

# *The impact of salinity perturbations on the future uptake of heat by the Atlantic Ocean*

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**<sub>1</sub> The impact of salinity perturbations on the future  
<sub>2</sub> uptake of heat by the Atlantic Ocean**

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3 Anthropogenic ocean heat uptake is a key factor in determining climate  
4 change and sea-level rise. There is considerable uncertainty in projections  
5 of freshwater forcing of the ocean, with the potential to influence ocean heat  
6 uptake. We investigate this by adding either -0.1 Sv or +0.1 Sv freshwater  
7 to the Atlantic in global climate model simulations, simultaneously impos-  
8 ing an atmospheric CO<sub>2</sub> increase. The resulting changes in the Atlantic merid-  
9 ional overturning circulation are roughly equal and opposite ( $\pm 2$ Sv). The im-  
10 pact of the perturbation on ocean heat content is more complex, although  
11 it is relatively small ( $\sim 5\%$ ) compared to the total anthropogenic heat up-  
12 take. Several competing processes either accelerate or retard warming at dif-  
13 ferent depths. Whilst positive freshwater perturbations cause an overall heat-  
14 ing of the Atlantic, negative perturbations produce insignificant net changes  
15 in heat content. The processes active in our model appear robust, although  
16 their net result is likely model- and experiment-dependent.

## 1. Introduction

17 The rate at which the global ocean takes up heat is a key factor in determining how  
18 societies will experience future climate change. This rate sets how fast the surface of  
19 the Earth can warm in response to the radiative changes in the atmosphere, and the  
20 changing ocean heat content (OHC) currently contributes about half of the total sea-level  
21 rise [*Church et al.*, 2013]. There is significant uncertainty in model projections of the  
22 future global mean sea-level, partly due to differences in how they model the evolution of  
23 the heat content of the ocean [*Church et al.*, 2013].

24 The long term heat uptake by the ocean is set by the rate at which water from the surface  
25 mixed layer communicates with the deep ocean below the thermocline. The Atlantic is  
26 a particularly interesting area from this point of view, due to the locally-intense vertical  
27 mixing of water properties associated with the deep-water formation regions in the far  
28 north Atlantic and the associated overturning circulation.

29 *Mauritzen et al.* [2012] demonstrate the current communication of surface heat anom-  
30 lies to the deep Atlantic, in part through density-compensated flows of warm, relatively  
31 saline waters. The future hydrological forcing of the ocean as the climate changes is uncer-  
32 tain, and the changing salinity of surface waters may affect how this density-compensated  
33 transport operates. Climate models also project a range of possible reductions in North  
34 Atlantic deep-water formation and the associated meridional overturning flow under cli-  
35 mate change scenarios, primarily driven by the increased heat-flux forcing of the surface  
36 ocean [*Gregory et al.*, 2005]. There are thus two closely linked but conceptually sepa-  
37 rate influences on the communication of surface heating to depth in the North Atlantic:

38 changes in the volume of deep-water formed, and the role of salinity in allowing the tem-  
39 perature of deep-water of a given density to vary. A number of previous studies (e.g.  
40 *Gregory* [2000]; *Knutti and Stocker* [2000]; *Levermann et al.* [2005]; *Yin et al.* [2010];  
41 *Kienert and Rahmstorf* [2012]) have noted an increase in ocean heat content and sea-level  
42 when freshwater is added to the North Atlantic, associated with a decrease in the strength  
43 of the Atlantic meridional overturning circulation (AMOC) (although *Bouttes et al.* [2014]  
44 argue that the sea-level changes are not primarily due to the AMOC change directly) but  
45 the sensitivity of this effect to salinity perturbations of both signs under climate change  
46 conditions has not been systematically investigated.

47 In this study, we conduct a suite of idealised coupled climate model experiments to  
48 investigate how uncertainty in the hydrological forcing of the North Atlantic affects its  
49 ability to take up heat under a climate change scenario.

## 2. Setup

### 2.1. Model Description

50 FAMOUS XFXWB (hereafter referred to as FAMOUS)[*Smith, 2012; Smith et al., 2008*],  
51 is a low resolution version of the widely used Hadley Centre atmosphere-ocean gen-  
52 eral circulation model (HadCM3)[*Gordon et al., 2000*]. The ocean component is based  
53 on the Cox-Bryan model [*Pacanowski et al., 1990*], run at a resolution of 2.5° latitude  
54 by 3.75° longitude, with 20 vertical levels. The atmosphere is based on the primitive  
55 equations with a resolution of 5° latitude by 7.5° longitude with 11 vertical levels. We  
56 use FAMOUS because it is much faster and computationally cheaper than HadCM3 for  
57 multi-centennial climate simulations. FAMOUS incorporates a number of differences from

58 HadCM3. The FAMOUS bathymetry does not have the deeper overflow channels that  
59 were added to HadCM3 to improve representation of deep-water flow from the Greenland-  
60 Iceland-Norwegian seas, and instead Iceland has been removed to facilitate more north-  
61 ward ocean heat transport. The ocean model in FAMOUS uses a rigid lid formulation,  
62 so surface freshwater fluxes are parameterised via the addition or removal of salt. The  
63 local surface salinity, rather than a global reference value, is used to calculate these fluxes,  
64 and a small annual adjustment is made to remove the resulting numerical drift in global  
65 average salinity.

66 Despite its relatively coarse resolution, previous studies have shown that FAMOUS  
67 produces a simulation of North Atlantic climate variability and AMOC that is well in line  
68 with other models [*Hawkins et al.*, 2011; *Smith and Gregory*, 2009; *Balan Sarojini et al.*,  
69 2011]. Characteristics of the Atlantic density compare reasonably well with observations  
70 [*Levitus et al.*, 1998] (figs S1, S2), although the North Atlantic deep water is biased warm  
71 and salty and the Antarctic bottom water is too cold and fresh. FAMOUS's global climate  
72 sensitivity to CO<sub>2</sub> increase (0.91 W/m<sup>2</sup>/K) is similar to that of HadCM3 (1.32 W/m<sup>2</sup>/K),  
73 and the sensitivity of the AMOC to buoyancy perturbation, and the associated impact on  
74 surface climate, also fit well with what is seen in higher resolution model intercomparisons  
75 [*Smith and Gregory*, 2009; *Stouffer et al.*, 2006].

## 2.2. Experiment Design

76 This study is primarily based around three idealised climate change experiments with  
77 FAMOUS. They are not meant to represent projections of realistic changes within the  
78 climate system, but to demonstrate which processes are important to Atlantic heat uptake

79 and the levels of sensitivity that they have to the different forcings. From a well-spun-up  
80 ( $\sim 4000$  years, [Eby *et al.*, 2013]), unperturbed preindustrial control state (CTR), the three  
81 climate-change experiments see a 1%/year increase in atmospheric CO<sub>2</sub> concentrations for  
82 140 years until 1160ppmv is realised (four times the initial preindustrial value), at which  
83 point CO<sub>2</sub> concentrations are held constant until the end of the simulation. Details of  
84 the forcings used in all the experiments in this study are listed in table 1. Two other  
85 experiments (CTRW and CTRS), where the freshwater forcing is applied without an  
86 increase in atmospheric CO<sub>2</sub>, will be mentioned later.

87 One of the experiments (CO2) has only this CO<sub>2</sub> forcing. The others are forced with  
88 an additional freshwater forcing in the North Atlantic for the duration of the experiment.  
89 In experiment CO2W, this forcing consists of 0.1Sv of freshwater, evenly distributed over  
90 the surface of the Atlantic between 50°N and 70°N. This forcing is one of the idealised  
91 protocols used in the THCmip study [Stouffer *et al.*, 2006], and was designed to ensure  
92 that the deep-water formation zones are uniformly covered while allowing a significant  
93 size of water flux to be used without causing numerical problems. Experiment CO2S is  
94 the same, except that the sign of the forcing is opposite, removing freshwater from the  
95 ocean and creating a positive surface salinity anomaly.

96 0.1Sv is the approximate magnitude of the change in freshwater flux to the North  
97 Atlantic due to changes in precipitation, evaporation and river inflow in both HadCM3 and  
98 FAMOUS around 2100 in business-as-usual type climate change experiments [Wood *et al.*,  
99 1999; Hawkins *et al.*, 2011]. It is well below the threshold at which the AMOC in FAMOUS



100 enters a different stability regime [*Hawkins et al.*, 2011], so our three experiments are  
101 unlikely to be pushed into radically different climate regimes by the perturbation.

102 Analysis of 14 of the models in the CMIP5 multi-model ensemble database [*Taylor et al.*,  
103 2012] (ACCESS1-0 ACCESS1-3 CNRM-CM5 CSIRO-Mk3-6-0 FGOALS-g2 GFDL-CM3  
104 HadGEM2-ES GFDL-ESM2G GFDL-ESM2M IPSL-CM5A-LR IPSL-CM5A-MR MPI-  
105 ESM-LR MPI-ESM-MR MPI-ESM-P) shows a change of  $0.058 \pm 0.012$  Sv (one standard  
106 dev) in area-integral P-E between 50-70°N in the Atlantic around the time of CO<sub>2</sub> doubling  
107 under 1% CO<sub>2</sub> (this does not include the influence of runoff from land). Our anomaly  
108 forcing range of  $\pm 0.1$  Sv thus comfortably brackets the spread of uncertainty in current  
109 model projections.

110 Each experiment is 250 years long; our intention is to study the long term (century-  
111 scale) processes in play as the system adjusts to the climate change forcing and this period  
112 is long enough to allow the anomalies we are interested in to become significant without  
113 requiring the expense of achieving a new equilibrium state. In general, our results are  
114 expressed as averages over the last decade of this period, and are indicative of the relative  
115 magnitudes of different quantities rather than projections relevant to a specific real-world  
116 time period.

### 3. Results

117 Experiment CO2 has a global, annual average surface temperature warming of 6.5°C  
118 after the 250 years of simulation. The global OHC (quantified as the volume integral of  
119 potential temperature (in °C) converted to heat using a fixed volumetric heat capacity of  
120  $3.95 \times 10^6$  J/°C/m<sup>3</sup>) rises from  $20.5 \times 10^{24}$  J to  $26.4 \times 10^{24}$  J over this period (fig 1a). At the

121 end of the experiment, this extra heat is predominantly found in the ocean between 40°S  
122 and 40°N above 1000m, although it penetrates deeper in the north Atlantic.

123 The strength of the AMOC declines in experiment CO2, as found in nearly all other  
124 models [*Gregory et al.*, 2005; *Meehl et al.*, 2007]. The strength of the AMOC maximum  
125 reduces from  $\sim 19\text{Sv}$  to  $\sim 12\text{Sv}$  (fig 2), and although still located near 30°N, it shallows from  
126 1000m to 500m. The shape of the streamfunction changes too: in the CTR simulation  
127 the influence of the main cell stretches almost to the ocean floor at 30°N whilst the cell  
128 is confined to the top 2000m by the end of CO2 (fig S3) (see table 1 for a summary of  
129 AMOC changes in the various experiments).

130 The experiments with additional freshwater forcing show further changes in the strength  
131 of the AMOC (fig 2), although they do not affect the shape of the streamfunction so  
132 much (fig S3). Compared to CO2, CO2W has an additional weakening of 3Sv in the  
133 AMOC maximum over the first 50 years of the simulation, stabilising a little higher at  
134  $\sim 10\text{Sv}$  for the rest of the run (-2 Sv relative to CO2, Table 1). The AMOC maximum in  
135 CO2S does not weaken as much as in CO2, and after the first 50 years of simulation it  
136 stabilises at  $\sim 13\text{Sv}$  (+2 Sv relative to CO2, Table 1). In CTR the AMOC maximum has  
137 decadal variability with a standard deviation of  $\sim 1\text{Sv}$ , although this variability is visibly  
138 suppressed in CO2 (fig 2a). The AMOC anomalies compared with CO2 in the last decade  
139 of CO2S and CO2W are  $\sim 2\text{Sv}$ . We judge these to be significant changes, because they  
140 are larger than 98% of the decadal anomalies from the mean in the last thousand years  
141 of the spinup to CTR. The timing of the changes in AMOC strength in these simulations  
142 is consistent with the slow southward advection timescales of density perturbations in

143 the North Atlantic in FAMOUS, propagating from the deep-water formation zones in the  
144 north Atlantic along the western boundary, reaching the mid-latitudes a few decades after  
145 the start of the experiments.

146 In their AMOC response, CO2S and CO2W show approximately equal and opposite  
147 differences with respect to CO2 due to the additional freshwater anomalies imposed (fig  
148 2). The effect of this freshwater forcing on the OHC are, however, not symmetrical (figs  
149 1, S4). The largest impact in each experiment is in the Atlantic (fig S2), where the major  
150 anomalies are found in the western boundary currents, suggesting a close association  
151 with the AMOC. There are small signals in the rest of the global ocean although they  
152 spread over a wide area and thus have some influence on the global OHC (fig 1a). By the  
153 end of the experiment, CO2W has an additional uptake of  $\sim 0.12 \times 10^{24}$  J in the Atlantic  
154 between 0-80°N compared to CO2, mostly found in the subtropics (south of 40°N). In  
155 CO2S there is an increase of  $0.03 \times 10^{24}$  J in subpolar OHC (north of 40°N) which is partly  
156 counteracted by a reduction in subtropical OHC, resulting in a total Atlantic OHC that  
157 is almost unchanged compared to CO2. The decadal variability in Atlantic OHC in CTR  
158 has a standard deviation of  $\sim 0.02 \times 10^{24}$  J, mostly concentrated in the subtropical region.  
159 The Atlantic OHC anomaly in the last decade of CO2W compared to CO2 is larger than  
160 any decadal anomaly from the mean in the last thousand years of the spinup to CTR, so  
161 we judge it to be significant. In CO2S, the subpolar anomaly taken alone is significant  
162 compared to the decadal variability in CTR, although the net change in the Atlantic is  
163 not. Heat content changes for each region at the end of each experiment are summarised  
164 in table 1.

165 The imposition of the additional freshwater anomalies results in a characteristic pattern  
166 of Atlantic OHC anomalies in CO2W and CO2S (fig 3 - note that this figure shows  
167 temperature anomalies with respect to CO2, and the cool anomalies shown are still warmer  
168 than in CTR). The additional freshwater forcing applied in CO2W and CO2S influences  
169 the deep-water formation and convective mixing that occurs between the cold surface  
170 and warmer mid-depth waters in GIN seas and south of Greenland in FAMOUS. This  
171 results in a warm anomaly relative to CO2 between 100m and 1500m in subpolar waters  
172 in CO2W as deep-water formation is hindered (fig 3b), and a cool anomaly in CO2S as  
173 deep-water formation is enhanced (fig 3d).

174 The main influence on subtropical OHC is uptake of heat into the thermocline. As north  
175 Atlantic deepwater formation is reduced in CO2W, the AMOC slows and the subtropical  
176 thermocline deepens. This deeper, warmer thermocline may be directly due to AMOC  
177 adjustment processes (e.g. [*Johnson and Marshall, 2002*]) or may be associated with  
178 some other feedback such as enhanced tropical heating by the atmosphere. The heat  
179 uptake associated with this is the dominant feature of the OHC response in CO2W (fig  
180 3a) relative to CO2. The opposite effect can be seen in CO2S as the AMOC strengthens  
181 and the thermocline shallows (fig 3c), relative to CO2. This process is associated more  
182 with a change in the volume of deep-water formed and dynamic change in the AMOC  
183 rather than density compensation, where the changed salinity of water of a given density  
184 implies a change in temperature.

185 Relative to CO2, CO2W has a significant additional surface cooling in the subpolar  
186 region as the AMOC weakens further (fig 3b), a well-known result of the reduction in

187 ocean heat transport associated in the AMOC that mitigates the surface warming over  
188 northern Europe and the Arctic caused by the increase in atmospheric  $p\text{CO}_2$ . This anoma-  
189 lous surface cooling (down to  $\sim 100\text{m}$ ) is advected southwards by the gyre circulation to  
190 the subtropics in CO2W. The strengthening of the AMOC in CO2S results in a surface  
191 warming of smaller size (fig 3c,d) than the cooling in CO2W.

192 The ocean temperature anomalies below 2000m in CO2S and CO2W originate north of  
193 the Greenland-Scotland ridge. In the Arctic regions in CO2 there are increases in both  
194 vertical and isopycnal diffusion to depth of the surface  $\text{CO}_2$ -forced warming signal. This is  
195 enhanced in CO2S, with surface warming being carried more rapidly to yet deeper regions,  
196 whilst in CO2W it is hindered, with some areas of the deep Arctic feeling no influence  
197 of the surface warming at all. The deep, cool anomaly in fig 3a,c does not represent an  
198 absolute cooling of the deep ocean in this experiment, and is the absence of the slow  
199 warming signal in the baseline CO2 simulation. These deeper anomalies, below the depth  
200 of direct influence of the AMOC, are more likely to be due to waters whose density is the  
201 same as in CO2, but whose temperature is different, now that the freshwater forcing has  
202 changed their salinity.

203 The enhanced communication of the surface  $\text{CO}_2$ -forced warming signal to depth in  
204 CO2S is likely the reason that the strengthening of the AMOC (compared to CO2) present  
205 in this experiment does not cause a surface warming anomaly of equal magnitude (but  
206 opposite sign) to the surface cooling in CO2W linked to the AMOC slowdown. It is also  
207 the reason for the difference in total Atlantic OHC changes between CO2W and CO2S.  
208 Both experiments have a change in subtropical thermocline heat uptake in line with the

209 dynamical changes in AMOC strength forced by their freshwater anomalies, but in CO2S  
210 this cool OHC anomaly is cancelled by communication of the warmer surface signal to  
211 depth further north, a mechanism that is inhibited in CO2W by the lower surface densities.

212 In a further set of experiments, the same freshwater forcing anomalies were imposed  
213 without an increase in atmospheric CO2 (experiments CTRW and CTRS). CTRW and  
214 CTRS were branched from a later point in CTR than the CO2 experiments, but the drift  
215 in ocean state in CTR is small and does not significantly affect the analysis here (where  
216 anomalies are quoted for CTRS and CTRW they have been calculated with respect to  
217 their parallel control. In the figures, only the portion of CTR parallel to the main CO2  
218 experiments has been shown.). Without the influence of warming surface heat fluxes in  
219 these experiments, the large-scale ocean stratification does not significantly change and the  
220 AMOC does not shallow as it does in CO2, CO2S and CO2W. CTRW and CTRS have a  
221 similar sensitivity of the maximum AMOC strength to the freshwater anomalies as CO2W  
222 and CO2S (table 1), and the same set of processes described above act to influence the  
223 Atlantic OHC . However, both CTRS and CTRW have a small net increase in Atlantic  
224 OHC (fig: 1b). The sensitivities of the Atlantic OHC to CO2 and pure hydrological  
225 forcing are thus seen to combine in a non-linear fashion. In CTRS, the weaker underlying  
226 stratification and deeper influence of the AMOC that is being perturbed by the freshwater  
227 anomalies mean that the communication to depth of surface temperature anomalies in the  
228 far north is enhanced. In CO2S this deep warming signal acts to cancel out the cooling in  
229 the thermocline, but in CTRS it becomes the dominant feature and results in an overall  
230 increase in Atlantic OHC.

#### 4. Discussion

231 We have shown, for an idealised scenario in one coupled climate model, how sensitive  
232 North Atlantic heat content is to uncertainty in changes in the hydrological forcing of  
233 the ocean. How representative (and useful) are our results in terms of projections of the  
234 real climate system? The major processes described above are fundamental in our under-  
235 standing of ocean behaviour, but are they being modelled appropriately and interacting  
236 in the right way in the climate model used here?

237 As stated in section 2.1, although FAMOUS displays some biases common to many  
238 lower-resolution climate models its representation of a number of features key to this  
239 study have been shown to be reasonable, so its faults should not present major problems  
240 in relating our results to the real climate system. However, the details of the underlying  
241 ocean stratification and shape of the AMOC are critical variables in the sensitivity of  
242 the ocean response to the forcing used in our experiment. As shown in figure S1, the  
243 basic ocean stratification in FAMOUS for the present day is fairly realistic. How realistic  
244 the sensitivity of that stratification and the AMOC structure to climate change cannot,  
245 however, be estimated. There is a wide spread in the projections of how the maximum  
246 strength of the AMOC will change from different climate models [*Cheng et al.*, 2013] and  
247 only relatively short and sparse observations of its current state; uncertainty in the shape  
248 of the overturning is yet higher.

249 Compared to experiment CO2, CO2W has an additional weakening of 2Sv in the  
250 AMOC, and an additional  $0.12 \times 10^{24}$ J of heat stored in the Atlantic at the end of the  
251 experiment. If it is assumed that all of the change in Atlantic OHC in CO2W is due to

252 the reduction in the volume of deep-water formed, rather than density-neutral changes in  
253 temperature, this relationship can be used to suggest that up to  $0.4 \times 10^{24} \text{J}$  of the  $1.8 \times 10^{24} \text{J}$   
254 of heat taken up by the Atlantic in experiment CO2 (in which the AMOC weakens by  
255  $7 \text{Sv}$ ) could be attributed to the influence of the weakening AMOC in that experiment.  
256 Although our experiments suggest that changes in OHC cannot be linearly scaled against  
257 changes in AMOC strength for all cases, in deriving this figure we are comparing two  
258 experiments where the AMOC weakens and shallows, and the same component processes  
259 (subtropical thermocline deepening and a reduction in high latitude deep convection) are  
260 acting in the same sense in both. Accepting the limitations of this simple linear extrapo-  
261 lation and allowing that some of the OHC anomaly in CO2W will be due to factors other  
262 than the AMOC change, our experiments suggest that on the order of 10% of the total  
263 Atlantic heat uptake under climate change could be attributed to the declining AMOC.  
264 The accuracy of projections of local sea level rise in the Atlantic, and other quantities de-  
265 pendent on OHC may thus be significantly dependent on the knowing the true sensitivity  
266 of the AMOC to climate change.

267 Experiment CO2 has a  $\sim 6 \times 10^{24} \text{J}$  increase in global OHC, compared to CTR. In com-  
268 parison, the freshwater perturbations used here produce at most a  $\sim 0.33 \times 10^{24} \text{J}$  change  
269 in global ocean heat content. Even in the Atlantic, where their influence is largest, OHC  
270 changes due to the freshwater forcing anomaly alone (the anomaly between CO2W and  
271 CO2) represent only 7% of the change due to the  $\text{CO}_2$  forcing (the anomaly between CO2  
272 and CTR). Based on the idealised experiments here then, we can conclude that uncer-  
273 tainty in changes in the hydrological forcing of the North Atlantic is not a major factor in



274 calculations of the large scale heat budget of the ocean. Such forcings may, however, be  
275 important in developing an understanding of the behaviour of the Atlantic heat budget  
276 at the process level. The distinctive three-layer anomaly pattern seen in fig 3 should also  
277 be robust, even if the details of the cancellation of the anomalies at different levels varies  
278 for different models and climate change scenarios, and may be useful in detection and  
279 attribution studies.

280 One possible source of additional freshwater input to the North Atlantic in the future  
281 might come from accelerated mass loss from the Greenland ice-sheet. Studies reviewed  
282 by *Church et al.* [2013] suggest that this might be a source of up to an additional 0.02Sv  
283 of freshwater by the year 2100. Although this is 5 times smaller than the idealised per-  
284 turbations used in this study, it would be put into the ocean in a more concentrated area;  
285 *Smith and Gregory* [2009] show that the sensitivity of the AMOC to freshwater perturba-  
286 tion is very sensitive to the location in which it is added. The multi-model comparison of  
287 *Swingedouw et al.* [2014] suggests that a 0.1Sv input from Greenland in the second half of  
288 the next century would cause an additional reduction in AMOC strength of  $1.1\pm 0.6$ Sv.  
289 There is uncertainty in both the magnitude and timing of future mass loss by the ice-sheet,  
290 along with variations in the simulated location and sensitivities of deep convection sites  
291 and AMOC behaviours across different climate models. Robust, quantitative assessments  
292 of the impact of Greenland ice-sheet mass loss on the ocean heat budget through ocean  
293 circulation changes are thus impossible, but our experiments here suggest that this impact  
294 is unlikely to be large.

## 5. Conclusions

295 Perturbations to the surface freshwater forcing of the North Atlantic have the potential  
296 to significantly impact the future uptake of heat to the ocean, affecting both the volume  
297 of deep-water formed and the temperature of deep-water of a given density.

298 Our idealised, constant  $\pm 0.1\text{Sv}$  perturbations in freshwater forcing alter the heat stored  
299 in the Atlantic under a 1% per year  $\text{CO}_2$  increase scenario by  $\sim 7\%$ . This represents an  
300 additional global uptake of approx  $0.33 \times 10^{24}\text{J}$  over 250 years when the Atlantic freshwater  
301 perturbation is positive, around 5% of the global increase in ocean heat content due to  
302 the increase in  $\text{CO}_2$ .

303 In our experiments, freshwater perturbations cause a net increase in Atlantic heat con-  
304 tent as deep-water formation slows and the subtropical thermocline deepens, in line with  
305 thermocline adjustment theories of the Atlantic overturning circulation. Salt perturba-  
306 tions produce only small net changes in ocean heat content compared to the baseline  $\text{CO}_2$   
307 increase scenario, as enhanced deep-water formation results in a decrease in the depth of  
308 the subtropical thermocline but the surface  $\text{CO}_2$  warming signal is mixed more effectively  
309 to depth in the less-stratified Arctic ocean.

310 The processes found to be active in model appear robust, although their net result on  
311 ocean heat content depends on details such as the basic ocean stratification, depth of  
312 influence of the Atlantic overturning circulation and the forcings used. More research is  
313 required across a range of models to determine the true sensitivity of the real ocean heat  
314 uptake to surface salinity forcings.

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328 tals. Data from the model simulations used in this study are available on request from  
329 the authors.

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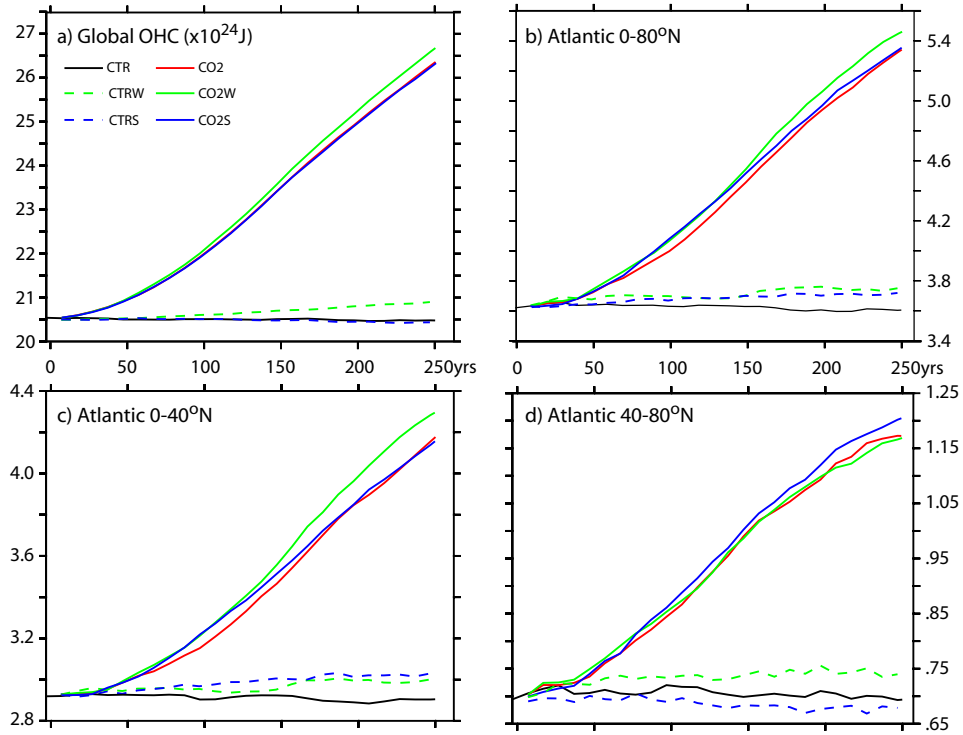
Expt Name	CO <sub>2</sub> increase	Freshwater anomaly	AMOC max anomaly (Sv)	Atlantic OHC anomaly (10 <sup>24</sup> J)			
				0-40N	40-80N	0-80N	
CO2 (wrt CTR)	1%	0	-7	0 - 100m	0.05	0.03	0.08
				100-1500m	0.89	0.36	1.25
				1500m - bottom	0.30	0.08	0.38
				full depth	1.24	0.47	<b>1.71</b>
CO2W (wrt CO2)	1%	+	-2	0 - 100m	0.00	-0.01	-0.01
				100-1500m	0.21	0.02	0.23
				1500m - bottom	-0.08	-0.02	-0.10
				full depth	0.13	-0.01	<b>0.12</b>
CO2S (wrt CO2)	1%	-	+2	0 - 100m	0.00	0.00	0.00
				100-1500m	-0.17	-0.02	-0.19
				1500m - bottom	0.15	0.04	0.20
				full depth	-0.02	0.03	<b>0.01</b>
CTRW (wrt CTR)	0	+	-5	0 - 100m	0.00	-0.01	-0.01
				100-1500m	0.16	0.06	0.21
				1500m - bottom	-0.09	-0.02	-0.11
				full depth	0.06	0.03	<b>0.10</b>
CTRS (wrt CTR)	0	-	+3	0 - 100m	0.00	0.00	0.00
				100-1500m	-0.04	-0.03	-0.07
				1500m - bottom	0.13	0.00	0.13
				full depth	0.09	-0.03	<b>0.06</b>

**Table 1.** Details of experiment setup and summary of final Atlantic ocean heat content (OHC)

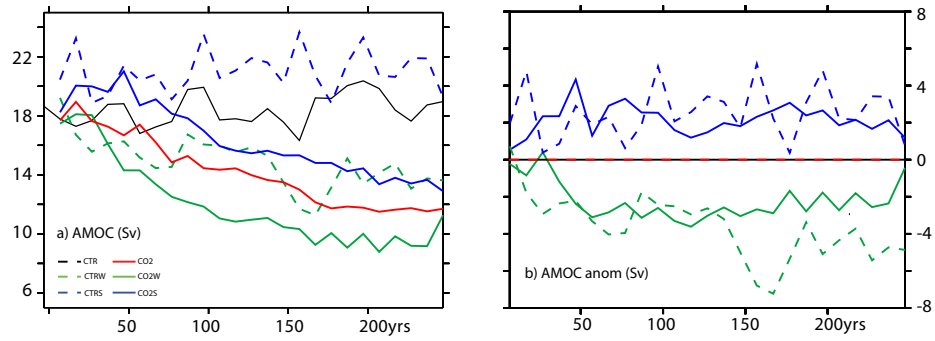
anomalies in various depth ranges, averaged over the final decade of the experiment. Anomalies for CO2S and CO2W are expressed with respect to CO2, those for CO2, CTRW and CTRS are with respect to the time-mean of the parallel CTR. Atlantic meridional overturning circulation (AMOC) anomalies are averaged over the last 30 years of the transient experiments

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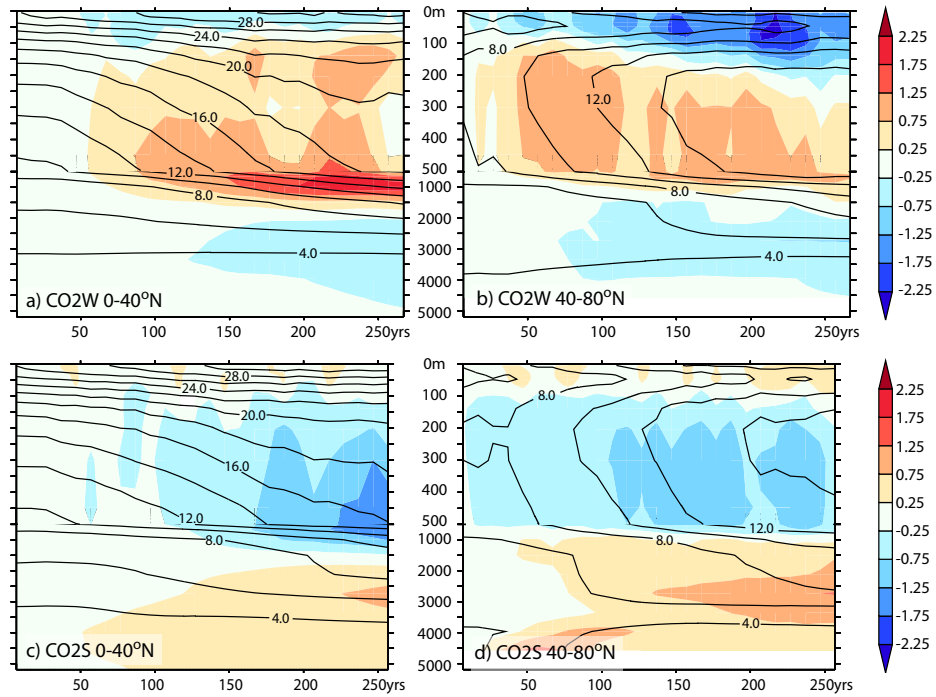


**Figure 1.** Evolution of depth integrated ocean heat content ( $10^{24}$  J) for different parts of the ocean over the course of the simulations.



**Figure 2.** Evolution of the maximum strength of the Atlantic overturning over the course of the simulations. a) absolute values; b) anomalies due to the freshwater forcing. Anomalies

for CO2W and CO2S are expressed with respect to CO2; anomalies for CTRW and CTRW are expressed with respect to the time-mean of their parallel CTR.



**Figure 3.** Evolution of horizontal average Atlantic potential temperature ( $^{\circ}\text{C}$ ). Contours are absolute temperature values; solid colours are anomalies relative to experiment CO2