



# 1 **Topological constraints by the Greenland-Scotland Ridge on AMOC and**

## 2 **climate**

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## ABSTRACT

14 Changes in the geometry of ocean basins have been influential in driving climate change  
15 throughout Earth's history. Here, we focus on the emergence of the Greenland-Scotland  
16 Ridge (GSR) and its influence on the ocean state, including large-scale circulation, heat  
17 transport, water mass properties and global climate. Using a coupled atmosphere-ocean-  
18 sea ice model, we consider the impact of introducing the GSR in an idealized Earth-like  
19 geometry, comprising a narrow Atlantic-like basin and a wide Pacific-like basin. Without  
20 the GSR, deep-water formation occurs near the North Pole in the Atlantic basin, associated  
21 with a deep meridional overturning circulation (MOC). By introducing the GSR, the volume  
22 transport across the sill decreases by 64%, and deep convection shifts south of the GSR,  
23 dramatically altering the structure of the high-latitude MOC. Due to compensation by the  
24 subpolar gyre, the northward ocean heat transport across the GSR only decreases by  $\sim 30\%$ .  
25 As in the modern Atlantic ocean, a bidirectional circulation regime is established with warm  
26 Atlantic water inflow and a cold dense overflow across the GSR. In sharp contrast to the large  
27 changes north of the GSR, the strength of the Atlantic MOC south of the GSR is unaffected.  
28 Outside the high-latitudes of the Atlantic basin, the surface climate response is surprisingly  
29 small, suggesting that the GSR has little impact on global climate. Our results suggest that  
30 caution is required when interpreting paleoproxy and ocean records, which may record large  
31 local changes, as indicators of basin-scale changes in the overturning circulation and global  
32 climate.

## 33 **1. Introduction**

34 Changes in the distribution of land masses driven by continental drift and plate tectonics play  
35 a fundamental role in shaping the geometry of ocean basins. The presence of meridional barriers  
36 in the ocean has a profound effect on large-scale ocean circulation, localization of deep water  
37 formation and oceanic heat transport that help modulate global energy transports and control the  
38 Earth's energy budget (Toggweiler and Bjornsson 2000; Enderton and Marshall 2009; Ferreira  
39 et al. 2010). Changes in ocean basin geometry have been invoked as a key factor in setting the  
40 mean climate of the Earth, explaining some of the major transitions in global climate over the past  
41 50 million years (Barker and Burrell 1977; Haug and Tiedemann 1998).

42 Bathymetry, such as oceanic ridges, also influence ocean basin geometry and global ocean cir-  
43 culation. Topographic features can steer major ocean currents, enhance vertical mixing over rough  
44 topography (Polzin et al. 1997; De Lavergne et al. 2017) providing a major energy source for  
45 driving the meridional overturning circulation and serve as solid barriers limiting water exchange  
46 between adjacent ocean basins (Gille and Smith 2003). Hence, exploring the impacts of changes  
47 in bathymetry on global ocean circulation allows us to separate the effects of bathymetry from  
48 changes in e.g. CO<sub>2</sub> in driving major climate events (e.g., the glaciation of Antarctica: Drake  
49 Passage opening versus declining atmospheric CO<sub>2</sub> (DeConto and Pollard 2003)).

50 In the North Atlantic region, oceanic gateway changes induced by the emergence of the  
51 Greenland-Scotland Ridge (GSR) have been argued to play an important role in the evolution  
52 of high latitude surface climate and North Atlantic ocean circulation throughout the Cenozoic (the  
53 past 65 million years to present) (Wright and Miller 1996; Stürz et al. 2017). The GSR presents a  
54 zonal barrier separating the North Atlantic from the Nordic Seas and Arctic Ocean and constricting  
55 the exchange of water between these ocean basins (Fig. 1). At present, the main sills of the GSR

56 are less than 500 m deep, with deeper channels at water depths of  $\sim 840$  m, where dense waters  
57 formed in the Nordic Seas can escape. As these cold, dense waters flow over the sill, they mix with  
58 warmer, lighter Atlantic water to form North Atlantic Deep Water (NADW) (Hansen and Øster-  
59 hus 2000), which constitutes the lower branch of the Atlantic Meridional Overturning Circulation  
60 (AMOC). Therefore even minor changes in sill-depth may significantly impact deep-water ex-  
61 change and NADW volume transport, potentially causing basin-wide changes in circulation (e.g.,  
62 Roberts and Wood 1997).

63 Paleo-oceanographic data suggests that variations in the depth of the GSR correlates with major  
64 changes in North Atlantic and global climate (see Uenzelmann-Neben and Gruetzner 2018, for  
65 a review). In particular, during the late Eocene to early Miocene period (35 - 16 Ma), the GSR  
66 gradually deepened, but was interrupted by episodes of topographic uplift, thus impacting the  
67 water exchange between the Norwegian Seas and North Atlantic (Wright and Miller 1996; Davies  
68 et al. 2001; Via and Thomas 2006). Evidence from  $\delta^{13}\text{C}$ -records in the deep North Atlantic have  
69 suggested that, at times when the GSR was higher, the production of Northern Component Water  
70 (NCW; a precursor of the modern NADW) ceased, with limited deep-water exchange across the  
71 GSR. As the GSR deepened, it resulted in high NCW fluxes into the deep North Atlantic marking  
72 the onset of the modern AMOC. This onset of the interhemispheric northern-sourced circulation  
73 cell has also been suggested to play a role in the glaciation of Antarctica at the Eocene-Oligocene  
74 Transition (EOT), about 33.5 Myr ago, prompting the EOT global cooling (Abelson et al. 2008;  
75 Abelson and Erez 2017). However, there is much uncertainty associated with the timing of tectonic  
76 gateway changes and their actual impact on AMOC evolution (see Fig. 3 in Ferreira et al. 2018).

77 It is widely accepted that the presence of the GSR is important for the surface climate conditions  
78 and the formation of dense water in the Nordic Seas, based on both numerical model simulations  
79 and observational data (e.g., Roberts and Wood 1997; Iovino et al. 2008). Meanwhile, its effect

80 on large-scale ocean circulation is not well established. In some models, even modest changes in  
81 the sill-depth produce large changes in Atlantic overturning circulation, with major consequences  
82 for ocean heat transport (OHT) and North Atlantic surface climate (e.g., Roberts and Wood 1997),  
83 while in others changes in the sill does not affect AMOC strength (Robinson et al. 2011; Born  
84 et al. 2009).

85 Using a coupled climate model with Miocene boundary conditions ( $\sim 20$ -15 Ma), Starz et al.  
86 (2017) investigated the long-term subsidence history of the GSR. As the GSR deepened, the Arctic  
87 Ocean gradually transitioned from a freshwater dominated environment towards the establishment  
88 of a bi-directional flow regime characterizing the modern North Atlantic-Arctic water exchange.  
89 This lead to an "Atlantification" of the Nordic Seas and Arctic Ocean and increased sea surface  
90 temperature (SST) and salinity (SSS) in the high northern latitudes. A similar mechanism has been  
91 invoked to explain the observed high-latitude warming of the Mid-Pliocene warm period (MPWP,  
92 3.3-3.0 Ma) (Robinson et al. 2011; Hill 2015). Reconstructions of ocean temperature from marine  
93 proxy data indicate that, during the MPWP, global mean temperature was warmer by 2 – 3°C  
94 compared to modern (Dowsett et al. 2010), while summer temperatures in the Arctic were about  
95 8°C warmer (Brigham-Grette et al. 2013). As a consequence, the North Atlantic equator-to-pole  
96 SST gradient was reduced to about 18°C (modern is  $\sim 27^\circ\text{C}$ ). The favoured hypothesis for the  
97 Mid-Pliocene warmth is that the AMOC was stronger (inferred from the observed SST pattern),  
98 contributing to enhanced meridional heat transport in the North Atlantic ocean (Dowsett et al.  
99 1992; Raymo et al. 1996; Dowsett et al. 2009). Using a climate model simulation with Pliocene  
100 boundary conditions, Robinson et al. (2011) argued that such changes in ocean heat transport could  
101 be explained by depth variations in the GSR and showed that lowering the sill by 800 m leads to  
102 a stronger overturning circulation in the subpolar region, increased high-latitude OHT and higher  
103 Arctic SSTs, consistent with the proxy data.

104 Common for many of these earlier studies (both proxy and model based) is that they assume  
105 AMOC strength and poleward ocean heat transport are correlated with the height of the GSR  
106 through the production of dense water in the Nordic Seas and its control on the overflow strength  
107 (Dowsett et al. 1992; Wright and Miller 1996; Raymo et al. 1996; Dowsett et al. 2009; Robinson  
108 et al. 2011). There are two main issues with this view. Firstly, from a modern observational  
109 point of view there has been no conclusive evidence linking AMOC variability to the formation  
110 of Nordic Seas overflow water (Hansen and Østerhus 2007; Olsen et al. 2008; Medhaug et al.  
111 2012; Lozier et al. 2017). Rather, AMOC variability is largely attributed to deep water formation  
112 south of the GSR in the subpolar North Atlantic (Olsen et al. 2008). Secondly, it assumes that  
113 the OHT across the GSR is nearly completely dominated by the AMOC, which conflicts with a  
114 number of studies (e.g., Wunsch 2005; Ferrari and Ferreira 2011; Tett et al. 2014; Årthun et al.  
115 2019) showing a decoupling between northern high-latitude climate and the AMOC at 26°N. This  
116 calls into questions the role of the GSR in controlling AMOC strength and high-latitude OHT, in  
117 particular on geological time-scales, where changes in sill-depth are invoked as the primary driver  
118 for changes in large-scale ocean circulation and global climate.

119 In this study, we explore the impact of introducing the GSR in a coupled ocean-atmosphere-sea  
120 ice general circulation model (GCM) with idealised land-sea geometry, focusing on the response  
121 of the Atlantic overturning circulation. Rather than investigating a specific time in Earth's history,  
122 we consider the fundamental role of ocean-basin geometry (induced by changes in bathymetry) in  
123 shaping the ocean circulation and mean climate. This builds on previous work by Ferreira et al.  
124 (2010). We note, however, that the present study provides valuable insight into understanding  
125 Cenozoic climate evolution, where Arctic-Atlantic gateway changes are likely to have played a  
126 major role (Stärz et al. 2017; Hutchinson et al. 2019).

127 In contrast to earlier model studies, we focus on a detailed understanding of how the presence  
128 of the GSR influences the location of deep water formation and its role in shaping the overturning  
129 circulation and ocean heat transport. Special attention is paid to the relationship between the  
130 AMOC and the northward transport of heat across the GSR and water mass transformation. This  
131 allows us to study and review how long-term changes in sill-depth, inferred from proxy-records,  
132 can impact high-latitude climate as well as global surface climate.

## 133 **2. Model and experiment**

134 We use the coupled atmosphere-ocean-sea ice configuration of the Massachusetts Institute of  
135 Technology General Circulation Model (MITgcm) (Marshall et al. 1997) with idealized land-sea  
136 geometry, which has been used previously to investigate fundamental aspects of large-scale ocean  
137 circulation and its impact on mean climate (e.g., Marshall et al. 2007; Ferreira et al. 2010, 2011;  
138 Rose et al. 2013). The model uses the rescaled pressure coordinate  $p^*$  for the compressible atmo-  
139 sphere and the rescaled height coordinate  $z^*$  for the Boussinesq ocean (Adcroft and Campin 2004).  
140 The atmosphere, ocean and sea-ice components are configured on the same cubed-sphere grid at  
141 a low-horizontal resolution C32 (32x32 points per face), yielding a resolution of roughly  $2.8^\circ$  at  
142 the equator. The cubed-sphere grid allows for better representation of dynamics in high latitude  
143 regions by avoiding problems with the converging meridian at the poles.

144 The atmospheric model is a five-level primitive equation model of intermediate complex-  
145 ity based on the simplified parameterizations primitive-equation dynamics (SPEEDY) scheme  
146 (Molteni 2003). This method comprises a four-band radiation scheme, a parameterization of moist  
147 convection, diagnostic clouds, and a boundary layer scheme.

148 The flat-bottomed ocean is 3 km deep with 30 vertical levels, increasing from 10 m in the surface  
149 layers to 200 m in the deep ocean. Effects of mesoscale eddies are parameterized as an advective

150 process (Gent and McWilliams 1990) and isopycnal diffusion (Redi 1982), both with a transfer  
151 coefficient of  $1200 \text{ m}^2 \text{ s}^{-1}$ . For the vertical mixing the non-local K-Profile Parameterization (KPP)  
152 scheme (Large et al. 1994) is used, which deals with the different mixing processes in the ocean  
153 interior and surface boundary layer.

154 The sea ice model is a two-and-a-half-layer thermodynamic model based on Winton (2000) with  
155 prognostics variables including sea ice area, snow and ice thickness, brine pockets and sea ice  
156 salinity. There is no ice dynamics, but sea ice deformation is crudely represented by a horizontal  
157 thickness diffusivity of  $2000 \text{ m}^2 \text{ s}^{-1}$ .

158 The land model is a simple two layer model with prognostic temperature, liquid groundwater,  
159 and snow height. Precipitation that falls on land (as snow or rain) is evenly distributed along the  
160 coast as run-off. There is no orographic effect from land masses and no continental ice. Land  
161 albedo is set to 0.10, plus a contribution from snow, if present, which varies from 0.25 to 0.80  
162 depending on snow height, surface temperature and snow age. Orbital forcing and  $\text{CO}_2$  levels are  
163 at present-day values.

#### 164 *a. Description of experiments*

165 To test the impact of the Greenland-Scotland Ridge on ocean circulation and climate, two bathy-  
166 metric configurations of the model are considered. These are illustrated in Fig. 2. The reference  
167 configuration (*noridge*) comprises two  $45^\circ$ -wide strips of land, set  $90^\circ$  apart, extending from the  
168 North Pole to  $40^\circ\text{S}$  and separates the ocean into a small "Atlantic-like" and a large "Pacific-like"  
169 basin, with a zonally unbound Southern Ocean. For simplicity we refer to these as the Atlantic  
170 and Pacific basin respectively. The second configuration (*ridge*) is similar to *noridge*, with the  
171 only difference that an oceanic ridge, extending across the small basin between  $61^\circ - 65^\circ\text{N}$ , is  
172 introduced. The ridge has a uniform sill-depth of 500 meters, roughly corresponding to the av-



173 erage depth of the modern GSR. Thereby, the *ridge* configuration mimics the effect of the GSR,  
174 by separating the Atlantic basin into a semi-enclosed (polar) basin at high latitudes, representing  
175 the Nordic Seas and Arctic Ocean, and a larger basin representing the subtropical and subpolar  
176 Atlantic. In contrast, *noridge* presents an Atlantic ocean without zonal boundaries (i.e. ridges).

177 Despite the simplified geometry, the model captures the general features associated with the  
178 large-scale ocean circulation in the subpolar and high-latitude Atlantic: shallow wind-driven gyres  
179 at mid-latitudes; a surface current resembling the North Atlantic Current (NAC) transporting warm  
180 and saline water from the subtropics to the polar regions; and deep water formation at northern  
181 high latitudes (Fig. 6 and 9). It also reproduces a key asymmetry between the two basins: deep  
182 water formation in the small basin drives a strong and deep meridional overturning circulation (i.e.  
183 an AMOC), while deep water formation is absent in the large basin (see also, Ferreira et al. 2010).

184 The *noridge* was initialized with global temperature and salinity fields from the "Double-Drake"  
185 simulation in Ferreira et al. (2010) and integrated forward until reaching a steady state solution  
186 after 4000 years. The initial conditions for *ridge* were obtained by adding the GSR to the *noridge*  
187 simulation and then run for 4000 years, allowing the ocean state to adjust to the altered bathymetry.  
188 This is sufficient time for both surface and deep waters to equilibrate. Note, that the local ocean  
189 adjustment to the GSR is relatively fast (within the first 200 years). In the following, we compare  
190 the last 100 years of the *ridge* experiment to the last 100 years of the *noridge* solution.

### 191 **3. Results**

#### 192 *a. Mean climate without the Greenland-Scotland Ridge (noridge)*

193 The mean climate of *noridge* is depicted in Fig. 3 (left) showing the annual mean SST and sea  
194 ice thickness. The equilibrium solution of *noridge* is warm with a global mean SST of 20.7°C

195 (Table 1) and a relatively weak equator-pole SST gradient of only  $19.0^{\circ}\text{C}$  (in the Atlantic basin)  
196 owing to the strong northward heat transport associated with the MOC. Consequently, the northern  
197 high latitudes are completely ice-free. In addition, there is an east-west asymmetry in high-latitude  
198 SST and SSS, with warmer and saltier surface waters on the eastern side of the basin, associated  
199 with the northward warm boundary current (analogous to the North Atlantic Drift), and extension  
200 of colder and fresher waters on the western side and in the interior of the basin. In the Southern  
201 Hemisphere, a strong zonal current driven by the intense westerly winds suppresses poleward  
202 ocean heat transport (i.e. the Drake Passage effect; Toggweiler and Samuels (1995)) allowing a  
203 large and thick ( $\sim 10$  m) ice cap to form. We note that the surface climate of *noridge* has many  
204 similarities to the climate of the MPWP, where Arctic sea ice cover was dramatically reduced, or  
205 absent, and the North Atlantic-Arctic SST gradient was reduced to  $\sim 18^{\circ}\text{C}$  (Dowsett et al. 2010).  
206 Therefore, the MPWP is often used as an analogue for future warming scenario.

207 In Fig. 4 we show the residual-mean overturning circulation integrated over the Atlantic basin,  
208 i.e. the total meridional volume transport defined as the sum of the Eulerian meridional velocity,  
209 and the eddy-induced velocity parameterized with the Gent-McWilliams scheme. In the upper  
210 ocean (0-500 m), the circulation is dominated by shallow overturning cells, whose horizontal  
211 structure is related to the wind-driven circulation in the subtropical and subpolar gyres. The deep  
212 overturning circulation is characterized by a clockwise circulation, associated with deep water  
213 formation in the northern high latitudes, extending from below the wind-driven layer to the abyssal  
214 ocean. In the absence of a zonal barrier in the Atlantic basin, the MOC stretches to the North Pole,  
215 where surface water cools and sinks, returning southward in the deep overturning branch with  
216 a maxima of  $23.2$  Sv at  $26^{\circ}\text{N}$ . We note the absence of an abyssal overturning cell associated  
217 with Antarctic Bottom Water (AABW) formation. This is due to insufficient brine release in the  
218 Southern Ocean possibly related to the absence of an Antarctic continent. As a consequence, the

219 overturning circulation in the Atlantic basin is dominated by deep water formation in the North  
220 Atlantic.

221 The meridional ocean heat transport in the Atlantic basin, shown in Fig. 5, is northward ev-  
222 erywhere in the basin, with a cross-equatorial transport of 0.7 PW, that can be attributed to the  
223 MOC transporting warm water north and returning cold water at depth. The Atlantic OHT peaks  
224 at about 20°N with a maximum of 1.18 PW, to be compared to the observed ~1.2 PW (Trenberth  
225 and Caron 2001), and is associated with a "mixed" circulation spanning both the shallow Ekman-  
226 driven subtropical cell and the deep overturning cell (Ferrari and Ferreira 2011). We also notice a  
227 relatively large interdecadal variability (shading) in the subtropical OHT, which is due to stronger  
228 variability in the MOC at these latitudes (not shown). At subpolar and high northern latitudes,  
229 the OHT is dominated by the strong MOC transporting roughly 16 Sv across 70°N (just north of  
230 the GSR) corresponding to a heat transport of 0.22 PW (Table 1) maintaining the northern high  
231 latitudes warm and ice free. By comparison, the observed modern transport across the GSR is only  
232 about 8.5 Sv (Østerhus et al. 2005).

### 233 *b. Effect of the GSR on meridional overturning circulation*

234 When the GSR is introduced, the transport of warm and salty Atlantic water over the GSR by the  
235 MOC is reduced by 64% (from 16 Sv to 5.7 Sv at 70°N). As a consequence, the spatial structure  
236 of the MOC changes dramatically, but mostly at subpolar and high latitudes (Fig. 4; right). The  
237 downwelling branch of the deep overturning cell shifts southwards, and the streamfunction decays  
238 sharply immediately to the south of the GSR. A small overturning circulation remains north of the  
239 ridge, associated with a dense overflow from the polar basin, balanced by a warm inflow at the  
240 surface. Despite significant changes in the structure of the overturning at high latitudes, the MOC  
241 south of 60°N is generally not affected; the volume transport at 26°N decreases from 23.2 Sv to

242 22.1 Sv, i.e. a 5% reduction. This suggests that the net deep water production in the North Atlantic  
243 and high latitudes remains unchanged, while the main difference between *noridge* and *ridge* is in  
244 the localization of deep water formation and MOC structure.

245 The impact of the GSR on the horizontal circulation is illustrated in Fig. 6b, showing the  
246 barotropic streamfunction featuring a basin-wide anticyclonic gyre in the subtropics (STG), a cy-  
247 clonic subpolar gyre (SPG) at sub-arctic latitudes (between 40° and 70°N) and an anticyclonic  
248 gyre at high latitudes driven by the polar easterlies north of 57°N. For *noridge* the wind-forced  
249 barotropic volume transport at the GSR latitudes is relatively small (less than 5 Sv), indicating  
250 that the total transport is dominated by the MOC (Table 1). In *ridge*, however, the barotropic flow  
251 intensifies associated with a poleward expansion of the SPG. Note, that there are no major changes  
252 in zonal wind stress, i.e. Fig. 6a, which implies that the changes in the barotropic streamfunction  
253 are a direct consequence of the topography changes. As a result, the barotropic component of  
254 the volume transport over the GSR increases (from <5 Sv to ~8 Sv), while the transport by the  
255 MOC decreases (Fig. 4). Hence, there is a partial conversion of the flow from an overturning to a  
256 barotropic flow with the introduction of the GSR. Nonetheless, the net (barotropic+MOC) volume  
257 transport over the GSR decreases as the MOC weakens and so does the OHT.

258 To better understand the changes in the gyre circulation, horizontal velocities averaged over the  
259 top 100 m for the North Atlantic are shown in Fig. 6c and d. In both cases, the surface flow in the  
260 polar basin is dominated by a warm cyclonic boundary current entering at the eastern side of the  
261 basin. Due to the zonal barrier in *ridge*, however, a substantial part of the poleward flow is steered  
262 along the GSR (following constant  $f/H$  contours), seen as an enhanced westward barotropic flow  
263 (at the GSR latitudes). This is consistent with observations and models showing that a substantial  
264 part of the NAC is steered by the complex bottom topography and recirculates south of the GSR  
265 (Bower et al. 2019; Stärz et al. 2017).

266 *c. Hydrographic changes in the North Atlantic*

267 As suggested by the MOC response, the GSR also affects the distribution of water mass proper-  
268 ties (i.e. potential temperature and salinity) in the northern Atlantic basin. A cooling and freshen-  
269 ing ( $3 - 6^{\circ}\text{C}$  and  $0.1-0.7$  psu respectively) is simulated over most of the water column north of the  
270 GSR, as the shallow sill weakens the flow of warm, salty subtropical waters (Fig. 7). As a result,  
271 the polar basin becomes more stratified, notably because a strong polar halocline can develop.  
272 However, the polar basin remains too warm for sea ice to form. South of the GSR, the surface  
273 water becomes warmer and saltier, as a result of the changes in the barotropic circulation shown  
274 above (Fig. 6b). In addition, a small increase in temperature and salinity ( $\sim 1^{\circ}\text{C}$  and  $\sim 0.1$  psu)  
275 can be seen at mid-depth in the subtropical Atlantic associated with changes in the properties of  
276 the NADW.

277 Because of the high-latitude cooling, the density increases dramatically north of the GSR, while  
278 it decreases slightly south of the GSR, resulting in an upward sloping of isopycnals from south to  
279 north (Fig. 7; bottom row). This is due to the fact that stratification is dominated by temperature  
280 in this warm state. In a colder climate, with a small thermal expansion coefficient, density might  
281 decrease from the freshening. The density structure reveals a modern-like bi-directional circulation  
282 regime with Atlantic water inflow in the surface-subsurface and a dense ( $\sigma_{\theta} > 27.6$ ) southward  
283 outflow above the sill. Here, the enhanced density contrast across the GSR help drive the overflow,  
284 which in turn, is balanced by the inflow (Hansen and Østerhus 2000). Hence, by introducing the  
285 GSR, the polar basin is transformed into a reservoir of dense and less ventilated deep water with  
286 longer residence time due to limited exchange with the North Atlantic.

287 *d. Ocean heat transport changes*

288 The changes in the North Atlantic ocean circulation are also reflected in the meridional OHT.  
289 Northward OHT decreases throughout the Atlantic basin (Fig. 5) when introducing the ridge.  
290 South of the GSR the reduction in OHT (by 0.08 PW (7%) at 20°N) is consistent with the small  
291 weakening (~5%) of the deep MOC (Fig. 4). The most prominent differences in OHT occur  
292 in the subpolar and high latitude regions, where the changes in the MOC are largest. However,  
293 despite the weak MOC in this region, the OHT across the GSR only decreases by 32% (from 0.22  
294 PW to 0.15 PW at 70°N), which implies a disconnect between changes in the MOC and high-  
295 latitude OHT. We note that there are no compensating effects by the atmospheric heat transport.  
296 Further north, the poleward OHT decays rapidly because less warm Atlantic water reaches the  
297 high latitudes.

298 To help understand the changes in OHT between the two states, we consider the scaling argu-  
299 ment by Ferrari and Ferreira (2011) which states that the heat transport associated with a closed  
300 circulation scales as  $\sim \rho_0 c_p \Psi \Delta_z \theta$ , where  $\Psi$  is the strength of the overturning circulation and  $\Delta_z \theta$   
301 the vertical temperature gradient it encounters.  $\rho_0$  is a reference density and  $c_p$  is the heat capacity  
302 of sea water. In the absence of a zonal barrier (*noridge*), the volume transport across 70°N is high  
303 ( $\Psi = 16$  Sv; Table 1), and the MOC spans the entire water column with a temperature gradient of  
304 about 4°C. The scaling yields a poleward OHT of roughly 0.26 PW, slightly more than the actual  
305 OHT in Fig. 5. In this case, the poleward heat transport is dominated by the deep MOC. For *ridge*,  
306  $\Psi$  is only 5.7 Sv (over the GSR), but is compensated by a larger temperature contrast of almost  
307 8°C between the surface inflow and the deep outflow over the sill (Fig. 7). Thus the scaling gives  
308 an OHT of 0.18 PW close to the simulated 0.15 PW. The larger temperature contrast is a result of  
309 the stronger gyre circulation (Fig. 6b) providing warmer water at the surface, while the outflow is

310 colder thereby facilitating a more efficient heat transport. Hence, the presence of the GSR com-  
311 pletely changes the dynamics of the North Atlantic-Arctic circulation; as the deep MOC vanishes  
312 poleward of 50°N, the heat transport across the GSR is dominated by a shallow circulation (driven  
313 by a combination of surface winds and buoyancy forcing). This is consistent with earlier model  
314 studies suggesting that the SPG accounts for most of the OHT across the GSR in the modern day  
315 climate (Spall 2001; Born et al. 2009; Ferrari and Ferreira 2011; Li and Born 2019).

316 *e. High-latitude and global surface climate response*

317 The GSR has a large impact on the surface climate conditions in the North Atlantic region  
318 through the aforementioned ocean circulation changes (Fig. 3; right and 7). The high latitude  
319 surface ocean cools and freshens by 2 – 4°C and 1-2 psu respectively, with the largest changes  
320 occurring on the western side of the polar basin. The average SST in the polar basin decreases to  
321 10.3°C (Table 1), and the Atlantic equator-to-pole SST gradient increases to 22.8°C (compared  
322 to 19.0°C in *noridge*). This is consistent with proxy data and previous model simulations (e.g.,  
323 Robinson et al. 2011; Stärz et al. 2017), suggesting that shallow sill-depths produce lower Arctic  
324 surface temperatures and greater North Atlantic temperature gradients. Meanwhile, the polar basin  
325 remains too warm for sea ice to form. The effect on global surface climate, however, is small, as  
326 illustrated by Fig. 8, showing the global surface air temperature (SAT) difference between *ridge*  
327 and *noridge*. While the presence of the GSR has a strong local effect at northern high latitudes,  
328 the surface climate outside the polar basin does not change much. We note, however, a small  
329 warming over the Southern Ocean and reduced sea-ice thickness (Fig. 3b), which is likely due to  
330 the weakening of the AMOC and northward OHT (Fig. 5). Ultimately, the simulated changes in  
331 SAT are linked to the spatial pattern of the OHT changes (Fig. 5). At low-latitudes the uniform  
332 shift in the transport profile results in the same heat flux convergence and hence same flux to the

333 atmosphere. Consequently, there is no change in surface air temperature. In contrast, the surface  
334 climate changes at subpolar and high latitudes are attributed to a southward shift in the location of  
335 OHT convergence. The fact that surface climate response is confined to the northern high-latitudes  
336 is conflicting with earlier proxy-based studies which suggest that deepening of the GSR triggered  
337 global cooling at the EOT (e.g., Abelson et al. 2008; Abelson and Erez 2017).

338 In the present climate, it is widely accepted that variations in air-sea fluxes over the subpolar  
339 North Atlantic have a significant effect on AMOC strength (e.g., Lozier et al. 2017; Sévellec  
340 et al. 2017). To better understand the link between the simulated changes in high-latitude surface  
341 climate and the MOC, we compare the surface density fluxes (Fig. 10) to the mixed layer depth  
342 (Fig. 9) reflecting the location of deep convection. Overall, there is a net buoyancy loss (i.e.  
343 densification) over the North Atlantic dominated by ocean heat loss, while the freshwater fluxes,  
344 mainly associated with E-P and run-off along the boundaries (in the absence of sea ice), contributes  
345 only with a small buoyancy gain. In *noridge* deeper mixed layers are found in the northwestern  
346 part of the basin, where there is strong ocean heat loss. As the Atlantic inflow weakens in *ridge*,  
347 the surface water gets colder and the ocean-atmosphere heat flux is reduced poleward of 70°N  
348 (Fig. 10). At the same time, the imprint of the freshwater forcing, by precipitation and run-off, is  
349 more pronounced, although its effect on surface density is small compared to the changes in heat  
350 flux. The reduced buoyancy loss weakens deep convection in the polar basin, also illustrated by the  
351 shallower mixed layers in Fig. 9. This results in poorly ventilated deep waters below the sill-depth.  
352 Meanwhile, deep convection is enhanced south of the GSR as indicated by the increased buoyancy  
353 loss between 60°N and 70°N, which is uniquely attributed to warmer SSTs along the GSR that  
354 favours stronger ocean-atmosphere heat fluxes and deeper mixed layers. Hence, when the GSR  
355 is present, deep convection occurs both in the polar basin (albeit weaker) and the subpolar North  
356 Atlantic. A similar response was found in a recent modelling study, using late Eocene boundary



357 conditions (Hutchinson et al. 2019), demonstrating that deep water formation shifted to the south  
 358 of the GSR in response to a shoaling of the sill from 500 m to 25 m. They also suggest that  
 359 the Arctic-Atlantic freshwater transport may be critical in determining the location of deep water  
 360 formation, including the preferred basin of sinking (i.e. Atlantic versus Pacific sinking).

361 *f. Water mass transformation estimates*

362 To understand how variations in the density fluxes over the North Atlantic and polar regions  
 363 shown in Fig. 10 influence the structure and strength of the overturning circulation we estimate  
 364 the surface-forced water mass transformation (WMT) based on the approach by Walin (1982).  
 365 This approach is particularly useful because it provides a way to estimate the rate of deep water  
 366 formation based solely on the surface forcing conditions. Walin (1982) showed that the transfor-  
 367 mation of surface waters by density fluxes between two isopycnals  $\sigma$  and  $\sigma + \delta\sigma$  (an outcrop  
 368 region), is equivalent to a diapycnal volume flux across the outcropping isopycnal. Adopting the  
 369 notation from Brambilla et al. (2008) and Speer and Tziperman (1992) the WMT over a year can  
 370 be written as:

$$F(\sigma) = \frac{1}{\Delta T \Delta \sigma} \int_{year} dt \iint_{area} dA \delta(\sigma - \sigma') D(x, y, t) \quad (1)$$

371  $F(\sigma)$  is the annual mean water mass transformation function equal to the diapycnal volume flux  
 372 ( $\text{m}^3 \text{s}^{-1}$ ) over the isopycnal outcrop region with the area  $dA$  and  $D(x, y, t)$  is the surface density  
 373 flux sampled at the surface density  $\sigma$  by the delta function  $\delta$ . Negative/positive values indicate  
 374 that waters become lighter/denser. The total surface density flux ( $D$ ;  $\text{kg m}^{-2} \text{s}^{-1}$ ) consists of a  
 375 thermal ( $D_{HF}$ ) and a haline component ( $D_{FW}$ ):

$$D(x, y, t) = D_{HF} + D_{FW} = -\frac{\alpha Q_{HF}}{c_p} + \beta S Q_{FW} \quad (2)$$

376 where  $c_p$  ( $\text{J kg}^{-1} \text{K}^{-1}$ ) is the specific heat of sea water,  $\alpha$  and  $\beta$  are the thermal expansion and  
 377 haline contraction coefficients respectively (calculated from UNESCO formulas; see McDougall  
 378 1987),  $Q_{HF}$  ( $\text{W m}^{-2}$ ) the net heat flux (positive for ocean heat gain), and  $Q_{FW}$  ( $\text{kg m}^{-2} \text{s}^{-1}$ ) the  
 379 net freshwater flux into the ocean associated with evaporation-precipitation, run-off and melt-  
 380 ing/freezing of sea ice.  $S$  is the sea surface salinity and all variables are functions of space and  
 381 time.

382 From the surface forced WMT function it is possible to estimate the water mass *formation*,  
 383 which is defined as the divergence of the WMT in Eq. (1):

$$Md\sigma = -[F(\sigma + d\sigma) - F(\sigma)] \quad (3)$$

384 corresponding to the water that accumulates or is lost over a year between two isopycnals  $\sigma + d\sigma$   
 385 and  $\sigma$ . The formation per unit density becomes:  $M(\sigma) = dF/d\sigma$ .

386 In practice we obtain the annual-mean surface forced WMT by calculating the surface density  
 387 flux in each grid cell for the interval  $28.4 \leq \sigma_2 \leq 36.4 \text{ kg m}^{-3}$  with  $\Delta\sigma = 0.1$  and a time-step of  
 388  $\Delta t = 5$  days for a total of  $\Delta T = 10$  years taken from the end of the integration. The WMT function  
 389 is then integrated over the Atlantic basin from  $40^\circ - 90^\circ \text{N}$ .

390 The surface forced basin-integrated WMT (Eq. 1) in the entire North Atlantic region ( $40^\circ -$   
 391  $90^\circ \text{N}$ ) is shown in Fig. 11 (upper panel) for the two experiments. The transformation rates for  
 392 *noridge* and *ridge* shows the same general features and compare well with previous studies (e.g.,  
 393 Speer and Tziperman 1992; Brambilla et al. 2008). Negative WMT (associated with a *buoyancy*  
 394 *gain*) is found at low densities in the interval  $31.0 \leq \sigma_2 \leq 33.5$  (i.e. subtropical water), while posi-

395 tive WMT (associated with *buoyancy loss* from surface cooling) occurs at higher surface densities  
396 in the range  $33.6 \leq \sigma_2 \leq 36.4$  (i.e. subpolar water). In *noridge*, however, there is no transformation  
397 for densities greater than  $\sigma_2 > 36.1$ . Both cases have comparable peak WMT (27.7 Sv and 27.9 Sv  
398 crossing the 35.3 and 35.2 isopycnal for *noridge* and *ridge* respectively). This is the magnitude  
399 of convection in the Atlantic basin due to surface exchanges, and implies that the production of  
400 dense water is virtually constant regardless of the GSR. We note that these values are comparable  
401 with the transport in the upper branch of the MOC in Fig. 4 (see also Brambilla et al. 2008).

402 The spatial distribution of the WMT, however, is quite different in the two configurations. To  
403 illustrate this in more detail, the total WMT in the North Atlantic is separated into contributions  
404 from the subpolar and polar regions (Fig. 11; middle and lower panel). This reveals a much  
405 weaker WMT in the polar basin for *ridge* (7 Sv versus 15.2 Sv in *noridge*), occurring at higher  
406 surface densities because the surface waters are colder. It is partly compensated by larger WMT  
407 in the subpolar North Atlantic, with a maximum of 26.4 Sv at the 35.2 isopycnal. This reflects an  
408 increase in deep water formation south of the GSR, consistent with the enhanced surface density  
409 fluxes and deeper mixed layers (Fig. 9 and 10). Hence, for *ridge* most of the dense water is  
410 formed in the subpolar North Atlantic between  $40^\circ$  and  $70^\circ$ N (although  $\sim 10\%$  of the total WMT  
411 still occurs north of the GSR; i.e., 2.8 kg/s of the net density input, 28 kg/s, occurs north of  $70^\circ$ N).  
412 In summary, the surface forced WMT is in good agreement with the overturning transport (i.e.  
413 Fig. 4), which both show that the GSR significantly impact the location of deep water formation,  
414 but does not affect the total deep water production over the Atlantic basin. As a consequence, the  
415 maximum transport by the MOC shows little change.

416 *g. Water mass formation*

417 The surface forced water mass *formation* can be estimated from the divergence of the WMT  
418 (Fig. 11), where a negative (positive) slope is associated with water mass formation (destruction)  
419 (see Eq. 3). We compare the implied isopycnal transport from the surface fluxes to the overturning  
420 streamfunction in density coordinates (i.e.  $\text{MOC}_\sigma$ ) in Fig. 12.

421 The overturning circulation in latitude-density space depicts the gradual densification of surface  
422 waters as they move northwards. The densification is consistent with a positive WMT (i.e. buoy-  
423 ancy loss) in the subpolar and polar region for densities greater than 33.5 and is similar in strength  
424 in both cases. Here the negative slope ( $dF/d\rho < 0$ ) for  $\sigma_2 > 35.3$  in Fig. 11 gives a formation  
425 of about 27.6 Sv for *noridge* and 23 Sv in *ridge* and is balanced by a destruction of interme-  
426 diate water masses in the range  $32.6 \leq \sigma_2 \leq 35.2$  where  $dF/d\rho > 0$ . These values are in good  
427 agreement with the simulated volume transport by  $\text{MOC}_\sigma$ , which shows a maximum strength  
428 of 25.5 Sv for *noridge* and 23.3 for *ridge*, slightly higher than the overturning in latitude-depth  
429 space (i.e.  $\text{MOC}_z$ ). In addition, the overturning maximum is shifted toward the subpolar region  
430 ( $\sim 50^\circ\text{N}$ ), reflecting the transport by the horizontal gyre circulation, which tends to average to zero  
431 in latitude-depth space and therefore has little imprint on  $\text{MOC}_z$ .

432 For *noridge*  $\text{MOC}_\sigma$  reveals a relatively low density ( $35.6 \leq \sigma_2 \leq 35.8$ ) outflow from the polar  
433 basin, indicating that it is formed by local conversion of lighter water originating from the subpolar  
434 region. Meanwhile, the water mass characteristics of the southward flow changes very little with  
435 latitude, which reflects an interior circulation that is largely adiabatic. In *ridge*, on the other  
436 hand, the southward transport over the GSR at  $60^\circ\text{N}$  consist of about 6 Sv of intermediate density  
437 waters being recirculated in the SPG and a very dense overflow ( $\sigma_2 > 36.0$ ) of  $\sim 3$  Sv formed by  
438 strong surface cooling in the polar basin. At  $50^\circ\text{N}$ , however, there is no transport of these extreme

439 waters suggesting that the dense overflow mixes downstream of the GSR to form waters of lighter  
440 densities. As a result, there is a sharp peak (18 Sv) in the southward flux across 50°N with density  
441  $\sigma_2=35.7$ .

442 We note that the estimated WMT by Walin (1982) only consider transformation due to air-  
443 sea fluxes, but neglects the WMT associated with mixing processes in the ocean interior and  
444 therefore we do not expect a perfect fit between the WMT estimates and  $\text{MOC}_\sigma$ . Vertical mixing is  
445 particularly important for the transformation of dense-water as it flows over the GSR (Hansen and  
446 Østerhus 2000), where it roughly doubles in volume due to turbulent entrainment with ambient  
447 waters (Beaird et al. 2013). As a consequence, it is estimated that the overflows supply about one-  
448 third of the total volume transport of the AMOC (Hansen et al. 2004). Using a similar entrainment  
449 rate implies that the dense overflow from the polar basin (see insert in Fig. 12b), would contribute  
450 with about 6 Sv to the total overturning transport. Still, this is small compared to the air-sea  
451 transformation in the subpolar North Atlantic (40° – 70°N) with an average rate of about 15 Sv,  
452 which is the main source of deep water feeding into the MOC (Fig. 11).

#### 453 *h. Sensitivity to the background climate state*

454 To test the sensitivity of the circulation response to the background climate state, we perform an  
455 additional experiment where the GSR is introduced in a colder climate. The main goal is to test if  
456 the model responds differently under cold climate conditions (when sea ice is present), where the  
457 climate system might be more sensitive to perturbations (e.g., Bitz et al. 2007). This is achieved  
458 by reducing the solar constant in *noridge* by  $6 \text{ W m}^{-2}$ , from the default  $S_0 = 1366 \text{ W m}^{-2}$  (*warm*  
459 *state*) to  $S_0 = 1360 \text{ W m}^{-2}$  (*cold state*), mimicking a reduction in atmospheric  $\text{pCO}_2$  (equivalent  
460 to slightly less than a factor of 2 reduction).

461 By lowering the solar constant the climate cools and the model (i.e. *cold noridge*) reaches  
462 a new steady state with a global mean SST of 18.2°C (a reduction of 2.5°C). The cooling is  
463 amplified at the poles, causing sea ice to form in the Northern Hemisphere (although mainly in the  
464 Pacific basin). The MOC in the Atlantic basin is shallower and slightly weaker (20.1 Sv at 26°N)  
465 compared to the warm case, and a weak AABW cell emerges.

466 When the GSR is introduced in this colder climate state (i.e. *cold ridge*), the ocean circulation  
467 response is virtually identical to the warm case (Fig. 13): the main downwelling region shifts  
468 south of the GSR, the volume transport by the MOC across the GSR is greatly reduced (~65%)  
469 and the poleward OHT weakens by 29%. Meanwhile, the overall AMOC strength is not affected,  
470 which emphasizes the small impact of the GSR on global overturning circulation strength.

471 The surface climate response is also similar, although the magnitude of the SST and SSS changes  
472 in the Atlantic basin are slightly stronger for the cold case (i.e. the polar basin cools by up to 6°C  
473 compared to 4°C in the warm case). We attribute this to the appearance of a small (seasonal) sea  
474 ice cover in the polar basin which amplifies the cooling and freshening of the high latitude surface  
475 ocean. However, the changes are too small to have an impact on deep water formation, and thus  
476 MOC strength and global climate.

## 477 **4. Discussion**

### 478 *a. The relationship between GSR height and AMOC strength*

479 Earlier proxy-based studies (e.g., Wright and Miller 1996; Davies et al. 2001; Via and Thomas  
480 2006), suggest that changes to the GSR during the Cenozoic had a major impact on the Atlantic  
481 overturning circulation and global climate. For example, Abelson et al. (2008) suggested that the  
482 global cooling during the EOT, culminating in the glaciation of Antarctica, is linked to the deep-

483 ening of the GSR thereby triggering the onset of the AMOC. According to this hypothesis, the  
484 AMOC strength is directly correlated to the height of the GSR: deepening of the sill leads to a  
485 stronger overflow and higher NADW fluxes; while a shallower sill limits deep water exchange and  
486 prevents North Atlantic sinking. Similarly, our results does show a connection between variations  
487 in sill-depth and deep water production in the "Nordic Seas": the average production rate of dense  
488 water north of the GSR decreases from  $\sim 7$  Sv to  $\sim 3$  Sv between *noridge* and *ridge*. This leads to  
489 a weaker (but denser) overflow across the GSR, consistent with other paleo-climate model simu-  
490 lations (e.g., Robinson et al. 2011; Hill 2015). However, the total amount of deep water formed is  
491 unchanged as a result of changes south of the GSR (Fig. 11), and the AMOC at  $26^\circ\text{N}$  decreases  
492 by less than 2 Sv. This implies that variations in the strength of the Nordic Seas overflow does  
493 not necessarily translate into large changes in the AMOC, as suggested by modern observational  
494 studies (e.g., Olsen et al. 2008; Tett et al. 2014; Moffa-Sanchez et al. 2015).

495 On the other hand, the AMOC might be more strongly affected if the GSR is significantly shal-  
496 lower, by restricting the North Atlantic-Arctic water exchange. To test the effect of a shallower  
497 GSR on the AMOC response we performed an experiment comparing the *ridge* configuration  
498 (with a sill-depth of 500 m) to a shallow GSR of  $\sim 100\text{m}$ . This experiment was run for 200 years.  
499 With a shallower GSR we found that the inflow of warm and salty Atlantic water to the polar  
500 basin is drastically reduced, and thus prevents deep water from forming in the polar basin (not  
501 shown). Consequently, the maximum AMOC at  $26^\circ\text{N}$  decreases from 22.1 Sv in *ridge* to 18.5  
502 Sv in *ridge100*. This weakening of the AMOC (by 3.6 Sv) is consistent with an absence of dense  
503 overflow water formed in the polar basin (roughly 3 Sv in *ridge*). Nevertheless, despite there being  
504 virtually no overflow, the AMOC remains relatively strong as a result of active deep water forma-  
505 tion south of the GSR. These results support our conclusion that GSR sill-depth changes are not

506 likely the primary driver of AMOC variations and changes in global climate during the Cenozoic  
507 (see also Hutchinson et al. 2019).

508 *b. The impact of dense overflows*

509 One explanation for the weak MOC response to changes in the sill-depth is that our model does  
510 not realistically simulate the GSR overflow. In reality, the dense-water overflow occurs in nar-  
511 row and deep channels, through the Denmark Strait ( $\sim 840$  m) and Faroe Bank Channel ( $\sim 630$   
512 m), which are poorly represented in most present-day ocean models (Heuzé and Årthun 2019).  
513 Using a coarse resolution ocean model, Roberts and Wood (1997) showed that even small topo-  
514 graphic changes in these deep channels can have a large impact on overflow transports and AMOC  
515 strength. Therefore, we might expect that the absence of such channels in our idealized setup, re-  
516 sults in too weak overflow transports. On the other hand, the overflow response to topographic  
517 changes can be sensitive to the model's vertical resolution and coordinate system (Griffies et al.  
518 2000; Riemenschneider and Legg 2007). Hence, the larger number of vertical levels in our model  
519 (30 levels compared to 20 in Roberts and Wood 1997) allows for a substantial overflow transport,  
520 evident in Fig. 12, despite the lack of deep overflow channels. We note, however, that the sim-  
521 ulated overflow of 3 Sv when the GSR is present, is only about half of the observed estimates  
522 (Hansen and Østerhus 2000). This is probably a combination of the idealized representation of the  
523 GSR bathymetry and low horizontal resolution. Admittedly, at coarse resolution ( $2.8^\circ$ ) we may  
524 not fully resolve the narrow boundary currents at high latitudes, which could affect the poleward  
525 penetration of Atlantic water across the GSR (Heuzé and Årthun 2019; Spence et al. 2008). As a  
526 consequence, we may underestimate the amount of deep water formed north of the GSR.

527 From Fig. 12 we see that no dense water mass is found south of the GSR, and is most likely  
528 due to strong vertical mixing at the GSR as our model lacks an implicit overflow parameterization.



529 This is a common issue in coarse resolution models, causing excessive convective entrainment and  
530 deep waters that are too light (Willebrand et al. 2001; Ezer and Mellor 2004; Danabasoglu et al.  
531 2010; Yeager and Danabasoglu 2012). Hence, the representation of overflow water may have a sig-  
532 nificant impact on the MOC volume transport. By implementing an overflow parameterization in a  
533 fully-coupled climate model, Yeager and Danabasoglu (2012) found that the AMOC increases by  
534 4-6 Sv in the subpolar North Atlantic between  $40^{\circ} - 60^{\circ}\text{N}$ , while the maximum AMOC transport  
535 at  $37^{\circ}\text{N}$  decreases. Meanwhile, other studies using a parameterization based on hydraulic con-  
536 straints (e.g., Legutke and Maier-Reimer 2002; Kösters et al. 2005; Born et al. 2009) only show  
537 a small (less than 2 Sv), effect of Nordic Seas overflows on North Atlantic overturning strength.  
538 Despite the weak response in overturning, Kösters et al. (2005) show that the GSR overflow has  
539 a large impact on North Atlantic climate, by increasing the northward ocean heat transport and  
540 warming the Nordic Seas. In our simulations as well, the overflows play a relatively small role in  
541 determining the strength of the Atlantic overturning circulation. Instead, we argue that the over-  
542 turning circulation in our model is more sensitive to changes in surface buoyancy fluxes over the  
543 North Atlantic region, setting the location and strength of NADW formation. This is further sup-  
544 ported by model simulations from Stärrz et al. (2017), who show that a substantial NADW flow can  
545 exist, even for extremely shallow sill-depths ( $<50$  m) with virtually no overflow. Alternatively,  
546 other mechanisms are likely to influence the strength of the AMOC (see review by Kuhlbrodt  
547 et al. 2007), for example: wind-driven upwelling of NADW in the Southern Ocean (Toggweiler  
548 and Samuels 1995; Marshall and Speer 2012). As such, the fact that surface winds and upwelling  
549 in the Southern Ocean remain constant in our model simulation (not shown), may contribute to the  
550 relatively stable AMOC.

551 *c. Implications for paleoclimate*

552 The prevailing explanation for past changes in North Atlantic and high-latitude climate involves  
553 changes in the AMOC and northward OHT ranging from relatively abrupt centennial-to-millennial  
554 time-scales (e.g., Broecker et al. 1985; Rahmstorf 2002; Clark et al. 2002; Lynch-Stieglitz 2016),  
555 to very slow multi-million year time-scales associated with bathymetry changes and plate-tectonics  
556 (e.g., Wright and Miller 1996; Davies et al. 2001; Via and Thomas 2006). Such changes in past  
557 climate are inferred from indirect measurements (i.e. proxies) that record a local climate signal,  
558 reflecting past changes in surface climate conditions, circulation strength or deep water properties,  
559 but is often interpreted as representing changes in large-scale ocean circulation. For example, it  
560 is generally assumed that the high SSTs found in the North Atlantic and Arctic ocean during the  
561 MPWP is evidence for a stronger AMOC compared to modern, driving a stronger OHT to the high  
562 latitudes (Dowsett et al. 1992; Raymo et al. 1996). In this view, the strength of the overturning  
563 circulation is inferred from the SST pattern recorded in the proxies, thus assuming that global  
564 AMOC strength is directly correlated with the poleward heat transport and surface temperature at  
565 high latitudes.

566 In contrast to this view, our simulations illustrate that large changes in high-latitude surface  
567 climate can occur without invoking changes in large-scale ocean circulation. A central element  
568 to this is the disconnect between the AMOC and high-latitude OHT, which has been highlighted  
569 by several studies (e.g., Spall 2001; Wunsch 2006; Ferrari and Ferreira 2011; Zhang et al. 2013;  
570 Nummelin et al. 2017; Li and Born 2019; Årthun et al. 2019), and questions the traditional view  
571 that the AMOC is the main driver of changes in high-latitude climate. This has implications for the  
572 interpretation of paleo-proxy records; while proxies could indicate a large change in high-latitude  
573 surface climate, water mass properties, or ventilation they do not necessarily imply a large change

574 in the AMOC on a global scale. For example, several models have shown that OHT across the  
575 GSR may actually increase despite an AMOC slowdown (e.g., Zhang et al. 2013; Årthun et al.  
576 2019), as a result of changes in the horizontal circulation. Similarly, our results emphasize the  
577 importance of the shallow overturning by the subpolar gyre in controlling high-latitude climate.  
578 In fact, the gyre maintains a substantial heat transport to the polar regions, despite a weakening of  
579 the heat transport by the deep MOC. Hence, an improved understanding of the processes affecting  
580 the horizontal gyre circulation is critical for understanding past climate variability at high northern  
581 latitudes (see also Li and Born 2019, for a discussion).

## 582 **5. Conclusion**

583 Using a coupled atmosphere-ocean-sea ice model with idealized Earth-like geometry, we explore  
584 the role of the Greenland-Scotland Ridge (GSR) in shaping the modern ocean circulation and its  
585 control on deep-water formation, ocean heat transport and high latitude surface climate.

586 When the GSR is absent, deep-water formation occurs near the North Pole in the Atlantic basin  
587 and a deep meridional overturning circulation (MOC) extends well into the high latitudes. This is  
588 associated with a strong northward ocean heat transport that warms the high latitudes and weakens  
589 the equator-to-pole SST gradient.

590 By introducing an idealized GSR, the main location of deep-water formation shifts southward  
591 causing the structure of the overturning circulation at subpolar and high latitudes to change dra-  
592 matically. The meridional volume transport by the MOC across the GSR decreases by more than  
593 64% and the poleward ocean heat transport is reduced by  $\sim 30\%$ . However, the overall strength of  
594 the MOC south of the GSR remains largely unaffected and the AMOC at  $26^\circ\text{N}$  decreases by only  
595 2 Sv (a 5% decrease). This relatively weak response is due to enhanced deep water production  
596 south of GSR, as a result of surface buoyancy changes. This suggests that AMOC strength may

597 be decoupled from the flow across the GSR. As the subpolar North Atlantic (south of the GSR)  
598 becomes the main region of deep water formation and the polar basin becomes less ventilated, the  
599 overall rate of water mass transformation in the Atlantic remains unchanged, while a modern-type  
600 bidirectional flow regime is established with warm inflow at the surface and a cold, dense overflow  
601 above the sill.

602 As a result, the water column north of the GSR cools and freshens, precipitation accumulates  
603 because of reduced exchange, while the subpolar region (south of the GSR) becomes warmer  
604 and saltier. The surface temperature and salinity in the polar basin decreases, which leads to a  
605 weakening of the equator-to-pole temperature gradient in the Atlantic basin. These results are  
606 consistent with paleo-proxies and previous modelling efforts that indicate that Arctic surface cli-  
607 mate is sensitive to changes in the height of the GSR (Dowsett et al. 2010; Robinson et al. 2011;  
608 Brigham-Grette et al. 2013; Stürz et al. 2017), and is likely due to reductions in the northward  
609 ocean heat transport.

610 The smaller sensitivity of the OHT across 70°N, compared to the MOC, shows that the horizontal  
611 circulation by the subpolar gyre maintains a significant transport of heat to the high latitudes. Our  
612 results emphasize that the shallow horizontal circulation is critical for controlling mid-to-high  
613 latitude OHT, as shown by previous studies (e.g., Ferrari and Ferreira 2011; Li and Born 2019).  
614 However, despite relatively large changes in ocean circulation, heat transport and surface climate  
615 at high-latitudes, the large-scale global climate shows no change. This result is independent of the  
616 background climate state in which the GSR is added. Contrary to previous suggestions (Wright and  
617 Miller 1996; Abelson et al. 2008), our results suggest that variations in GSR height alone cannot  
618 explain dramatic planetary climate change. At least, further amplification by internal feedbacks  
619 (e.g. CO<sub>2</sub>) needs to be invoked.

620 Our results have potential implications for the interpretation of paleo-proxies by highlighting  
621 a potential disconnect between the AMOC and high-latitude surface climate response, which is  
622 often inferred in paleoclimate studies (e.g., Dowsett et al. 1992; Raymo et al. 1996; Rahmstorf  
623 2002). In contrast, we show that large changes in water mass properties and surface climate at  
624 high latitudes can be simulated without large AMOC changes on hemispheric and global scales.  
625 Therefore we propose, that caution be taken when inferring a direct relationship between indirect  
626 and non-local measurements of overturning strength, high latitude SST/SSS changes recorded in  
627 proxies and large-scale climate.

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851 **LIST OF TABLES**

852 **Table 1.** Climate variables for *noridge* and *ridge* experiments. Mean SST (°C) and SSS  
853 (psu) in the polar basin averaged north of 70°N. AMOC<sub>max</sub> (Sv) is calculated  
854 as the maximum overturning streamfunction at 26°N below 500 m depth in the  
855 Atlantic basin.  $\Psi_{O,70N}$  and  $\Psi_{B,70N}$  is the maximum volume transport by the  
856 meridional overturning and barotropic streamfunction respectively across the  
857 GSR. OHT<sub>70N</sub> is the meridional northward OHT in petawatts (1 PW = 10<sup>15</sup>  
858 W) across 70°N in the Atlantic basin. . . . . 40



859 TABLE 1. Climate variables for *noridge* and *ridge* experiments. Mean SST ( $^{\circ}\text{C}$ ) and SSS (psu) in the polar  
 860 basin averaged north of  $70^{\circ}\text{N}$ .  $\text{AMOC}_{max}$  (Sv) is calculated as the maximum overturning streamfunction at  
 861  $26^{\circ}\text{N}$  below 500 m depth in the Atlantic basin.  $\Psi_{O,70N}$  and  $\Psi_{B,70N}$  is the maximum volume transport by the  
 862 meridional overturning and barotropic streamfunction respectively across the GSR.  $\text{OHT}_{70N}$  is the meridional  
 863 northward OHT in petawatts ( $1 \text{ PW} = 10^{15} \text{ W}$ ) across  $70^{\circ}\text{N}$  in the Atlantic basin.

Experiment	Global mean SST ( $^{\circ}\text{C}$ )	mean SST polar basin ( $^{\circ}\text{C}$ )	mean SSS polar basin (psu)	maximum AMOC (Sv)	$\Psi_{O,70N}$ (Sv)	$\Psi_{B,70N}$ (Sv)	$\text{OHT}_{70N}$ (PW)
noridge	20.7	11.5	34.55	23.2	16.0	<5	0.22
ridge	20.8	10.3	34.48	22.1	5.7	8.0	0.15

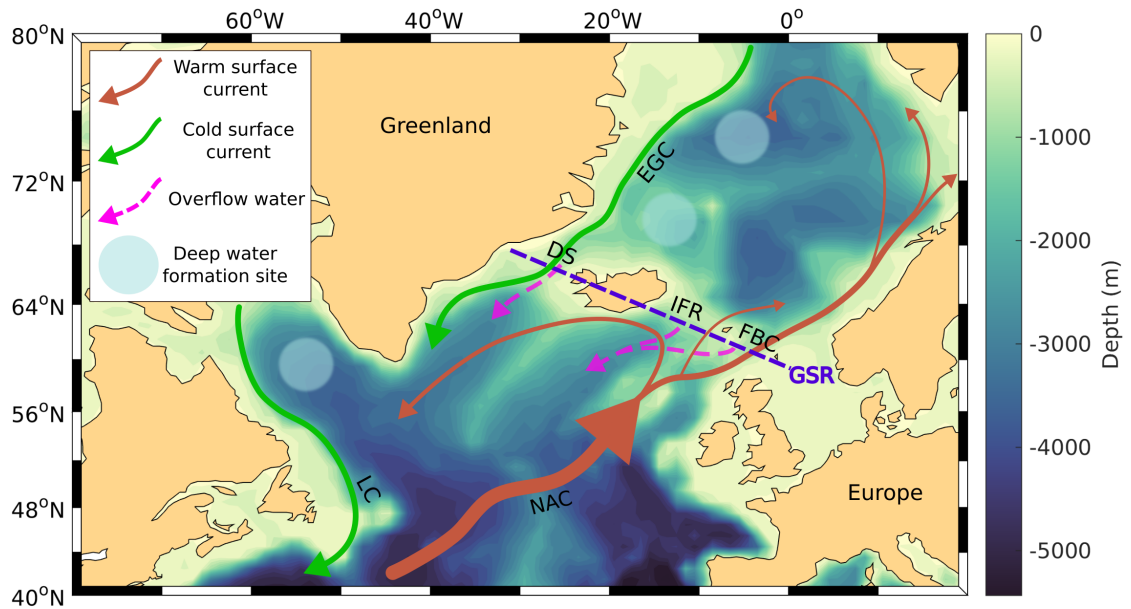
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865	<b>Fig. 1.</b>	Bathymetric map of the modern North Atlantic showing the main ocean currents; NAC = North Atlantic Current, EGC = East Greenland Current and LBC = Labrador Current. The dashed magenta line indicates the location of the Greenland-Scotland Ridge (GSR). Circles indicate the approximate location of deep water formation and dashed blue arrows shows the overflow through the Denmark Strait (DS), Faroe Bank Channel (FBC) and Iceland-Faroe Ridge (IFR). . . . . 43
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871	<b>Fig. 2.</b>	Land-sea distribution and bathymetry (m) for the <i>nordige</i> and <i>ridge</i> configuration. . . . . 44
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875	<b>Fig. 4.</b>	Residual meridional overturning circulation ( $MOC_z$ ) in Sverdrups ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) averaged over the Atlantic basin for <i>noridge</i> (left) and <i>ridge</i> (right). Contour lines are plotted at 2 Sv intervals, with solid (dashed) lines corresponding to clockwise (counterclockwise) circulation and the zero-contour indicated by the thick black line. . . . . 46
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879	<b>Fig. 5.</b>	Zonal mean northward ocean heat transport (in PW) over the Atlantic basin for the <i>noridge</i> (blue), <i>ridge</i> experiment (red) and the difference between the two (grey). Shading shows the interdecadal spread in OHT for each experiment, calculated as the difference between the maximum and minimum value over the last 100 years. The approximate location of the GSR is shown by the black dashed line. . . . . 47
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884	<b>Fig. 6.</b>	Wind stress and horizontal circulation in the Atlantic basin. (a) Zonal mean wind stress ( $\tau_x$ ; $\text{N m}^{-2}$ ) for <i>noridge</i> (dashed) and <i>ridge</i> (solid). The approximate position of the GSR is indicated by the black dashed lines. (b) Barotropic streamfunction (Sv) for <i>noridge</i> shown in contours with 5 Sv contour intervals where positive (negative) values corresponds to clockwise (counterclockwise) circulation. Shading shows the barotropic streamfunction anomaly calculated as the difference between <i>ridge</i> and <i>noridge</i> . (c and d) Mean top 100 m horizontal velocities in $\text{cm s}^{-1}$ for <i>noridge</i> and <i>ridge</i> respectively. . . . . 48
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891	<b>Fig. 7.</b>	Zonal mean potential temperature (°C; top row), salinity (psu; middle row) and potential density referenced to the surface ( $\sigma_0$ in $\text{kg m}^{-3}$ ; bottom row) in the North Atlantic for <i>noridge</i> (left), <i>ridge</i> (middle) and difference between the two, i.e. <i>ridge-noridge</i> (right). . . . . 49
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894	<b>Fig. 8.</b>	Global annual mean surface air temperature anomaly (SAT; °C) calculated as the difference between the final 100 years of <i>ridge</i> and <i>noridge</i> . Contour lines are plotted at 0.5°C intervals. . . . . 50
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896	<b>Fig. 9.</b>	Annual-mean mixed layer depth (MLD; m) in the northern Atlantic. Absolute values for <i>noridge</i> are shown in <i>left</i> panel and anomalies of <i>ridge</i> relative to the <i>noridge</i> in the <i>right</i> panel. . . . . 51
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899	<b>Fig. 10.</b>	Zonal mean surface density flux out of the ocean (in $10^{-6} \text{ kg m}^{-2} \text{ s}^{-1}$ ) in the North Atlantic for <i>noridge</i> (dashed lines) and <i>ridge</i> (solid lines). The total density flux (blue) is decomposed into contributions from heat (red) and freshwater (yellow) fluxes. Negative values indicate densification (i.e. buoyancy loss). . . . . 52
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903	<b>Fig. 11.</b>	Annual mean water mass transformation (WMT) function ( $F(\sigma)$ ) in Sv for <i>noridge</i> (dashed) and <i>ridge</i> (solid). The WMT is integrated over different regions; (top) the entire North Atlantic from 40° – 90°N, (middle) subpolar North Atlantic between 40° – 70°N and (bottom)
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906 polar basin north of 70°N. The surface forced WMT is estimated from the spatial integral of  
 907 the surface density flux ( $D$ ) over 10 years in the Atlantic basin spanning the density range  
 908  $\sigma_2 = 28.4 - 36.4 \text{ kg m}^{-3}$  with a density bin width of  $\Delta\sigma = 0.1$ . Here  $F(\sigma) \approx 0$  for  $\sigma < 30$ .  
 909 Negative values imply a WMT to lower density classes (i.e. *buoyancy gain*) and positive  
 910 values represents transformation from lower to greater densities (i.e. *buoyancy loss*). . . . . 53

911 **Fig. 12.** Meridional overturning circulation streamfunction in latitude-density space ( $\text{MOC}_\sigma$ ; Sv) for  
 912 (a) *noridge* and (b) *ridge*. The density bins are the same as used for the computation of the  
 913 WMT, with a reference density at 2 km depth. Contour lines are plotted at 2 Sv intervals.  
 914 The insert in (a) and (b) shows a zoom on the region around the GSR. (c and d) Mean  
 915 volume flux (Sv) by  $\text{MOC}_\sigma$  for *noridge* and *ridge* across different latitudes in the Atlantic  
 916 basin. Positive values indicate northward flow. . . . . 54

917 **Fig. 13.** Meridional overturning streamfunction anomaly (*ridge-noridge*) and ocean heat transport  
 918 (PW) in the Atlantic basin for the COLD case, where the solar constant is reduced to 1360  
 919  $\text{W m}^2$ . . . . . 55



920 FIG. 1. Bathymetric map of the modern North Atlantic showing the main ocean currents; NAC = North  
 921 Atlantic Current, EGC = East Greenland Current and LBC = Labrador Current. The dashed magenta line  
 922 indicates the location of the Greenland-Scotland Ridge (GSR). Circles indicate the approximate location of  
 923 deep water formation and dashed blue arrows shows the overflow through the Denmark Strait (DS), Faroe Bank  
 924 Channel (FBC) and Iceland-Faroe Ridge (IFR).

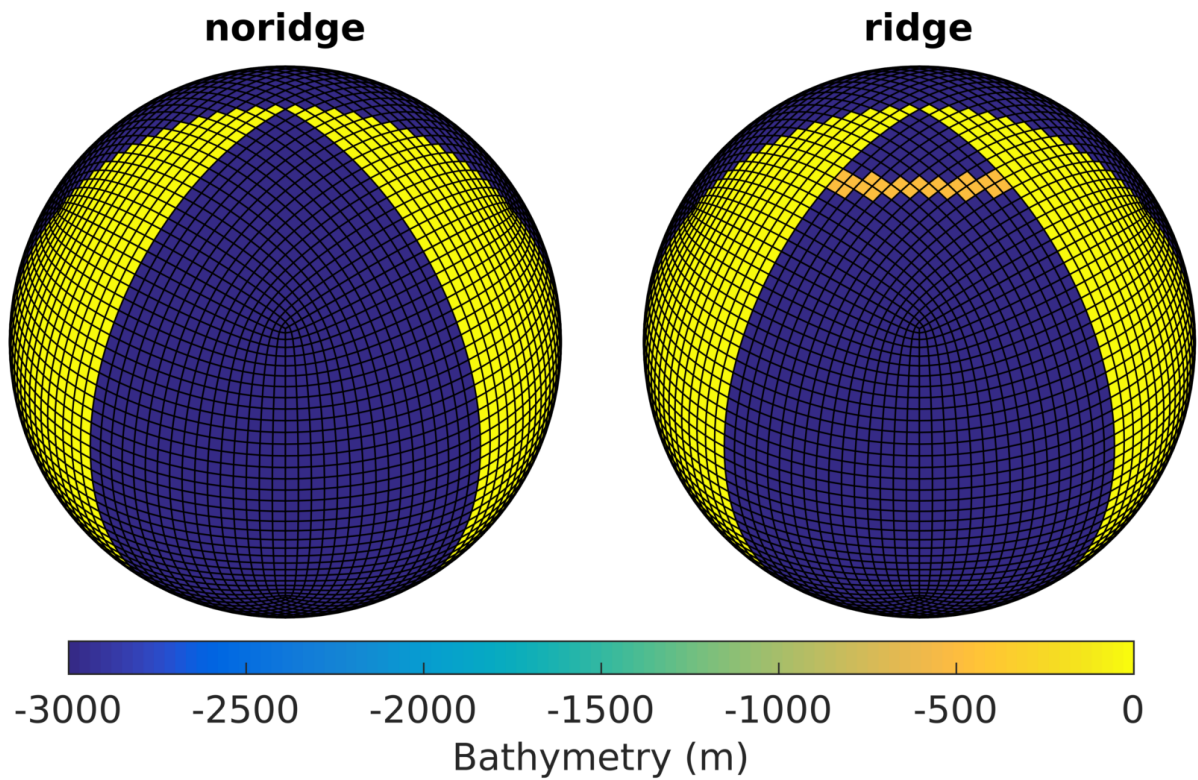
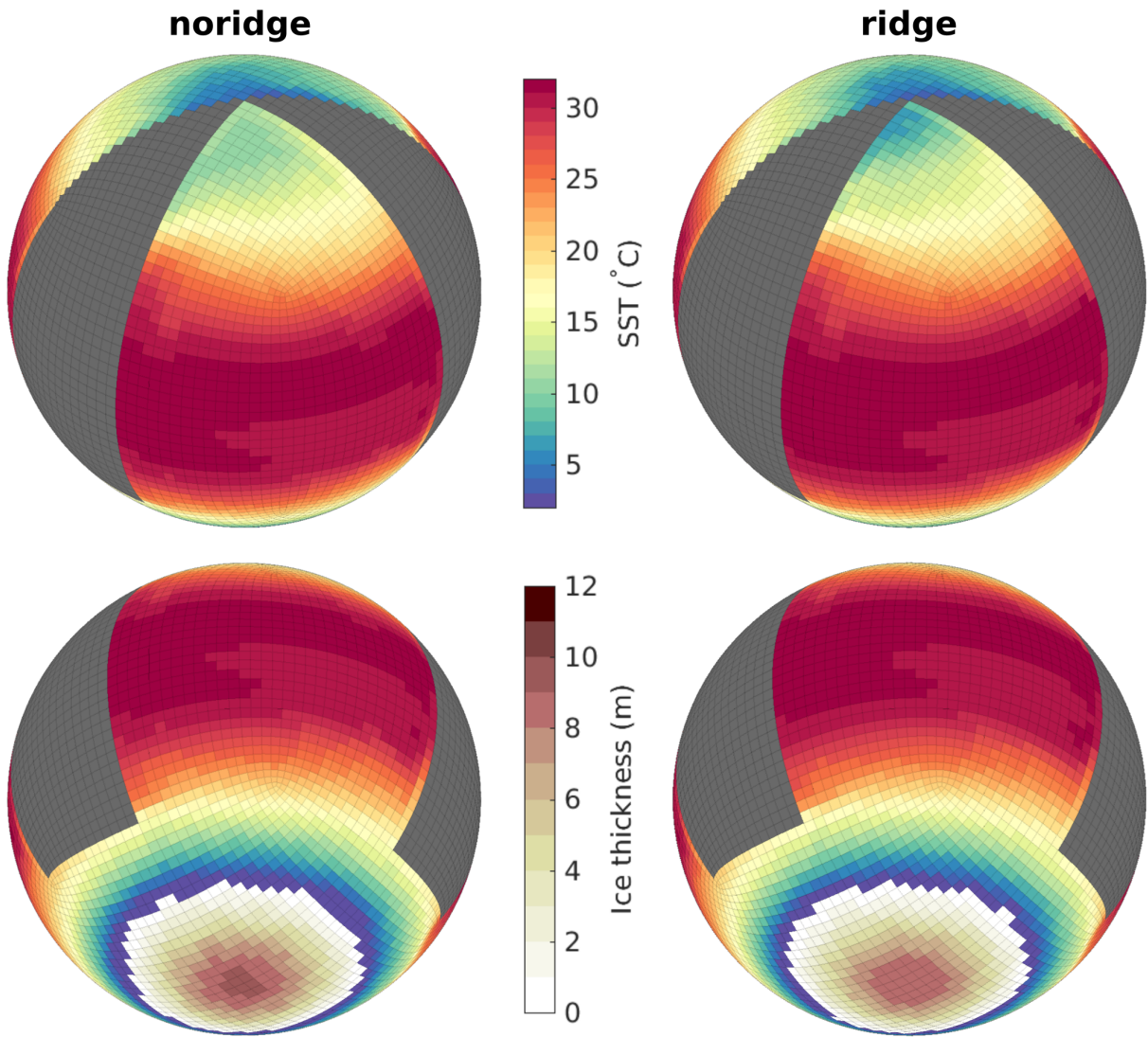
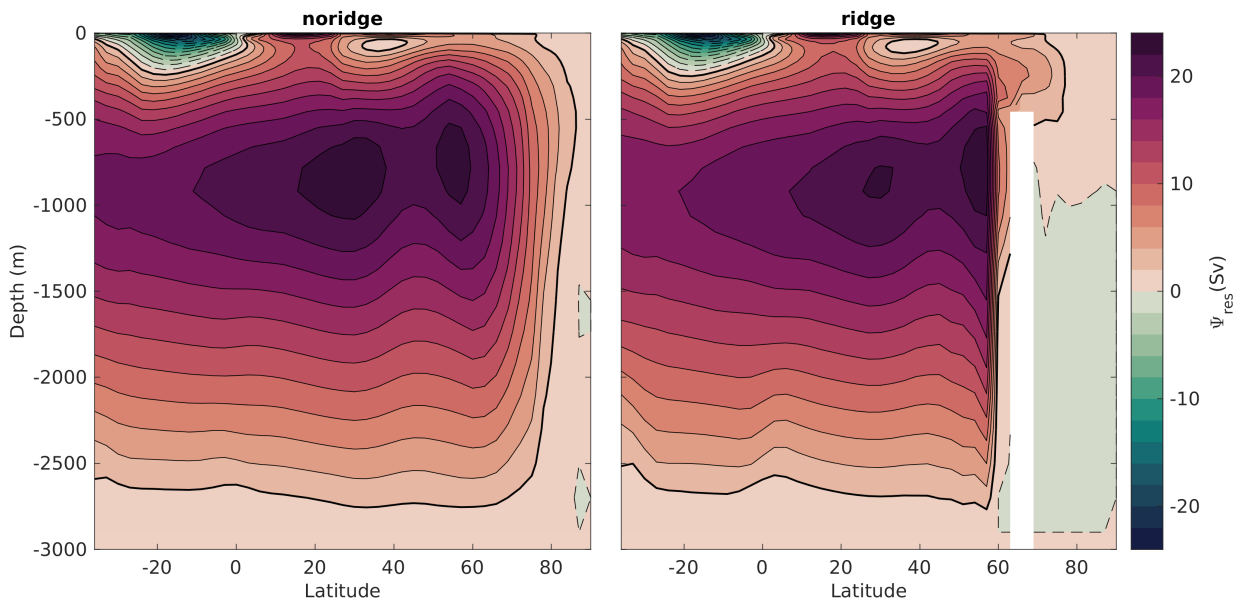


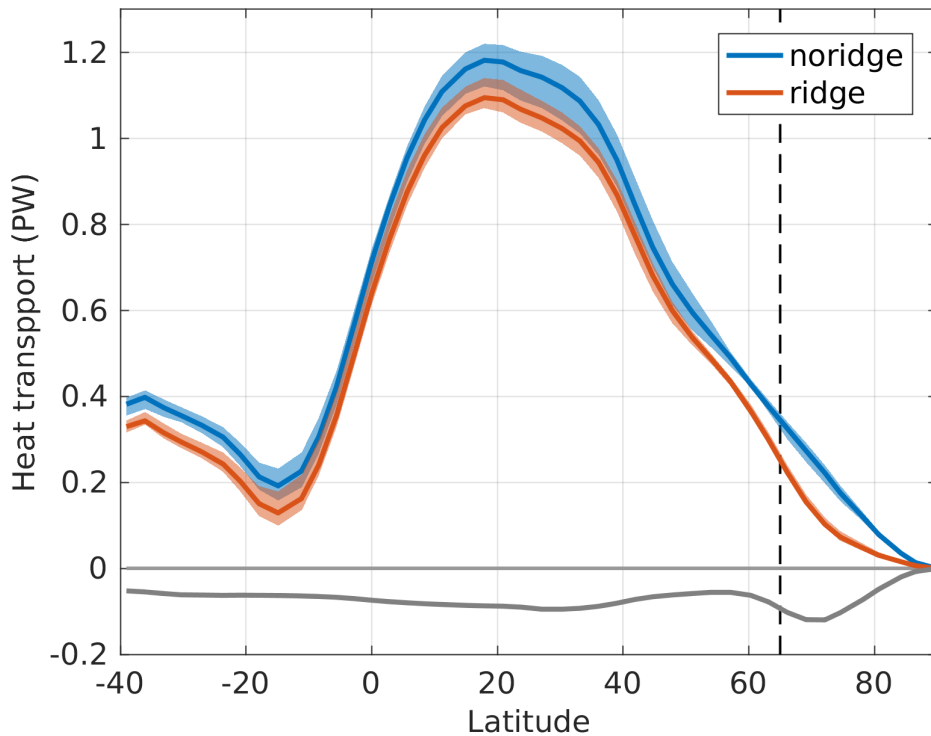
FIG. 2. Land-sea distribution and bathymetry (m) for the *noridge* and *ridge* configuration.



925 FIG. 3. Annual mean sea surface temperature (SST; °C) and sea ice thickness (in meters) over the small basin  
 926 for the *noridge* (left) and *ridge* (right) experiments. Values are based on the final 100 years at the end of the  
 927 integration.

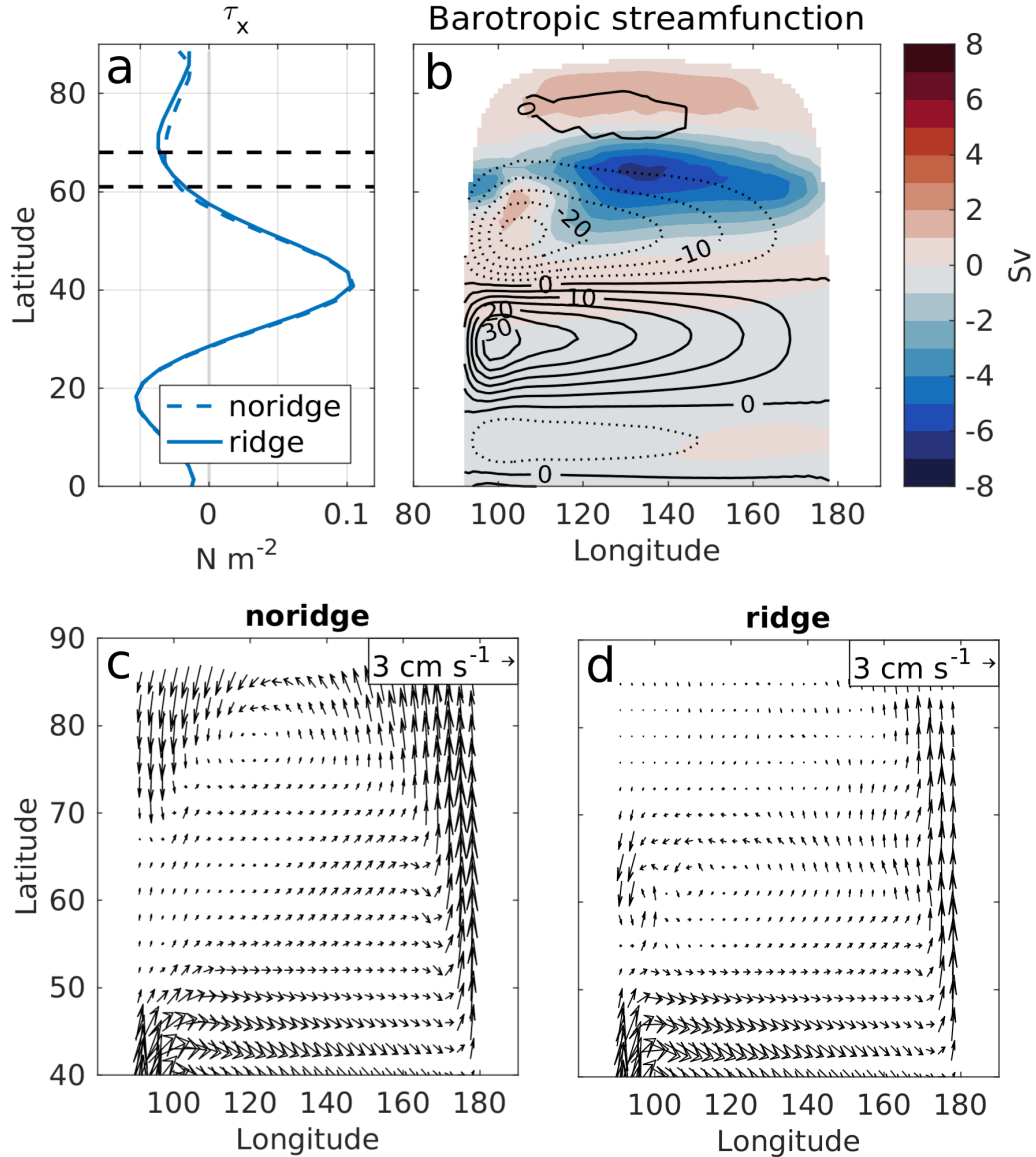


928 FIG. 4. Residual meridional overturning circulation ( $MOC_z$ ) in Sverdrups ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) averaged over  
 929 the Atlantic basin for *noridge* (left) and *ridge* (right). Contour lines are plotted at 2 Sv intervals, with solid  
 930 (dashed) lines corresponding to clockwise (counterclockwise) circulation and the zero-contour indicated by the  
 931 thick black line.

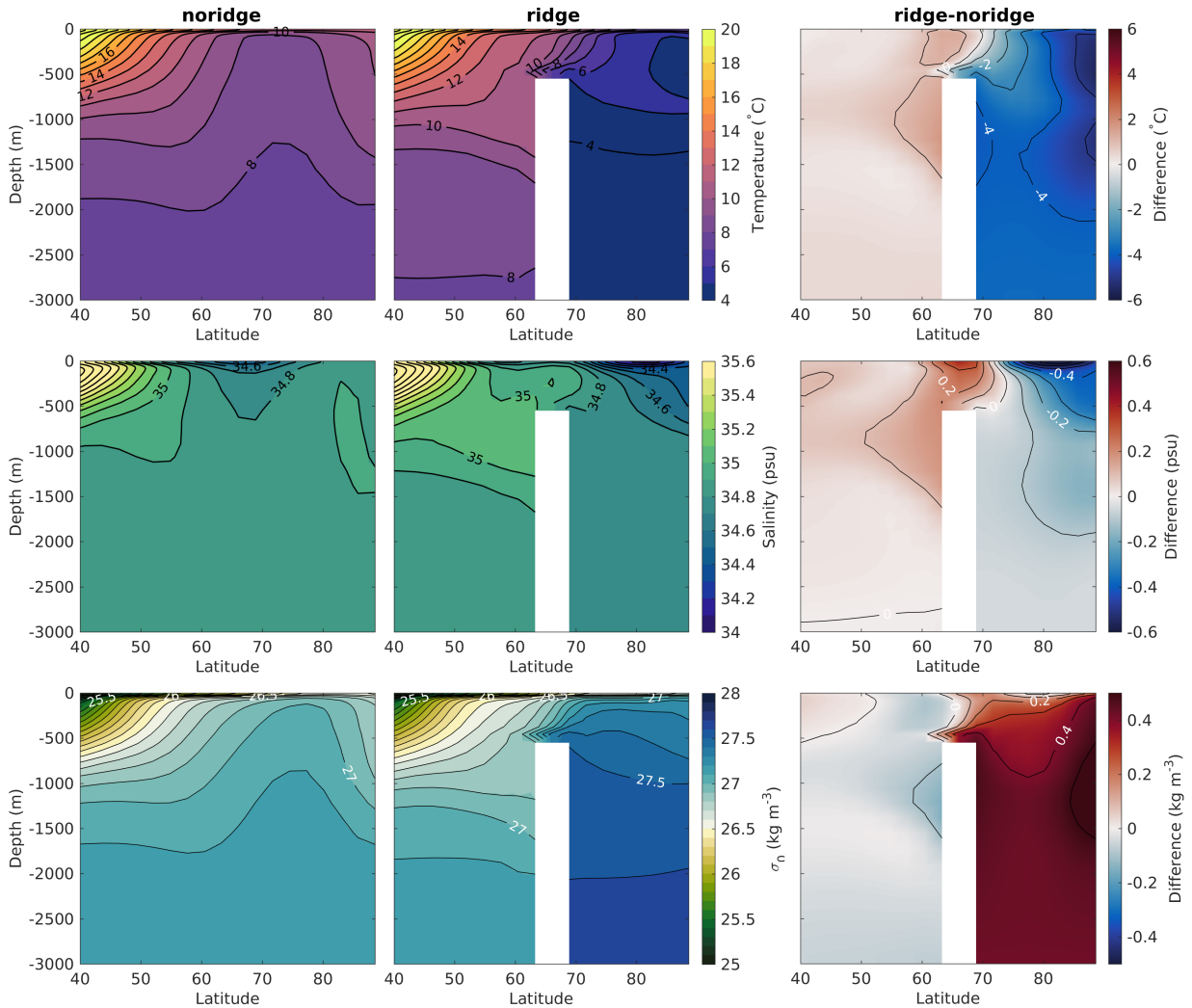


932 FIG. 5. Zonal mean northward ocean heat transport (in PW) over the Atlantic basin for the *noridge* (blue),  
 933 *ridge* experiment (red) and the difference between the two (grey). Shading shows the interdecadal spread in  
 934 OHT for each experiment, calculated as the difference between the maximum and minimum value over the last  
 935 100 years. The approximate location of the GSR is shown by the black dashed line.

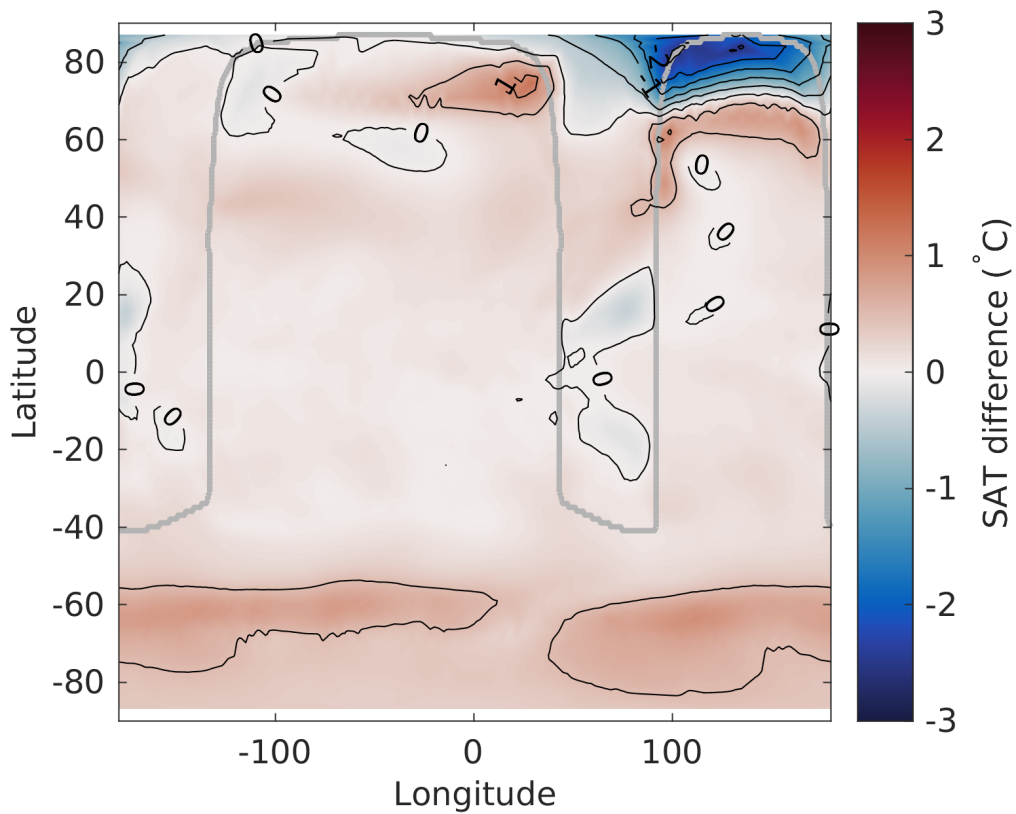




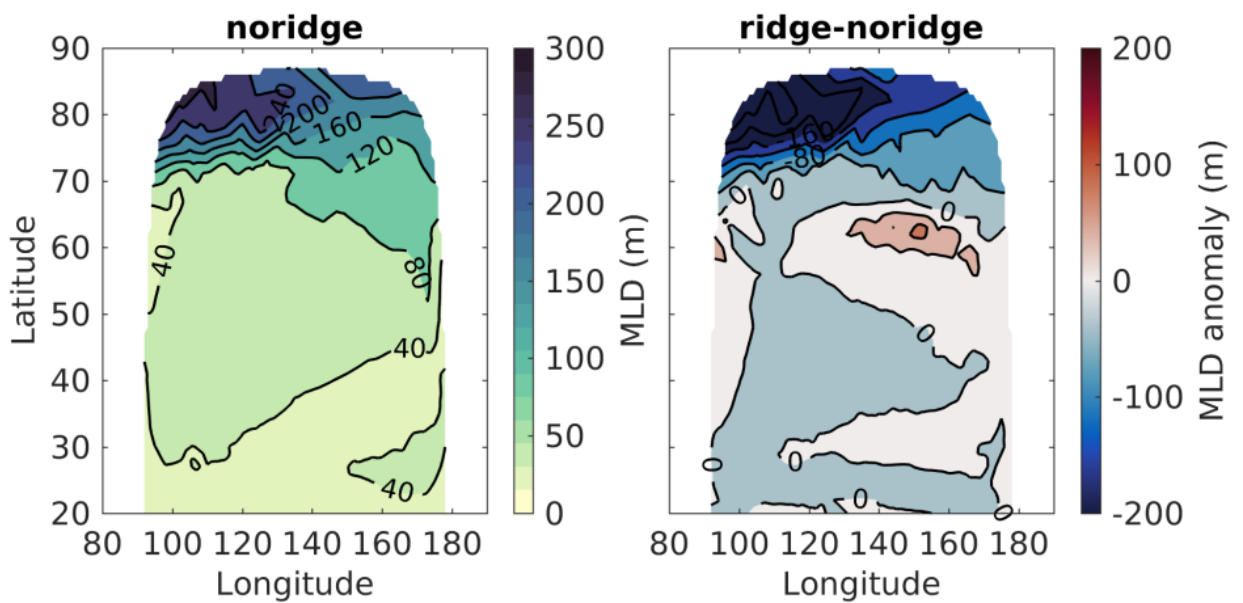
936 FIG. 6. Wind stress and horizontal circulation in the Atlantic basin. (a) Zonal mean wind stress ( $\tau_x$ ;  $\text{N}$   
 937  $\text{m}^{-2}$ ) for *noridge* (dashed) and *ridge* (solid). The approximate position of the GSR is indicated by the black  
 938 dashed lines. (b) Barotropic streamfunction (Sv) for *noridge* shown in contours with 5 Sv contour intervals  
 939 where positive (negative) values corresponds to clockwise (counterclockwise) circulation. Shading shows the  
 940 barotropic streamfunction anomaly calculated as the difference between *ridge* and *noridge*. (c and d) Mean top  
 941 100 m horizontal velocities in  $\text{cm s}^{-1}$  for *noridge* and *ridge* respectively.



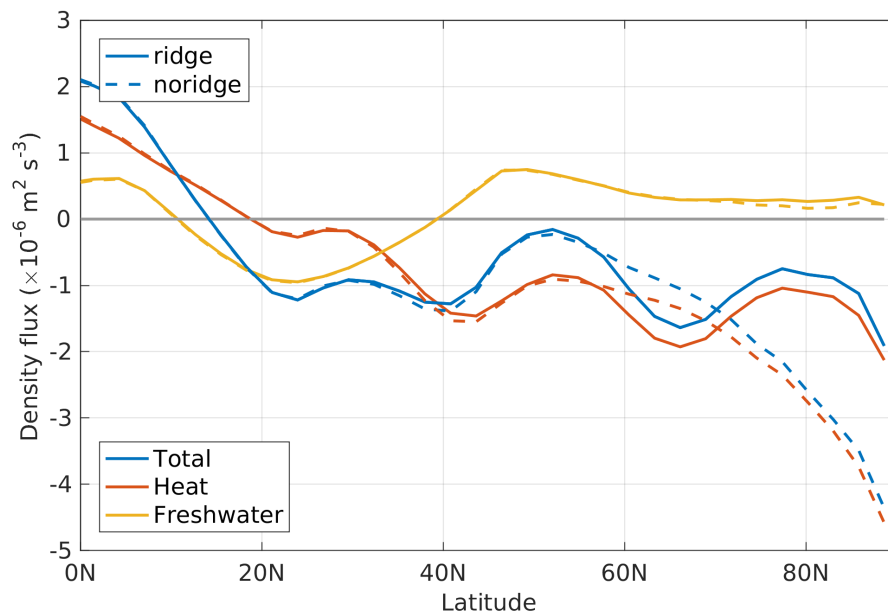
942 FIG. 7. Zonal mean potential temperature ( $^{\circ}\text{C}$ ; top row), salinity (psu; middle row) and potential density  
 943 referenced to the surface ( $\sigma_0$  in  $\text{kg m}^{-3}$ ; bottom row) in the North Atlantic for *noridge* (left), *ridge* (middle) and  
 944 difference between the two, i.e. *ridge-noridge* (right).



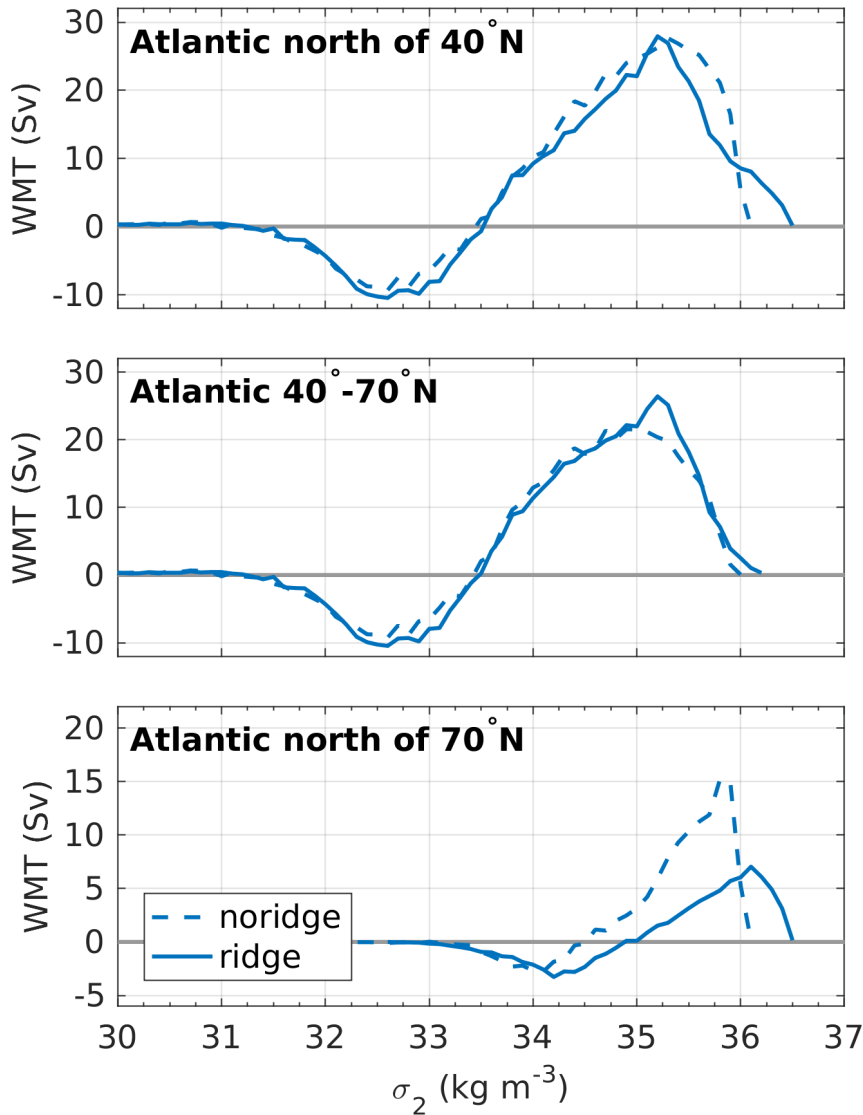
945 FIG. 8. Global annual mean surface air temperature anomaly (SAT; °C) calculated as the difference between  
 946 the final 100 years of *ridge* and *noridge*. Contour lines are plotted at 0.5°C intervals.



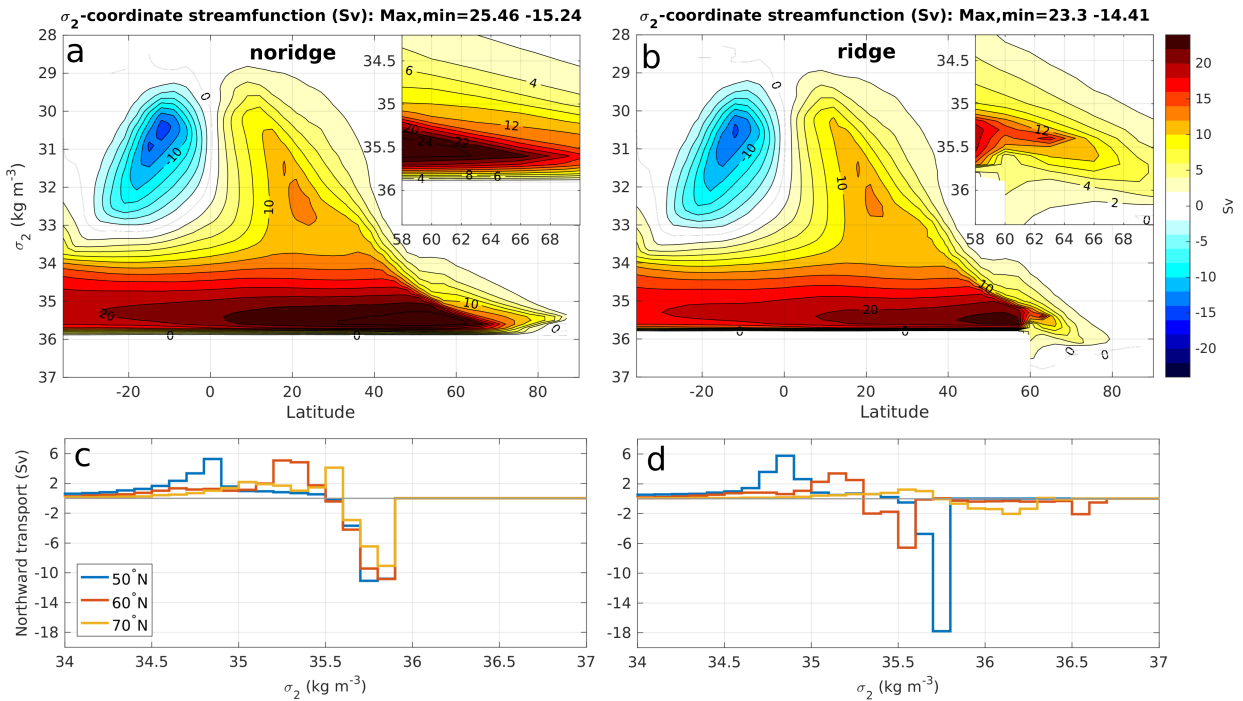
947 FIG. 9. Annual-mean mixed layer depth (MLD; m) in the northern Atlantic. Absolute values for *noridge* are  
 948 shown in *left* panel and anomalies of *ridge* relative to the *noridge* in the *right* panel.



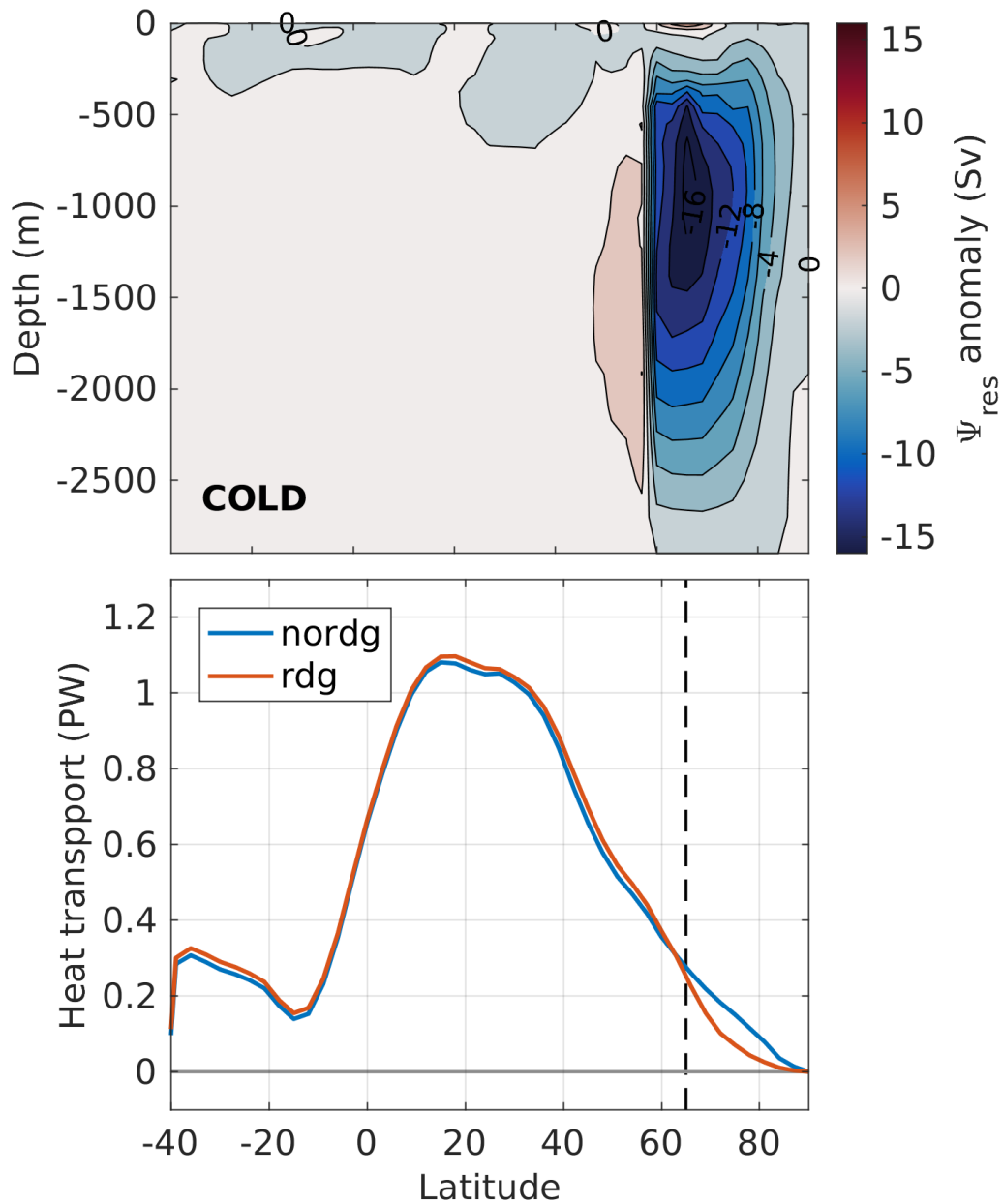
949 FIG. 10. Zonal mean surface density flux out of the ocean (in  $10^{-6} \text{ kg m}^{-2} \text{ s}^{-1}$ ) in the North Atlantic for  
 950 *noridge* (dashed lines) and *ridge* (solid lines). The total density flux (blue) is decomposed into contributions  
 951 from heat (red) and freshwater (yellow) fluxes. Negative values indicate densification (i.e. buoyancy loss).



952 FIG. 11. Annual mean water mass transformation (WMT) function ( $F(\sigma)$ ) in Sv for *norige* (dashed) and  
 953 *ridge* (solid). The WMT is integrated over different regions; (top) the entire North Atlantic from  $40^\circ - 90^\circ\text{N}$ ,  
 954 (middle) subpolar North Atlantic between  $40^\circ - 70^\circ\text{N}$  and (bottom) polar basin north of  $70^\circ\text{N}$ . The surface  
 955 forced WMT is estimated from the spatial integral of the surface density flux ( $D$ ) over 10 years in the Atlantic  
 956 basin spanning the density range  $\sigma_2 = 28.4 - 36.4 \text{ kg m}^{-3}$  with a density bin width of  $\Delta\sigma = 0.1$ . Here  $F(\sigma) \approx 0$   
 957 for  $\sigma < 30$ . Negative values imply a WMT to lower density classes (i.e. *buoyancy gain*) and positive values  
 958 represents transformation from lower to greater densities (i.e. *buoyancy loss*).



959 FIG. 12. Meridional overturning circulation streamfunction in latitude-density space ( $MOC_{\sigma}$ ; Sv) for (a)  
 960 *noridge* and (b) *ridge*. The density bins are the same as used for the computation of the WMT, with a reference  
 961 density at 2 km depth. Contour lines are plotted at 2 Sv intervals. The insert in (a) and (b) shows a zoom on  
 962 the region around the GSR. (c and d) Mean volume flux (Sv) by  $MOC_{\sigma}$  for *noridge* and *ridge* across different  
 963 latitudes in the Atlantic basin. Positive values indicate northward flow.



964 FIG. 13. Meridional overturning streamfunction anomaly (*ridge-noridge*) and ocean heat transport (PW) in  
 965 the Atlantic basin for the COLD case, where the solar constant is reduced to  $1360 \text{ W m}^2$ .