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Tropical versus high latitude freshwater influence on the Atlantic circulation

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Abstract We investigate the model sensitivity of the Atlantic meridional overturning circulation (AMOC) to anomalous freshwater flux in the tropical and northern Atlantic. Forcing in both locations leads to the same qualitative response: a positive freshwater anomaly induces a weakening of the AMOC and a negative freshwater anomaly strengthens the AMOC. Strong differences arise in the temporal characteristics and amplitude of the response. The advection of the tropical anomaly up to the deep water formation area leads to a time delayed response compared to a northern forcing. Thus, in its transient response, the AMOC is less sensitive to a constant anomalous freshwater flux in the tropics than in the north. This difference decreases with time and practically vanishes in equilibrium with constant freshwater forcing. The equilibrium response of the AMOC shows a non-linear dependence on freshwater forcing in both locations, with a stronger sensitivity to positive freshwater forcing. As a consequence, competitive forcing in both regions is balanced when the negative forcing is about 1.5 times larger than the positive forcing. The relaxation time of the AMOC after termination of a freshwater perturbation depends significantly on the AMOC strength itself. A strong overturning exhibits a faster relaxation to its unperturbed state. By means of a set of complementary experiments (pulse-perturbations, constant and stochastic forcing) we quantify these effects and discuss the corresponding time scales and physical processes.

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1 Introduction

The Atlantic meridional overturning circulation (AMOC) plays a major role in the transport and distribution of heat in the climate system (Ganachaud and Wunsch 2003; Talley et al. 2003). In global warming scenarios, model experiments generally predict a decrease in AMOC strength (Gregory et al. 2005) mainly caused by changes in heat flux in the North Atlantic that affect the formation of deep water. Other studies suggest that the freshwater balance is strongly perturbed in case of global warming (Church et al. 2001; Winguth et al. 2005). The freshwater input in the North Atlantic in particular is expected to be enhanced as a result of enhanced atmospheric moisture transport, melting of sea ice (Johannessen et al. 2002) and of the Greenland ice sheet (Rignot and Kanagaratnam 2006) and increase in the river discharge into the Arctic (Peterson et al. 2002). These perturbations would decrease the surface density in deep convection regions, and thus reduce deep water formation and slow down the AMOC. Furthermore, some abrupt climate changes in the last glacial period are thought to be linked to disruptions of the oceanic circulation following abrupt releases of freshwater into the North Atlantic (see Rahmstorf 2002 for a review). For these reasons, a large number of model studies have focused on changes in the North Atlantic freshwater budget (recently for example Vellinga and Wood 2002; Zhang and Delworth 2005; Dahl et al. 2005; Stouffer et al. 2006).

Besides direct changes in high latitudes, several recent model studies have suggested that freshwater anomalies in the tropics can also substantially influence the AMOC (e.g. Latif et al. 2000; Thorpe et al. 2001; Vellinga et al. 2002; Vellinga and Wu 2004; Mignot and Frankignoul 2005). In all these studies, salinity or freshwater anomalies created in the tropics could be traced up to the convection sites, where they altered deep water formation and thus AMOC strength. In two of these studies (Latif et al. 2000; Mignot and Frankignoul 2005), the tropical freshwater anomalies were linked to El Niño/Southern Oscillation (ENSO) events. The latter originates in the tropical Pacific and is the strongest global inter-annual mode of coupled climate variability (Philander 1990). It is known to have an influence on the freshwater balance of the tropical Atlantic (TA; Hastenrath 1976; Kousky et al. 1984). Evidence of an enhanced transport of water vapor out of the TA into the tropical Pacific during warm ENSO (El Niño) events and a reduction during cold events (La Niña) was further documented by Schmittner et al. (2000). Thus, further changes in ENSO may have an important influence on the AMOC. A comparison of 12 coupled atmosphere–ocean Global Climate Models (AOGCMs) by Doherty and Hulme (2002) found predominantly a tendency toward more cold (La Nina like) events in a global warming scenario. On the other hand, a comparison of simulated ENSO events in 20 AOGCMs by Collins and the CMIP Modelling Groups (2005) found no significant trend toward either mean El Niño-like or La Niña-like conditions in the future. Thus, the evolution of ENSO in global warming scenarios is still under debate, and with the linkage described above, so are associated changes in AMOC.

In other models, the freshwater anomaly in the tropics was not due to an influence of ENSO but due to internal Atlantic variability (Vellinga and Wu 2004) or related to a global warming scenario (Thorpe et al. 2001). In any case, the relevance of all these modeling results indicating an influence of tropical salinity anomalies on the AMOC was emphasized by observations of Curry et al. (2003) showing an increase in upper ocean salinity in the tropical and subtropical Atlantic over recent decades.

In order to better understand the sensitivity of the AMOC to an anomalous freshwater flux in the tropics, we investigate this issue in a series of model experiments that consist in perturbing the system with freshwater anomalies either in the TA or in high latitudes. The free surface formulation of our ocean model allows us to operate with actual positive and negative freshwater fluxes rather than equivalent salt fluxes, as is usually done in ocean models using the ‘‘rigid lid’’ approximation. Recent model studies have already shown the sensitivity of the oceanic circulation to the location of the perturbation (e.g. Rahmstorf 1996; Manabe and Stouffer 1997). In these, a stronger sensitivity to freshwater forcing in the northern North Atlantic as compared to midlatitudes was found. We focus here more systematically on the mechanisms and time scales of signal transport in the Atlantic. The high computational efficiency of CLIMBER-3α allows a large number of long-term simulations, and sensitivity studies can thus be performed not only for transients but also for the equilibrium state.

After describing the model and the experimental setup (Sect. 2) we discuss the response of the AMOC to different forcing scenarios in Sects. 3, 4 and 5 and conclude in Sect. 6.

2 Model and experiments

The coupled atmosphere–ocean earth system model of intermediate complexity CLIMBER-3α is used for a series of sensitivity experiments to investigate the dependence of the AMOC strength on surface freshwater anomalies. The model consists of the oceanic general circulation model (GCM) MOM-3 coupled to an interactive atmosphere (Petoukhov et al. 2000) and a dynamic–thermodynamic sea ice module (Fichefet and Maqueda 1997). The horizontal resolution of the ocean model is $3.75 \times 3.75°$ with 24 vertical levels. A detailed description of the model can be found in Montoya et al. (2005). We use the model version with a background vertical diffusivity of 0.1×10^{-4} m^2 s^{-1}. All

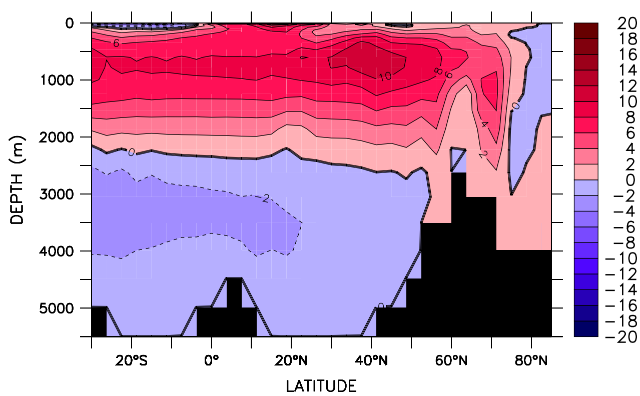


Fig. 1 Zonally integrated Atlantic overturning streamfunction in Sv for the control run

described experiments start from the same stable equilibrium simulation which was integrated for more than 5,000 years. Figure 1 gives the zonally integrated stream function in the Atlantic basin for this starting point. It is in good qualitative agreement with observations (Talley et al. 2003; Ganachaud and Wunsch 2000) in spite of a rather weak maximum value: about 12 Sv of warm water flow northward in the upper ocean, sink in the Greenland, Iceland and Norwegian (GIN) Seas and Irminger Sea, and return southward at depth (below about 700 m) as North Atlantic Deep Water (NADW). Note that there is no deep water formation in the Labrador Sea because of its narrowness in the coarse resolution model, as detailed in Montoya et al. (2005). The NADW cell reaches a depth of roughly 2,300 m, somewhat shallow compared to observations. The 4 Sv inflow of Atlantic Antarctic Bottom Water at the bottom is on the contrary relatively thick (e.g. Tally et al. 2003). In order to capture the interhemispheric flow of the AMOC, we use the maximum of the stream function at the equator between 500 and 3,000 m depth M_0 as our main diagnostic. In the unperturbed equilibrium we get $M_0^{\mathrm{ctrl}} = 8.3$ Sv (compare Fig. 1).

We design a series of experiments dedicated to compare the AMOC response to surface freshwater forcing in the TA and on the convection sites in the northern North Atlantic (*CS*). In TA, freshwater anomalies are applied in the western TA between 20°S and 20°N (blue region in Fig. 2) and are compensated by anomalies of opposite sign in the tropical Pacific (red region). Note that this forcing mimics the ENSO-associated pattern (Schmittner et al. 2000) but results will be interpreted more generally as the effect of tropical freshwater anomalies. In CS, freshwater forcing is applied to the convection sites in the GIN Seas (green region in Fig. 2) which corresponds to the classical 'water hosing experiments' as carried out for example by Vellinga and Wood (2002), Dahl et al. (2005) and Stouffer et al. (2006). In order to be comparable with TA, the forcing is compensated in the same area in the tropical Pacific (red region in Fig. 2). The two forcing locations are used in the following model experiments designed to document different time characteristics:

1. Long-term responses are investigated by applying a constant forcing with different amplitudes (Sect. 3).
2. Internal time scales of the meridional transport are extracted through stochastic forcing (Sect. 4).
3. The characteristic decay time for buoyancy anomalies in the model are finally obtained from a series of pulse forcing experiments (Sect. 5).

3 Response of the AMOC to constant freshwater forcing

In order to simulate long-term climate shifts, a scenario with constant forcing applied respectively in TA and CS regions has been integrated for more than 1,200 years until a new equilibrium state was reached (Fig. 3). Eight different amplitudes of freshwater anomaly are used (± 0.05, ± 0.1, ± 0.2 and ± 0.4 Sv).

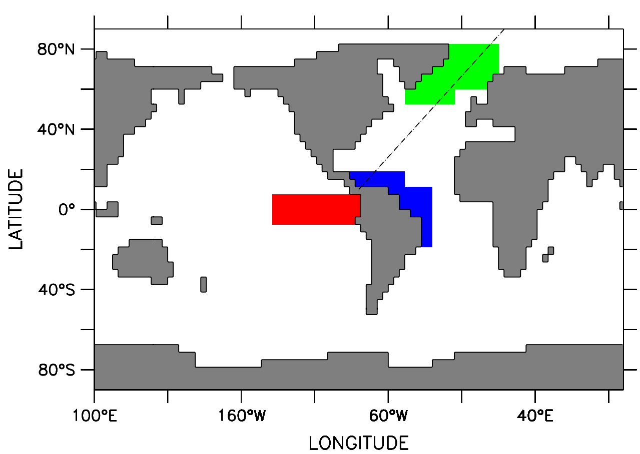


Fig. 2 Forcing patterns for model experiments: balanced freshwater exchange between the GIN Seas convection sites (CS, *green region*) and the tropical Pacific (*red region*). Balanced freshwater exchange between the tropical Atlantic (TA, *blue region*) and Pacific (*red region*). The *dashed line* marks a section for the salinity profile in Fig. 7

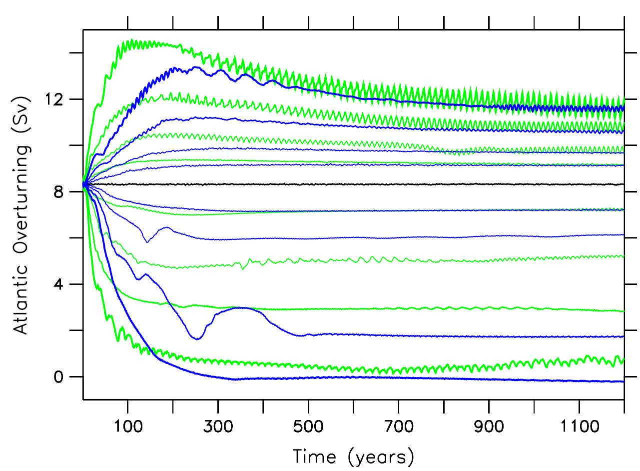


Fig. 3 AMOC response to a constant freshwater forcing on the convection sites (*green*) and in the TA (*blue*). The forcing pattern is scaled to integral amplitudes of freshwater forcing in the Atlantic of $F = \pm 0.4, \pm 0.2, \pm 0.1, \pm 0.05$ Sv. The equilibrium run for $F = 0$ Sv is given as *black curve*

For TA, positive (negative) forcing could be interpreted as shifts toward constant La Niña-like (El Niño-like) conditions. Note that the freshwater anomaly induced in the tropical Atlantic by the El Niño of 1983 was in the order of 0.15 Sv. For CS, the anomalous forcing can be interpreted as a permanent increase in local runoff, precipitation or Greenland melting. Anomalous freshwater flux from Greenland melting can become important on centennial timescales. Coupled climate–land–ice models predict that half of the Greenland ice sheet could disappear by the year 2500 (e.g. Huybrechts and de Wolde 1999), corresponding to an average increase in freshwater flux of 0.1 Sv. In light of recent observations of an acceleration of the melting (Rignot and Kanagaratnam 2006) this number might well be exceeded. Amplitudes of 0.2 and 0.4 Sv represent strong forcings for both regions. In the present study the latter are conducted in order to investigate the whole range of the AMOC responses.

3.1 Transient response

In our experiments, a change in the AMOC (and thus in M_0 in Fig. 3) is apparent after about 10 years of anomalous forcing, with reduced overturning when freshwater is added to the Atlantic and an increase in M_0 for a negative freshwater forcing. For both regions, stronger freshwater forcing leads to stronger responses of M_0. After 100–300 years an extremum is reached and the maximum overturning relaxes to its new equilibrium value.

The mechanism by which freshwater anomalies over the convection sites affect the AMOC is well known: alteration of the surface density in this region inhibits the deep water formation and reduces the north–south density gradient (Rahmstorf 1995; Levermann and Griesel 2004; Griesel and Morales-Maqueda 2006) thereby weakening the AMOC. In case of tropical freshwater forcing, the anomalies first have to be transported to the northern North Atlantic before they can affect the deep water formation there. The transient response of M_0 to the CS forcing within the first 100 years is therefore faster and steeper compared to TA, in agreement with the results of Manabe and Stouffer (1997).

For negative and very weak positive freshwater amplitudes, the responses of TA and CS converge on centennial timescales. In case of strong positive freshwater forcing, differences between the two regions remain for longer times and wear off on millennial timescales (not shown). We will discuss this long-term response in more detail in Sect. 3.3.

3.2 Transport mechanism

In order to illustrate the mechanism of tropical freshwater anomaly impact on the AMOC, we use Lagrangian particles, released at the surface at the same location as the TA freshwater anomaly and follow them along their path for 20 years in the control run. We show these trajectories in the upper panels of Fig. 4. Consistent with other modeling studies (e.g. Mignot and Frankignoul 2005) only a small fraction of the particles takes a direct surface route along the Gulf Stream and its extension, the North Atlantic Current. The majority circulate at least once in the subtropical gyre before crossing the Atlantic at levels of 100–200 m depth. Particles reach the high latitudes in subsurface (about 200–500 m depth).

The general picture of the trajectories remains the same during the whole course of experiment TA as the AMOC strengthens or weakens (not shown). The main difference as compared to the control run is the speed of the transport and thus the amount of particles that reach the convection sites. In order to quantify this, we released the same amount of particles as in Fig. 4 at the end of the TA experiments using F_{TA}=–0.2 Sv and F_{TA}=+0.2 Sv, respectively. Within 20 years, the stronger AMOC (F_{TA}=–0.2 Sv) transports 30% more particles than the control run north of 60°N while the weaker AMOC (F_{TA}=+0.2 Sv) transports 60% less. This has strong impacts on the sensitivity of the AMOC to TA freshwater perturbations as opposed to perturbations applied in the high latitudes and explains differences in the transient response discussed in Sect. 3.1.

Note that the time scale of the northward transport can be expected to be model-dependent and to vary with changing parameters, as coarse models typically underestimate current velocities. However, the results from our model are in good agreement with results from the Bergen Climate Model (Furevik et al. 2003), a higher resolution state-of-the-art coupled ocean–atmosphere GCM (Fig. 4, lower panel).

To complement the Lagrangian trajectories analysis, we explore the transport of passive Eulerian tracers in the control run. The latter are advected and diffused, similar to the freshwater anomalies, but neither impact the density nor the circulation. The distribution of the passive tracer content (gray shade), integrated over the upper 450 m, 20 years after the release in the TA forcing region (TA) is given in Fig. 4 (upper left). A zonal average of the depth distribution is given in the upper right. The tracer released in the tropics reaches the convection sites (north of 60°N) with a time delay and exceeds 5% of the amount added every month in

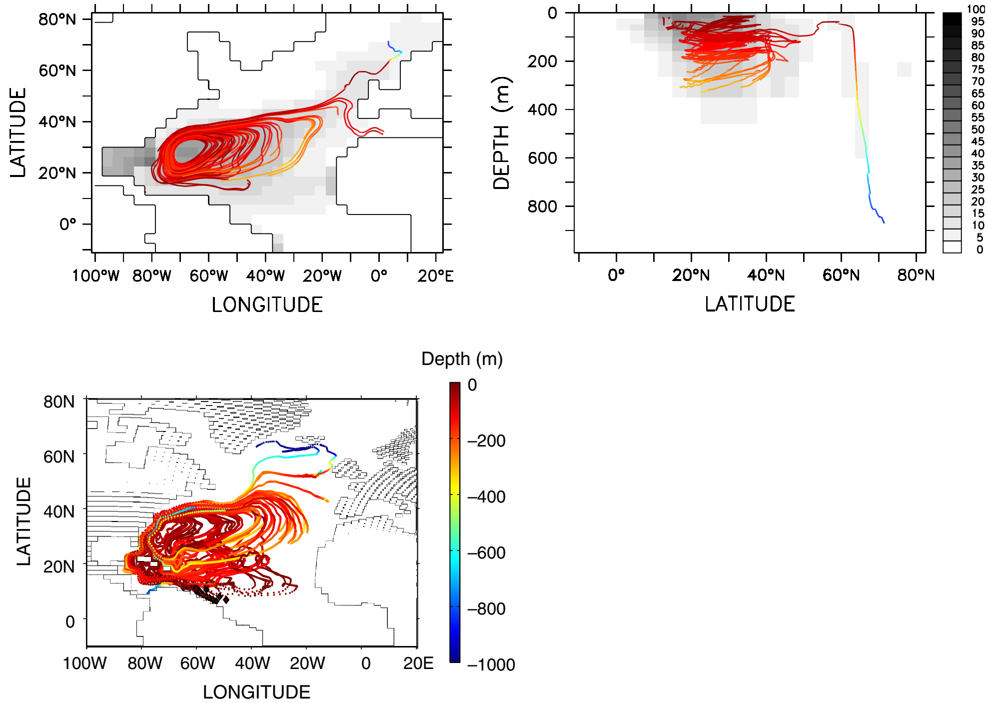


Fig. 4 Particle trajectories after 20 years of control integration in CLIMBER-3α (*top*, left and right) and in the Bergen Climate Model (*bottom*). The *color code* represents the trajectory depth in meters. The particles, released in the surface at the Caribbean Sea, circulate in the subtropical gyre and subduct to deeper levels before crossing the Atlantic and eventually reaching the North Atlantic deep water formation regions. The concentration of Eulerian tracers after 20 years is displayed in the two upper panels (*grey shade*, arbitrary units). The lower panel is a slightly modified version of Fig. 21 in Mignot and Frankignoul (2005)

the tropics after 5 years (Fig. 5). This is thus the minimal time needed for an anomaly created in the tropical region to reach the deep water formation and constitutes a lower limit for the timescale of the response of the large scale overturning to the tropical forcing. Note that unlike Lagrangian tracers, Eulerian tracers are influenced not only by advection but also by small scale processes such as mixing and diffusion, so that they give an indication on the propagation timescale of salinity anomalies in the model. Yet, the correspondence of grey shadings on Fig. 4 with the Lagrangian trajectories suggest that advection is the primary transport mechanism for freshwater anomalies from the tropics to high latitudes.

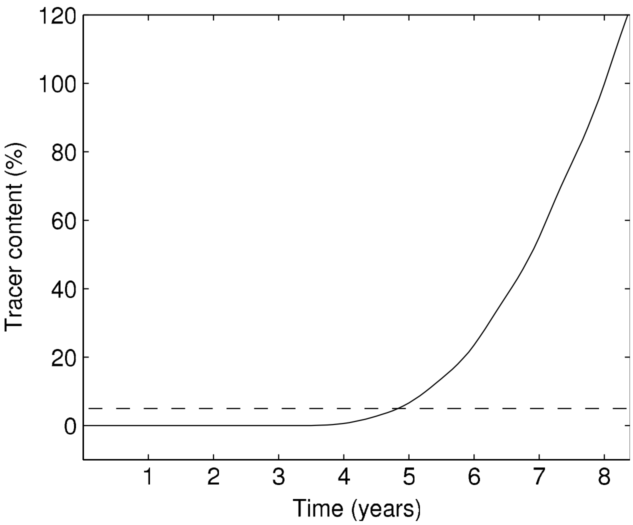


Fig. 5 Passive Eulerian tracer content north of 60°N in the control run. After about 5 years, the tracer content notedly rises and exceeds 5% of the amount added every month in the tropics

3.3 Long-term response

We now compare the amplitude of the long-term response of the AMOC (M_0^{1000}, year 1000 in Fig. 3) for the different constant forcing scenarios to the amplitude of the short-term response after 30 years (M_0^{30}). Figure 6 shows the values of M_0^{30} (left) and M_0^{1000} (right) as a function of the forcing amplitude F. The short-term response M_0^{30} shows the results discussed above for the transient case: for both forcing regions, the overturning M_0^{30} decreases with increasing F, as additional freshwater reduces the deep water formation in the North Atlantic. A higher sensitivity of the AMOC to the CS forcing is visible as a steeper slope of $M_0(F_{CS})$ (green crosses) compared to the TA case (blue crosses).

Differences in sensitivity of the AMOC to the two forcing regions, decrease toward the equilibrium. The long-term response M_0^{1000} (Fig. 6, right) decreases with increase in F for both forcing regions. For $F < 0.2$ Sv, the sensitivity of M_0^{1000} to freshwater forcing increases with increase in F, evident in a steepening slope of M_0^{1000} toward higher values of F. This non-linear dependence can be understood by the salt advection

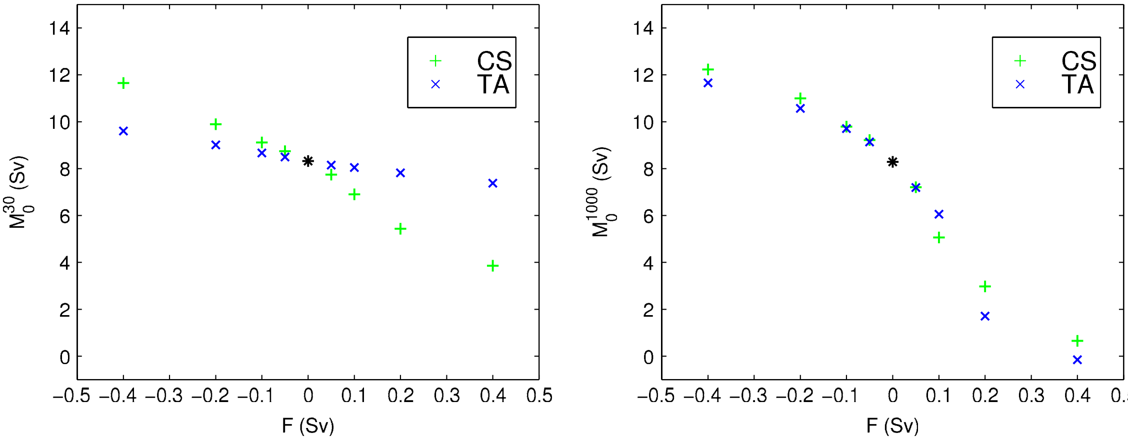


Fig. 6 *Left*: maximum values of the Atlantic overturning M_0 for constant freshwater forcing F in the northern North Atlantic (*green*) and in the TA (*blue*) after 30 years of integration. *Right*: same as left panel but after 1,000 years of integration. Compare Fig. 3 for time series. The figure illustrates that the transient response on multi-decadal time scales is linear in the freshwater forcing, while the near-equilibrium is non-linear

feedback first described by Stommel (1961) and further investigated by e.g. Rahmstorf (1996): in equilibrium, stronger freshwater input to the North Atlantic weakens the overturning which subsequently inhibits the evacuation of the freshwater anomaly, acting as a positive feedback. Conversely, freshwater export out of the North Atlantic enhances the overturning and thus speeds up the removal of the anomaly itself. This feedback mechanism is qualitatively and quantitatively the same whether the constant anomaly is prescribed in the TA or in the North Atlantic itself, because in equilibrium the anomaly has filled the transport channel from the tropics to the North Atlantic and a constant amount is supplied there.

For high freshwater forcings $F \geq 0.2$ Sv we find higher (but still weak) values of M_0^{1000} in the CS case in the long-term response. In this regime, the residual overturning circulation is different in CS and TA because of different density structures in the Atlantic. A comparison of North Atlantic salinity cross-sections for the two cases (Fig. 7) for $F = 0.4$ Sv at year 1000 shows that only the upper 150 m of the water column north of 45 °N are lighter in CS compared to TA. The rest of the basin is more efficiently freshened in the TA experiment. The CS forcing produces a capping of the north Atlantic region with freshwater, which persists throughout the run. This capping inhibits an efficient ventilation of the deep ocean, leaving it more saline than in the TA case (Mignot et al. 2006). The result is a reversal of the sensitivity between CS and TA for strong freshwater forcings.

Our results concerning the long-term response are qualitatively consistent with Rahmstorf (1996) for negative and weak positive freshwater amplitudes. For strong net freshwater inputs, however, his model shows a steep breakdown of the AMOC that does not recover in response to a reduction of the forcing (hysteresis effect). In CLIMBER-3α, the AMOC is only suppressed but recovers when reducing the forcing. No substantial hysteresis effect is found. Indeed, the CLIMBER-3α model does not exhibit a stable off-state or multiple equilibrium states of the AMOC in any freshwater induced regime (Levermann et al. 2005; Mignot et al. 2006).

3.4 Competitive freshwater forcing

One possible effect of global warming is the intensification of the atmospheric meridional moisture transport (Rahmstorf and Ganopolski 1999; Latif et al.

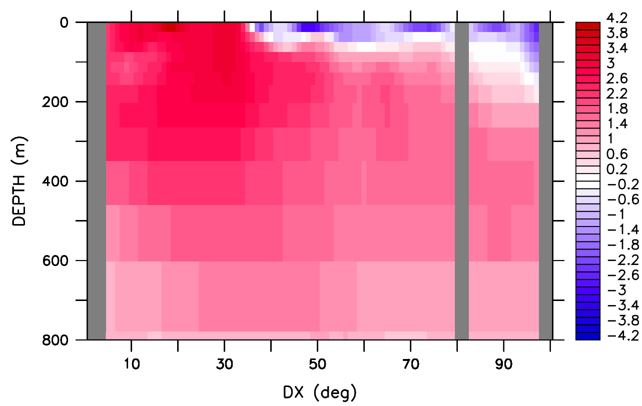


Fig. 7 North Atlantic salinity difference cross section *CS–TA* between (80°W, 10°N) and (20°E, 90°N) for $F = 0.4$ Sv at t = 1,000 years. A capping of the upper 150 m north of 45°N is apparent for the *CS* case. The section is marked as *dashed line* in Fig. 2

2000) that would result in an increase of the freshwater input into the northern North Atlantic and reduction in the tropical areas. In order to mimic this effect, we applied an anomalous forcing simultaneously in TA and CS regions, with opposite signs. The aim is to study how the freshening of the northern North Atlantic is balanced by a negative freshwater forcing in the tropics. The amplitude of the tropical forcing is fixed to F_{TA} = –0.1 Sv and the ratio $R = -F_{CS}/F_{TA}$ is varied between 0.125 and 2 so that F_{CS} = 0.0125, 0.025, 0.05, 0.066, 0.1, 0.2 Sv, respectively. The values of M_0^{1000} are given in Fig. 8 as circles. For decreasing R, M_0^{1000} approaches the value for F_{TA} = – 0.1 Sv of the former experiment (cross), as a consequence of a decreasing influence of the CS forcing. The forcing with equal amplitudes (R = 1) yields M_0^{1000} close to the value for F_{CS} = 0.05 Sv of the former experiment, consistent with a stronger sensitivity of the AMOC to the positive freshwater forcing. The influence of both forcings on M_0^{1000} is balanced for $R = \frac{2}{3}$, which underlines the nonlinear nature of the sensitivity of the AMOC to freshwater forcing and the stronger sensitivity to positive freshwater forcing.

4 Time scales of signal propagation

In order to examine internal time scales of the AMOC, the model is forced with a 1,000-years signal of yearly random values. Six sequences with Gaussian amplitude

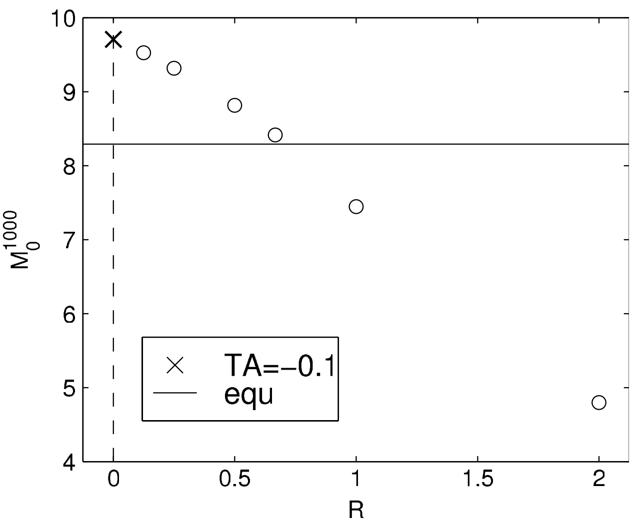


Fig. 8 Competitive freshwater forcing of opposed sign in the TA and in the northern North Atlantic. The *circles* give M_0^{1000} as a function of the ratio $R = -F_{CS}/F_{TA}$ with F_{TA} = – 0.1 Sv in all cases. The *black cross marks* the value for R = 0 i.e. F_{CS} = 0.0 Sv. The impact of the freshwater anomalies is balanced for $R = \frac{2}{3}$ where $M_0(R)$ crosses the equilibrium value of M_0 (*black line*) of the former experiment

distribution and zero mean are generated and used as forcing amplitudes for the spatial patterns CS and TA, respectively. We analyze the average cross-correlation between forcing and response of the AMOC of these six runs in order to increase the signal-to-noise ratio. The forcing patterns are scaled so as to exhibit a temporal variance of 0.1 Sv. The response of the AMOC to this white noise has a red noise frequency characteristic for both forcing regions (not shown), consistent with the integrating properties of the ocean (e.g. Frankignoul 1985; Deshayes and Frankignoul 2005).

The cross-correlation between forcing (CS and TA) and overturning maximum at 60°N (M_{60}) displayed in the left panel of Fig. 9, has a strong and distinct peak at T_{CS} = 2 years and T_{TA} = 17 years time lag, respectively. This corresponds to the time needed for changes in the surface forcing to alter the overturning in the North Atlantic. The time difference of 15 years between T_{CS} and T_{TA} is due to the transport of anomalies from the tropics to the North Atlantic for case TA. It should be noted that the general model dependence of current velocities, mentioned in Sect. 3.2, does affect this time lag. Furthermore, a correlation analysis is a linear transformation and thus has limited applicability to nonlinear systems. In this sense, the time scales found above have to be considered as averages between the different responses to positive and negative forcing which will be discussed in Sect. 5.

The right panel of Fig. 9 displays the cross-correlation of the stochastic forcing sequence and the maximum of the overturning stream function measured at different latitudes for experiment CS. The maximum correlation is shifted toward larger lags with decreasing latitude. Thus, the circulation anomaly originates near the deep water formation sites and is propagating toward the south. This enables us to deduce an approximate signal travel time of circulation adjustment at depth from the convection sites to the equator of about 4 years and a corresponding signal propagation speed of ~5 cm/s. This time scale is consistent with observations of Curry et al. (1998) and it is close to the 6 years found by Mignot and Frankignoul (2005) in the Bergen Climate Model. Furthermore, it supports that the adjustment is primarily caused by coastally trapped waves, consistent with observations of Yang (1999) and theoretical studies of Johnson and Marshall (2002, 2004). Dong and Sutton (2002) find a similar adjustment speed in the temperature signal after the collapse of the AMOC as the response to a strong freshwater anomaly in the North Atlantic. For forcing in the tropics we can find similar propagation speeds from 60°N to 20°N, though the picture is complicated by interac-

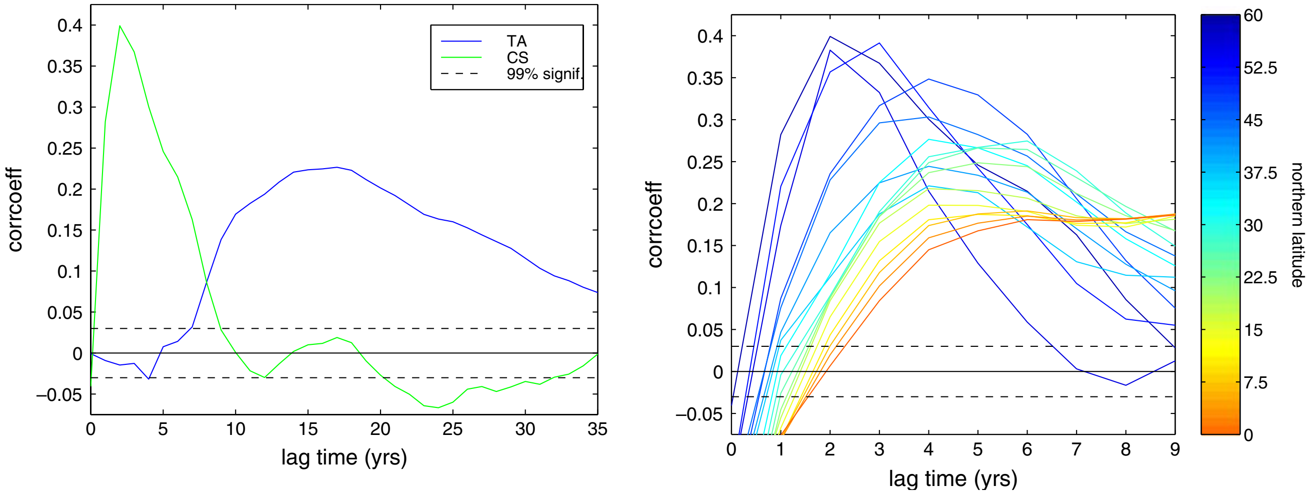


Fig. 9 *Left*: average cross-correlation of the stochastic forcing and the maximum overturning at 60°N (M_{60}) for the two experiments CS (*green*) and TA (*blue*). M_{60} lags for positive time lags. *Right*: average cross-correlation of the stochastic forcing and maximum overturning at different latitudes between 0°N and 60°N for the CS case. The *color scale* on the left gives the northern latitude at which the overturning maximum is considered. The 99% significance level for one-tailed tests is given as *red lines*

tions with the northward propagating freshwater anomalies and the circulation of these anomalies in the subtropical gyre (not shown).

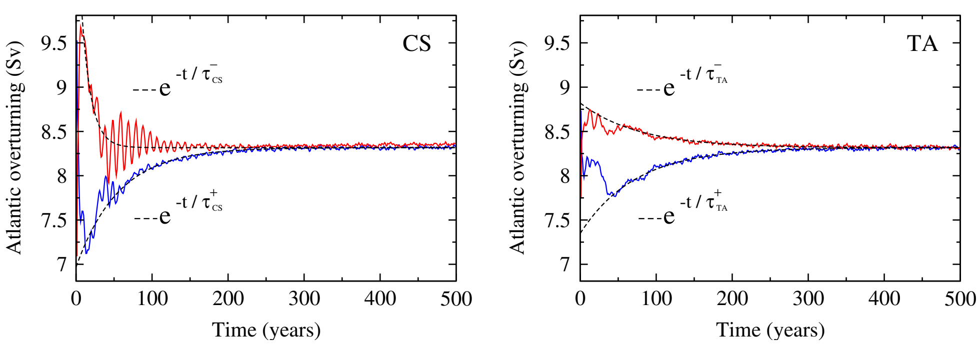
5 Damping of freshwater anomalies

In order to quantify the damping in the system, we finally investigate the response of the AMOC to anomalous pulses of freshwater. Both positive and negative freshwater anomalies are applied for 1 year with amplitudes of 2 Sv. The main feature of the response of the overturning to a pulse of negative (positive) freshwater anomalies is a steep increase (decrease) due to the impact of the anomalies on the deep water formation as described above (Fig. 10). The response then decays exponentially, with a time scale corresponding to the time needed to disperse the anomaly in the system.

Both for CS and TA, the responses to positive and negative anomalies are not symmetric: for a positive freshwater anomaly, the extremum is reached at a later time than in the opposite experiment with negative flux. As mentioned in Sect. 3.1, this is due to the freshwater anomaly slowing down the AMOC and partly inhibiting further advection of the anomaly in the system, which leads to a delay of the effect (salt advection feedback). In the opposite case, once the negative freshwater anomaly reaches the North Atlantic, it enhances the overturning and speeds up the transport, so that the maximum is reached earlier. This behavior is robust to different durations and amplitudes of the forcing, as long as the overturning strength is effectively changed by the imposed anomaly.

Differences between CS and TA can be found on top of this general asymmetry. The initial response, indicated by the first maximum amplitude, is stronger for CS, consistent with a stronger sensitivity to the

Fig. 10 Response to a 1-year freshwater pulse (*blue*) and freshwater extraction (*red*) of amplitude 2 Sv for the experiments CS (*left*) and TA (*right*). The decay time constants are τ^-_{CS} = 13 years, τ^+_{CS} = 56 years for the CS case and τ^+_{TA} = τ^-_{TA} = 75 years for TA



forcing (Sect. 3.3). Furthermore, in the CS case, the experiment with negative freshwater anomaly forcing has a faster exponential decay (τ_{CS}^{-} = 13 years) than the corresponding experiment with positive forcing (τ_{CS}^{+} = 56 years). This asymmetry is due to the salt advection feedback: when the overturning is reduced, it takes longer to evacuate the freshwater anomaly from the convection sites. In the TA case, this asymmetry vanishes and the exponential decay time $\tau_{TA}^{\pm}$ = 75 years for both positive and negative forcing is longer than in the CS case. This can be explained by the fact that the anomalies reach the northern North Atlantic with a time delay, leading to a smoothed, time-stretched response compared to the CS case. The smaller amplitude of the response in the TA case prevents an asymmetry because the anomaly dispersion time scale itself is a function of the overturning strength. A vanishing of the asymmetry can therefore also be expected for low amplitudes of the imposed freshwater anomalies in the CS case, if the influence on the overturning strength of the latter becomes negligible.

6 Conclusions

A series of experiments was performed with the coupled atmosphere–ocean climate model of intermediate complexity CLIMBER-3α in order to analyze and compare the response of the AMOC to surface freshwater forcing in the tropical Atlantic on the one hand and in the northern North Atlantic on the other hand. For both forcing regions, an anomalous input of freshwater to the Atlantic leads to a weaker AMOC, while a freshwater anomaly of negative sign strengthens the circulation.

Freshwater forcing on the convection sites of the northern North Atlantic directly influences the deep water formation in this region, which induces a fast response of the AMOC. A time scale of 2 years was found for this response by correlation analysis. The overturning circulation in the rest of the Atlantic basin adjusts as the anomalous circulation signal propagates southward, reaching the equator within about 4 years. For the TA forcing, the anomaly propagates up to the deep convection areas, where it influences the AMOC. Such a mechanism was found in several other model studies (Thorpe et al. 2001; Vellinga et al. 2002; Vellinga and Wu 2004; Mignot and Frankignoul 2005). The northward path was shown to be advective and to consist mainly of a circulation in the northern subtropical gyre, a subduction and a subsequent north-east passage at levels of 100–200 m depth. Freshwater anomalies applied in the tropics thus reach the deep convection sites at mid-depth (200–500 m) and influence the AMOC in the North Atlantic after about 15 years.

The different influence of freshwater anomalies applied in the two regions causes differences in sensitivity and temporal characteristics of the transient response of the AMOC. In our simulations a weaker sensitivity of the AMOC to the tropical forcing is evident with a lagged response compared to a North Atlantic freshwater forcing, in agreement with Manabe and Stouffer (1997). The difference in overturning strength between the two cases decreases toward the long-term equilibrium of constant anomalous freshwater forcing. This effect is qualitatively and quantitatively similar in both experiments and can be explained by the salt advection feedback (Stommel 1961). While the dependence of the AMOC strength on freshwater input is linear in the transient, a non-linearity is found in near-equilibrium.

Results from freshwater pulse forcing experiments support the difference in sensitivity described above. The evacuation of the freshwater anomalies from the region where they influence the AMOC is accelerated or inhibited dependent on the AMOC strength. Differences in relaxation behavior between the experiments with positive and negative forcing therefore arise if the forcing amplitude is high enough to change the AMOC strength significantly.

In our model, a direct competition of the two forcings with opposite sign is balanced in equilibrium, when the negative freshwater forcing in the tropics is about 1.5 times larger than the positive forcing in the northern North Atlantic. This is a direct consequence of the non-linearity of the equilibrium response which describes a weaker sensitivity of the AMOC for smaller amplitudes of the freshwater forcing in both regions.

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