

Effects of fluctuating daily surface fluxes on the time-mean oceanic circulation

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Abstract

The effect of fluctuating daily surface fluxes on the time-mean oceanic circu-16 lation is studied using an empirical flux model. The model produces fluctuating 17 fluxes resulting from atmospheric variability and includes oceanic feedbacks on 18 the fluxes. Numerical experiments were carried out by driving an ocean general 19 circulation model with three different versions of the empirical model. It is 20 found that fluctuating daily fluxes lead to an increase in the Meridional Over-21 turning Circulation (MOC) of the Atlantic of about 1 Sv and a decrease in the 22 Antarctic Circumpolar Current (ACC) of about 32 Sv. The changes are ap-23 proximately 7% of the MOC and 16% of the ACC obtained without fluctuating 24 daily fluxes. 25

The fluctuating fluxes change the intensity and the depth of vertical mixing. 26 This, in turn, changes the density field and thus the circulation. Fluctuating 27 buoyancy fluxes change the vertical mixing in a non-linear way: They tend 28 to increase the convective mixing in mostly stable regions and to decrease the 29 convective mixing in mostly unstable regions. The ACC changes are related to 30 the enhanced mixing in the subtropical and the mid-latitude Southern Ocean 31 and reduced mixing in the high-latitude Southern Ocean. The enhanced mixing 32 is related to an increase in the frequency and the depth of convective events. 33 As these events bring more dense water downward, the mixing changes lead 34 to a reduction in meridional gradient of the depth-integrated density in the 35 Southern Ocean and hence the strength of the ACC. The MOC changes are 36 related to more subtle density changes. It is found that the vertical mixing in a 37 latitudinal strip in the northern North Atlantic is more strongly enhanced due 38 to fluctuating fluxes than the mixing in a latitudinal strip in the South Atlantic. 39

- 40 This leads to an increase in the density difference between the two strips, which
- 41 can be responsible for the increase in the Atlantic MOC.
- 42 Keywords : Fluctuating daily fluxes, Vertical mixing, Meridional Overturning Cir-
- 43 culation, Antarctic Circumpolar Current, Air-Sea interaction.

44 1 Introduction

The issue of what determines the strength of the global Meridional Overturning 45 Circulation (MOC) has drawn the attention of many researchers. The prevailing 46 view is that the circulation is driven partly by the diapycnal mixing of heat that 47 lightens water masses in the deep ocean and causes them to rise uniformly in low 48 latitudes (Munk and Wunsch 1998), and partly by wind-driven upwelling induced by 49 the strong westerly circumpolar winds in the Southern Ocean (Webb and Suginohara 50 2001, Toggweiler and Samuels 1995). Both the diapycnal mixing and the wind-driven 51 upwelling focus on the mechanisms that allow deep dense water masses to return to 52 the surface. The surface buoyancy forcing, though not considered as a driver of the 53 MOC capable for providing energy supply, is necessary for setting up the flow by 54 controlling the rate and site of the deep water formation (Kuhlbrodt et al. 2007). 55

The major factors which control the MOC in the above picture are the diapycnal 56 mixing, the upwelling due to wind forcing and the rate and the site of deep water 57 formation set up by the surface buoyancy forcing. All these factors are directly or 58 indirectly related to the air-sea fluxes. So far, the analyses have mainly focused on 59 the effects of *climatological mean* components of the wind forcing in providing the 60 energy required for diapycnal mixing or in inducing wind-driven upwelling (Munk 61 and Wunsch 1998, Webb and Suginohara 2001, Toggweiler and Samuels 1995). This 62 paper aims at a detailed picture that can isolate the effect of fluctuating day-to-day 63 *fluxes* from that of the mean fluxes. 64

65

Generally, the role of air-sea fluxes in determining the stratification and the circu-

lation of the oceans has been known for long time. Such a role has been investigated 66 within theoretical frameworks (Walin 1982 and Tziperman 1986) and with respect to 67 change in convection (Rahmstorf 1995, Kuhlbrodt and Monahan 2003, Swingedouw 68 et al. 2007). Walin (1982) studied the relation between sea-surface heat flux and 69 thermal circulation in the ocean. Tziperman (1986) derived a relation between the 70 interior stratification and the air-sea heat fluxes and used this relation to study the 71 buoyancy driven circulation. The role of surface flux anomalies in triggering convec-72 tion was studied by Rahmstorf (1995). Using a simple box model, Kuhlbrodt and 73 Monahan (2003) showed that the variability of surface fluxes is important for the 74 open ocean convection and deep water formation in the Labrador Sea. Swingedouw 75 et al. (2007) found a linear relationship between density changes in the convection 76 sites and the strength of the Atlantic MOC. 77

Even though the previous studies support the important role of day-to-day 78 anomalies of air-sea fluxes, it is generally difficult to obtain a quantitative esti-79 mation of the impact of all fluctuating fluxes on the MOC in the framework of 80 GCMs. For instance, it is obvious that an evaporation anomaly can lead to the 81 formation of water denser than 1028 kg/m^3 , while a precipitation anomaly can lead 82 to the formation of water lighter than 1028 kg/m^3 . With these anomalies, water 83 mass production denser than 1028 kg/m^3 can occur. Without these anomalies, but 84 with the same time-mean buoyancy forcing, the water mass production denser than 85 1028 kg/m^3 would have been zero. However, what is less clear is the net effect of all 86 buoyancy anomalies on the oceanic circulation. 87

The effect of fluctuating fluxes can be strongly non-linear. For example, consider buoyancy anomalies occuring in a mostly stable region. In this case, positive anomalies (e.g. due to a precipitation event or a downward positive heat flux anomaly) may not significantly affect the statistics of convective events (since the water column is already stable), whereas negative anomalies (e.g. due to an evaporation event or a negative heat flux anomaly) could significantly increase convective events, resulting in non-linear responses to fluctuating fluxes.

Given the potential and complexity of daily air-sea fluxes in changing the water 95 mass production and from that the interior stratification and circulation, the effect 96 of daily fluxes is investigated using a coupled system, specially developed for this 97 purpose. The system consists of an ocean GCM and an empirical global flux model 98 which describes the day-to-day flux variations in a realistic manner. The advantage 99 of this system is that it allows a separation of effects of fluctuating air-sea fluxes 100 from that of the climatological mean fluxes. Such a separation is difficult within a 101 fully coupled atmosphere and ocean GCM. Numerical experiments were carried out 102 using the hybrid coupled model. As will be shown, the fluctuating daily fluxes affect 103 not only the Atlantic MOC, but also the Antarctic Circumpolar Current (ACC). 104 The models and the numerical experiments are described in Section 2 and 3. The 105 results of the experiments are presented in Section 4. Discussion and conclusions 106 are given in the final section. 107

108 2 Model Description

The empirical flux model and the OGCM used in this study are briefly introduced below. A more detailed description can be found in von Storch et al. (2005) and Marsland et al. (2003).

¹¹² 2.1 The empirical model of daily air-sea fluxes: EMAD

The flux model, referred to as EMAD (Empirical Model of Atmospheric Dynamics), is designed to generate air-sea flux anomalies relative to given climatological mean fluxes. Based on the assumption that deviations from a given mean state of the coupled system are small, the dynamics of the flux anomalies and the response of these fluxes to anomalous sea surface condition are considered to be linear and described by

$$\mathbf{x}_{t+1}^{'} = \mathcal{A}\mathbf{x}_{t}^{'} + \mathcal{C}\mathbf{n}_{t+1} + \mathcal{B}\mathbf{y}_{t}^{'}.$$
 (1)

 \mathbf{x}' comprises anomalies of all fluxes required to drive the OGCM. These are the net 119 heat flux, the zonal and meridional momentum flux, the freshwater flux, the short-120 wave radiation which penetrates into the sea water, and the conductive and residual 121 heat flux required to describe the sea ice formation and depletion. \mathbf{y}' represents 122 anomalies of oceanic variables at the sea surface, such as the SST, the sea ice cover 123 and the sea ice thickness, that can affect the fluxes. \mathbf{n} is a multivariate white noise 124 with zero mean and unit variance. \mathcal{A} describes the linear dynamics of the fluxes, 125 $\mathcal B$ the linear response of fluxes to the ocean surface condition, and $\mathcal C$ the covariance 126 structure of the residual that is not depicted by \mathcal{A} and \mathcal{B} . The time step of Eq.(1) 127

128 is one day.

 \mathcal{A}, \mathcal{B} and \mathcal{C} are matrices obtained by fitting Eq.(1) on to the daily output of a 129 200-year control integration performed with the fully coupled ECHO-G Atmosphere 130 Ocean General Circulation Model (AOGCM) (Legutke and Voss 1999; Raible et 131 al. 2001). The fitting is done first for \mathcal{A} and \mathcal{B} in EOF-spaces represented by the 132 leading EOFs of fluxes and then for \mathcal{C} in the physical space. We use 100 EOFs for 133 each flux and 100 EOFs for SST in the water module (further explained below) and 134 50 EOFs for each of the same variables in the ice module. The physical space has 135 the Gaussian grid of T30 resolution. 136

The EMAD model consists of a water module and an ice module for the separate 137 treatment of the fluxes over permanently open water and sea surface where ice can be 138 formed. The formulation of the two modules is necessary to deal with the additional 139 fluxes which are required to drive the sea-ice model. Different from the water module, 140 the state vector \mathbf{x}' in the ice module contains the conductive and residual heat and 141 distinguishes the fluxes of net heat, fresh water and momentum over ice and water. 142 Depending on the sea ice fraction within a grid cell, either the fluxes over ice or the 143 fluxes over water or both will be used to drive the ocean. 144

¹⁴⁵ Without the last term, the model equation (1) mimics the linear dynamics of flux ¹⁴⁶ anomalies driven by the atmospheric variability. The last term with \mathcal{B} describes the ¹⁴⁷ oceanic feedbacks on the fluxes. Since \mathcal{B} is derived from a coupled model integration ¹⁴⁸ which is essentially statistically stationary, the interaction described by $\mathcal{B}\mathbf{y}'$ acts to ¹⁴⁹ keep the ocean in the given mean state. This means that, if the sea surface condition is moved away from the given sea surface state, there would be non-zero anomalies of sea surface variables and from that a non-zero \mathbf{y}' , which generates anomalous fluxes $\mathcal{B}\mathbf{y}'$ that drive the sea surface back to the given state.

The just described feedback mechanism involves essentially the interaction be-153 tween the heat flux and the SST anomalies. It functions in a way as if the heat flux 154 is described by a restoring condition. However, in contrast to the traditional restor-155 ing formulation which uses a constant restoration time, \mathcal{B} implies a dependence of 156 the restoration time on spatial scales. By formulating in EOF-space, \mathcal{B} captures the 157 restoring time scales for modes with different spatial scales in the ECHO-G integra-158 tion. In particular, large-scale SST anomalies are allowed to exist over a longer time 159 period, while small-scale SST anomalies will be damped out quickly. The need for 160 such a scale-dependent restoration was first pointed out by Rahmstorf and Wille-161 brand (1995). The present formulation can be considered as an empirical approach 162 that captures the scale-dependent feedback of SST on heat flux in the ECHO-G inte-163 gration. Due to the scale dependence, the $\mathcal{B}y'$ -term does not act as a rigid restoring. 164 One does not obtain exactly the same SST when using $\mathcal{B}\mathbf{y}'$ to nudge SSTs to the 165 same climatological mean SST (see Section 4.1). 166

¹⁶⁷ By collecting all fluxes into vector \mathbf{x}' and all relevant sea surface variables in ¹⁶⁸ vector \mathbf{y}' , the model equation (1) ensures that the fluxes and the oceanic variables ¹⁶⁹ are physically coherent. When coupling EMAD to an OGCM, the fluxes of heat, ¹⁷⁰ fresh water and momentum will not respond independently to a given anomalous ¹⁷¹ state of the sea surface. The noise term acts to excite the EOF-modes described by the deterministic part of EMAD (i.e. by A- and B-terms). The matrix C ensures that the distributions of the total variances of the fluxes match those obtained from the coupled ECHO-G model.

Despite the extremely simple form, the model equation (1) is able to describe 176 various types of air-sea interactions. If $\mathcal{A}\mathbf{x}' + \mathcal{C}\mathbf{n}$ dominates $\mathcal{B}\mathbf{y}'$ for a certain flux, 177 this flux would force the ocean by and large stochastically. This could be the case 178 for wind stress anomalies over high-latitudes, where the influence of the SST on 179 the wind stress is weak. If the \mathcal{A} -term and the \mathcal{B} -term have similar strength, the 180 flux would be affected both by the stochastic forcing and by the oceanic feedback. 181 If the \mathcal{B} -term dominates, the flux would be essentially determined by the oceanic 182 conditions. The relative importance of the various types of air-sea interactions is 183 given by the amplitudes of elements of \mathcal{A} and \mathcal{B} . The result of different types of 184 air-sea interactions can be identified by studying the lagged correlation functions 185 between the flux and SST (Frankignoul et al. 1998; von Storch 2000). 186

The ability of EMAD in reproducing the second moments of fluxes found in the coupled ECHO-G is considered in von Storch et al. (2005). In particular, it was shown that EMAD produces variances of fluxes, whose strength and distribution are in general comparable to that found in ECHO-G. The various types of interactions, as can be identified using the lagged correlation functions between the SST and the fluxes (Frankignoul et al. 1998 ; von Storch 2000), are by and large reproduced when coupling EMAD to an OGCM. Finally, EMAD is able to act realistically to anomalous sea surface condition, such as those related to an ENSO event or to aPolynya.

To give the reader an idea of how the EMAD-fluxes look like, Figure 1 shows 196 anomalies of wind stress (arrows) and heat flux (colour shading), obtained by forc-197 ing EMAD with the anomalous sea surface conditions derived from the coupled 198 ECHAM5/MPI-OM AOGCM (Jungclaus et al. 2005). Also shown in Figure 1 is a 199 snapshot of the anomalies of the same fluxes from the NCEP reanalysis for an arbi-200 trary day. Similar to the fluxes of the reanalysis, the EMAD-fluxes have maxima at 201 the mid- and high-latitudes. The amplitudes of EMAD-anomalies are slightly larger 202 than those of the reanalysis. The structure of the EMAD-anomalies are somewhat 203 smoother than that of the NCEP-fluxes, reflecting the fact that EMAD describes 204 the leading EOF-modes excited by white noise forcing. Figure 1 and the previous 205 validation (von Storch et al. 2005) suggest that EMAD is capable of producing the 206 basic features of the fluctuating day-to-day fluxes. 207

208 2.2 The Ocean General Circulation Model: MPI-OM

The OGCM used in this study is the Max-Planck Institute Ocean Model (MPI-OM). It is a z-coordinate model based on primitive equations for a Boussinesq fluid on a rotating sphere. It is formulated on the horizontal Arakawa C grid with the north pole located at northern Greenland and south pole close to Weddell Sea. It has horizontal resolution varying from 20 km in the main sinking regions associated with the MOC to about 350 km in the tropics. For the present study, the model configuration with 40 vertical levels (Haak et al. 2003) is used. The model contains a free surface and a state-of-the-art sea ice model with viscous-plastic rheology and snow. Overflow over the sills and off continental shelves are represented by a bottom boundary layer slope convection scheme.

Tracer diffusion is isoneutral and dianeutral and is described by the diffusion 219 tensor \mathcal{K} (Redi 1982), which is a function of the neutral density gradient and hor-220 izontal and vertical diffusion coefficients K_H and K_V . The scheme is numerically 221 implemented following Griffies (1998). The effect of horizontal tracer mixing by 222 advection due to the unresolved mesoscale eddies is parameterized after Gent et 223 al. (1995). The horizontal eddy viscosity is parameterized using a scale-dependent 224 biharmonic formulation. The vertical eddy viscosity follows Pacanowski and Phi-225 lander (1981). It utilizes an eddy coefficient which is represented in the same way 226 as the vertical eddy diffusivity coefficient K_V (see Eq.(2) below), except that the 227 Richardson-number dependent part is proportional to $(1 + C_{RD}R_i)^{-2}$, rather than 228 $(1 + C_{RD}R_i)^{-3}$ as given in Eq.(3). The vertical diffusion, as described by K_V or 229 in short K, plays an important role in the present study and is further described 230 below. 231

K is a function of convective mixing, Richardson number (Ri) dependent mixing,
wind-induced mixing and background diffusivity (Marsland et al. 2003) and is given
by

$$K = \begin{cases} K_{conv} & \text{if statically unstable} \\ K_{Ri} + K_{wind} + K_{back} & \text{if statically stable} \end{cases}$$
(2)

235 with

$$K_{Ri} = D_{VO}(1 + C_{RD}R_i)^{-3}, (3)$$

where R_i is the Richardson number, $D_{VO} = 2 \times 10^{-3} m^2 s^{-1}$ and $C_{RD} = 5$ are model constants. According to Eq.(3), the maximum value of K_{Ri} is $2 \times 10^{-3} m^2 s^{-1}$. The diffusion related to convection K_{conv} is set to $10^{-1} m^2/s$. Thus, static instability is removed by switching on an extremely strong mixing. In the surface layer, the wind-induced mixing K_{wind} over ice free regions is given by

$$K_{wind} = W_T V_{10}^3 \tag{4}$$

where V_{10} is the local 10 m wind speed and W_T equals $5 \times 10^{-4} m^{-1} s^2$. Below the surface, K_{wind} depends on the stability of the water column and decays exponentially with *e*-folding depth being 40 m. The diffusion related to other unresolved processes, such as internal waves, is described by $K_{back} = 10^{-5} m^2/s$. This set of parameters is used in the integration of MPI-OM coupled to the ECHAM5 AGCM (Jungclaus et al. 2005) that produces a realistic oceanic state.

The vertical diffusion coefficient K can vary spatially and temporally, depending on the static stability and wind forcing. Since K_{back} is unchanged in the experiments performed and since K_{wind} is confined to the first 40 meters of the ocean (depending on the stability), the changes in K below 40 meters are related to the changes in K_{Ri} and / or K_{conv} . Fluctuating fluxes can change both K_{Ri} and K_{conv} .

Fluctuating fluxes can affect K_{conv} by turning convection on and off in a nonlinear way, depending on the background static stability. In the regions where

the stratification is mostly stable and K_{conv} is mostly turned off, a positive buoy-254 ancy forcing (induced e.g. by additional precipitation events or additional heat flux 255 anomalies) makes the ocean more stable and hence will leave K_{conv} switched off. By 256 contrast, a negative buoyancy forcing will reduce static stability, and hence K_{conv} 257 might be switched on more often. The net effect is an increase in the convective 258 mixing. Examples of mostly stable oceans are the tropical and subtropical oceans. 259 On the other hand, in regions where the stratification is mostly unstable and K_{conv} 260 is mostly switched on, a negative buoyancy forcing will not affect the convective mix-261 ing much, since the convective mixing is already switched on. A positive buoyancy 262 anomaly on the contrary can increase the static stability, making K_{conv} switched on 263 less often. The net effect is a decrease in K. Examples of mostly unstable oceans 264 are the GIN (Greenland Iceland Norwegian) Seas and the high-latitude Southern 265 Ocean in the MPI-OM model. 266

The above described changes in convective events are not inconsistent with previous numerical experiments in which the convection at single grid points can be switched on and off by flux anomalies and be crucial for maintaining deep water formation (Rahmstorf 1995, Kuhlbrodt and Monahan 2003).

Fluctuating fluxes can also change the Ri-dependent mixing, since a fluctuating buoyancy flux can affect the stratification of the water column and a fluctuating wind stress forcing can alter the shear of the current. The change in the Ri-dependent mixing is expected to be more pronounced in the tropics. In these regions, the static stability of the ocean is so high that static instability rarely occurs and, when ²⁷⁶ it occurs, it will be confined to a shallow surface layer.

In the present study, the vertical diffusion coefficient K is stored on monthly 277 basis. Since the maximum value of K_{Ri} of $2 \times 10^{-3} m^2 s^{-1}$ is much smaller than 278 $K_{conv} = 10^{-1} m^2/s$, large changes in K must be related to changes in the number of 279 convective events occurring within a month. Generally, a large increase (decrease) 280 in K indicates an increase (a decrease) in the number of convective events, and from 281 that an increase (a decrease) in the formation of dense water masses. In this sense, 282 changes in K can be used as a crude measure of changes in water mass formation 283 due to convection. As shown by the mean convection depth in Figure 10a, the true 284 deep water formation, reaching about 1000 meter depth on average, occurs only in 285 GIN seas and off the Antarctic coast in the Atlantic sector in the version of the 286 MPI-OM model used here. 287

288 2.3 The Coupled Model: EMAD/MPI-OM

To couple the EMAD with the MPI-OM, the EMAD fluxes, which are on the T30-Gaussian grid, are interpolated into the curvilinear grid of the MPI-OM model. The coupling takes place once a day.

²⁹² When coupling EMAD to the MPI-OM model, one needs a set of fields of climato-²⁹³ logical mean fluxes and a set of fields of climatological mean sea surface conditions. ²⁹⁴ Both sets were derived from the last 50 years of a 600-year integration with the ²⁹⁵ ECHAM5/MPI-OM coupled AOGCM (Jungclaus et al. 2005). The climatology ²⁹⁶ contains the annual cycle on a daily basis. Given an oceanic state at time t, the anomalous sea surface condition \mathbf{y}' is derived by subtracting the actual oceanic state from the given climatological mean state. With this \mathbf{y}' and the anomalous flux forcing \mathbf{x}' at t - 1, EMAD produces the anomalous flux forcing at t. Adding this to the climatological mean forcing gives the net flux forcing at time t which is used to produce \mathbf{y}' at time t + 1.

It should be noted that since EMAD is only an approximation of ECHAM5, and since the $\mathcal{B}\mathbf{y}'$ -term is not a rigid restoring, the climatological mean state produced by the MPI-OM model coupled to EMAD generally does not match the climatological mean state produced by the fully coupled ECHAM5/MPI-OM. As a consequence, the time-mean of \mathbf{y}' is not zero. This non-zero time-mean of \mathbf{y}' can feed back to the fluxes and produce non-zero time-mean of \mathbf{x}' , whereby complicating the interpretation of the experiments to be introduced in Section 3. We will return to this issue later.

Apart from the feedbacks described by \mathcal{B} , there is no relaxation of salinity or temperature in the ocean. The only procedure used to prevent the ocean drifting away from the given climatological mean state is to restore the sea ice cover and sea ice thickness to that found in the integration with the ECHAM5/MPI-OM. The restoring time constant is chosen as 39 days.

314 **3** Numerical Experiments

To study the effect of fluctuating daily fluxes, three experiments were carried out. In the experiment BH, MPI-OM was driven by the climatological mean fluxes of heat, fresh water and momentum plus an additional heat flux anomaly, H', which was obtained from $\mathcal{B}\mathbf{y}'$ with \mathbf{y}' representing SST anomalies. This particular form of $\mathcal{B}\mathbf{y}'$ contains the SST feedbacks that prevent large climate drifts. This is shown by an additional experiment in which the MPI-OM model was driven by the fixed climatological fluxes only. The ocean drifts to a warmer climate and produces a global mean surface temperature which is about 4 °C more (not shown) than that found in the coupled ECHAM5/MPI-OM run. The drift disappears when the $\mathcal{B}\mathbf{y}'$ term is installed.

In the second experiment ABC, in which all the three terms in Eq.(1) are included, the MPI-OM model was forced with the same climatological mean fluxes plus fluctuating fluxes produced by EMAD. In the third experiment AB2C, MPI-OM was coupled to EMAD with the variance of white noise doubled.

The three experiments are summarized in Tab.1. For each experiment, a spin-up run of about 600 years was carried out. The spin-up runs started from the same initial state obtained from the coupled ECHAM5/MPI-OM model (Jungclaus et al. 2005), after the ocean has reached a more or less statistically stationary state. Following the respective spin-up runs, the experiments were carried out for 200 years. The analysis given below is based on these 200-year integrations.

If all the three experiments produce the same climatological mean state (i.e. the same mean sea surface conditions and the same mean surface fluxes) and if this state is identical to that produced by the ECHAM5/MPI-OM model, the time-means of fluctuating fluxes in the three experiments will be zero. In this case, the difference between the experiments BH and ABC would describe the effect of fluctuating dayto-day fluxes, and that between the experiments BH and AB2C would describe the
effect of enhanced fluctuating fluxes.

Unfortunately, the three experiments do not produce exactly the same climato-342 logical mean state of the ECHAM5/MPI-OM model. Consequently, the time-mean 343 of fluctuating fluxes in experiments BH, ABC and AB2C are not zero. Moreover, 344 they differ from each other, since the climatological state in experiment BH can 345 differ from that obtained from experiment ABC or AB2C. Due to these differences, 346 the changes from experiment BH to ABC or from BH to AB2C are induced not only 347 by fluctuating fluxes included in experiment ABC and AB2C, but also by the dif-348 ferences in the time-mean fluxes. A consideration of these time-mean fluxes reveals 349 some notable differences in the time-mean zonal wind stress (Figure 2). For instance, 350 there is an increase in zonal wind stress in the North Atlantic (at 30° W, 50° N) and 351 a northward shift of the mean zonal wind stress pattern over the Southern Ocean 352 from experiment BH to experiment ABC and AB2C. The increase in zonal wind 353 stress in the North Atlantic could be relevant, since the wind-driven gyre partici-354 pates in the meridional salt transport and can therefore affect the MOC (Marti et 355 al. 2008). The shift in the Southern Ocean could contribute to the ACC differences 356 from experiment BH to ABC and to AB2C. 357

To assess the relevance of these non-zero time-mean fluxes, a supplementary experiment, referred to as BH*, is carried out. Experiment BH* is identical to experiment BH, except that the difference between the time-mean zonal wind stress of experiment BH (Figure 2a) and that of experiment ABC (Figure 2b) is added to the climatological mean wind forcing. It will be shown that the time-mean fluxes do not significantly affect the MOC changes and contribute only to a small part of ACC changes. Experiment BH* is integrated for 350 years.

³⁶⁵ 4 Changes induced by fluctuating day-to-day fluxes

366 4.1 Time Evolutions

This subsection describes the time evolutions of the oceanic states in different experiments. The consideration is confined to the globally integrated sea surface temperature (SST) and two circulation indices, the Atlantic MOC-index and the ACC-index. The globally integrated SST is considered to describe the effect of the SST-feedback over time. The MOC-index and the ACC-index are chosen, since they characterize major global-scale circulations.

In all experiments, the SST time series are essentially statistically stationary 373 (Figure 3a). The SST decreases slightly from experiment BH to experiment ABC 374 and AB2C. The respective time-mean values are 18.57 °C, 18.15 °C and 18.12 °C. 375 The spatial distribution of SST changes reveals decreases over most of the subtrop-376 ical and mid-latitude oceans and increases partially over the North Atlantic from 377 experiment BH to ABC and AB2C. The decrease is partly related to the increase in 378 frequency and depth of convective events in the subtropical and mid-latitude oceans 379 (see Section 4.4), which bring cold dense water down. The fluctuating fluxes also 380 enhance the SST variability. 381

Figure 3b shows the time series of the MOC-index, defined as the maximum of the Atlantic meridional overturning streamfunction near 30°N at about 1220 m. The experiment BH (dotted line), which does not include fluctuating fluxes, reveals little variability in the MOC. Stronger variations are obtained by including fluctuations (solid line) in experiment ABC. The variations are strongest in experiment AB2C (dashed line) where the variance of the stochastic forcing is doubled.

Not only the variability but also the time mean of the MOC-index changes from 388 experiment to experiment. This is further summarized in Tab.2. The smallest value 389 of about 17 Sv is obtained from experiment BH. Inclusion of fluctuations leads to 390 an increase of about 18 Sv in experiment ABC. Experiment AB2C, in which the 391 strongest MOC of about 22 Sv is found, further confirms that the 1-Sv increase 392 from experiment BH to experiment ABC is caused by the fluctuating component in 393 the fluxes. The time-mean of the MOC-index of a 200-year time series of the coupled 394 ECHAM5/MPI-OM simulation is also comparable to that of experiment ABC (last 395 row in Tab.2). 396

Figure 3c shows the time series of the ACC-index defined as the mass transport through the Drake Passage. There is an enhancement of variability through fluctuations in experiment ABC and AB2C (solid line and dashed lines respectively). Concerning the time-mean (see also Tab.2.), a mean transport of about 200 Sv is obtained in experiment BH. This value is too high relative to the observed value of about 120 to 150 Sv (Nowlin and Klinck 1986 ; Cunningham et al. 2003). The transport reduces to about 148 Sv in experiment ABC and to about 122 Sv in AB2C. The strength of the net mass transport through the Drake Passage reflects well the strength of the zonal current in the entire Southern Ocean (not shown). Tab.2 shows that a stronger MOC corresponds to a weaker ACC. The correspondence concerns only the time-mean values. The variability of the ACC-index is not correlated to that of the MOC-index.

To have an idea about how much of the changes listed in Tab.2 are caused by 409 the differences in the time-mean zonal wind stress shown in Figure 2, the time-410 mean values of the MOC-index and the ACC-index are calculated from the last 200 411 years of experiment BH^{*}. They amount to 17.2 Sv and 180 Sv, respectively. The 412 first number suggests that the increase in zonal wind stress in the North Atlantic 413 from Figure 2a to Figure 2b is not responsible for the 1-Sv MOC increase found by 414 comparing experiment ABC with experiment BH. Instead, the 1-Sv increase is likely 415 caused by the fluctuating fluxes included in experiment ABC. The second number 416 suggests that the ACC change from experiment BH to ABC is partly due to the 417 northward shift in the time-mean zonal wind stress shown in Figure 2b. However, 418 if the effect of time-mean wind stress and that of fluctuating fluxes can be linearly 419 superimposed, the effect due to the time-mean zonal wind stress is smaller than that 420 of the fluctuating fluxes. 421

For the comparison, the values of the MOC- and ACC-indices in ECHAM5/MPI-OM are shown in Tab.2 (last row). If the climatological mean state in the ECHAM5/MPI-OM model is identical to that in the hybrid EMAD/MPI-OM model, the values obtained from experiment ABC would be close to those shown in the last row. One finds a good agreement for the MOC-index, but not as good an agreement for the A27 ACC-index. This further confirms that the circulation in the MPI-OM model is more sensitive to the time-mean zonal wind stresses in the Southern Ocean than those in the North Atlantic.

430 4.2 Changes in the Atlantic Meridional Overturning Circulation

The different formulations of the surface fluxes result in different Atlantic meridional 431 overturning circulations (Figure 4). The difference concerns not only the strength 432 that is described by the MOC-index in Figure 3 and Tab.2., but also the structure. 433 When the MPI-OM is driven by the climatological fluxes plus the oceanic feed-434 backs (experiment BH), an overturning cell of around 2800 m and 17 Sv maximum 435 strength is obtained (Figure 4a). In experiment ABC (Figure 4b), the overturning 436 cell is stronger and extends a couple of hundreds of meters down to the deep ocean 437 compared to that of BH. When the stochastic forcing is doubled, the overturning 438 circulation further strengthens and deepens (Figure 4c). The deepening of the over-439 turning cell is accompanied by the weakening of the Antarctic Bottom Water cell 440 and the retreat of the Antarctic Bottom Water (AABW). In experiment AB2C, 441 the penetration of AABW into the abyssal North Atlantic is severely blocked. The 442 spatial structure of the Atlantic MOC in experiment ABC is comparable to that 443 produced by the ECHAM5/MPI-OM model (not shown) and that obtained from an 444 ensemble of coupled AOGCMs (Stouffer et al. 2006). 445

^{446 &}lt;sup>′</sup>L

To make sure that the 1-Sv increase in MOC is statistically significant, a t-test

⁴⁴⁷ is carried out. The null hypothesis that the maximum overturning in experiment
⁴⁴⁸ BH equals that in experiment ABC is considered. The null hypothesis is rejected
⁴⁴⁹ with 1% risk.

To show that the above described structural changes do not result from the different time-mean zonal wind stress, the mean Atlantic overturning streamfunction obtained from the last 200 years of experiment BH* is shown in Figure 5. Both the strength and the structure are comparable to the streamfunction obtained from experiment BH.

455 4.3 Changes in the density fields

To understand whether and to what extent fluctuating fluxes change the mean cir-456 culation via changing density structures, consider first the situation at the surface. 457 Figure 6a shows the zonal-mean meridional profiles of surface density in the Atlantic 458 sector. The large differences at the high northern latitudes result from different 459 climatological mean states in the Arctic: the Arctic becomes more saline from ex-460 periment BH to experiment ABC and AB2C (not shown). In the North Atlantic 461 from about 40° N to 75° N, it is difficult to relate changes in the meridional density 462 gradient to the MOC changes found in experiments BH, ABC and AB2C. In the 463 south from 30° S to 60° S, the meridional gradient in experiment BH is stronger than 464 that in experiment ABC and AB2C, as indicated by the dotted line (BH) which 465 is below the solid (ABC) and dashed (AB2C) lines north of about 45° S and above 466 them, south of 45° S. The change from experiment ABC to AB2C (solid and dashed 467

lines) is less clear. Following the previous studies (e.g. Russell et al. 2006) suggesting that the ACC is related to the density gradient in deeper layers, in particular
to the density gradient integrated over the sill depth, density changes in the oceanic
interior were considered.

Indeed the effects of fluctuating fluxes can be traced down to the deep ocean. 472 Figure 7a shows the in-situ density in experiment BH at 1365 m. The density field 473 is characterized by higher density in the Atlantic and the Southern Oceans than in 474 the Pacific and the Indian Oceans. To describe the changes in the density gradient 475 induced by fluctuating fluxes, the differences between experiments ABC and BH and 476 between AB2C and BH are shown in Figure 7b and c, respectively. The dominant 477 feature of the density changes is the zonally oriented density increases centered near 478 40°S and density decreases further south. The amplitudes of the density increases are 479 larger than those of decreases. These changes lead to a reduction in the meridional 480 density gradient in the Southern Ocean and hence a weakening of the ACC. 481

The density changes in the North Atlantic are more subtle. From the difference AB2C-BH shown in Figure 7c, one can identify a few isolines in the Atlantic north of 40°N that reveal strong tilt in the north-south direction. These isolines suggest an increase in the zonal density gradient that are by geostrophic relation consistent with the large increase of the MOC of more than 4 Sv from experiment BH to AB2C. However, this feature does not show up clearly in the difference ABC-BH (Figure 7b).

489

A further search for a clear relation between changes in density distribution and

changes in the Atlantic MOC leads to the consideration of the depth-integrated in-490 situ density. This quantity was shown to be related to the strength of the MOC 491 in previous studies (Hughes and Weaver, 1994 and Thorpe et al 2001). Figure 6b 492 shows the zonal mean of depth-integrated in-situ density in the Atlantic sector. The 493 increase of the density equatorward of about 40° from experiment BH to ABC and 494 AB2C is related to the change in the climatological mean state, which becomes colder 495 in experiment ABC and AB2C, relative to that in experiment BH. The cooling is 496 partly due to the increase in frequency and depth of convective events (see Section 497 4.4). Regarding the meridional gradient, the meridional gradients between 40° S and 498 40° N and between 40° N and 60° N do not change much. However, small changes 499 in density difference between northern North Atlantic and the South Atlantic are 500 possible. After calculating the density difference between different latitudinal strips 501 in the North and South Atlantic, we found that the density difference between the 502 northern strip extending from 55° N to 60° N and the southern strip extending from 503 45° S to 50° S increases with the MOC from 0.11 kg/m³ in experiment BH to 0.12 504 kg/m^3 in experiment ABC and to 0.13 kg/m^3 in experiment AB2C. A similar north-505 south density difference was considered in studies by Rahmstorf (1996) and Thorpe 506 et al (2001). 507

In the Southern Ocean, the meridional gradient of the depth-integrated density (Figure 6b) is reduced. Expressed in terms of the density difference between 40°S and 60°S, one finds decreases from 1.02 kg/m^3 in experiment BH (dotted) to 0.77 kg/m³ in experiment ABC (solid) and to 0.63 kg/m³ in experiment AB2C (dashed). The reduction of meridional gradient is more strongly related to the increase in density around 40°S than to the decrease in density around 60°S. These changes in meridional density distribution are related to the changes in ACC in different experiments by the geostrophic relation.

516 4.4 Changes in the vertical mixing

In this section, the way surface fluxes alter the density in the deep ocean is examined. 517 It will be shown that surface fluxes change the density via vertical mixing. Before 518 dealing with these mixing changes, consider first the time-mean mixing, as described 519 by the time-mean vertical diffusion coefficient K for experiment BH at 285 m depth 520 in Figure 8a and at 900 m depth in Figure 9a. In GIN Seas and along and near 521 the Antarctic coast in the Atlantic sector and south of the South America, large 522 time-mean values of K and mean convection depth (Figure 10a) are found, which is 523 also the case in the coupled ECHAM5/MPI-OM model. These regions are the most 524 unstable regions of the ocean model. Elsewhere, the modelled ocean is much more 525 stable. 526

⁵²⁷ A maximum of K is also found at 285 m in the north Pacific just west of the ⁵²⁸ date line between 50°N and 60°N (Figure 8a). This maximum disappears at 900 m ⁵²⁹ in Figure 9a. The map of the mean depth of convection (Figure 10a) suggests that ⁵³⁰ the convection related to this maximum is shallower than 600-650 m.

Generally, K is smaller than 0.002 m^2/s in most part of the ocean at 900 m and decreases with depth to values smaller than $10^{-4} m^2 s^{-1}$ below 2500 m. These values should be compared with the observational range of 10^{-6} to $10^{-3} m^2/s$ reported from interior oceanic regions (Ledwell et al. 1993; Moum et al. 2002, Gregg et al. 2003, Ledwell et al. 2000, Sloyan 2005).

⁵³⁶ Consider now the changes in the vertical mixing due to different surface fluxes. ⁵³⁷ In the following, the effect of the fluctuating fluxes is indicated by the difference in K⁵³⁸ (ABC-BH) obtained from experiment ABC and BH. The effect of the fluctuations ⁵³⁹ with enhanced variance is obtained by comparing the difference AB2C-BH with ⁵⁴⁰ the difference ABC-BH. The possible effect of the time-mean zonal wind stress on ⁵⁴¹ changes in K is small and will be discussed at the end of this section.

Figure 8b shows the difference in K (ABC-BH) at 285 m depth, induced by the 542 fluctuations. Outside the tropics, where large changes of K are found, there is a 543 striking correspondence between the distribution of the time-mean frequency and 544 depth of convective events shown in Figure 8a and Figure 10a and the distribution 545 of the changes in K from experiment BH to ABC shown in Figure 8b in the MPI-546 OM model: The strong decreases in K are found in the regions where the time-547 mean values of K are large, indicating frequent occurrence of convective events due 548 to mostly unstable stratification. These regions consist of the GIN Seas, an area 549 centered near 50° W and 35° N in the North Atlantic, the areas south of the South 550 American continent and west and east of the Antarctic Peninsula. The increases 551 in K, on the other hand, are found in regions where the time-mean values of K552 are generally small and convective events are less frequent due to mostly stable 553 stratification. In the North Atlantic, an area with increases in K is found between 554

⁵⁵⁵ 40°N to 60°N, south of the GIN Seas. There, the time-mean values of K are generally ⁵⁵⁶ small, apart from the areas close to the Irminger Sea which will be discussed at the ⁵⁵⁷ end of this section. In the Southern Ocean, increases in K are found mainly in the ⁵⁵⁸ latitude band from 20°S to 50°S. This correspondence between the time-mean values ⁵⁵⁹ of K and changes in K can also be seen in experiment AB2C (Figure 8c).

At 900 m (Figure 9b), the correspondence between the time-mean mixing and 560 the changes in the convective mixing is also noticeable poleward of about 40° . In 561 particular, the decreases in K are mainly located in regions with large values of the 562 time-mean mixing shown in Figure 9a. Overall, the magnitudes of mixing changes 563 are much smaller at 900 m (Figure 9b,c) than at 285 m (Figure 8b,c). The areas 564 with enhanced mixing in experiment ABC (Figure 9b) is enlarged when the strength 565 of stochastic fluctuations is doubled in experiment AB2C (Figure 9c): The patchy 566 structure over the Southern Ocean in Figure 9b becomes more uniform in Figure 567 9c. Mixing structures similar to Figure 9 but with smaller amplitudes can be found 568 down to about 2500 to 3000 m. Further below, the mixing signal is much less zonally 569 oriented. 570

Following the definition of K, the above described changes in K are related to changes in the *frequency* of convective events. Changes in the mean *depth* of convective events are described in Figure 10b and c. The largest changes in the mean convection depth are about 300 m. A comparison of Figure 10b,c with Figure 8b,c suggests that, apart from the tropical oceans, an increase (a decrease) in the frequency of convective events corresponds to an increase (a decrease) in the depth of convective events. Since the regions with small (large) time-mean values of *K* and shallow (deep) mean convection represent regions which are often stably (unstably) stratified, one can conclude that the fluctuating fluxes tend to increase the convective mixing over mostly stable ocean and decrease convective mixing over mostly unstable ocean. Moreover, an increase (a decrease) in convective mixing is accomplished by an increase (a decrease) in both frequency and depth of convective events.

The above described mixing changes can be responsible for the density changes 583 described in Section 4.3. As the enhanced vertical mixing is related to an increase 584 in the frequency and the depth of convective events and these events generally bring 585 dense water down, the increase in vertical mixing in the midlatitudes found in ex-586 periments ABC and AB2C can lead to an increase in density there. In the Atlantic, 587 the increase in the vertical mixing near the northern strip 55°N-60°N is stronger 588 than that in the southern strip 45° S- 50° S. This can lead to the increase of the den-589 sity difference between the two strips from experiment BH to experiment ABC and 590 AB2C, which is related to the respective MOC increases. The large-scale increase 591 in the vertical mixing in the Southern Ocean can be responsible for the increase 592 in density near 40° S, which results in a weaker meridional density gradient and a 593 weaker ACC. 594

There exists a few spots where the change in K is not clearly related to the mean stratification, for instance west of Svalbard and also in the Arctic.

⁵⁹⁷ Note that the distribution of changes in the convection depth (Figure 10b, c) ⁵⁹⁸ compares less well with changes in K at 900 m (Figure 9b, c) than with changes at 285 m (Figure 8b, c). This is because Figure 10 represents the mean convection depth, rather than the depth of individual convective events. Apart from a few exceptions at high-latitudes, convective events are mostly shallower than a few hundred meters. The changes in convection depth due to fluctuating fluxes are generally smaller than 300 m. Thus, Figure 10b and c reflect mainly the changes related to convections shallower than 900 m.

Also in tropical oceans, the changes in convective activity (Figure 10b,c) do 605 not correspond to changes in K (Figure 8b,c and Figure 9b,c) in tropical oceans. 606 There, one finds increases in the convective depth, even though the vertical mixing 607 is reduced. This is because convective events are confined to the upper 60 m in the 608 tropical and subtropical Atlantic and Indian Oceans and to the upper 100 m in the 609 tropical and subtropical Pacific, but can reach a few hundred meters in the extra-610 tropical regions. As the convective events triggered by the fluctuating daily fluxes 611 are confined to a shallow surface layer, the decrease in K in the tropics below, say, 612 100 m, as seen in Figure 8b,c and Figure 9b,c is likely caused by the *Ri*-dependent 613 mixing, rather than the convective mixing. 614

The above described mixing changes are mainly due to fluctuating fluxes. The effect of time-mean fluxes is mostly secondary. This is shown by the difference in the mean convection depth found in the experiments ABC and BH* (Figure 11). Different from Figure 10b which shows changes due to fluctuating fluxes *and* the difference in the time-mean zonal wind stress, the effect due to different time-mean zonal wind stress is eliminated in Figure 11. Since Figure 10b is very close to

Figure 11, the effect due to different time-mean zonal wind stress must be small. An 621 exception is the change near the Irminger Sea. In this region, the mechanism which 622 alters the MOC through the wind-driven gyre (Marti et al. 2008) can be at work. 623 The stronger time-mean zonal wind stress in experiment ABC can transport more 624 salt northward, whereby enhancing the surface density and triggering more often 625 convective events. This is likely the reason, why large increases in the frequency 626 and depth of convective events are found in the Irminger Sea (Figure 8b and Figure 627 10b), where, if the effect of fluctuating fluxes dominated, a decrease in the frequency 628 and depth of convective events was expected. 629

5 Conclusions and discussion

The MPI-OM model coupled to the empirical flux model EMAD is used to isolate 631 the effect of fluctuating fluxes. It is found that fluctuating daily fluxes can produce 632 a 1-Sv-increase in the strength of the Atlantic MOC, which is about 7% of the 633 MOC obtained without the fluctuating fluxes, and a 32-Sv-reduction of the ACC, 634 which is about 16% of the ACC obtained without fluctuating fluxes. These changes 635 exclude (with the aid of experiment BH*) the effect of non-zero time-mean fluxes 636 that cannot be completely excluded from the experiments. The MOC changes are 637 related to the change in the meridional density difference between two latitudinal 638 strips in the northern North Atlantic and in the South Atlantic, defined in a way 639 similar to that in Rahmstorf (1996) and Thorpe et al (2001). The ACC changes are 640 related to a reduction in meridional gradient in the depth-integrated density. 641

These density changes are likely caused by changes in vertical mixing induced 642 by fluctuating surface fluxes. In the MPI-OM model, large changes in the vertical 643 mixing are related to the changes in the convective mixing. Fluctuating fluxes alter 644 convective mixing in a non-linear way. In the mostly unstable regions, e.g. in645 the GIN Seas and near the Antarctic Peninsula, positive buoyancy anomalies can 646 restrain convective events, whereas negative buoyancy anomalies do not significantly 647 affect the convective behaviour, leading to an overall reduction in the convective 648 mixing. In mostly stable regions, e.g. in the subtropical and mid-latitude oceans, 649 large negative buoyancy anomalies can trigger additional convective events, whereas 650 positive buoyancy anomalies do not significantly change the convective behaviour, 651 leading to an overall strengthening of the convective mixing. 652

The conclusions about the effect of fluctuating fluxes on the convective mixing 653 are drawn within the framework of a coarse resolution version of the MPI-OM. In 654 another OGCM, which produces a different stratification with a different distribution 655 of convective activity, the effect of fluctuating fluxes on the convective mixing, which 656 is closely related to the mean stratification, can be different. As a consequence, the 657 exact numbers concerning the changes in the MOC and ACC due to the fluctuating 658 fluxes can depend on the model used. Nevertheless, the mechanism through which 659 fluctuating daily fluxes alter the convective mixing should operate in other models, 660 and probably also in nature. 661

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752 FIGURE CAPTIONS

- Figure 1: Daily snapshots of wind stress and heat flux anomalies from a) EMAD model and b) NCEP/NCAR Reanalysis. The units are in *Pascal* and W/m^2 respectively.
- Figure 2: Time-mean zonal wind stress in *Pa* obtained in a) experiment BH, b)
 experiment ABC and c) experiment AB2C.
- Figure 3: Yearly time series of a) globally integrated sea surface temperature in ${}^{o}C$, b) the maximum of Atlantic meridional overturning streamfunction located near 30°N in Sv (MOC-index) and c) Drake Passage mass transport in Sv(ACC-index) obtained from experiment BH (dotted), ABC (solid) and AB2C (dashed).
- Figure 4: Spatial structure of the time-mean Atlantic MOC in Sv in different experiments.
- Figure 5: Spatial structure of the time-mean Atlantic MOC in Sv in experiment BH*.
- Figure 6: Meridional profiles of the zonal mean of surface density and the zonal mean of depth-integrated in-situ density in the Atlantic sector obtained from experiment BH (dotted), ABC (solid) and AB2C (dashed). The depth integration starts from 1220 m for experiments BH, ABC and AB2C. The Atlantic sector covers the oceanic region from 70°W to 10°E and hence includes the Drake passage. Unit is kg/m^3 .

773	Figure 7: Horizontal distributions of in situ density at 1365 m in kg/m^3 for a) the
774	time-mean in experiment BH and b) the difference between experiments ABC
775	and BH and c) between experiments AB2C and BH.
776	Figure 8: Horizontal distributions of K at 285 m in m^2/s for a) the time-mean in
777	experiment BH and b) the difference between experiments ABC and BH and
778	c) between experiments AB2C and BH.
779	Figure 9: Same as Figure 8, but for K at 900 m.
780	Figure 10: Horizontal distributions of mean convection depth in m for a) the time-
781	mean in experiment BH and b) the difference between experiments ABC and
782	BH and c) between experiments AB2C and BH.
783	Figure 11: Horizontal distribution of the change in the mean convection depth in
784	m between experiments ABC and BH [*] .

785 TABLE CAPTIONS

- Table 1 : Experiments done with MPI-OM driven by different types of daily fluxes
- $_{787}$ at the sea surface. $\bar{\mathbf{x}}$ denotes the climatological fluxes and \mathbf{x}' the flux anomalies
- 788 predicted by different versions of EMAD model.
- Table 2 : 200-year means of MOC-index and ACC-index in Sv in different experiments.



Figure 1: Daily snapshots of windstress and heatflux anomalies from a) EMAD model and b) NCEP/NCAR Reanalysis. The units are in *Pascal* and W/m^2 respectively.





40°\$

80°Ø

Figure 2: Time-mean zonal wind stress in *Pa* obtained in a) experiment BH, b) experiment ABC and c) experiment AB2C.

0° Longitude

100°W

100°E



Figure 3: Yearly time series of a) globally integrated sea surface temperature in ${}^{o}C$, b) the maximum of Atlantic meridional overturning streamfunction located near 30°N in Sv (MOC-index) and c) Drake Passage mass transport in Sv (ACC-index) obtained from experiment BH (dotted), ABC (solid) and AB2C (dashed).



Figure 4: Spatial structure of the time-mean Atlantic MOC in Sv in different experiments.



Figure 5: Spatial structure of the time-mean Atlantic MOC in Sv in experiment BH*.



Figure 6: Meridional profiles of the zonal mean of surface density and the zonal mean of depth-integrated in-situ density in the Atlantic sector obtained from experiment BH (dotted), ABC (solid) and AB2C (dashed). The depth integration starts from 1220 m for experiments BH, ABC and AB2C. The Atlantic sector covers the oceanic region from 70°W to 10°E and hence includes the Drake passage. Unit is kg/m^3 .







-0.28-0.24 -0.2 -0.16-0.12-0.08-0.04 **0** 0.04 0.08 0.12 0.16 0.2 0.24 0.28



Figure 7: Horizontal distributions of a) the time-mean in situ density at 1365 m in experiment BH and b) the difference between density in experiments ABC and BH and c) between experiments AB2C and BH. The unit is kg/m^3 .





Figure 8: Horizontal distributions of K at 285 m in m^2/s for a) the time-mean in experiment BH and b) the difference between experiments ABC and BH and c) between experiments AB2C and BH.



Figure 9: Same as Figure 8 but for K at 900 m.





Figure 10: Horizontal distributions of mean convection depth in m for a) the time-mean in experiment BH and b) the difference between experiments ABC and BH and c) between experiments AB2C and BH.



Figure 11: Horizontal distribution of the change in the mean convection depth in m between experiments ABC and BH*.

Table 1: Experiments done with MPI-OM and different versions of EMAD model. $\bar{\mathbf{x}}$ denotes the climatological fluxes and \mathbf{x}' the flux anomalies predicted by different versions of the EMAD model.

name	fluxes used	characteristics
BH	$\mathbf{\bar{s}}_{\mathbf{H}}$ $\mathbf{\bar{x}} + \mathbf{x}'$ no fluctuations,	
	$\mathbf{x}_{t}^{'} = \mathcal{B}\mathbf{y}_{t}^{'}, (\mathbf{x}^{'} = H^{'}, \mathbf{y}^{'} = SST^{'})$	with SST-feedback on heat flux
ABC	$ar{\mathbf{x}} + \mathbf{x}^{'},$	with fluctuations + feedback
	$\mathbf{x}_{t}^{'} = \mathcal{A}\mathbf{x}_{t-1}^{'} + \mathcal{B}\mathbf{y}_{t}^{'} + \mathcal{C}\mathbf{n}_{t}$	
AB2C	$ar{\mathbf{x}} + \mathbf{x}',$	with feedback + $2 \times$ fluctuations
	$\mathbf{x}_{t}^{'} = \mathcal{A}\mathbf{x}_{t-1}^{'} + \mathcal{B}\mathbf{y}_{t}^{'} + 2 imes \mathcal{C}\mathbf{n}_{t}$	

Experiment name	MOC-index	ACC-index
ВН	17.1	200
ABC	18.3	148
AB2C	21.8	122
ECHAM5/MPI-OM	18.0	186

Table 2: 200-year means of MOC-index and ACC-index in Sv in different experiments.