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Anthropogenic ocean heat uptake is a key factor in determining climate change and sea-level rise. There is considerable uncertainty in projections of freshwater forcing of the ocean, with the potential to influence ocean heat uptake. We investigate this by adding either -0.1 Sv or +0.1 Sv freshwater to the Atlantic in global climate model simulations, simultaneously imposing an atmospheric CO_2 increase. The resulting changes in the Atlantic meridional overturning circulation are roughly equal and opposite (±2Sv). The impact of the perturbation on ocean heat content is more complex, although it is relatively small (∼5%) compared to the total anthropogenic heat uptake. Several competing processes either accelerate or retard warming at different depths. Whilst positive freshwater perturbations cause an overall heating of the Atlantic, negative perturbations produce insignificant net changes in heat content. The processes active in our model appear robust, although their net result is likely model- and experiment-dependent.

1. Introduction

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

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

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

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

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

2. Setup

2.1. Model Description

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

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

Despite its relatively coarse resolution, previous studies have shown that FAMOUS produces a simulation of North Atlantic climate variability and AMOC that is well in line with other models [*Hawkins et al.*, 2011; *Smith and Gregory*, 2009; *Balan Sarojini et al.*, 2011]. Characteristics of the Atlantic density compare reasonably well with observations [*Levitus et al.*, 1998] (figs S1, S2), although the North Atlantic deep water is biassed warm and salty and the Antarctic bottom water is too cold and fresh. FAMOUS's global climate sensitivity to CO_2 increase (0.91 $W/m^2/K$) is similar to that of HadCM3 (1.32 $W/m^2/K$), and the sensitivity of the AMOC to buoyancy perturbation, and the associated impact on surface climate, also fit well with what is seen in higher resolution model intercomparisons [*Smith and Gregory*, 2009; *Stouffer et al.*, 2006].

2.2. Experiment Design

This study is primarily based around three idealised climate change experiments with FAMOUS. They are not meant to represent projections of realistic changes within the climate system, but to demonstrate which processes are important to Atlantic heat uptake and the levels of sensitivity that they have to the different forcings. From a well-spun-up (∼4000 years, [*Eby et al.*, 2013]), unperturbed preindustrial control state (CTR), the three climate-change experiments see a 1%/year increase in atmospheric CO_2 concentrations for 140 years until 1160ppmv is realised (four times the initial preindustrial value), at which point CO_2 concentrations are held constant until the end of the simulation. Details of the forcings used in all the experiments in this study are listed in table 1. Two other experiments (CTRW and CTRS), where the freshwater forcing is applied without an increase in atmospheric CO_2, will be mentioned later.

One of the experiments (CO2) has only this CO_2 forcing. The others are forced with an additional freshwater forcing in the North Atlantic for the duration of the experiment. In experiment CO2W, this forcing consists of 0.1Sv of freshwater, evenly distributed over the surface of the Atlantic between 50°N and 70°N. This forcing is one of the idealised protocols used in the THCmip study [*Stouffer et al.*, 2006], and was designed to ensure that the deep-water formation zones are uniformly covered while allowing a significant size of water flux to be used without causing numerical problems. Experiment CO2S is the same, except that the sign of the forcing is opposite, removing freshwater from the ocean and creating a positive surface salinity anomaly.

0.1Sv is the approximate magnitude of the change in freshwater flux to the North Atlantic due to changes in precipitation, evaporation and river inflow in both HadCM3 and FAMOUS around 2100 in business-as-usual type climate change experiments [*Wood et al.*, 1999; *Hawkins et al.*, 2011]. It is well below the threshold at which the AMOC in FAMOUS enters a different stability regime [*Hawkins et al.*, 2011], so our three experiments are unlikely to be pushed into radically different climate regimes by the perturbation.

Analysis of 14 of the models in the CMIP5 multi-model ensemble database [*Taylor et al.*, 2012] (ACCESS1-0 ACCESS1-3 CNRM-CM5 CSIRO-Mk3-6-0 FGOALS-g2 GFDL-CM3 HadGEM2-ES GFDL-ESM2G GFDL-ESM2M IPSL-CM5A-LR IPSL-CM5A-MR MPI-ESM-LR MPI-ESM-MR MPI-ESM-P) shows a change of 0.058 $\pm$0.012 Sv (one standard dev) in area-integral P-E between 50-70°N in the Atlantic around the time of CO_2 doubling under 1% CO_2 (this does not include the influence of runoff from land). Our anomaly forcing range of $\pm$0.1 Sv thus comfortably brackets the spread of uncertainty in current model projections.

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

3. Results

Experiment CO2 has a global, annual average surface temperature warming of 6.5°C after the 250 years of simulation. The global OHC (quantified as the volume integral of potential temperature (in °C) converted to heat using a fixed volumetric heat capacity of $3.95\text{x}10^6$ J/°C/m^3) rises from $20.5\text{x}10^{24}$J to $26.4\text{x}10^{24}$J over this period (fig 1a). At the end of the experiment, this extra heat is predominantly found in the ocean between 40°S and 40°N above 1000m, although it penetrates deeper in the north Atlantic.

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

The experiments with additional freshwater forcing show further changes in the strength of the AMOC (fig 2), although they do not affect the shape of the streamfunction so much (fig S3). Compared to CO2, CO2W has an additional weakening of 3Sv in the AMOC maximum over the first 50 years of the simulation, stabilising a little higher at ∼10Sv for the rest of the run (-2 Sv relative to CO2, Table 1). The AMOC maximum in CO2S does not weaken as much as in CO2, and after the first 50 years of simulation it stabilises at ∼13Sv (+2 Sv relative to CO2, Table 1). In CTR the AMOC maximum has decadal variability with a standard deviation of ∼1Sv, although this variability is visibly suppressed in CO2 (fig 2a). The AMOC anomalies compared with CO2 in the last decade of CO2S and CO2W are ∼2Sv. We judge these to be significant changes, because they are larger than 98% of the decadal anomalies from the mean in the last thousand years of the spinup to CTR. The timing of the changes in AMOC strength in these simulations is consistent with the slow southward advection timescales of density perturbations in the North Atlantic in FAMOUS, propagating from the deep-water formation zones in the north Atlantic along the western boundary, reaching the mid-latitudes a few decades after the start of the experiments.

In their AMOC response, CO2S and CO2W show approximately equal and opposite differences with respect to CO2 due to the additional freshwater anomalies imposed (fig 2). The effect of this freshwater forcing on the OHC are, however, not symmetrical (figs 1, S4). The largest impact in each experiment is in the Atlantic (fig S2), where the major anomalies are found in the western boundary currents, suggesting a close association with the AMOC. There are small signals in the rest of the global ocean although they spread over a wide area and thus have some influence on the global OHC (fig 1a). By the end of the experiment, CO2W has an additional uptake of $\sim$0.12x10^{24}J in the Atlantic between 0-80°N compared to CO2, mostly found in the subtropics (south of 40°N). In CO2S there is an increase of 0.03x10^{24}J in subpolar OHC (north of 40°N) which is partly counteracted by a reduction in subtropical OHC, resulting in a total Atlantic OHC that is almost unchanged compared to CO2. The decadal variability in Atlantic OHC in CTR has a standard deviation of $\sim$0.02x10^{24}J, mostly concentrated in the subtropical region. The Atlantic OHC anomaly in the last decade of CO2W compared to CO2 is larger than any decadal anomaly from the mean in the last thousand years of the spinup to CTR, so we judge it to be significant. In CO2S, the subpolar anomaly taken alone is significant compared to the decadal variability in CTR, although the net change in the Atlantic is not. Heat content changes for each region at the end of each experiment are summarised in table 1.

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

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

Relative to CO2, CO2W has a significant additional surface cooling in the subpolar region as the AMOC weakens further (fig 3b), a well-known result of the reduction in ocean heat transport associated in the AMOC that mitigates the surface warming over northern Europe and the Arctic caused by the increase in atmospheric $p\mathrm{CO}_2$. This anomalous surface cooling (down to ∼100m) is advected southwards by the gyre circulation to the subtropics in CO2W. The strengthening of the AMOC in CO2S results in a surface warming of smaller size (fig 3c,d) than the cooling in CO2W.

The ocean temperature anomalies below 2000m in CO2S and CO2W originate north of the Greenland-Scotland ridge. In the Arctic regions in CO2 there are increases in both vertical and isopycnal diffusion to depth of the surface CO_2-forced warming signal. This is enhanced in CO2S, with surface warming being carried more rapidly to yet deeper regions, whilst in CO2W it is hindered, with some areas of the deep Arctic feeling no influence of the surface warming at all. The deep, cool anomaly in fig 3a,c does not represent an absolute cooling of the deep ocean in this experiment, and is the absence of the slow warming signal in the baseline CO2 simulation. These deeper anomalies, below the depth of direct influence of the AMOC, are more likely to be due to waters whose density is the same as in CO2, but whose temperature is different, now that the freshwater forcing has changed their salinity.

The enhanced communication of the surface CO_2-forced warming signal to depth in CO2S is likely the reason that the strengthening of the AMOC (compared to CO2) present in this experiment does not cause a surface warming anomaly of equal magnitude (but opposite sign) to the surface cooling in CO2W linked to the AMOC slowdown. It is also the reason for the difference in total Atlantic OHC changes between CO2W and CO2S. Both experiments have a change in subtropical thermocline heat uptake in line with the dynamical changes in AMOC strength forced by their freshwater anomalies, but in CO2S this cool OHC anomaly is cancelled by communication of the warmer surface signal to depth further north, a mechanism that is inhibited in CO2W by the lower surface densities.

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

4. Discussion

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

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

Compared to experiment CO2, CO2W has an additional weakening of 2Sv in the AMOC, and an additional 0.12x10^{24}J of heat stored in the Atlantic at the end of the experiment. If it is assumed that all of the change in Atlantic OHC in CO2W is due to the reduction in the volume of deep-water formed, rather than density-neutral changes in temperature, this relationship can be used to suggest that up to 0.4x10^{24}J of the 1.8x10^{24}J of heat taken up by the Atlantic in experiment CO2 (in which the AMOC weakens by 7Sv) could be attributed to the influence of the weakening AMOC in that experiment. Although our experiments suggest that changes in OHC cannot be linearly scaled against changes in AMOC strength for all cases, in deriving this figure we are comparing two experiments where the AMOC weakens and shallows, and the same component processes (subtropical thermocline deepening and a reduction in high latitude deep convection) are acting in the same sense in both. Accepting the limitations of this simple linear extrapolation and allowing that some of the OHC anomaly in CO2W will be due to factors other than the AMOC change, our experiments suggest that on the order of 10% of the total Atlantic heat uptake under climate change could be attributed to the declining AMOC. The accuracy of projections of local sea level rise in the Atlantic, and other quantities dependent on OHC may thus be significantly dependent on the knowing the true sensitivity of the AMOC to climate change.

Experiment CO2 has a $\sim$6x10^{24}J increase in global OHC, compared to CTR. In comparison, the freshwater perturbations used here produce at most a $\sim$0.33x10^{24}J change in global ocean heat content. Even in the Atlantic, where their influence is largest, OHC changes due to the freshwater forcing anomaly alone (the anomaly between CO2W and CO2) represent only 7% of the change due to the CO_2 forcing (the anomaly between CO2 and CTR). Based on the idealised experiments here then, we can conclude that uncertainty in changes in the hydrological forcing of the North Atlantic is not a major factor in calculations of the large scale heat budget of the ocean. Such forcings may, however, be important in developing an understanding of the behaviour of the Atlantic heat budget at the process level. The distinctive three-layer anomaly pattern seen in fig 3 should also be robust, even if the details of the cancellation of the anomalies at different levels varies for different models and climate change scenarios, and may be useful in detection and attribution studies.

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

5. Conclusions

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

Our idealised, constant ± 0.1Sv perturbations in freshwater forcing alter the heat stored in the Atlantic under a 1% per year CO_2 increase scenario by $\sim$7%. This represents an additional global uptake of approx $0.33\text{x}10^{24}$J over 250 years when the Atlantic freshwater perturbation is positive, around 5% of the global increase in ocean heat content due to the increase in CO_2.

In our experiments, freshwater perturbations cause a net increase in Atlantic heat content as deep-water formation slows and the subtropical thermocline deepens, in line with thermocline adjustment theories of the Atlantic overturning circulation. Salt perturbations produce only small net changes in ocean heat content compared to the baseline CO_2 increase scenario, as enhanced deep-water formation results in a decrease in the depth of the subtropical thermocline but the surface CO_2 warming signal is mixed more effectively to depth in the less-stratified Arctic ocean.

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

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Expt Name	CO_2 increase	Freshwater anomaly	AMOC max anomaly (Sv)	Atlantic OHC anomaly (10^{24}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	**1.71**
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	**0.12**
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	**0.01**
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	**0.10**
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	**0.06**

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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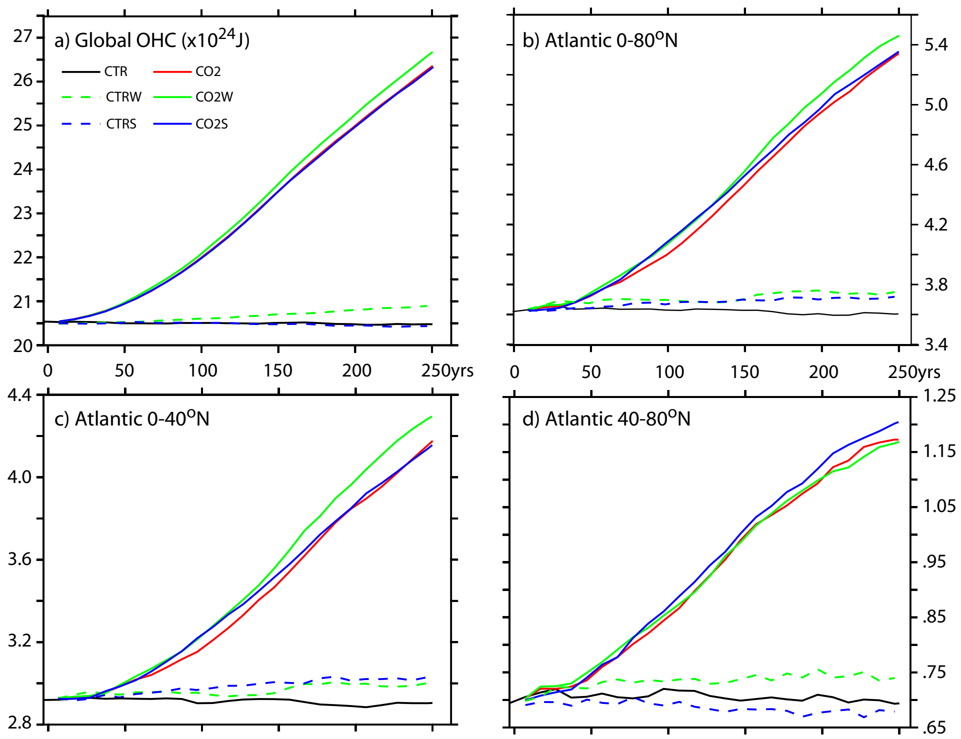

Figure 1. Evolution of depth integrated ocean heat content (x10^{24}J) for different parts of the ocean over the course of the simulations.

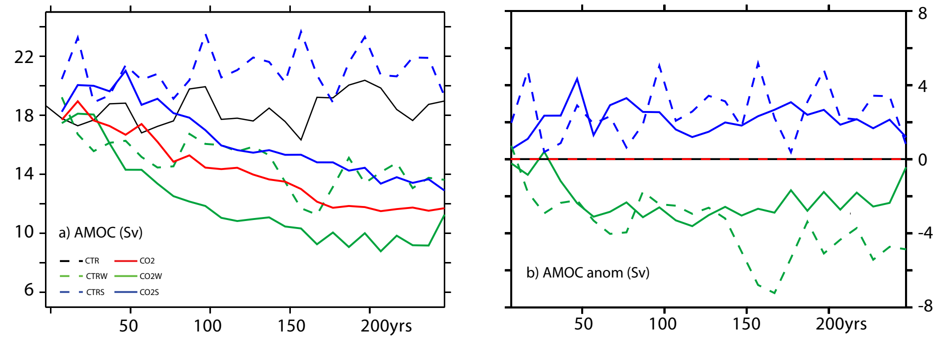

Figure 2. Evolution of the maximum strength of the Atlantic overturning over the course of the simulations. a) absolute values; b) anomalies due to the freshwater forcing. Anomalies for CO2W and CO2S are expressed with respect to CO2; anomalies for CTRS and CTRW are expressed with respect to the time-mean of their parallel CTR.

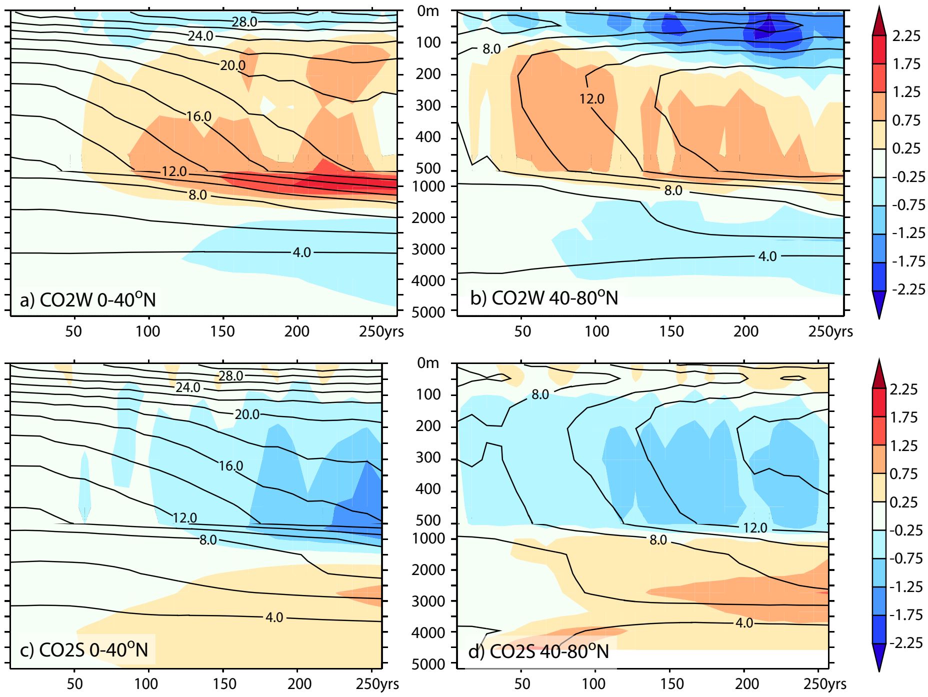

Figure 3. Evolution of horizontal average Atlantic potential temperature (°C). Contours are absolute temperature values; solid colours are anomalies relative to experiment CO2