. Results for Net Biosphere Production (NBP) in $\text{kg C m}^{-2} \text{ s}^{-1}$. Panels represent the same as in Fig. 2. Green colors represent uptake of CO_2 into the land and red colors represent emission of CO_2 to the atmosphere.

340 weakening would result into a collapse of the Amazonian rainforests or an increase in the
 341 Sahara desert since the model is used without a dynamic vegetation model. In the ocean,

342 a decrease in NPP (Fig. S6) and surface nutrient concentrations (Fig. S11) occurs. The
 343 changes in NPP can have effects on the entire food web and thereby have a negative im-
 344 pact on ecosystems and ecosystem functions. If the trend of the surface ocean becom-
 345 ing more depleted of nutrients (Fig. 7) continues, this might drive a large decline in NPP
 346 for the coming centuries. Another important effect of the AMOC weakening is increased
 347 ocean acidification (i.e. a decrease in pH; Fig. S16). Lower pH values increase the stress
 348 on calcifying organisms and reduces the uptake capacity of the ocean, which might in-
 349 crease the AMOC-pCO₂ feedback strength on longer timescales.

350 In many climate and carbon cycle variables we see a similar response in spatial pat-
 351 tern, but sometimes with a slightly different amplitude (Fig. 10c, d). In the terrestrial
 352 biosphere, the main differences are seen in the boreal forests in Scandinavia and Rus-
 353 sia (box 1 in Fig. 10), and in the Siberian permafrost regions (box 2). The difference in
 354 the boreal forests can be explained by looking at the temperature differences between
 355 the HOS and CTL simulations. In SSP1-2.6, the northern hemisphere cools more, which
 356 causes increased GPP reduction in the boreal forests. For the permafrost region we find
 357 a stronger response in SSP5-8.5, because in SSP1-2.6 there is not much permafrost melt
 358 in the CTL simulation; therefore the additional cooling in the HOS simulation does not
 359 have a large effect on the permafrost melt. In the ocean, we find the largest changes in
 360 the subpolar North Atlantic and the Arctic sea-ice regions (boxes 7 and 8 in Fig. 10).
 361 In the subpolar region there is a relatively stronger decrease in SSS and SST (Fig. S7
 362 and S8) in SSP1-2.6 compared to 5-8.5 leading to a larger increase in solubility of CO₂
 363 and therefore more uptake. Because of the increased cooling, and lower background tem-
 364 peratures in SSP1-2.6, sea-ice cover does not diminish over the simulation whereas in SSP5-
 365 8.5 we see in both simulations a strong reduction in sea-ice cover (Fig. S4). This is the
 366 reason why we see a stronger reduction in the Arctic in SSP1-2.6.

367 4 Summary and discussion

368 In this study, we have investigated the carbon cycle response to a weakening of the
 369 Atlantic Meridional Overturning Circulation (AMOC) under climate change scenarios.
 370 We did this by forcing a state-of-the-art Earth System Model, the Community Earth Sys-
 371 tem Model v2 (CESM2), on a nominal 1° resolution with emissions from two different
 372 SSP scenarios (SSP1-2.6 and SSP5-8.5) and an additional freshwater flux in the North
 373 Atlantic to artificially decrease the AMOC. To our knowledge, this is the first study that
 374 utilizes a model of this high complexity with a horizontal resolution of 1° to study the
 375 effects of an AMOC weakening on the carbon cycle. We find a positive feedback in both
 376 emission scenarios of 0.44 ppm Sv⁻¹ and 1.3 ppm Sv⁻¹ for SSP1-2.6 and SSP5-8.5, re-
 377 spectively. The response in SSP1-2.6 is driven by both the land and ocean carbon reser-
 378 voirs, whereas in SSP5-8.5 it is driven solely by the ocean. The response is small, being
 379 the effect of many compensating effects over both the land and the ocean. Looking at
 380 regional response patterns, both emission scenarios show similar behavior in many cli-
 381 mate and carbon cycle variables. In absolute numbers, the response is stronger in SSP5-
 382 8.5, but when the high CO₂ concentrations are taken into account, the relative response
 383 is actually weaker in SSP5-8.5 compared to SSP1-2.6.

384 Our simulations show the climate response to an AMOC weakening, such as a south-
 385 ward shift of the ITCZ and the bipolar seesaw, similar to many previous studies (Obata,
 386 2007; Zickfeld et al., 2008; Orihuela-Pinto et al., 2022). The AMOC weakening in our
 387 simulations follows a very similar trajectory as in Orihuela-Pinto et al. (2022), which used
 388 an older version of CESM (i.e. v1.2) under pre-industrial boundary conditions. In our
 389 study, the AMOC weakening results in a small increase in atmospheric CO₂ concentra-
 390 tions. This small effect, especially on the multi-decadal to centennial timescales assessed
 391 here, was also found in more idealized models (e.g. Zickfeld et al., 2008; Nielsen et al.,
 392 2019; Gottschalk et al., 2019), but as described in Gottschalk et al. (2019) the relative
 393 response of the ocean and land reservoirs are dependent on climatic boundary conditions

394 and the used model. Here, we have used a member of the newest generation of Earth Sys-
395 tem Models with a relatively high spatial resolution (i.e. nominal $1^\circ \times 1^\circ$ ocean grid).
396 When considering studies with induced AMOC weakening we find, integrated over the
397 entire ocean, a similar response as in Zickfeld et al. (2008), and spatially as in Obata (2007),
398 though local differences remain which can be attributed to the use of a higher resolu-
399 tion, and a more complex model in our study. It is also possible to collapse the AMOC
400 without an additional freshwater forcing. In Nielsen et al. (2019) they used such an al-
401 ternative method under Pleistocene conditions, which resulted in a much slower response
402 in the ocean compared to our simulations. The response of the terrestrial biosphere, es-
403 pecially the changes related to the southward shift of the ITCZ, is also similar to that
404 of previous studies using static vegetation (e.g. Obata, 2007; Nielsen et al., 2019). In Köhler
405 et al. (2005) a dynamic vegetation model is used, and they show that an AMOC collapse
406 affects vegetation type. This leads to reduced carbon storage in the high latitudes and
407 increased carbon storage in the Northern Hemisphere midlatitudes. This dynamic be-
408 havior is not captured in our simulations and unfortunately, it is not possible to assess
409 what the effect of dynamic vegetation would be based on Köhler et al. (2005) since they
410 consider Pleistocene conditions.

411 The result that the pattern of the carbon cycle response to an AMOC weakening
412 is independent of the cumulative CO_2 emissions on multi-decadal to centennial timescales
413 has been shown before. In Zickfeld et al. (2008), for example, the marine carbon cycle
414 remains independent on the used emission scenario for the first 200 years of their sim-
415 ulation, and for the terrestrial carbon cycle this is 150 years. After this period the dif-
416 ferent emissions start to diverge, though the qualitative behavior remains similar. In our
417 simulations, globally integrated variables show little change as a response to the AMOC
418 weakening. However, on regional scales the effects of an AMOC weakening can be large,
419 e.g. SATs can decrease or increase by more than 3°C locally (Fig. 2) and some regions
420 become much drier and other see a large increase in precipitation (Fig. S2). These chang-
421 ing climate conditions, on top of already greenhouse gas driven climate change, require
422 climate adaptation which might be difficult to achieve in such a short time frame (i.e.
423 decades). The climate changes associated to an AMOC weakening also cause changes
424 in the carbon cycle. Such changes can increase, for example, desertification and reduce
425 (but also increase) crop yields. This may lead locally to increased food stress, potentially
426 leading to more frequent and more severe famines. The changes in the ocean can lead
427 to more frequent marine heatwaves in the Southern Hemisphere due to the warming, and
428 reduced (global) NPP due to changing nutrient distributions, which might impact food
429 web dynamics and ecosystem function. However, due to the cooling effect of the bipolar
430 seesaw we would also expect a (relative) reduction in marine heatwaves in the
431 Northern Hemisphere. These effects show that an AMOC collapse can have local effects
432 that have a beneficiary impact or a detrimental impact on the terrestrial and marine bio-
433 spheres.

434 Interestingly, the relative effects on multi-decadal timescales are independent to the
435 (cumulative) greenhouse gas emissions. This means that the uncertainty around the ef-
436 fects of a possible AMOC collapse or weakening is not related to past emissions. How-
437 ever, in a future climate without AMOC weakening, emissions do have an influence on
438 when the AMOC might collapse. Furthermore, the small positive feedback found in this
439 study might make the AMOC more likely to tip earlier. Even though on these timescales
440 the relative effects are not dependent on the greenhouse gas emissions, this might be dif-
441 ferent on intermediate (multi-centennial to millennial) timescales. Because the ocean cir-
442 culation is associated with timescales on the intermediate timescales, we can expect the
443 most important effects to occur in this time frame. We find, for example, that the sur-
444 face ocean is becoming more depleted of nutrients (Fig. 7), which might depress NPP
445 for centuries.

446 Other long term effects that might be relevant are tipping cascades (e.g. Dekker
 447 et al., 2018), meaning that a collapse of the AMOC could set off an other tipping ele-
 448 ment in the Earth System. In our simulations, we find decreasing temperatures in the
 449 Northern Hemisphere due to the AMOC weakening, which reduces the probability of tip-
 450 ping for example melting of the Greenland Ice Sheet, Arctic sea ice, and Northern Hemi-
 451 spheric permafrost. However, due to the bipolar seesaw, the Southern Hemisphere be-
 452 comes warmer, which might increase the probability of tipping the Antarctic Ice Sheets.
 453 Another tipping point connected to the AMOC is the die off of the Amazonian rainfor-
 454 est. Because we do not use a dynamic vegetation model in this study, we cannot inves-
 455 tigate whether the AMOC weakening in our simulations would lead to such a die off.

456 By using a low and a high emission scenario we have tried to cover uncertainties
 457 regarding future emissions. However, we have only used one Earth System Model, which
 458 means that the results presented here could be model dependent. Especially ocean pro-
 459 ductivity shows large spread in the CMIP6 ensemble, which can influence the uptake ca-
 460 pacity of the ocean. Another bias in Earth System Models is a too stable AMOC, mean-
 461 ing we need a large freshwater flux in the North Atlantic Ocean to weaken the AMOC.
 462 This flux is generally too high to represent for example Greenland Ice Sheet melt, but
 463 necessary to achieve a weakened AMOC. This large freshwater flux also leads to fresh-
 464 ening of the surface ocean in the subpolar gyre which influences the carbonate chemistry
 465 and carbon uptake capacity unrealistically. We have not taken this effect into account
 466 explicitly, but it could potentially result in reduced uptake capacity of the ocean, and
 467 therefore more CO₂ in the atmosphere, increasing the feedback strength.

468 Finally, we have shown in a relatively high resolution, state-of-the-art Earth Sys-
 469 tem Model, that the spatial pattern of the carbon cycle response to an AMOC weaken-
 470 ing is not dependent on cumulative CO₂ emissions. As a follow up study it would be in-
 471 teresting to see what happens on multi-centennial and longer timescales, and what the
 472 pCO₂ response would be under an AMOC recovery. Though not analyzed thoroughly,
 473 NPP in the ocean shows large decreases due to the AMOC weakening. This could ef-
 474 fect food web dynamics in the ocean with possible (detrimental) changes in fishery yields,
 475 food securities and income. These ecosystem and socio-economic effects are worth in-
 476 vestigating, to see how a change in the climate system cascades through ecosystems to
 477 socio-economic systems.

478 Appendix A Open Science

479 Yearly output for the most important variables, data necessary to replicate the fig-
 480 ures, and the scripts used for creating the figures can be downloaded from [https://doi](https://doi.org/10.5281/zenodo.8376701)
 481 [.org/10.5281/zenodo.8376701](https://doi.org/10.5281/zenodo.8376701) (Boot et al., 2023b).

482 Acknowledgments

483 This study has been supported by the Netherlands Earth System Science Centre (NESSC),
 484 which is financially supported by the Ministry of Education, Culture and Science (OCW;
 485 grant no. 024.002.001). The authors want to thank Michael Kliphuis (IMAU, Utrecht
 486 University) for performing the CESM2 simulations, which were performed at at SURF-
 487 sara in Amsterdam on the Dutch Supercomputer Snellius within NWO-SURF project
 488 17239.

489 References

490 Behrenfeld, M. J., O'Malley, R. T., Siegel, D. A., McClain, C. R., Sarmiento, J. L.,
 491 Feldman, G. C., ... Boss, E. S. (2006). Climate-driven trends in contem-
 492 porary ocean productivity. *Nature*, 444(7120), 752–755. Retrieved from
 493 <https://doi.org/10.1038/nature05317> doi: 10.1038/nature05317

- 494 Boot, A., von der Heydt, A. S., & Dijkstra, H. A. (2022). Effect of the At-
 495 lantic Meridional Overturning Circulation on atmospheric CO_2
 496 variations. *Earth System Dynamics*, *13*(3), 1041–1058. Retrieved from
 497 <https://esd.copernicus.org/articles/13/1041/2022/> doi: 10.5194/
 498 esd-13-1041-2022
- 499 Boot, A., von der Heydt, A. S., & Dijkstra, H. A. (2023a). Effect of Plankton
 500 Composition Shifts in the North Atlantic on Atmospheric pCO_2 . *Geophys-*
 501 *ical Research Letters*, *50*(2), e2022GL100230. Retrieved from [https://](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2022GL100230)
 502 agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2022GL100230 doi:
 503 <https://doi.org/10.1029/2022GL100230>
- 504 Boot, A., von der Heydt, A. S., & Dijkstra, H. A. (2023b, September). *GBC_2023:*
 505 *v1.1_25_09_23 (v1.1)*. [dataset]. Zenodo. doi: 10.5281/zenodo.8376701
- 506 Buckley, M. W., & Marshall, J. (2016, mar). Observations, inferences, and mecha-
 507 nisms of the Atlantic Meridional Overturning Circulation: A review. *Reviews*
 508 *of Geophysics*, *54*(1), 5–63. Retrieved from [https://doi.org/10.1002/](https://doi.org/10.1002/2015RG000493)
 509 [2015RG000493](https://doi.org/10.1002/2015RG000493) doi: <https://doi.org/10.1002/2015RG000493>
- 510 Burke, E. J., Zhang, Y., & Krinner, G. (2020). Evaluating permafrost physics
 511 in the Coupled Model Intercomparison Project 6 (CMIP6) models and their
 512 sensitivity to climate change. *The Cryosphere*, *14*(9), 3155–3174. Re-
 513 trieved from <https://tc.copernicus.org/articles/14/3155/2020/> doi:
 514 [10.5194/tc-14-3155-2020](https://doi.org/10.5194/tc-14-3155-2020)
- 515 Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G., & Saba, V. (2018). Ob-
 516 served fingerprint of a weakening Atlantic Ocean overturning circulation.
 517 *Nature*, *556*(7700), 191–196. Retrieved from [https://doi.org/10.1038/](https://doi.org/10.1038/s41586-018-0006-5)
 518 [s41586-018-0006-5](https://doi.org/10.1038/s41586-018-0006-5) doi: [10.1038/s41586-018-0006-5](https://doi.org/10.1038/s41586-018-0006-5)
- 519 Cheng, W., Weijer, W., Kim, W. M., Danabasoglu, G., Yeager, S. G., Gent,
 520 P. R., ... Zhang, J. (2018). Can the Salt-Advection Feedback Be De-
 521 tected in Internal Variability of the Atlantic Meridional Overturning Circu-
 522 lation? *Journal of Climate*, *31*(16), 6649–6667. Retrieved from [https://](https://journals.ametsoc.org/view/journals/clim/31/16/jcli-d-17-0825.1.xml)
 523 journals.ametsoc.org/view/journals/clim/31/16/jcli-d-17-0825.1.xml
 524 doi: <https://doi.org/10.1175/JCLI-D-17-0825.1>
- 525 Cuffey, K. M., & Clow, G. D. (1997, nov). Temperature, accumulation, and ice
 526 sheet elevation in central Greenland through the last deglacial transition.
 527 *Journal of Geophysical Research: Oceans*, *102*(C12), 26383–26396. Retrieved
 528 from <https://doi.org/10.1029/96JC03981> doi: [https://doi.org/10.1029/](https://doi.org/10.1029/96JC03981)
 529 [96JC03981](https://doi.org/10.1029/96JC03981)
- 530 Danabasoglu, G. (2019). *NCAR CESM2 model output prepared for CMIP6 CMIP*
 531 *esm-hist*. Earth System Grid Federation. Retrieved from [https://doi.org/10](https://doi.org/10.22033/ESGF/CMIP6.7575)
 532 [.22033/ESGF/CMIP6.7575](https://doi.org/10.22033/ESGF/CMIP6.7575) doi: [10.22033/ESGF/CMIP6.7575](https://doi.org/10.22033/ESGF/CMIP6.7575)
- 533 Danabasoglu, G., Lamarque, J.-F., Bacmeister, J., Bailey, D. A., DuVivier, A. K.,
 534 Edwards, J., ... Strand, W. G. (2020). The Community Earth System Model
 535 Version 2 (CESM2). *Journal of Advances in Modeling Earth Systems*, *12*(2),
 536 e2019MS001916. Retrieved from <https://doi.org/10.1029/2019MS001916>
 537 doi: <https://doi.org/10.1029/2019MS001916>
- 538 Dekker, M. M., von der Heydt, A. S., & Dijkstra, H. A. (2018). Cascading transi-
 539 tions in the climate system. *Earth System Dynamics*, *9*(4), 1243–1260. Re-
 540 trieved from <https://esd.copernicus.org/articles/9/1243/2018/> doi: [10](https://doi.org/10.5194/esd-9-1243-2018)
 541 [.5194/esd-9-1243-2018](https://doi.org/10.5194/esd-9-1243-2018)
- 542 Ditlevsen, P., & Ditlevsen, S. (2023). Warning of a forthcoming collapse of the
 543 Atlantic meridional overturning circulation. *Nature Communications*, *14*(1),
 544 4254.
- 545 Eyring, V., Bony, S., Meehl, G. A., Senior, C. A., Stevens, B., Stouffer, R. J., &
 546 Taylor, K. E. (2016, may). Overview of the Coupled Model Intercompari-
 547 son Project Phase 6 (CMIP6) experimental design and organization. *Geosci.*
 548 *Model Dev.*, *9*(5), 1937–1958. Retrieved from <https://gmd.copernicus.org/>

- 549 [articles/9/1937/2016/https://gmd.copernicus.org/articles/9/1937/](https://gmd.copernicus.org/articles/9/1937/2016/gmd-9-1937-2016.pdf)
 550 [2016/gmd-9-1937-2016.pdf](https://gmd-9-1937-2016.pdf) doi: 10.5194/gmd-9-1937-2016
- 551 Gottschalk, J., Battaglia, G., Fischer, H., Frölicher, T. L., Jaccard, S. L., Jeltsch-
 552 Thömmes, A., ... Stocker, T. F. (2019). Mechanisms of millennial-scale
 553 atmospheric CO₂ change in numerical model simulations. *Quaternary Sci-*
 554 *ence Reviews*, 220, 30–74. Retrieved from [http://www.sciencedirect.com/](http://www.sciencedirect.com/science/article/pii/S0277379118310473)
 555 [science/article/pii/S0277379118310473](http://www.sciencedirect.com/science/article/pii/S0277379118310473) doi: [https://doi.org/10.1016/](https://doi.org/10.1016/j.quascirev.2019.05.013)
 556 [j.quascirev.2019.05.013](https://doi.org/10.1016/j.quascirev.2019.05.013)
- 557 Henson, S. A., Laufkötter, C., Leung, S., Giering, S. L. C., Palevsky, H. I., & Ca-
 558 van, E. L. (2022). Uncertain response of ocean biological carbon export in a
 559 changing world. *Nature Geoscience*, 15(4), 248–254. Retrieved from [https://](https://doi.org/10.1038/s41561-022-00927-0)
 560 doi.org/10.1038/s41561-022-00927-0 doi: 10.1038/s41561-022-00927-0
- 561 Jackson, L. C., Kahana, R., Graham, T., Ringer, M. A., Woollings, T., Mecking,
 562 J. V., & Wood, R. A. (2015). Global and European climate impacts of a slow-
 563 down of the AMOC in a high resolution GCM. *Climate Dynamics*, 45(11),
 564 3299–3316. Retrieved from <https://doi.org/10.1007/s00382-015-2540-2>
 565 doi: 10.1007/s00382-015-2540-2
- 566 Katavouta, A., & Williams, R. G. (2021). Ocean carbon cycle feedbacks in CMIP6
 567 models: contributions from different basins. *Biogeosciences*, 18(10), 3189–3218.
 568 Retrieved from <https://bg.copernicus.org/articles/18/3189/2021/> doi:
 569 10.5194/bg-18-3189-2021
- 570 Köhler, P., Joos, F., Gerber, S., & Knutti, R. (2005). Simulated changes in vege-
 571 tation distribution, land carbon storage, and atmospheric CO₂ in response to
 572 a collapse of the North Atlantic thermohaline circulation. *Climate Dynamics*,
 573 25(7), 689. Retrieved from <https://doi.org/10.1007/s00382-005-0058-8>
 574 doi: 10.1007/s00382-005-0058-8
- 575 Kwiatkowski, L., Torres, O., Bopp, L., Aumont, O., Chamberlain, M., Christian,
 576 J. R., ... Ziehn, T. (2020). Twenty-first century ocean warming, acidifica-
 577 tion, deoxygenation, and upper-ocean nutrient and primary production decline
 578 from CMIP6 model projections. *Biogeosciences*, 17(13), 3439–3470. Re-
 579 trieved from <https://bg.copernicus.org/articles/17/3439/2020/> doi:
 580 10.5194/bg-17-3439-2020
- 581 Lawrence, D. M., Fisher, R. A., Koven, C. D., Oleson, K. W., Swenson, S. C., Bo-
 582 nan, G., ... Zeng, X. (2019, dec). The Community Land Model Version
 583 5: Description of New Features, Benchmarking, and Impact of Forcing Un-
 584 certainty. *Journal of Advances in Modeling Earth Systems*, 11(12), 4245–
 585 4287. Retrieved from <https://doi.org/10.1029/2018MS001583> doi:
 586 <https://doi.org/10.1029/2018MS001583>
- 587 Lenton, T. M., Held, H., Kriegler, E., Hall, J. W., Lucht, W., Rahmstorf, S., &
 588 Schellnhuber, H. J. (2008). Tipping elements in the Earth’s climate system.
 589 *Proceedings of the National Academy of Sciences*, 105(6), 1786–1793. Re-
 590 trieved from <https://www.pnas.org/doi/abs/10.1073/pnas.0705414105>
 591 doi: 10.1073/pnas.0705414105
- 592 Liu, Y., Moore, J. K., Primeau, F., & Wang, W. L. (2023). Reduced CO₂ up-
 593 take and growing nutrient sequestration from slowing overturning circulation.
 594 *Nature Climate Change*, 13(1), 83–90. Retrieved from [https://doi.org/](https://doi.org/10.1038/s41558-022-01555-7)
 595 [10.1038/s41558-022-01555-7](https://doi.org/10.1038/s41558-022-01555-7) doi: 10.1038/s41558-022-01555-7
- 596 Long, M. C., Moore, J. K., Lindsay, K., Levy, M., Doney, S. C., Luo, J. Y., ...
 597 Sylvester, Z. T. (2021). Simulations With the Marine Biogeochemistry Li-
 598 brary (MARBL). *Journal of Advances in Modeling Earth Systems*, 13(12),
 599 e2021MS002647. Retrieved from <https://doi.org/10.1029/2021MS002647>
 600 doi: <https://doi.org/10.1029/2021MS002647>
- 601 Lynch-Stieglitz, J. (2017). The Atlantic Meridional Overturning Circulation and
 602 Abrupt Climate Change. *Annual Review of Marine Science*, 9(1), 83–104.
 603 Retrieved from <https://doi.org/10.1146/annurev-marine-010816-060415>

- 604 doi: 10.1146/annurev-marine-010816-060415
- 605 Marchal, O., Stocker, T. F., & Joos, F. (1998, jun). Impact of oceanic reorganiza-
606 tions on the ocean carbon cycle and atmospheric carbon dioxide content. *Pa-*
607 *leoceanography*, 13(3), 225–244. Retrieved from [https://doi.org/10.1029/](https://doi.org/10.1029/98PA00726)
608 [98PA00726](https://doi.org/10.1029/98PA00726) doi: <https://doi.org/10.1029/98PA00726>
- 609 McKay, D. I. A., Staal, A., Abrams, J. F., Winkelmann, R., Sakschewski, B., Lo-
610 riani, S., . . . Lenton, T. M. (2022). Exceeding 1.5°C global warm-
611 ing could trigger multiple climate tipping points. *Science*, 377(6611),
612 eabn7950. Retrieved from [https://www.science.org/doi/abs/10.1126/](https://www.science.org/doi/abs/10.1126/science.abn7950)
613 [science.abn7950](https://www.science.org/doi/abs/10.1126/science.abn7950) doi: 10.1126/science.abn7950
- 614 Nielsen, S. B., Jochum, M., Pedro, J. B., Eden, C., & Nuterman, R. (2019, apr).
615 Two-Timescale Carbon Cycle Response to an AMOC Collapse. *Paleoceanogra-*
616 *phy and Paleoclimatology*, 34(4), 511–523. Retrieved from [https://doi.org/](https://doi.org/10.1029/2018PA003481)
617 [10.1029/2018PA003481](https://doi.org/10.1029/2018PA003481) doi: <https://doi.org/10.1029/2018PA003481>
- 618 Nohara, D., Yoshida, Y., Misumi, K., & Ohba, M. (2013, mar). Dependency of cli-
619 mate change and carbon cycle on CO₂ emission pathways. *Environmental Re-*
620 *search Letters*, 8(1), 14047. Retrieved from [https://dx.doi.org/10.1088/](https://dx.doi.org/10.1088/1748-9326/8/1/014047)
621 [1748-9326/8/1/014047](https://dx.doi.org/10.1088/1748-9326/8/1/014047) doi: 10.1088/1748-9326/8/1/014047
- 622 Obata, A. (2007, dec). Climate Carbon Cycle Model Response to Freshwater Dis-
623 charge into the North Atlantic. *Journal of Climate - J CLIMATE*, 20, 5962–
624 5976. doi: 10.1175/2007JCLI1808.1
- 625 O’Neill, B. C., Carter, T. R., Ebi, K., Harrison, P. A., Kemp-Benedict, E., Kok,
626 K., . . . Pichs-Madruga, R. (2020). Achievements and needs for the climate
627 change scenario framework. *Nature Climate Change*, 10(12), 1074–1084.
628 Retrieved from <https://doi.org/10.1038/s41558-020-00952-0> doi:
629 [10.1038/s41558-020-00952-0](https://doi.org/10.1038/s41558-020-00952-0)
- 630 Orihuela-Pinto, B., England, M. H., & Taschetto, A. S. (2022). Interbasin and in-
631 terhemispheric impacts of a collapsed Atlantic Overturning Circulation. *Nature*
632 *Climate Change*, 12(6), 558–565. Retrieved from [https://doi.org/10.1038/](https://doi.org/10.1038/s41558-022-01380-y)
633 [s41558-022-01380-y](https://doi.org/10.1038/s41558-022-01380-y) doi: 10.1038/s41558-022-01380-y
- 634 Rahmstorf, S. (2002). Ocean circulation and climate during the past 120,000 years.
635 *Nature*, 419(6903), 207–214. Retrieved from [https://doi.org/10.1038/](https://doi.org/10.1038/nature01090)
636 [nature01090](https://doi.org/10.1038/nature01090) doi: 10.1038/nature01090
- 637 Sarmiento, J. L., Hughes, T. M. C., Stouffer, R. J., & Manabe, S. (1998). Simulated
638 response of the ocean carbon cycle to anthropogenic climate warming. *Nature*,
639 393(6682), 245–249. Retrieved from <https://doi.org/10.1038/30455> doi:
640 [10.1038/30455](https://doi.org/10.1038/30455)
- 641 Schmittner, A. (2005). Decline of the marine ecosystem caused by a reduction in the
642 Atlantic overturning circulation. *Nature*, 434(7033), 628–633. Retrieved from
643 <https://doi.org/10.1038/nature03476> doi: 10.1038/nature03476
- 644 Schmittner, A., & Galbraith, E. D. (2008). Glacial greenhouse-gas fluctuations con-
645 trolled by ocean circulation changes. *Nature*, 456(7220), 373–376. Retrieved
646 from <https://doi.org/10.1038/nature07531> doi: 10.1038/nature07531
- 647 Smith, R., Jones, P. W., Briegleb, P. A., Bryan, O., Danabasoglu, G., Dennis,
648 M. L., . . . Yeager, S. G. (2010). The Parallel Ocean Program (POP) refer-
649 ence manual: Ocean component of the Community Climate System Model
650 (CCSM)..
- 651 Swingedouw, D., Bopp, L., Matras, A., & Braconnot, P. (2007, dec). Effect
652 of land-ice melting and associated changes in the AMOC result in little
653 overall impact on oceanic CO₂ uptake. *Geophysical Research Letters*,
654 34(23). Retrieved from <https://doi.org/10.1029/2007GL031990> doi:
655 <https://doi.org/10.1029/2007GL031990>
- 656 Tagliabue, A., Kwiatkowski, L., Bopp, L., Butenschön, M., Cheung, W., Lengaigne,
657 M., & Vialard, J. (2021). Persistent Uncertainties in Ocean Net Primary
658 Production Climate Change Projections at Regional Scales Raise Chal-

- 659 lenges for Assessing Impacts on Ecosystem Services. *Frontiers in Climate*,
 660 3. Retrieved from <https://www.frontiersin.org/articles/10.3389/fclim.2021.738224> doi: 10.3389/fclim.2021.738224
- 661
 662 Vellinga, M., & Wood, R. A. (2002). Global Climatic Impacts of a Collapse of the
 663 Atlantic Thermohaline Circulation. *Climatic Change*, 54(3), 251–267. Re-
 664 trieved from [https://doi.org/10.1023/A:](https://doi.org/10.1023/A:1016168827653)
 665 1016168827653 doi: 10.1023/A:1016168827653
- 666 Weijer, W., Cheng, W., Drijfhout, S. S., Fedorov, A. V., Hu, A., Jackson, L. C., ...
 667 Zhang, J. (2019, aug). Stability of the Atlantic Meridional Overturning Cir-
 668 culation: A Review and Synthesis. *Journal of Geophysical Research: Oceans*,
 669 124(8), 5336–5375. Retrieved from <https://doi.org/10.1029/2019JC015083>
 670 doi: <https://doi.org/10.1029/2019JC015083>
- 671 Weijer, W., Cheng, W., Garuba, O. A., Hu, A., & Nadiga, B. T. (2020, jun).
 672 CMIP6 Models Predict Significant 21st Century Decline of the Atlantic
 673 Meridional Overturning Circulation. *Geophysical Research Letters*, 47(12),
 674 e2019GL086075. Retrieved from <https://doi.org/10.1029/2019GL086075>
 675 doi: <https://doi.org/10.1029/2019GL086075>
- 676 Whitt, D. B., & Jansen, M. F. (2020, jul). Slower nutrient stream suppresses
 677 Subarctic Atlantic Ocean biological productivity in global warming. *Pro-*
 678 *ceedings of the National Academy of Sciences*, 117(27), 15504–15510. Re-
 679 trieved from <https://doi.org/10.1073/pnas.2000851117> doi: 10.1073/
 680 pnas.2000851117
- 681 Wilson, J. D., Andrews, O., Katavouta, A., de Melo Viríssimo, F., Death,
 682 R. M., Adloff, M., ... Ying, R. (2022). The biological carbon pump
 683 in CMIP6 models: 21st century trends and uncertainties. *Proceedings*
 684 *of the National Academy of Sciences*, 119(29), e2204369119. Retrieved
 685 from <https://www.pnas.org/doi/abs/10.1073/pnas.2204369119> doi:
 686 10.1073/pnas.2204369119
- 687 Yamamoto, A., Abe-Ouchi, A., & Yamanaka, Y. (2018). Long-term response of
 688 oceanic carbon uptake to global warming via physical and biological pumps.
 689 *Biogeosciences*, 15(13), 4163–4180. Retrieved from [https://bg.copernicus](https://bg.copernicus.org/articles/15/4163/2018/)
 690 [.org/articles/15/4163/2018/](https://bg.copernicus.org/articles/15/4163/2018/) doi: 10.5194/bg-15-4163-2018
- 691 Zhang, R., & Delworth, T. L. (2005). Simulated Tropical Response to a Substan-
 692 tial Weakening of the Atlantic Thermohaline Circulation. *Journal of Cli-*
 693 *mate*, 18(12), 1853–1860. Retrieved from [https://journals.ametsoc.org/](https://journals.ametsoc.org/view/journals/clim/18/12/jcli3460.1.xml)
 694 [view/journals/clim/18/12/jcli3460.1.xml](https://journals.ametsoc.org/view/journals/clim/18/12/jcli3460.1.xml) doi: [https://doi.org/10.1175/](https://doi.org/10.1175/JCLI3460.1)
 695 JCLI3460.1
- 696 Zhu, J., Zeng, X., Gao, X., & Zhang, H. (2022). Response of Terrestrial Net
 697 Primary Production to Quadrupled CO₂ Forcing: A Comparison between
 698 the CAS-ESM2 and CMIP6 Models. *Biology*, 11(12). Retrieved from
 699 <https://doi.org/10.3390/biology11121693> doi: [https://doi.org/10.3390/](https://doi.org/10.3390/biology11121693)
 700 biology11121693
- 701 Zickfeld, K., Eby, M., & Weaver, A. J. (2008, sep). Carbon-cycle feedbacks of
 702 changes in the Atlantic meridional overturning circulation under future at-
 703 mospheric CO₂. *Global Biogeochemical Cycles*, 22(3). Retrieved from
 704 <https://doi.org/10.1029/2007GB003118> doi: [https://doi.org/10.1029/](https://doi.org/10.1029/2007GB003118)
 705 2007GB003118

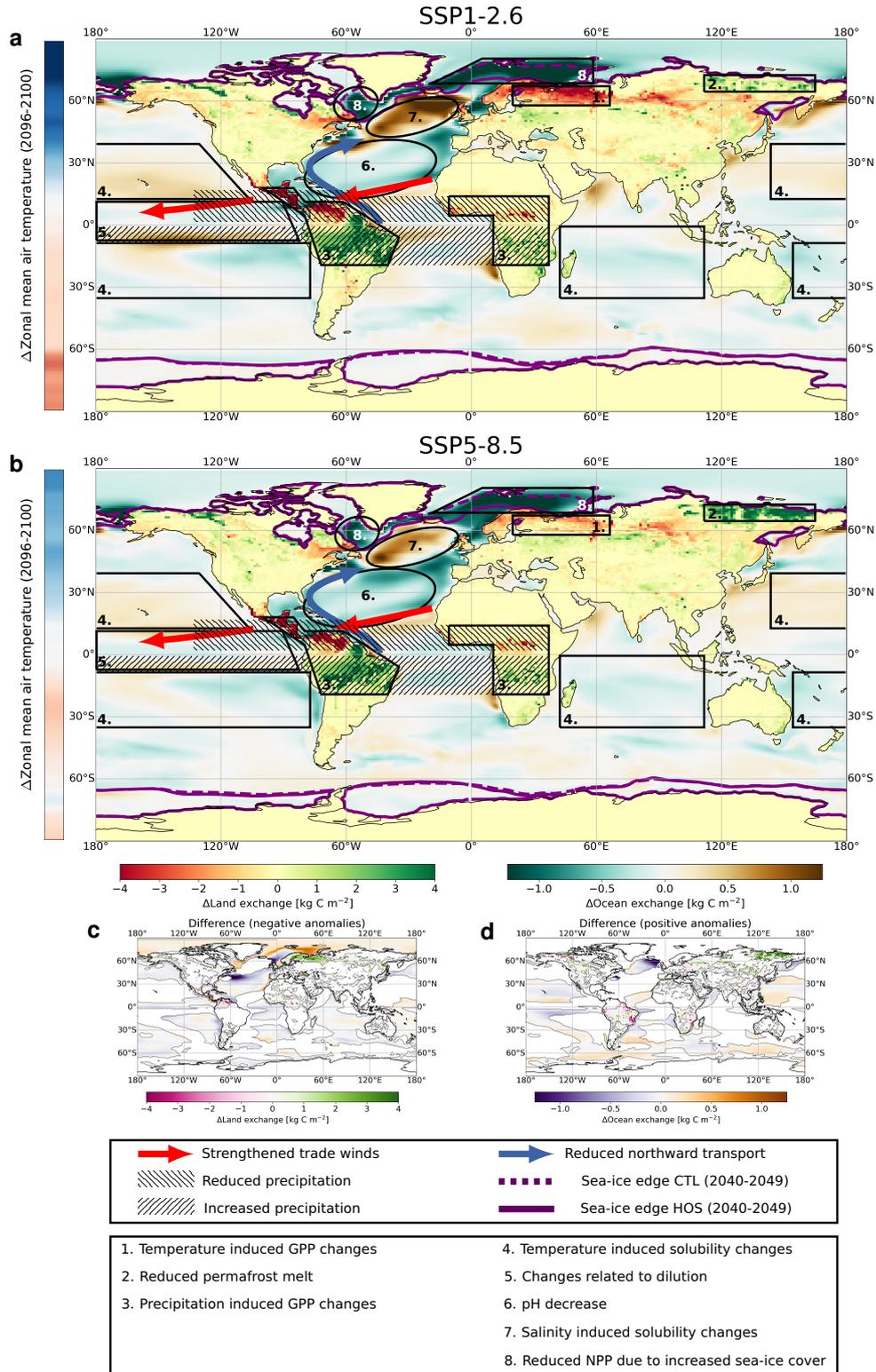


Figure 10. Summarizing figure with dominant mechanisms included for SSP1-2.6 (a) and SSP5-8.5 (b). (a) and (b) represent results from HOS minus the CTL simulations. The sea-ice edge is taken as where the ice fraction is 0.25 and denoted by the purple lines, where dashed lines represent the CTL simulations and solid lines the HOS simulations. The bar at the left shows the difference in zonal mean surface air temperature averaged over 2096-2100 between HOS and CTL. The scaling of this bar is between -2.5°C (dark blue) and 2.5°C (dark red). (c) The difference between SSP5-8.5 (b) and SSP1-2.6 (a) in the regions where (b) is negative. Negative values represent a higher negative anomaly in SSP5-8.5 compared to SSP1-2.6. (d) as in (c) but for positive anomalies. Positive values represent a higher positive anomaly in SSP5-8.5 compared to SSP1-2.6. The color bars in (c) and (d) apply to both subfigures.