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## Emergence of deep convection in the Arctic Ocean under a warming climate

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### Abstract :

The appearance of winter deep mixed layers in the Arctic Ocean under a warming climate is investigated with the HiGEM coupled global climate model. In response to a four times increase of atmospheric CO<sub>2</sub> levels with respect to present day conditions, the Arctic Basin becomes seasonally ice-free. Its surface becomes consequently warmer and, on average, slightly fresher. Locally, changes in surface salinity can be far larger (up to 4 psu) than the basin-scale average, and of a different sign. The Canadian Basin undergoes a strong freshening, while the Eurasian Basin undergoes strong salinification. These changes are driven by the spin up of the surface circulation, likely resulting from the increased transfer of momentum to the ocean as sea ice cover is reduced. Changes in the surface salinity field also result in a change in stratification, which is strongly enhanced in the Canadian Basin and reduced in the Eurasian Basin. Reduction, or even suppression, of the stratification in the Eurasian Basin produces an environment that is favourable for, and promotes the appearance of, deep convection near the sea ice edge, leading to a significant deepening of winter mixed layers in this region (down to 1000 m). As the Arctic Ocean is transitioning toward a summer ice-free regime, new dynamical ocean processes will appear in the region, with potentially important consequences for the Arctic Ocean itself and for climate, both locally and on larger scales.

## 23 1 Introduction

24 The mixed layer at the surface of the ocean connects the atmosphere with the deep ocean. It allows  
25 for the exchange of buoyancy and momentum, as well as gases including oxygen and carbon diox-  
26 ide, between the atmosphere and the ocean interior. In a few locations, the mixed layer episodically  
27 reaches depths of hundreds of metres, symptomatic of the formation of dense water. In the North  
28 Atlantic, deep convection events have been observed in the Labrador, Irminger and Greenland Seas  
29 (Marshall and Schott, 1999; Pickart et al., 2003). These regions are all characterized by the sea-  
30 sonally intermittent presence of sea ice, and the deepest mixed layers are often observed near the  
31 sea ice edge, where large horizontal temperature and salinity gradients can be found at the ocean  
32 surface. Juxtaposition of ice-free and ice-covered areas accentuates the intensity of air–sea exchange  
33 over the ocean, resulting in buoyancy loss and water mass transformation (Griffies et al., 2009;  
34 Våge et al., 2009; Germe et al., 2011). In contrast, the mixed layer in the Arctic Ocean is much  
35 shallower (Peralta-Ferriz and Woodgate, 2015); the near-perennial presence of sea ice and the strong  
36 surface stratification insulate the ocean from intense air-sea exchange, limiting buoyancy loss and  
37 momentum input into the upper ocean.

38

39 As part of the fifth phase of the Climate Model Intercomparison Project (CMIP5; Taylor et al.,  
40 2012), the projected climate conditions in the North Atlantic and the Arctic have been examined  
41 in a series of coupled climate models forced with a range of increasing greenhouse gas emission  
42 scenarios. Although the CMIP5 models exhibit a large spread in their results, they largely agree  
43 on the direction of change for several key climate features. These include a decline in the Atlantic  
44 Meridional Overturning Circulation (AMOC) (Cheng et al., 2013), a large shoaling of the Mixed  
45 Layer Depths (MLDs) in the North Atlantic and in particular in the Labrador Sea (Heuzé et al.,

2015), indicating a decrease in the intensity of deep convection, and a decline in Arctic sea ice cover (Stroeve et al., 2012), along with a northward migration of the sea ice edge into the Arctic Ocean and the Barents Sea. Given that the location of deep convection sites is inherently tied to the location of the sea ice edge, one might expect that new regions of deep convection could appear at higher latitudes (Rainville et al., 2011). We expect that changes in deep convection location and MLD will alter water mass properties and ultimately the deep branch of the AMOC (Rahmstorf, 2002; Heuzé, 2017) and ocean heat transport (Exarchou et al., 2015).

Today, the deepest layers of the Canadian and Eurasian Basins are mostly filled with dense waters formed by brine rejection during sea ice formation on the continental shelves, or transformed in the Barents Sea, or advected from the Greenland Sea through Fram Strait. These dense water masses are eventually advected to the Nordic Seas through Fram Strait (Jones et al., 1995; Lique et al., 2010). As the sea ice edge migrates northward, the expectation is that the relative contribution of deep waters formed locally in the Arctic Basin by open ocean convection will increase relative to those waters formed on continental shelves or in the Barents Sea. In the deep Greenland Sea, observations suggest that the amount of dense water originating from the deep Arctic has recently increased compared to the dense water formed locally in the Greenland Sea during deep convection events (Langehaug and Falck, 2012; Somavilla et al., 2013). Moreover, Langehaug and Falck (2012) show that the export of dense water from the Arctic through Fram Strait has switched from being an intermittent to a permanent feature over the past decade, which indicates a change in the deep pressure gradient across Fram Strait.

The aim of the present study is to seek evidence for the emergence of deep convection areas in the high Arctic, as the sea ice edge moves northward. Our analysis is based on simulations from the

70 global coupled climate model HiGEM (Shaffrey et al., 2009) in which  $CO_2$  levels in the atmosphere  
71 drastically alter the radiative balance. The model and simulations used are briefly described in  
72 Section 2. In Section 3, we examine the present-day and future mixed layers as simulated in the  
73 model runs. Changes in MLD are then related to changes in sea ice conditions, atmospheric forcing  
74 and ocean dynamics in Section 4. Climatic consequences associated with changes in MLD are  
75 discussed in Section 5. A summary and conclusions are given in Section 6.

## 76 **2 The numerical experiment**

77 The simulations used in this study were performed with the High-Resolution Global Environmental  
78 Model (HiGEM), which is an ocean–sea ice–atmosphere coupled model based on the Hadley Cen-  
79 tre Global Environmental Model version 1 (HadGEM1; Johns et al., 2006). A full description and  
80 basic evaluation of the HiGEM model can be found in Shaffrey et al. (2009). It uses a spherical  
81 latitude-longitude grid with an atmospheric horizontal resolution of  $0.83^\circ$  latitude  $\times$   $1.25^\circ$  longitude  
82 and 38 vertical levels, and a  $1/3^\circ \times 1/3^\circ$  resolution ocean with 40 unevenly-spaced levels in the  
83 vertical. For the ocean component, parameterizations include a scale-selective biharmonic scheme  
84 for the momentum dissipation, the isopycnal formulation of Griffies et al. (1998) with constant  
85 isopycnal diffusivity for the lateral mixing of tracers, and a biharmonic scheme to represent en-  
86 hanced horizontal mixing of temperature and salinity in the upper 20 m. Eddies are permitted at  
87 mid and low latitudes and are parameterized at high latitudes by the Gent and McWilliams (1990)  
88 adiabatic mixing scheme with a latitudinally varying thickness diffusion coefficient, and the adia-  
89 batic biharmonic scheme of Roberts and Marshall (1998). The mixed layer scheme is based on the  
90 Richardson number parametrization, subject to minimum depth-dependent background diffusivity  
91 and viscosity, following Kraus and Turner (1967). The sea-ice model is an adaptation of the Commu-  
92 nity Ice Code (CICE; Hunke and Dukowicz, 1997), that uses an elastic-viscous-plastic rheology and

93 a five-category ice thickness distribution, but an independent zero-layer thermodynamics scheme  
94 (McLaren et al., 2006). The ocean and the atmosphere are initialized from rest using data from the  
95 World Ocean Atlas 2001 (Boyer et al., 2005) and the ECMWF analysis, respectively.

96  
97 Two simulations are analyzed. The first is a 130 year control integration (labelled CTRL) in  
98 which greenhouse gases are kept constant at close to present-day concentrations (the concentrations  
99 of  $CO_2$ ,  $CH_4$ , and  $N_2O$  are 345 ppm, 1656 ppb, and 307 ppb, respectively). The second simulation  
100 (labelled  $4\times CO_2$ ) is initialized from the CTRL simulation at year 30 and then integrated for 100  
101 years. For this integration, the atmospheric concentration of  $CO_2$  is increased by 2% per year for  
102 70 years until levels reach 4 times that of the control run, and are then kept constant for a further  
103 30 years. The rate of  $CO_2$  increase in this simulation is roughly twice as large as that in the least  
104 conservative scenario used for the IPCC AR5 (RCP 8.5), but the increase is not as sustained and  
105 plateaus at a value that is 30% lower. The idealized forcing allows for a clearer assessment of the  
106 response to radiative forcing than a simulation with more complicated scenarios of future radiative  
107 forcing. It is clear that the two simulations used in this study are much shorter than the time  
108 required to bring the ocean–atmosphere system into an equilibrated state with no memory of the  
109 somewhat arbitrary initial conditions (Covey et al., 2006). With this potential shortcoming in mind,  
110 we note that a strong adjustment takes place over the first 20–30 years of the CTRL simulation,  
111 and we thus choose to discard the first 30 years of the simulation from our analyses.

112  
113 For the present study, we use 10 years of each simulation (years 120 to 129), in order to focus  
114 on the signal of interest here (i.e. the emergence of deep convection in the High Arctic). Monthly  
115 outputs over the 10 years are averaged month by month to create a climatological year for each  
116 simulation. In the following, we discuss particularly the averages for the months of March and

117 September, as these are representative of the periods when the two extrema of the seasonal cycles  
118 in MLD and sea ice extent occur. For all the variables presented in this paper, the significance of  
119 the difference between the two simulations is assessed by comparing the difference field with two  
120 standard deviations of the same variable, computed from 100 years of monthly or annual means in  
121 the *CTRL* run. To first order, this diagnostic allows us to distinguish between change driven by the  
122 radiative forcing and change that could potentially come from internal decadal variability of the  
123 ocean–atmosphere system.

### 124 **3 Change in mixed layer depth**

125 In this section, we examine the properties of the ocean surface and the mixed layer as simulated by  
126 the HiGEM model. The investigation of differences between the  $4\times CO_2$  and the *CTRL* simulations  
127 allows us to determine the ocean and sea ice responses to increasing levels of  $CO_2$  in the atmosphere.

128  
129 Fig. 1 shows the mean MLD in March and September and the mean location of the sea ice  
130 edge over the last 10 years of the *CTRL* and  $4\times CO_2$  simulations, as well as the differences be-  
131 tween the two runs. For each grid point, MLD is computed from monthly mean averages and the  
132 density is evaluated using the EOS80 formulation (Fofonoff, 1985), consistent with the equation  
133 of state used in HiGEM. MLD is then determined based on a density difference from the surface  
134 of  $\Delta\sigma = 0.03kg/m^3$ , following the criteria chosen in previous Arctic-focused studies (Toole et al.,  
135 2010; Jackson et al., 2012). Note that the results show little sensitivity to the choice of threshold,  
136 as long as this is expressed in term of density. Stratification in the Arctic is mostly determined by  
137 salinity, and thus a temperature criterion, as often used in other part of the globe, is not a suitable  
138 definition of MLD there.

140 For the *CTRL* simulation, a classical pattern stands out, with MLD in March reaching greater  
141 than 800m in typical deep convection areas (i.e. the Labrador, Irminger and Nordic Seas and the  
142 Rockall Basin), and MLD in September much shallower ( $\sim 50m$ ). As is often the case with CMIP-  
143 type climate models (Heuzé et al., 2013), HiGEM tends to produce most of its dense water masses  
144 through convection in the open ocean rather than through shelf processes and further cascading of  
145 dense water. The magnitude and general pattern of MLD are in agreement with climatology based  
146 on observations using the same density criterion (de Boyer Montégut et al., 2004) (not shown). The  
147 deepest winter MLDs are located close to the sea ice edge or to steep bathymetry (or both), while  
148 MLDs do not exceed  $\sim 50m$  under sea ice.

149

150 Compared to the *CTRL* run, winter MLD in the  $4\times CO_2$  run is generally much shallower. MLDs  
151 in the areas where deep convection occurs in the *CTRL* run reach 300 – 400m in the  $4\times CO_2$   
152 run, i.e roughly half the MLD found in the *CTRL* run. The strong shoaling of winter MLDs in  
153 the Labrador Sea in the  $4\times CO_2$  HiGEM run has been previously linked with the decrease of the  
154 AMOC (Thomas et al., 2012) and the weaker intensity of the Subpolar Gyre (Lique et al., 2015).  
155 In the Greenland Sea, the convective patch is displaced southward along the sea ice edge. This shift  
156 occurs at the same time as a strong intensification of the gyre in the Greenland Sea in the  $4\times CO_2$   
157 run compared to the *CTRL* run, which has been previously related to a change in wind stress curl  
158 (Lique et al., 2015). Whilst MLD decreases in most existing deep convection locations, the difference  
159 in MLD between the two runs for the representative month of March also reveals that new areas  
160 of convection appear North of Svalbard and in the high Arctic. In these regions, MLD deepens by  
161 up to 400m in the  $4\times CO_2$  run compared to the *CTRL* run on average over the last 10 years of the  
162 runs. In the Arctic, the ‘new’ locations with deep winter MLDs correspond to regions that are ice  
163 free year round in the  $4\times CO_2$  run as both summer and winter sea ice edges move northward. A

164 similar deepening of the winter MLD in the Nansen Basin under the IPCC RCP8.5 scenario was  
165 found in a different coupled model by Brodeau and Koenigk (2016), although these authors did not  
166 investigate the origin of this signal.

167

168 We examine the time series of maximum MLD detected in different parts of the North Atlantic  
169 and the Arctic in the two runs (Fig. 2), to get a sense of the episodic nature of the convective  
170 events that appear in this region in the  $4\times CO_2$  run. One should remember that the  $4\times CO_2$  run is  
171 relatively short, and that Year 100 corresponds to the time when the atmospheric concentration of  
172  $CO_2$  starts to plateau at its maximum value. In the Eurasian Basin, winter MLDs in the  $4\times CO_2$   
173 run first diverge significantly from the *CTRL* run in March of year 67, but the convective events  
174 only become recurrent and intense (with MLD greater than 500m) at the very end of the run, corre-  
175 sponding to the time when the sea ice cover has become seasonal and the winter sea ice extent has  
176 also greatly reduced (Fig. 2(a)). The maximum MLD detected in the Eurasian Basin varies between  
177 360m and 960m during the ten years analyzed in this study (years 120 to 129). It is also possible  
178 that the emergence of deep convection in the Nansen Basin could be a transient signal that would  
179 vanish again in a longer simulation (Brodeau and Koenigk, 2016). The time series of MLD in the  
180 Greenland Sea reveals a sharp transition (around Year 55) toward shallower MLD in the  $4\times CO_2$   
181 run, unlike in the Labrador and Irminger Seas where MLD gradually decreases as the level of  $CO_2$   
182 in the atmosphere increases.

183

184 In summer, MLDs in the  $4\times CO_2$  run generally become shallower where they were deepest in the  
185 *CTRL* run (Fig. 1). However, a deepening of the MLDs by  $\sim 40m$  is visible in the Arctic, principally  
186 North of the Kara and Barents Seas and Bering Strait. By the end of the  $4\times CO_2$  run, the Arctic is  
187 almost completely ice-free in summer (Fig. 2(a)). The ocean surface is thus in direct contact with

188 the atmosphere and the momentum input to the ocean increases, resulting in intensified wind-driven  
189 mixing (e.g. Rainville and Woodgate, 2009). A closer look at the monthly MLD seasonal cycle (using  
190 an EOF decomposition; not shown) reveals that the timing of the seasonal cycle is similar in the  
191 two simulations, with the shallowest MLD in summer (June to September) and the deepest MLD  
192 in winter (February to April).

#### 193 **4 Emergence of favourable conditions for deep convection in the Arctic**

194 We next examine the atmospheric, oceanic and sea ice conditions in the two runs, in order to un-  
195 derstand what circumstances lead to a deepening of Arctic MLD in the  $4\times CO_2$  run. Figs. 3, 4 and  
196 5 show the March and September Sea Surface Temperature (SST), Sea Surface Salinity (SSS) and  
197 surface potential density ( $\sigma$ ), respectively, averaged over the last 10 years of the *CTRL* and  $4\times CO_2$   
198 simulations, as well as the differences between the two runs. Fig. 6 shows the mean seasonal cycle  
199 of the same quantities averaged over the region north of  $65^\circ N$ .

200

201 The changes affecting the SST in the Arctic are closely linked to the presence or absence of sea  
202 ice. By the end of the  $4\times CO_2$  run, the Arctic sea ice cover has become seasonal (Fig. 6(d)), leading  
203 to a strong summer warming of the surface layer throughout the whole basin: SST rises by  $6-8^\circ C$  in  
204 September in the  $4\times CO_2$  run compared to the *CTRL* run (Fig. 3 and 6(a)). In turn, the strong  
205 summer warming results in a strong increase in the amplitude of the SST seasonal cycle (Fig. 6(a)),  
206 given that the SST remains at the freezing point in winter in the  $4\times CO_2$  run over much of the  
207 basin, owing to the presence of sea ice (Fig. 3).

208

209 Such an increase in the amplitude of the Arctic SST seasonal cycle under a warming climate has  
210 also been documented previously in a set of CMIP5 coupled climate models (Carton et al., 2015).

211 Interestingly, as seen in Holland and Bitz (2003), the growth in the amplitude of the SST seasonal  
212 cycle occurs in spite of a decrease in amplitude of the seasonal cycle in air surface temperature,  
213 implying accompanying changes in the seasonality of air-sea fluxes (Fig. 6(f)). Over large parts of  
214 the Arctic, sea-ice is still present in winter in the  $4\times CO_2$  run and so winter SST still approaches the  
215 freezing point throughout the run (Fig. 3). Because more heat has accumulated in summer though,  
216 this means that, locally and overall for the region, heat fluxes out of the ocean have to intensify  
217 to restore winter SST conditions. The intensification in the air-sea fluxes can be seen in Fig. 6(f),  
218 which shows greater heat fluxes into the ocean during the longer sea ice-free period in summer and  
219 greater fluxes out of the ocean in winter in the  $4\times CO_2$  run compared to the CTRL run. Part of  
220 these changes might be due to cloud feedbacks, which may not be properly captured in the model  
221 (Shaffrey et al., 2009). The annual mean heat flux out of the ocean, averaged over the Arctic Basin,  
222 increases by  $\sim 30\%$  in the  $4\times CO_2$  run. Graham and Vellinga (2013) argue that this strong change  
223 in surface heat flux is, however, likely compensated by an increase in the advective ocean heat  
224 transport from the North Atlantic, resulting in only moderate changes to the total Arctic Ocean  
225 heat storage.

226

227 Carton et al. (2015) also examine the change in the SSS seasonal cycle in CMIP5 models under  
228 different greenhouse gas emission scenarios. They find that, across the models examined, the ampli-  
229 tude of the Arctic basin-averaged SSS seasonal cycle tends to decrease, along with the annual-mean  
230 basin-averaged SSS. The decrease in annual-mean basin-averaged SSS is found to be larger for the  
231 model runs under IPCC scenarios with higher greenhouse gas emissions. These findings also hold for  
232 HiGEM. The basin-averaged annual mean SSS decreases by  $\sim 0.5$  psu in the  $4\times CO_2$  HiGEM run,  
233 and the amplitude of the basin-scale average SSS seasonal cycle also decreases under a warming  
234 climate, with a stronger decrease in winter and roughly no change in summer (Fig. 6(b)). Carton

235 et al. (2015) attribute the decrease of the amplitude of the basin-averaged SSS seasonal cycle to the  
236 decline of the sea ice cover. Indeed, the primary driver of the SSS seasonal cycle in the Arctic is the  
237 intensity of sea ice melting and formation processes (Ding et al., 2016), which is strongly reduced  
238 when the sea ice volume decreases, as a smaller volume of sea ice is formed and melted every year.

239

240 Importantly, the spatial pattern of the change in SSS (Fig. 4) reveals strong differences between  
241 the Canadian Basin (where a strong freshening is visible) and the Eurasian Basin (where the ocean  
242 surface is becoming saltier). This pattern of change in SSS also exhibits seasonality: the changes on  
243 the Eurasian side of the Arctic Basin are intensified in summer, while the freshening in the Cana-  
244 dian Basin is strongest in winter. This contrast between the two basins leads to compensation at  
245 the basin scale, which is why the basin-averaged and annual mean changes appear to be relatively  
246 small. On the Arctic shelves, the SSS changes between the *CTRL* and  $4\times CO_2$  runs are overall less  
247 intense, even reversing sign between winter and summer in some parts of the Chukchi Sea and the  
248 Canadian shelves.

249

250 The large scale pattern of SSS change seen in the interior of the basin is linked with a change  
251 of the ocean circulation (Fig. 7). The strong positive SSH anomaly in the Canadian Basin reveals  
252 that the anticyclonic Beaufort Gyre experiences a strong spin up. In the Beaufort Gyre, the SSH  
253 variations are a good proxy for the variations of freshwater content (Proshutinsky and Johnson,  
254 1997), and thus the SSH increase indicates a convergence and accumulation of freshwater in the  
255 gyre, similar to that recently observed in the Arctic, albeit to a lesser extent (e.g. Giles et al., 2012).  
256 In contrast, salty water is accumulated in the topographically constrained cyclonic circulation in  
257 the Eurasian Basin. The SSH change pattern (Fig. 7) reveals a strong decrease of the SSH in the  
258 Eurasian Basin, indicative of a stronger cyclonic gyre in that basin. This suggests that, in the

259 model, Atlantic Water entering the Arctic through Fram Strait tends to penetrate further around  
260 the Eurasian basin in the  $4\times CO_2$  run, while the branch of Atlantic Water recirculating just north  
261 of Fram Strait (e.g. Bourke et al., 1988) is stronger in the *CTRL* run. Atlantic Water in the  $4\times CO_2$   
262 run also tends to remain longer at the surface in the Arctic Basin and only subducts under the  
263 mixed layer or the halocline when it reaches the St. Anna Trough, located on the northern shelf of  
264 the Barents Sea, East of Franz Joseph Land. The increase in intensity of the two gyres within the  
265 Arctic Basin occurs whilst (a) the net volume exchange through the various Arctic gateways does  
266 not differ by more than 10%, and (b) there is no significant change in pattern or intensity of sea  
267 level pressure between the two runs (not shown). This suggest that the spin up of the gyre must  
268 be at least partly a response to intensification of the ocean surface stress resulting from reduction  
269 of the sea ice cover, although the impact of sea ice reduction on momentum transfer to the ocean  
270 surface is strongly model-dependent and even its sign is still a matter of debate (e.g. Martin et al.,  
271 2014; Tsamados et al., 2014; Martin et al., 2016).

272

273 The mean warming and freshening of the Arctic surface both contribute to the decrease of the  
274 surface density, by  $0.5\text{ kg/m}^3$  on average (Fig. 6(c)). The decrease in annual mean basin-averaged  
275 density is due to both surface warming in summer and freshening in winter (Fig. 6(a, b)). Examining  
276 the spatial pattern of the change in surface density (Fig. 5), we find a similar pattern to that of  
277 salinity (Fig. 4), with lighter water in the Canadian Basin and denser in the Eurasian Basin in the  
278  $4\times CO_2$  run compared to the *CTRL* run. The anomalies are of similar amplitude ( $\sim 3\text{ kg/m}^3$ ) and  
279 tend to compensate at the basin scale.

280

281 All these surface property changes, notably surface density, also affect the ocean stratification  
282 (Fig. 8). We define stratification as the density difference between a depth of 500m and the ocean

283 surface, in order to capture the change over the depth range of the halocline everywhere in the  
284 Arctic Basin; however, using a shallower level (200m as in Capotondi et al. (2012)) results in  
285 similar patterns of stratification change (not shown). As expected, the change in stratification is  
286 mostly driven by changes in surface density, apparent from the similar patterns seen in Figs 5 and  
287 8. Indeed, sub-surface changes in density at 500m do not exceed  $0.3 \text{ kg/m}^3$ . In the  $4\times\text{CO}_2$  run,  
288 stratification becomes very weak in the Eurasian Basin. This provides a preconditioning favourable  
289 for deep mixing to occur as soon as the sea ice edge retreats enough to allow for intense and localized  
290 cooling of the ocean surface. In contrast, the surface stratification in the Canadian Basin is strongly  
291 enhanced, which will tend to suppress vertical mixing and favour the formation of sea ice in winter  
292 (Davis et al., 2016).

## 293 **5 Potential impacts for the Arctic Basin and beyond**

294 Changes in MLD and stratification in the Arctic Basin under a warming climate are expected to  
295 have multiple profound effects, on the ocean and climate as well as on other components of the  
296 Earth system, both locally in the Arctic and on a global scale.

297 First, the changes highlighted in sections 3 and 4 will have a large impact on the primary  
298 production in the Arctic Basin, and also affect biogeochemical cycles, including the carbon cycle.  
299 Arrigo and van Dijken (2011) have shown that the net primary production in the Arctic Basin  
300 increased by 20% between 1998 and 2009 in response to the reduction in sea ice extent and the  
301 increase in duration of the open water season. Duarte et al. (2012) also suggested that these changes  
302 have resulted in a shift of the Arctic planktonic community towards smaller species. As the sea ice  
303 cover will continue to retreat further and for longer each year, allowing more input of momentum  
304 and light to the ocean, net primary production will likely continue to increase in the future. The  
305 change in stratification is also an important factor determining the amount of primary production:

306 stratification limits the nutrient supply from the deep layers and inhibits phytoplankton growth.  
307 Given that the changes in stratification (Fig. 8) simulated by the HiGEM model are of opposite  
308 sign in the Eurasian and Canadian Basins, future change in basin-wide net primary production  
309 will depend on regional conditions in these two basins. Dependency on regional scale changes likely  
310 explains some of the discrepancy (in sign and amplitude) amongst CMIP5 model forecasts of Arctic  
311 primary production under a warming climate (Vancoppenolle et al., 2013). Changes in the biology,  
312 together with those in MLD, will also impact the biological and physical carbon pumps in the Arctic  
313 (which currently accounts for roughly 10% of net global carbon uptake (Bates and Mathis, 2009;  
314 MacGilchrist et al., 2014)), although predicting the impacts of increasing  $CO_2$  once again requires  
315 an understanding of changes on a regional scale.

316 Second, the deepening of the mixed layer in some parts of the Arctic Basin will impact ven-  
317 tilation of the Arctic interior. This could consequently affect the export of dense water through  
318 Fram Strait into the Nordic Seas, and modify the downstream properties governing the overflows  
319 over the Greenland-Scotland Ridge. At the end of the  $4\times CO_2$  run, the bottom density in the deep  
320 Eurasian Basin has increased by  $0.25\text{--}0.5\text{ kg/m}^3$  compared to the *CTRL* run (Fig. 9). This density  
321 change results from an increase in salinity by  $0.5\text{--}1$  psu. In contrast, the density at the bottom of  
322 the Canadian Basin remains roughly similar to that in the *CTRL* run. The bottom waters on the  
323 shallow Arctic shelves evolve differently from the deep waters in the basins. In HiGEM, most water  
324 masses found at the bottom on the shallow Arctic shelves become lighter, due to both a freshening  
325 and a warming. This is in agreement with results from Heuzé et al. (2015), who find a consistent  
326 change over the Arctic shelves when looking at the CMIP5 models under the RCP8.5 emissions  
327 scenario. They estimate a multi-model mean decrease in bottom density of  $0.62\text{ kg/m}^3$  averaged  
328 over the shelves shallower than 1000m north of  $60^\circ\text{N}$ . In HiGEM, the Kara Sea exhibits a mean  
329 trend that is very different to that on other Arctic shelves. Water masses at the bottom of the

330 Kara Sea become denser and saltier (Fig. 9). This salinification is linked with the thick ( $\sim 2m$ ) sea  
331 ice that persists in winter in the region, suggesting strong brine rejection during sea ice formation.  
332 Denser waters found at the bottom of the Eurasian Basin are likely a consequence of the deepening  
333 of the MLD described in Section 3, which occurs near the continental slope in the Eurasian Basin;  
334 dense waters formed by convection over the slope in that location will cascade to the bottom of the  
335 basin. Another process that could explain this bottom density increase is export of dense waters  
336 formed in the Kara Sea and in the St. Anna Trough. It is, however, very likely that, in the  $4\times CO_2$   
337 run, the model overestimates the formation of dense water through open ocean convection and un-  
338 derestimates (or does not properly represent) the component formed through shelf-processes; this  
339 is a common bias of most coupled climate models (e.g., Heuzé et al., 2013).

340

341 The increase in bottom density remains confined to the Eurasian Basin in the HiGEM 130-year  
342 long  $4\times CO_2$  run. A section of density across Fram Strait, which is the only connection between the  
343 deep Arctic and the deep Nordic Seas, does not reveal any significant change between the two runs  
344 below 1000m (Fig. 10). In contrast, the upper layer becomes less dense, due to a freshening of the  
345 export to the North Atlantic on the western side of the strait and a warming of the Atlantic Water  
346 flowing into the Arctic on the eastern side of the strait. Note that the volume exported to the North  
347 Atlantic through the deeper part of Fram Strait is also roughly equal in the *CTRL* and  $4\times CO_2$  runs.

348

349 Based on observations of the deep Greenland Sea, Langehaug and Falck (2012) and Somavilla  
350 et al. (2013) have both suggested that the relative proportion of Arctic-origin dense water has in-  
351 creased compared to the dense water mass formed locally during convective events in the Greenland  
352 Sea. The Arctic-origin dense water they observe has formed on the Arctic shelves and subsequently  
353 filled the deep Arctic Basin. In the future, denser water formed in the Eurasian Basin and further

354 exported to the Greenland Sea could in principle modify the direct contribution of the Arctic Basin  
355 to the dense overflows in the Nordic Seas and thus to the deeper branch of the AMOC. We do  
356 not find evidence that deep hydrographic changes in the Arctic influence overflow properties in our  
357 short HiGEM experiment by the end of the  $4\times CO_2$  run, but we cannot rule out that such an effect  
358 could arise in the longer term.

## 359 **6 Summary and conclusions**

360 The depth of the mixed layer and its variations are important controls on a wide range of ocean  
361 processes, including upper ocean productivity, air-sea exchange processes and ventilation of the  
362 ocean interior. Examining the changes in MLD projected by CMIP5 models has thus been the focus  
363 of several recent studies that have pointed out a general tendency for shallowing in regions where  
364 deep mixed layers are presently observed in the North Atlantic (e.g., Heuzé et al., 2015) and the  
365 Southern Ocean (e.g., Sallée et al., 2013). However, these studies do not investigate the potential  
366 for emergence or migration of deep convection hotspots. Here we have used simulations from the  
367 high-resolution climate model HiGEM to show that new regions of deep convection may appear in  
368 the Arctic Basin under a warming climate.

369  
370 In response to a strong increase in atmospheric  $CO_2$  concentrations, the Arctic Basin becomes  
371 seasonally ice-free, and the Arctic ocean surface becomes consequently much warmer and slightly  
372 fresher on average. Although this is in agreement with previous CMIP5 assessments (e.g., Carton  
373 et al., 2015), we argue that limiting the examination to average basin-scale changes hides more  
374 extreme and important local changes. While surface warming is spatially quite homogeneous across  
375 the whole Arctic Basin, changes in salinity are much more spatially variable: a strong freshening  
376 is observed in the Canadian Basin but a strong salinification occurs in the Eurasian Basin, due to

377 the intensification of the surface circulation in both basins, associated with sea ice retreat. These  
378 local changes in sea surface salinity are ultimately responsible for changes in surface density and  
379 stratification that do not reflect the average conditions of the basin: stratification is strongly en-  
380 hanced in the Canadian Basin but strongly reduced in the Eurasian Basin. The drastic decrease  
381 in stratification in the Eurasian Basin results in conditions that promote the appearance of deep  
382 convection in that basin: the northward migration of the winter sea ice edge results in localized  
383 Eurasian air-sea buoyancy fluxes that induce severe local deepening of the mixed layers (down to  
384 500m) in that region, able to ventilate the deep Arctic.

385

386 A northward retreat of the position of the sea ice edge in winter is a necessary first step for  
387 deep convection to occur in the Arctic Basin. Although most of the CMIP5 models and HiGEM  
388 predict that the transition toward a summer ice-free Arctic will only occur during the second half  
389 of the century (Stroeve et al., 2012; Wang and Overland, 2012), it has also been suggested that  
390 these projections of future sea ice conditions are likely too conservative, even under the most pes-  
391 simistic emission scenario (Mahlstein and Knutti, 2012; Stroeve and Notz, 2015). This is due to the  
392 mis-representation of important feedbacks between the different components of the Arctic system  
393 (e.g., Lique et al., 2016) that results in an underestimation of the so-called 'Arctic amplification of  
394 climate change' (Serreze and Barry, 2011; Pithan and Mauritsen, 2014).

395

396 Unfortunately, the influence of new regions of deep mixed layers on the deep Arctic, and their  
397 subsequent effects on the Greenland Sea and possibly the Atlantic Meridional Overturning Cir-  
398 culation downstream, cannot be established with the current  $4\times CO_2$  model run. Owing to the  
399 high-resolution and computational requirements of HiGEM, and the fact that the largest changes  
400 were only observed at the end of the run, the available simulation is currently too short to inves-

401 tigate the consequences for the deep water with any confidence. The duration of the simulation is  
402 also too short to rule out the hypothesis that emergence of deep convection in the Eurasian Basin  
403 is a transient signal that would vanish as the surface temperature continues to increase (enhancing  
404 the surface stratification) associated with further sea ice retreat (Brodeau and Koenigk, 2016). Our  
405 results nonetheless indicate that ventilation of the Arctic by local deep mixing events is possible  
406 in a high  $CO_2$  future (and possibly sooner than forecast by the CMIP5 and HiGEM models, if  
407 the sea ice were to retreat sooner than predicted), and we expect that these changes would have  
408 important downstream ramifications for deep water formation processes and deep water properties  
409 in the North Atlantic. Probing the emergence of deep convection in the Arctic Basin in a variety  
410 of climate models at a variety of resolution forced with different emission scenario is needed to test  
411 the robustness of our findings based on the HiGEM model.

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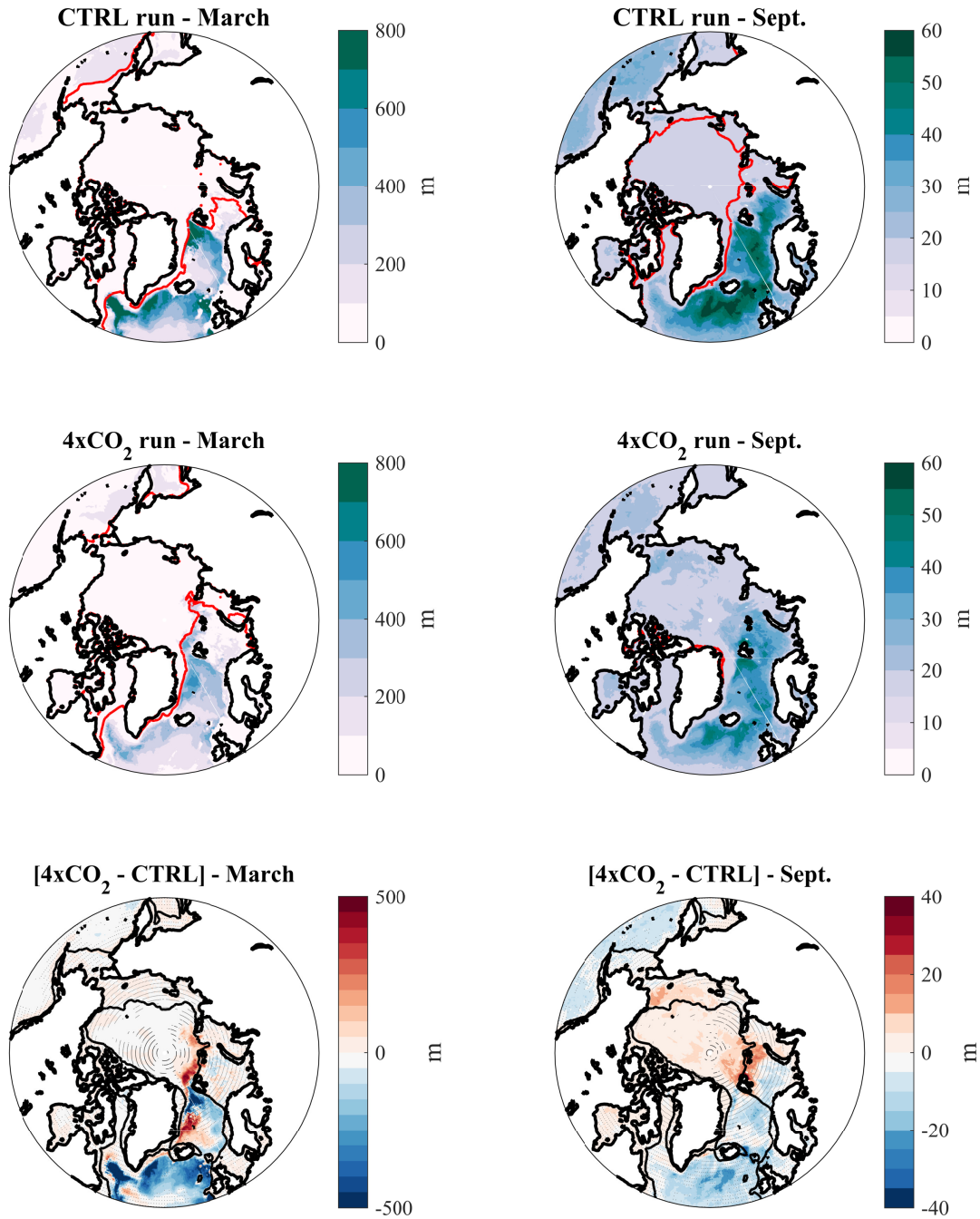


Fig. 1: Mean March and September mixed layer depth (in m) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The red contour corresponds to the location of the sea ice edge (defined as the 15% concentration contour), and the black contour on the bottom panels shows the 500m isobath.

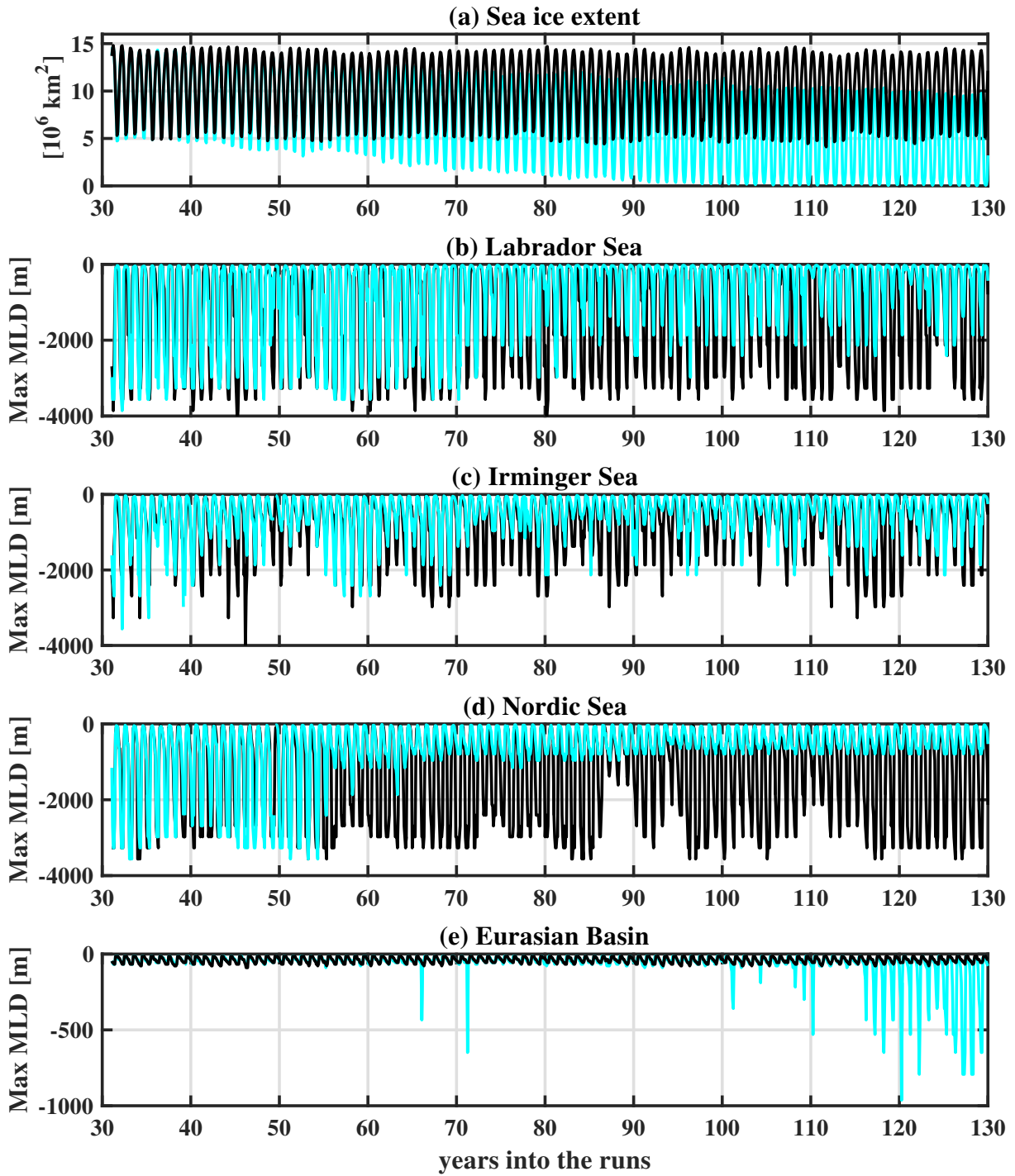


Fig. 2: Time series of (a) sea ice extent, and the maximum MLD detected in (b) the Labrador Sea, (c) the Irminger Sea, (d) the Nordic Sea, and (e) the Eurasian Basin for the last 100 years of the *CTRL* (black) and the  $4 \times CO_2$  (light blue) runs.

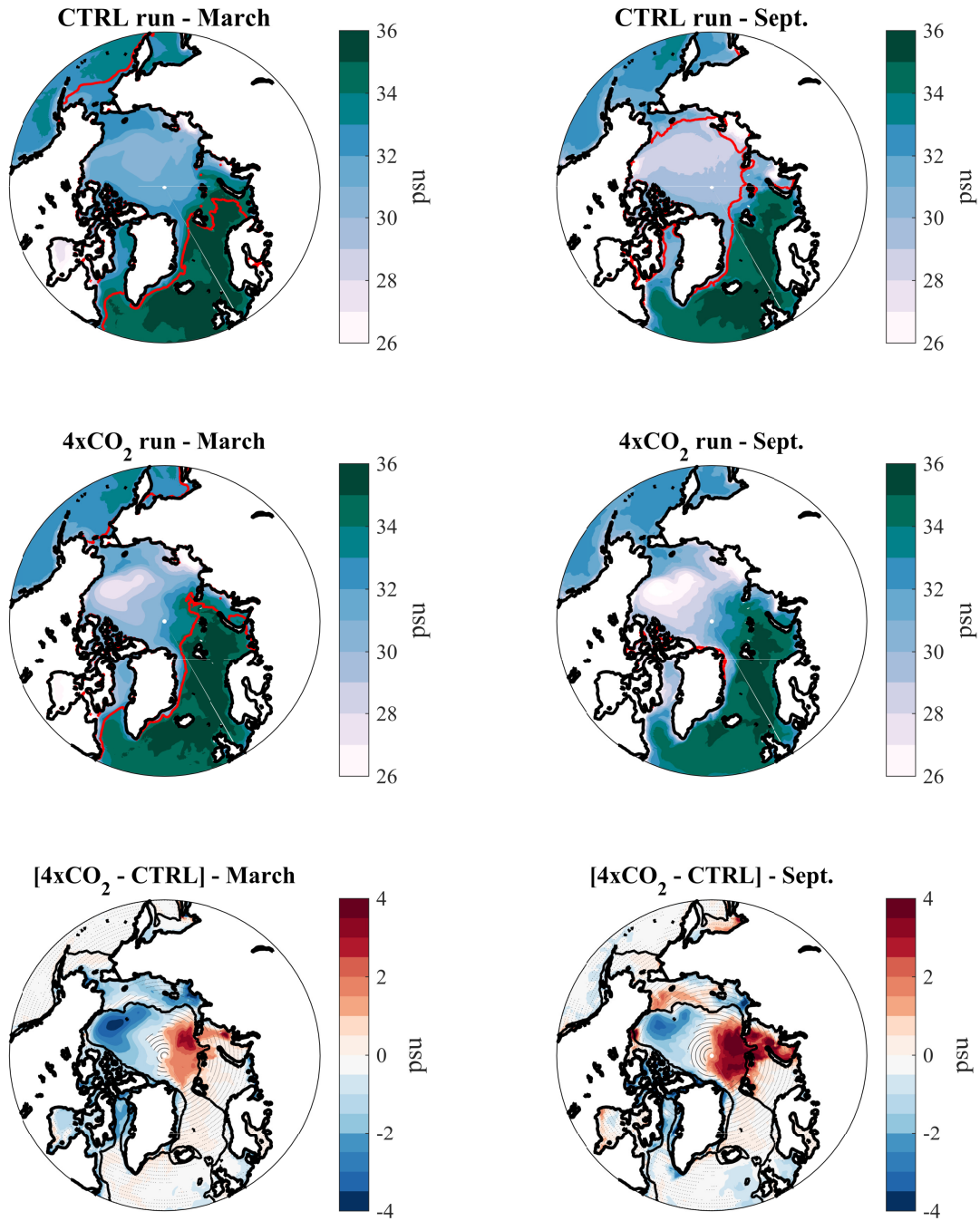


Fig. 3: Mean March and September Sea Surface Temperature (SST, in °C) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The red contour corresponds to the location of the sea ice edge (defined as the 15% concentration contour), and the black contour on the bottom panels shows the 500m isobath.

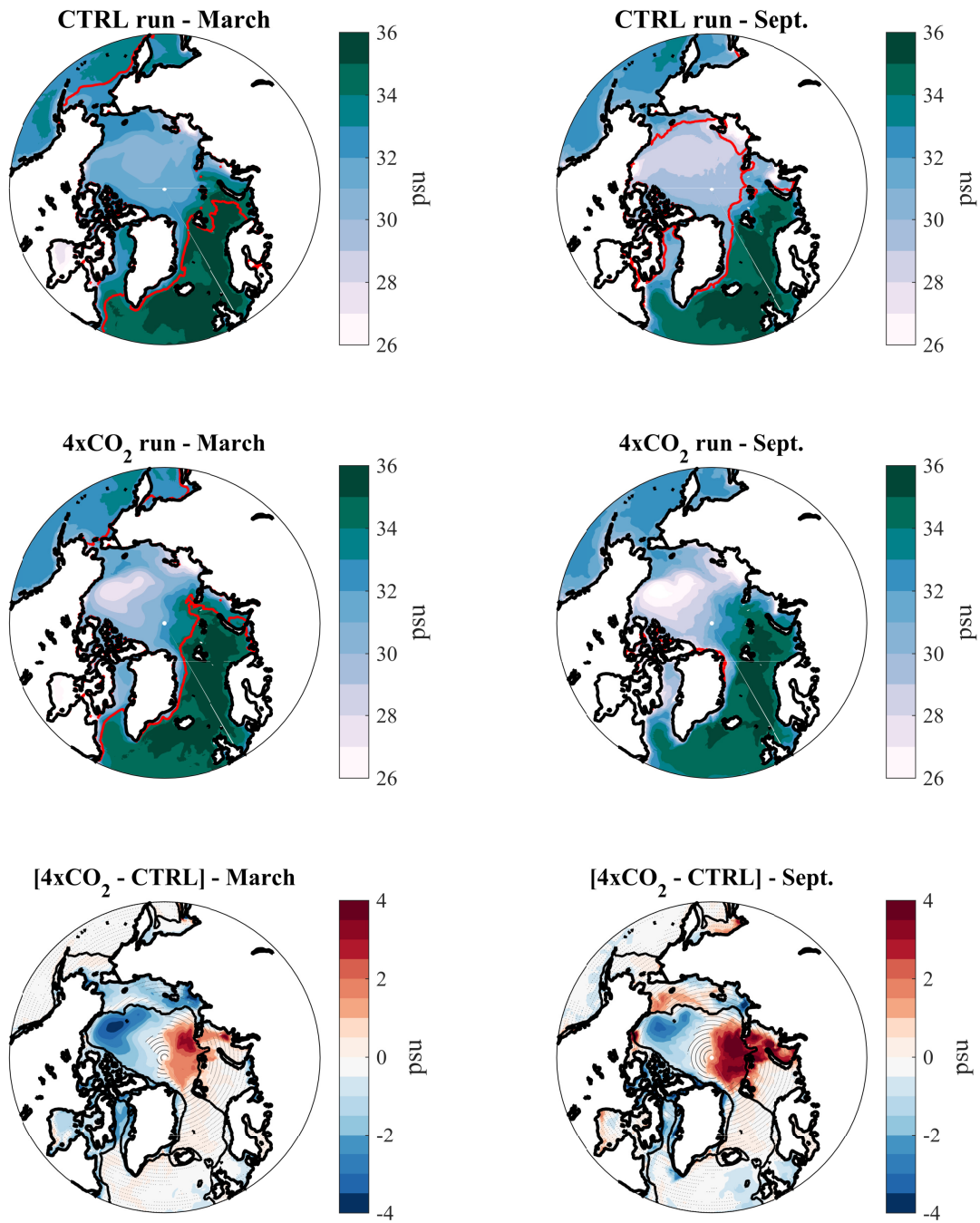


Fig. 4: Mean March and September Sea Surface Salinity (SSS, in psu) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The red contour corresponds to the location of the sea ice edge (defined as the 15% concentration contour), and the black contour on the bottom panels shows the 500m isobath.

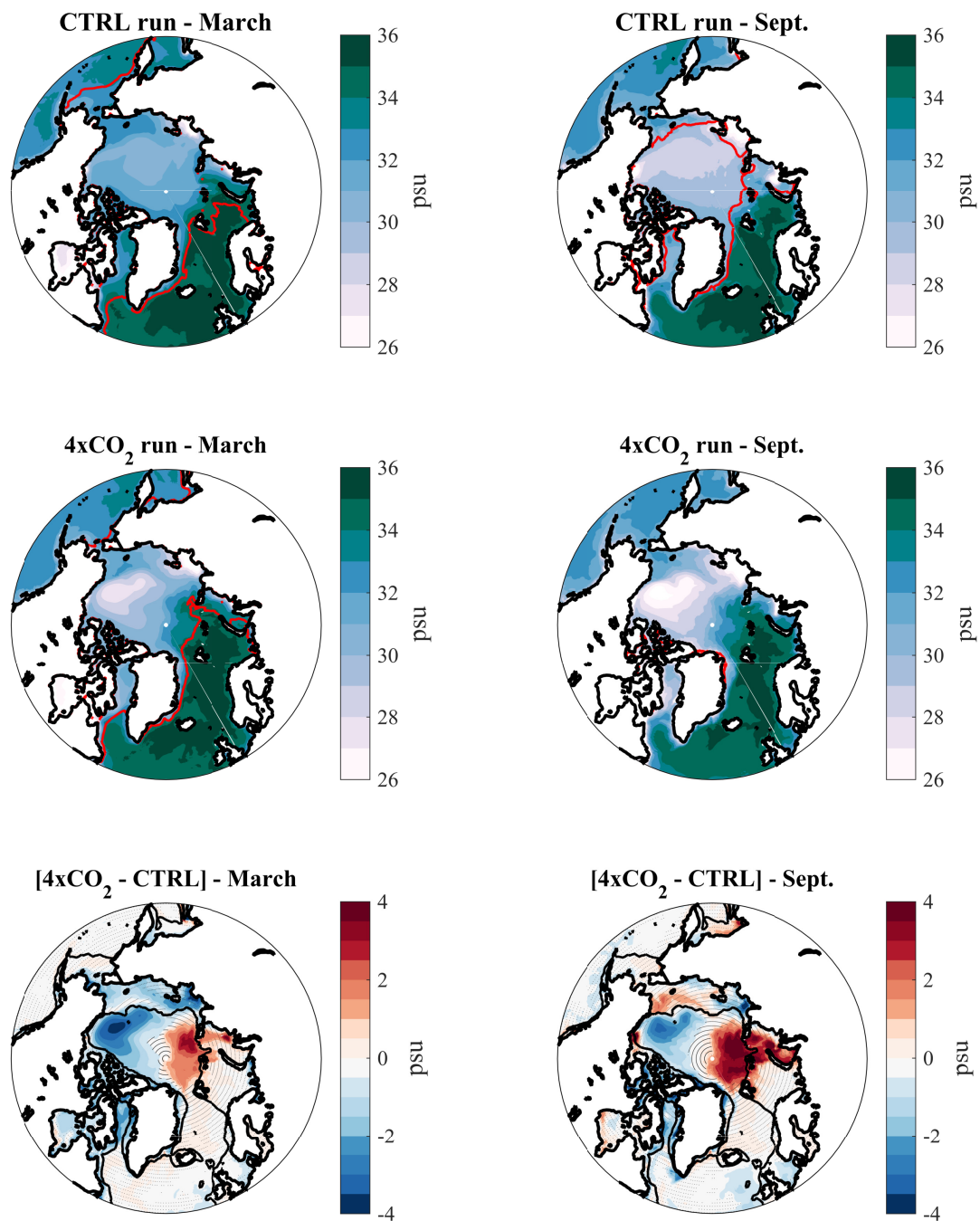


Fig. 5: Mean March and September sea surface potential density ( $\sigma$ , in  $kg/m^3$ ) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The red contour corresponds to the location of the sea ice edge (defined as the 15% concentration contour), and the black contour on the bottom panels shows the 500m isobath.

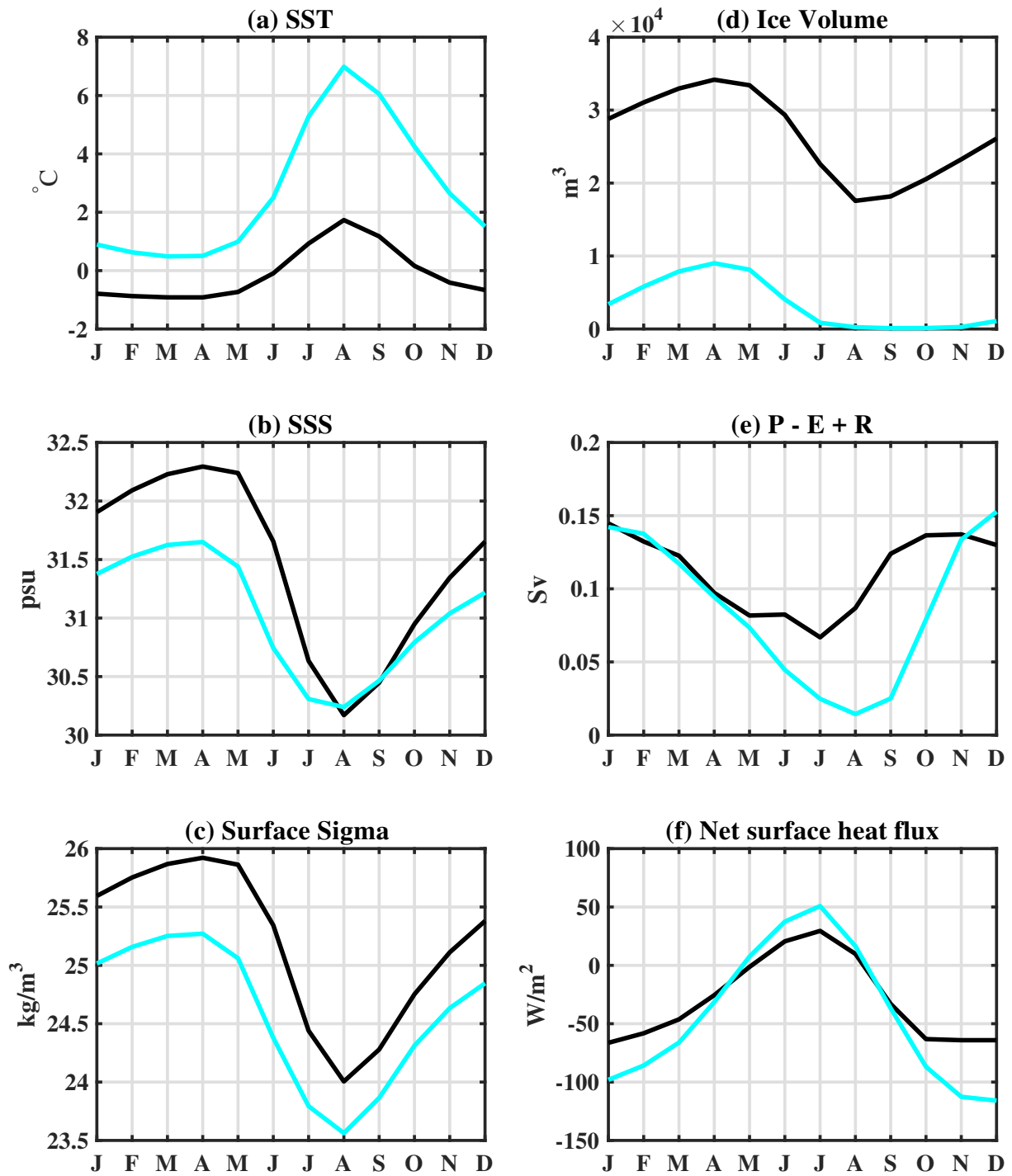


Fig. 6: Mean seasonal cycles of the SST (in °C), SSS (in psu), surface density (in  $kg/m^3$ ), ice volume (in  $km^3$ ), sum of precipitation (as water and snow) minus evaporation plus river runoff (in Sv) and net surface heat flux (in  $W/m^2$ ), averaged over the region north of 65°N for the last ten years of the CTRL (black) and the 4 × CO<sub>2</sub> (light blue) runs.

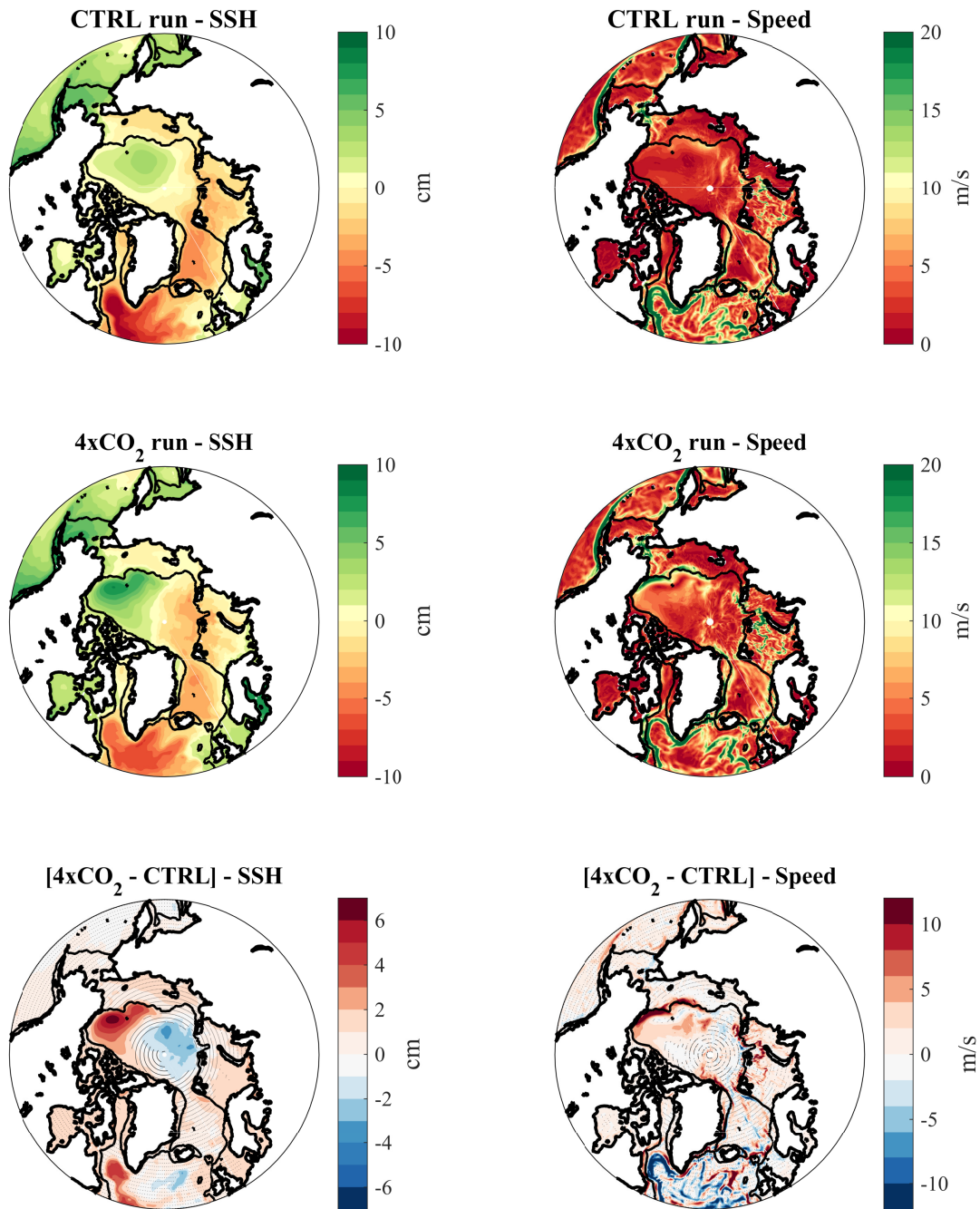


Fig. 7: 10-year averages of Sea Surface Height (SSH in *cm*, left column) and the mean speed over the first 100m (in *cm/s*, right column) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The black contour shows the 500m isobath.

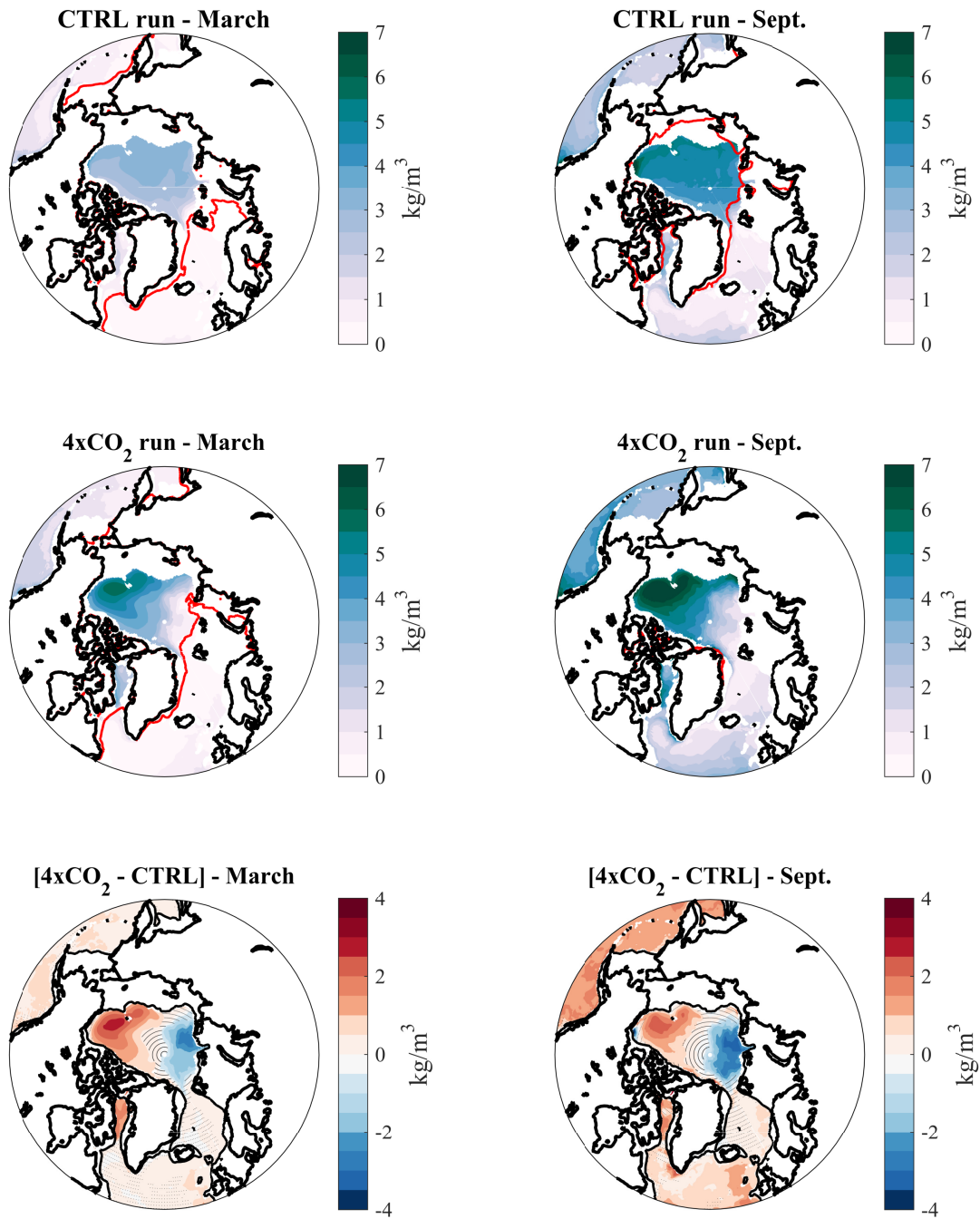


Fig. 8: Mean March and September stratification (defined as the density difference between 500m and the surface, in  $\text{kg/m}^3$ ) for the *CTRL* and the  $4 \times \text{CO}_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. The red contour corresponds to the location of the sea ice edge (defined as the 15% concentration contour), and the black contour on the bottom panels shows the 500m isobath.

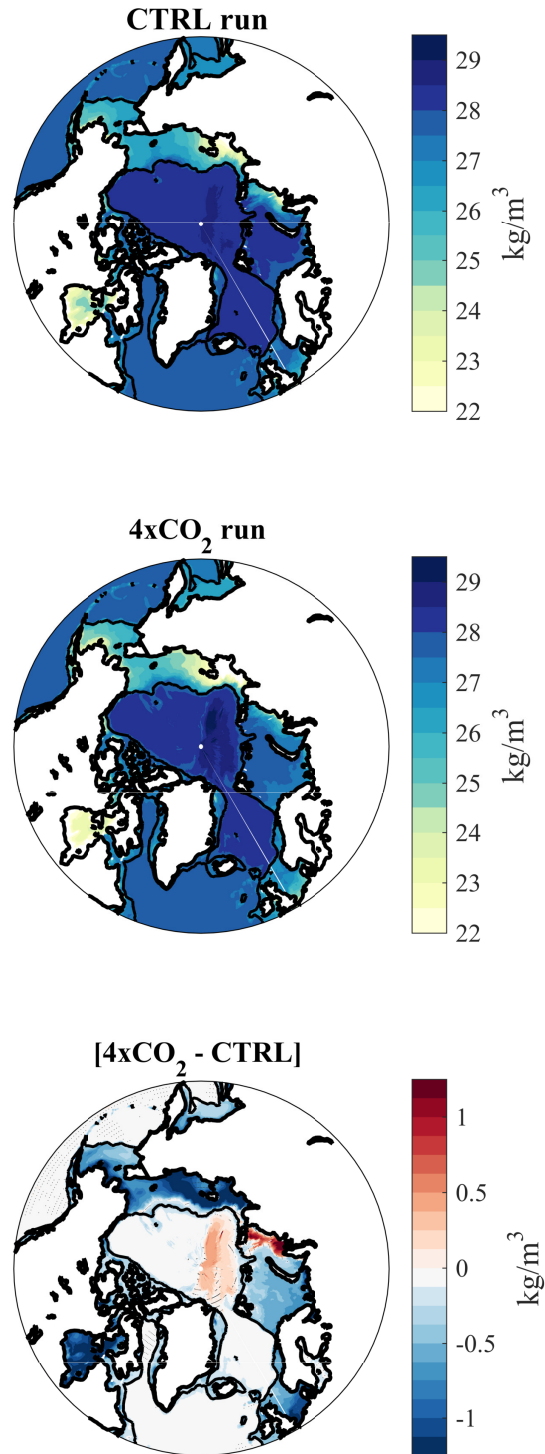


Fig. 9: 10-year average of the bottom density (in  $\text{kg/m}^3$ ) for the *CTRL* and the  $4 \times \text{CO}_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates areas where the difference between the two runs is not significant. Bottom density corresponds to the density of the last ocean model vertical level at each grid point, and is computed relative to the surface. The black contour shows the 500m isobath.

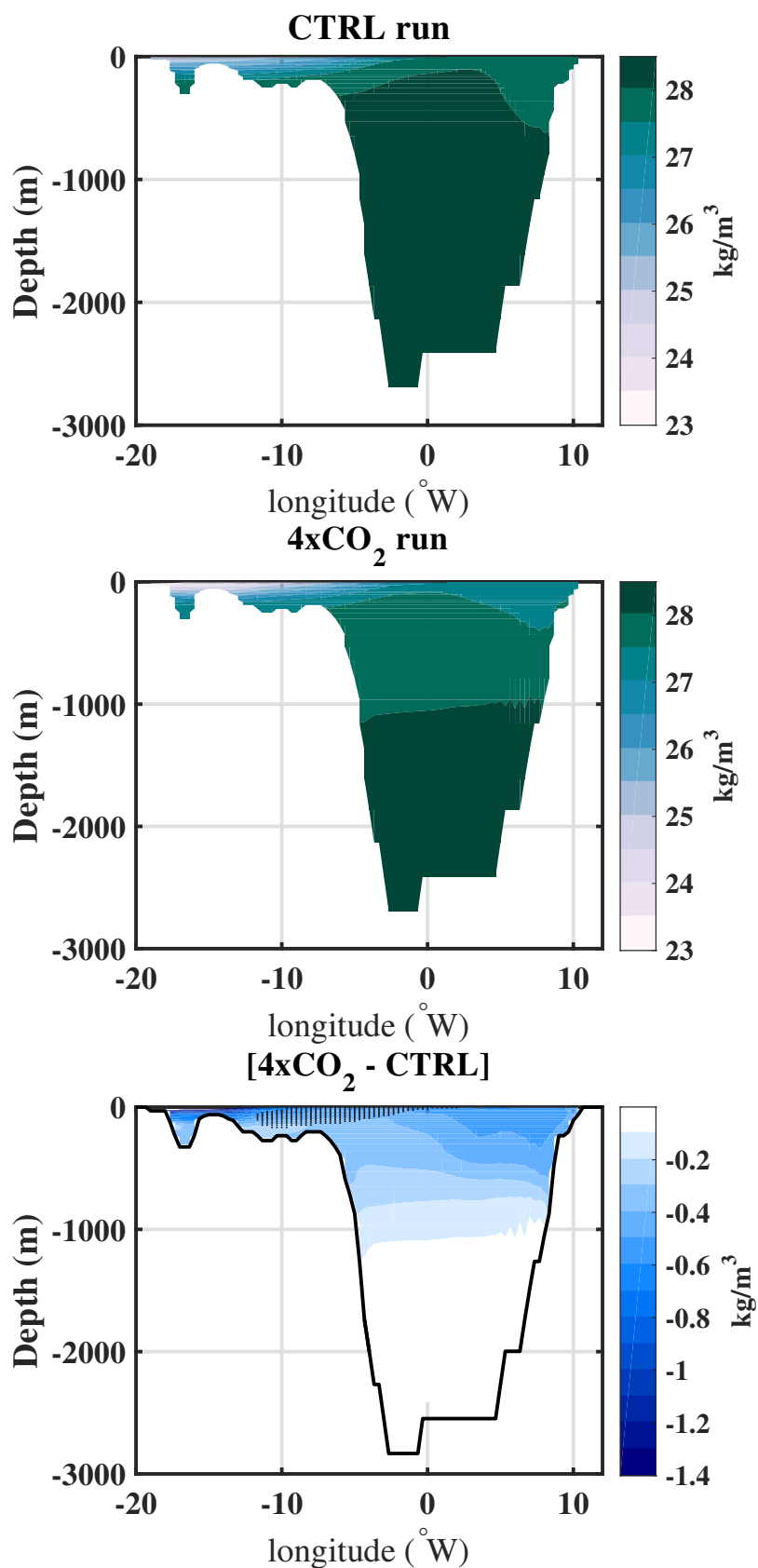


Fig. 10: Section across Fram Strait of the 10-year average density (in  $kg/m^3$ ) for the *CTRL* and the  $4 \times CO_2$  runs and the difference between the two runs. For the bottom row, black hatching indicates the parts of the strait where the difference between the two runs is not significant.