Distinct Modes of Internal Variability in the Global Meridional Overturning Circulation Associated with the Southern Hemisphere Westerly Winds

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ABSTRACT

The internal variability of the global meridional overturning circulation (GMOC) is examined in this study. Two distinct modes of the GMOC, which are closely linked to the Southern Hemisphere westerly winds (SWW) anomalies, are found to vary on multidecadal and centennial time scales. The dominant mode is characterized by Southern Ocean dynamics: strengthening and poleward shift of the SWW associated with a positive phase of the southern annular mode yield Ekman-induced northward mass transport, including a zonally asymmetric response in the Southern Ocean sea surface temperature and a cooling in the tropical Pacific Ocean due to large-scale upwelling. The second mode projects mainly onto the Atlantic meridional overturning circulation (AMOC). It is driven by a combination of SWW variation and buoyancy forcing. Based on the relationship between the two modes together with the wind perturbation experiments, the authors emphasize that the full AMOC response to the SWW change takes several centuries in their model. The sea surface temperature in Northern Hemisphere high latitudes is significantly affected in this mode, showing a large-scale warming. Their results from a mid-Holocene experiment imply that both modes are independent from the climate background conditions in the Holocene. Finally, the authors argue that the natural modes of GMOC are important to understand trends in ocean circulation, with consequences for heat and carbon budgets for past, present, and future climate.

1. Introduction

The winds represent one of the major forcing that modulates the global meridional overturning circulation (GMOC) (Toggweiler and Russell 2008), by transferring the atmospheric kinetic energy directly or indirectly to the ocean via Ekman transport and geostrophic flow (Wunsch and Ferrari 2004). It is estimated that about 70% of this transfer takes place over the Southern Ocean (SO) (Wunsch 1998). The most important pattern of climate variability in the Southern Hemisphere (SH) middle to high latitudes is related to the southern annular mode (SAM). Positive phase of the SAM is associated with the enhanced Southern Hemisphere westerly winds (SWW), resulting in increased equatorward Ekman transport (e.g., Lovenduski and Gruber 2005; Sen Gupta and England 2006).

Toggweiler and Samuels (1993, 1995) examine how the SWW can influence the GMOC. By performing a number of sensitivity experiments, they highlight the “Drake Passage effect”: the magnitude of the inflow and outflow from the Atlantic to the SO is linearly correlated with the strength of the SWW at latitude of the Drake Passage. This effect is further elaborated, showing that after applying stronger SWW, stronger upwelling at the latitude of Drake Passage increases the density gradient across the Antarctic Circumpolar Current (ACC) and thus produces a stronger ACC (McDermott 1996). On the contrary, a shutdown of the SWW is reported to induce the disappearance of the SO overturning (Ma et al. 2010).

Using idealized single-basin and twin-basin models, as well as ocean general circulation models (GCMs), several
studies (e.g., Gnanadesikan and Hallberg 2000; Hirabara et al. 2007) indicate that stronger SWW can induce more deep outflow through the South Atlantic, which needs to be compensated by more deep-water formation in the North Atlantic and thus increases the North Atlantic Deep Water (NADW) formation. The GMOC, on one hand, varies with the magnitude of SWW; on the other hand, it also depends on the latitude of the wind bands. A poleward shift of zero wind stress curl can result in a stronger Atlantic meridional overturning circulation (AMOC) and vice versa (Sijp and England 2008, 2009). Delworth and Zeng (2008) use a coupled GCM to simulate the impact of altered wind stress over the SH on the AMOC and conclude that an enhanced (reduced) and poleward (equatorward) shift of the SWW results in a strengthening (weakening) of the AMOC. However, using the fine-resolution eddy-permitting version of the same model, Farneti and Delworth (2010) demonstrate that poleward eddy fluxes associated with a strengthening of the SWW can largely compensate the enhanced equatorward Ekman transport and thus significantly reduce the AMOC anomalies.

The SWW affects not only the ocean circulation in the Atlantic but also the Indo-Pacific overturning circulation. A SH wind stress perturbation experiment (Klinger and Cruz 2009) estimates the Indo-Pacific volume transport anomaly in the SH as twice the amount in the South Atlantic. By contrast, the volume transport anomaly has a similar magnitude over both basins in the Northern Hemisphere (NH). However, changes of the SWW have the opposite effect on different ocean basins in the NH. A stronger AMOC due to an enhanced and poleward shift of the SWW normally accompanies a damped sinking in the North Pacific (Sijp and England 2009). Thus, a seesaw-like pattern between the Atlantic and the Pacific overturning can be expected, which might be amplified by the Atlantic–Pacific seesaw (Saenko et al. 2004), which involves a positive feedback between ocean circulation and salinity at low and high latitudes.

The GMOC response to the SWW varies temporally as well, in all ocean basins. Klinger and Cruz (2009) pointed out that a SH wind stress anomaly can generate the volume transport anomaly in the whole SH within one decade. While in the NH, this anomaly in the AMOC persists in their 80-yr simulation. They also argue that the Ekman transport initiated by the wind stress anomaly should be much larger in the Indo-Pacific basin than that in the Atlantic basin, because the former one is much broader. In another perturbation experiment mentioned above, Delworth and Zeng (2008) found that it takes 1–2 centuries for the AMOC to fully adjust to the wind stress anomaly. Both results are consistent with previous studies, which show that the GMOC anomalies on interannual to decadal time scales are confined to the basin in which they are generated but might have a global impact through the atmospheric and oceanic teleconnection (Johnson and Marshall 2004).

Modeling studies reveal that the AMOC have the variability on decadal and multidecadal to centennial time scale (e.g., Delworth et al. 1993; Park and Latif 2008), which is not induced by any external forcing but through internal variability in the climate system. The variability at these time scales generates specific sea surface temperature (SST) patterns. The most discussed one, identified in observational data, is the so-called Atlantic multidecadal oscillation (AMO; Kerr 2000), the basin-wide North Atlantic SST anomaly with a typical period of 65–80 yr (Enfield et al. 2001). Dima and Lohmann (2007) examined a hemispheric mechanism of this mode based on long-term instrumental data. They found that its periodicity is related to the AMOC adjustment to freshwater fluxes, the Atlantic SST response to it, the ocean–atmosphere feedback in the North Pacific, and the sea ice response to wind stress. Evidence of this mode also exists in proxy data (e.g., Mann et al. 1995; Lohmann et al. 2004). More recently, two distinct SST patterns have been found in the instrumental data: a global mode on centennial time scale and an Atlantic mode associated to multidecadal variability, which are potentially linked to variability of GMOC and AMOC, respectively (Dima and Lohmann 2010).

A previous study (McDermott 1996) suggests that longer time scales rather than interannual variability in SWW can influence the sinking in the NH and modulate the GMOC. In this study, we highlight the decadal and longer time-scale internal variability of the GMOC driven by SWW anomalies based on model simulations. To our knowledge, it is the first attempt to focus on the internal variability of this phenomenon compared to a large number of SWW perturbation experiments performed before. Our aims are 1) to identify the dominant modes of the GMOC internal variability at decadal and longer time scales, 2) to find and distinguish the forcing mechanisms that control the different modes, and 3) to analyze the influence of the corresponding modes on the climate. The layout of this paper is as follows: A brief introduction of experimental setup and methods used in this study are given in section 2. In section 3 the results of numerical experiments and statistical analyses are presented and the physical mechanisms to support our findings are proposed. Finally, discussion and conclusions complete the paper in sections 4 and 5.

2. Experimental setup and method

Several numerical experiments are performed with the coupled general circulation Community Earth System...
The atmospheric model is initialized by the mean climatology from an Atmospheric Model Intercomparison Project (AMIP)-style experiment, which is performed using observed monthly sea surface temperatures and sea ice cover for the time period 1978–99 (Roeckner et al. 2007). For our SWW perturbation experiments, the wind stress anomalies are applied only to the ocean surface where there is water but not sea ice. The atmosphere is not directly influenced by this change. However, some feedback between the atmosphere and the ocean cannot be precluded after the perturbation, which might have impact on the air–sea heat flux and moisture transport in the atmosphere. The boundary conditions for orbital parameters and greenhouse gases in these wind perturbation experiments are identical with CTL. To guarantee that
the atmosphere and the ocean, especially in the NH, experience the feedback from any change in the SH wind stress, each perturbation experiment is integrated for 600 yr, with continuous perturbation starting from the last year of the 3000-yr CTL experiment.

3. Results

a. Mean state of the GMOC and AMOC

Figure 2 shows the 2000-yr mean state of simulated GMOC and AMOC in both CTL and H6K experiments. In CTL (Fig. 2a), the model produces the Deacon cell (Deacon 1937) in a range 30–40 Sv (1 Sv = 10⁶ m³ s⁻¹) near 50°S, due to the Ekman divergence. The simulated Deacon cell is comparably larger than that in previous studies (e.g., Döös and Webb 1994; Speer et al. 2000), which can be related to stronger SWW in the simulation. In CTL, the annual-mean value of zonal-averaged surface wind \( u \) component has a maximum of 9.2 m s⁻¹ near 50°S (not shown), whereas it is about 7.3 m s⁻¹ in the observations averaged over the period 1948–2010 (Kalnay et al. 1996). Such stronger SWW might be partly attributed to the different boundary conditions in CTL compared with the present-day condition: for example, a lower greenhouse gas concentration on the preindustrial level, which can result in a relative cooling climate background and hence strong SWW (Toggweiler and Russell 2008). Production of the Antarctic Bottom Water (AABW) is 8.4 Sv, which is consistent with the observation data (8.1 Sv; Orsi et al. 1999). The AMOC (Fig. 2b) has a reasonable structure as well, with the deep-water formation between 30° and 60°N penetrating to a depth of 2500 m. The maximum value of the AMOC index (defined as the AMOC streamfunction maximum at 30°N in the North Atlantic), which has a 2000-yr mean annual value of 16.0 Sv with standard deviation of 1 Sv, is in good agreement with the estimates of global circulation from the hydrographic data (15 ± 2 Sv; Ganachaud and Wunsch 2000). Its spectrum analysis exhibits variability with multidecadal as well as centennial time scales (Fig. 3). Such multidecadal variability has been interpreted as an atmosphere–ocean coupled mode in a present-day simulation (Zhu and Jungclaus 2008), using the same coupled model setup but without interaction of the land surface model.

In the H6K experiment, the structures of the GMOC and AMOC are similar to those in CTL. However, compared to CTL, the intriguing feature of the AMOC in the mid-Holocene simulation (Fig. 2d) is that its maximum value at 30°N of the North Atlantic is reduced by 2.5 Sv. This reduction is induced by changes in the orbital parameters and high-latitude warming (Fischer and Jungclaus 2010). Meanwhile, the Pacific overturning is also weakened (Fig. 2c). The SWW in H6K shows no significant change compared to CTL. The annual-mean value of zonal-averaged surface wind \( u \) component reaches a maximum of 9.2 m s⁻¹ near 50°S, in a similar magnitude with that in CTL (not shown). Therefore, the overturning circulation in the SO has not much difference with that in CTL (Fig. 2c).

b. Global modes of the overturning circulation

The two most important modes and the corresponding PCs from the EOF analysis that is based on the GMOC streamfunction are shown in Fig. 3. They explain 29.38% and 12.22% of total variance, respectively, and are well separated. For both modes, oscillations with vigorous multidecadal to centennial time scale can be seen in multilaper method (MTM) spectral analysis (Park 1992) of the corresponding PCs (Figs. 3c,d).

The first mode (Fig. 3e) is mainly constrained to the SH. Instead of having the common feature of the Deacon cell (Fig. 2a), it shows prominent enhanced southward flow between 45° and 65°S, penetrating to a depth
of more than 4000 m. As a consequence, the AABW formation in the SO is interrupted to a certain extent and becomes slightly weakened. Meanwhile, north of 45°S, the upper cell, comprising southward advection of upper circumpolar deep water (UCDW) and northward movement of the Antarctic Intermediate Water (AAIW), becomes weakened. The second mode has a similar structure in the SO but resembles the AMOC pattern in the north (Fig. 3f), with the maximum overturning around 30°N, at a depth of 1000 m. In this mode, the southward flows extend from 60°N to 60°S and dominate almost the whole overturning cell above a 4000-m depth, except for a northward flow around 40°S, over the top 800 m.

To investigate the relationship between these modes and the SWW, the composite maps of SWW zonal component $u$ with the corresponding PCs are shown in Fig. 4. In both modes, the SWW is shifted poleward and enhanced during both the austral winter (June–August) and summer (December–February). The change is much stronger during the austral summer (Figs. 4b,d). More precisely, in the first mode, the wind anomalies exhibit a zonal symmetric structure and are centered over the adjacent area of the South Pacific and SO, which is bounded by the Drake Passage (Fig. 4b). Such a pattern resembles the wind anomalies during a positive phase of SAM (e.g., Lovenduski and Gruber 2005; Sen Gupta and England 2006). Time series of this mode has the maximum correlation with the SAM indices, which are defined as the difference of zonal-mean sea level pressure between 40° and 65°S (Gong and Wang 1999), when they are in phase. It can be as significant as 0.61 and 0.50 for the austral winter and austral summer SAM indices, respectively. This implies that the GMOC pattern in the first mode is primarily dominated by the SAM-associated SWW change. In the second mode, the SWW anomalies are distributed quasi uniformly over the SO and show the maximum over the Drake Passage (Fig. 4d), but with a much smaller magnitude compared to that in the first mode. It is not surprising that the in-phase wind anomalies in the second mode are relatively smaller, because the response of the AMOC to the SWW change needs at least one century (Delworth and Zeng 2008). A detailed explanation of the forcing mechanism for this mode will be given in section 3d.

As mentioned in the introduction, we concentrate mainly on internally induced GMOC variability, possibly linked to SWW anomalies. To exclude any factor that might influence the EOF patterns in the CTL experiment due to different boundary conditions, we apply the same analysis to the H6K experiment. The two
dominant EOF patterns (Figs. 5c,d) in the H6K experiment have a similar structure to those in CTL and explain 28.06% and 17.48% of total variance, respectively. Composite analyses using the PCs (Figs. 5a,b) with the wind field also demonstrate a poleward shift and strengthening of the SWW when the GMOC and AMOC anomalies occur (not shown). This indicates that the wind-driven GMOC modes can be found in both present-day and mid-Holocene climate conditions and thus is generated through internal interactions in the climate system.

c. Wind perturbation experiments

The GMOC and AMOC anomalous streamfunctions in the three wind perturbation experiments $W_{\times 1.5}$, $W_{PLWD}$, and $W_{PLWD\times 1.5}$ are illustrated in Fig. 6. Notably, the structure of the GMOC streamfunction anomalies (Figs. 6a,c,e) seems to be a combination of those in the two EOF patterns from CTL. The wind-induced divergence over the latitude band of the SWW is intensified significantly, which leads to stronger upwelling over the Drake Passage. North of this band, the southward flow
in the intermediate depth has slightly positive anomalies. In the AMOC (Figs. 6b,d,f), there is an increase in the NADW outflow, along with an increased NADW formation. The magnitude of the GMOC and AMOC streamfunction anomalies shows the maximum in the WPLWD × 1.5 experiment, resulting from both poleward shift and intensified SWW. The NADW outflow is approximately 3.6 Sv stronger at a depth of about 2200 m than that in CTL. Comparably, the anomalies are 1.9 and 1.3 Sv in the W × 1.5 and WPLWD experiments, respectively. This suggests that these two effects related to the SWW are additive: that is, the patterns derived from WPLWD × 1.5 can be obtained as a superposition of the corresponding structures induced by each forcing, for both the global and the Atlantic streamfunctions. It is worth mentioning that the strengthening of the NADW outflow and the associated overturning cell in the North Atlantic seems to be compensated by a more pronounced Indo-Pacific upwelling in WPLWD, which leads to no remarkable change in the GMOC streamfunction (Fig. 6c). This suggests that the Indo-Pacific overturning circulation is likely to be more sensitive to the shift of the SWW than its strengthening. The results reveal that the GMOC and AMOC are sensitive to both types of changes related to the SWW in our simulations.

d. The SO mode and the NH mode

As seen in Fig. 3, the first mode shows maximum values south of 30°S, whereby the feature in the NH is more pronounced in the second mode. Furthermore, the SWW anomalies (Fig. 4) reach the maximum in the adjacent area of the SO and the Indo-Pacific Ocean in

Fig. 4. Composite maps of the seasonal-mean surface zonal wind anomalies (m s⁻¹) with (a),(b) PC1 and (c),(d) PC2 during (left) austral summer (December–February) and (right) austral winter (June–August) in CTL. (e),(f) As in (c),(d), but for PC2 lagging the wind field for 305 yr. Values with significance higher than 95% using Student’s t test are considered.
the first mode and distribute quasi uniformly in the second mode. Thus, we infer that the first mode can be mostly projected onto the SO, especially the longitude band of the Indo-Pacific Ocean, and the second mode onto the Atlantic Ocean. To verify it, we apply the composite analysis to the Indo-Pacific overturning circulation and the AMOC using the two PCs, respectively.

When the first PC is projected onto the Indo-Pacific streamfunction, the overturning pattern (Fig. 7a) resembles that of the first mode. The stronger SWW cause an enhanced divergence centered on the maximum latitude band of the SWW. This leads to a 3-Sv increase of the upwelling centering around 55°S, close to where the Drake Passage is located. This increased upwelling can extend downward to a depth of more than 4000 m. North of 45°S, the enhanced divergence also leads to a decrease of the southward flow above 3000-m depth. In contrast, the anomaly in the NH is not pronounced, which implies that response of the Indo-Pacific overturning circulation to SWW change has some years lag with the direct response in the SO. However, when the first PC is projected onto the AMOC, the anomaly is moderate and it shows a slight decrease of the AMOC with a magnitude of less than 0.5 Sv at maximum (Fig. 7b). Comparably, the first mode is mainly identified in the SO.

Contrary to the first mode, the composite maps of the Indo-Pacific streamfunction and AMOC with PC2 (Figs. 7c,d) demonstrate that the second mode is primarily captured by the Atlantic Ocean. PC2 is highly correlated to the AMOC index (with maximum correlation of 0.47 at 0 lag; not shown). Therefore, spectral analysis of PC2 (Fig. 3d) shows a similar periodicity with that of the AMOC index. In the second mode, the NADW outflow is strengthened and NADW formation also has an increase of more than 1 Sv (Fig. 7d). Simultaneously, the upwelling in the North Pacific decreases by approximately 0.2 Sv. A seesaw-like change in the overturning circulation between two ocean basins can be obtained, which has been attributed to the SWW-induced decrease in the salinity contrast between the Indo-Pacific and the Atlantic basin in a previous study (Sijp and England 2009). Here, we refer to this mode as the NH mode.

The wind anomalies with PC2 (Fig. 4) are much weaker than those with PC1. Because of the rather long response time for the AMOC to the SWW change (e.g., Delworth and Zeng 2008), one can speculated that there might be some lag between PC2 and the wind anomalies.
To find the possible lag time, we apply the cross correlation between PC1 and PC2. The correlation coefficient at 0 lag is nearly 0 and there is no strong correlation in other lags. Then a 91-yr running mean is applied to both PCs prior to the correlation calculation, which enables us to concentrate mainly on the lag longer than one century. Interestingly, the correlation between the filtered PCs (Fig. 8a) shows a maximum value when PC1 leads PC2 for 305 yr, which indicates that the response time of the AMOC to the SWW might be around 3 centuries. The wind perturbation experiments provide us the possibility to examine this response time. The GMOC and AMOC anomalies in $W_{PLWD \times 1.5}$ within different integration intervals relative to the CTL experiment are illustrated in Fig. 9. The quick response of the SO to the SWW change can be clearly displayed in the GMOC anomalies (Figs. 9a–f), with the clear structure formed immediately after the perturbation. Comparably, the AMOC anomalies (Figs. 9g–l) are characterized by a propagation of SWW-induced NADW outflow strengthening from 30°S to 60°N. After about one decade, the NADW outflow becomes increased. However, in the first 50 yr, the SWW anomalies hardly have any influence on the AMOC north of 30°N. It takes at least one century for the anomalies to reach the North Atlantic (Fig. 9i), which supports the previous view (Delworth and Zeng 2008). However, the AMOC anomalies keep increasing until it reaches the maximum after 300-yr perturbation (Figs. 9j–l), after which the AMOC anomalies are in a similar magnitude (Fig. 6f). Such an interval strongly matches the 305-yr lag time in the cross correlation of PC1 and PC2. Therefore, we
conclude that it takes 3 centuries for a full adjustment of the AMOC to SWW change in our model. The cross correlation of PC1 with AMOC (Fig. 8b) shows the peak at the same lag with that of PC1 and PC2 and further supports our finding.

Based on this interval, we calculate the composite maps of PC2 with the wind field at 305-yr lag. The wind anomalies (Figs. 4e,f) resemble those with PC1 (Figs. 4a,b) but are still much smaller in the magnitude. This result suggests that the NH mode can also be controlled by other forcing factors. Thus, the cross correlations of surface downward heat flux and freshwater flux including the river runoff with PC2 are also shown (Figs. 8c,d). Here, the time series of these two indices are constructed according to Delworth and Greatbatch (2000) by averaging over the region 50°–70°N, 55°–15°W, where the NADW formation takes place. PC2 is highly correlated to both indices, showing the highest coefficients of 0.63 when the heat flux leads PC2 by 18 yr and −0.49 when the freshwater flux leads PC2 by 19 yr (Figs. 8c,d). It indicates that the second mode, which resembles the AMOC, is driven by a combined effect of different forcing. However, an entire understanding of the controlling mechanism for the AMOC is beyond the scope of this study.

e. Response of the SST and the sea ice to different modes

As the SWW strengthens, the SSTs associated with the two modes are affected differently (Fig. 10). Notably, SST change over the latitude band of the SWW has much similarity in both modes: the Atlantic sector experiences a temperature increase by as much as 0.5°C, whereas the Pacific sector is slightly colder (Figs. 10a,b). South of this band, the stronger upwelling results in a slight warming in the SO, which in turn decreases the sea ice thickness in the associated regions (Figs. 10c,d). However, north of this band, the SST response to the SWW change in the two modes seems to be the opposite. In the SO mode, the SST anomaly is mainly negative, except in the Labrador Sea and the Norwegian Sea (Fig. 10a). One notes the prominent zonal SST dipole in the SH resembles the SAM-associated SST anomaly pattern from the previous studies (Sen Gupta and England 2006, 2007). Meanwhile, intensified upwelling in the tropical and subtropical Pacific Ocean generates a strong cooling there, because of increased Ekman transport associated with stronger SWW. In the NH mode, warmer SST covers almost the whole North Atlantic and North Pacific (Fig. 10b), which is caused by increased northward
heat transport associated with the strengthened AMOC in the North Atlantic and less cold deep water upwelling due to the weakened overturning circulation in the North Pacific. This leads to a decrease in the sea ice thickness over the North Pacific (Fig. 10d).

4. Discussion

In this study, we have presented two distinct multi-decadal to centennial modes of the GMOC, which result from internal variability of the climate system in the earth system model COSMOS. The physical mechanism providing the necessary energy to sustain an overturning circulation (Kuhlbrodt et al. 2007) is a crucial aspect in the GMOC and AMOC. Besides diapycnal mixing (Jeffreys 1925; Munk and Wunsch 1998; Prange et al. 2003; Rahmstorf et al. 2005), the SWW-induced divergence (Toggweiler and Samuels 1998; Gnanadesikan et al. 2005; Schewe and Levermann 2010) is one of the driving mechanisms. We examined through the analyses of long-term integrations as well as sensitivity experiments that both the SO mode and the NH mode of the GMOC in our simulations are linked to natural variations of the SWW. This internal variability can generate a strengthening and deepening of the Deacon cell in the SO. Furthermore, strengthening or poleward shift in the SWW can cause considerable increase in the AMOC. Different internal AMOC patterns has been found in

Fig. 8. Cross correlation of (a) PC1 and PC2, (b) PC1 and AMOC streamfunction maximum at 30°N, (c) surface downward heat flux and PC2, and (d) net freshwater flux into the ocean including the river runoff with PC2. Here, time series of PC1 and PC2 have been filtered by a 91-yr running mean prior to the correlation calculation. The heat flux and freshwater flux time series are averaged over the region 50°–70°N, 55°–15°W and then filtered by the same strength with PCs. The dashed lines indicate the 95% significance level when 20 degrees of freedom are selected concerning the autocorrelation in the time series. Positive lags (in years) indicate the first time series lags the second one and vice versa.
FIG. 9. Annual-mean (a)–(f) GMOC and (g)–(l) AMOC streamfunction anomalies (Sv) in WPLWD×1.5, relative to CTL. The numbers of year indicate the integration time after the perturbation.
a previous study (Park and Latif 2008), where the authors propose a North Atlantic source of multidecadal variability and a SO controlled multicentennial variability of the AMOC in their multimillenial integration. They argue that both of them are associated with sea ice changes.

Results from the mid-Holocene experiment show that the two modes of the GMOC have a similar structure with that in the CTL experiment, although GMOC and AMOC streamfunctions have different magnitudes compared to their present values. The mechanisms found are valid for preindustrial as well as mid-Holocene conditions, indicating the independence of these modes on the climate background conditions during the Holocene. A natural next step would be the analysis of different climate states with strongly deviating background conditions (e.g., Ruhlemann et al. 2004).

The modes identified here may mask any changes and ongoing trends in the GMOC, and therefore it is important to understand the underlying mechanisms. Changes in the SWW can directly cause a rapid and pronounced local response in SO. In the SO mode, we can see that the intensified and poleward shifted SWW lead to a stronger local divergence through increased downward Ekman transport, which permits an increased northward flow at the surface north of the latitude band of the SWW and a deepening and strengthening of the Deacon cell over the latitude band of the Drake Passage. The wind anomaly structure associated with this mode resembles that in a positive phase of the SAM. Furthermore, the in-phase correlation between the SAM and time series of SO mode is high. This indicates that the SO mode can be linked to a SAM-like SWW change. Change in the SAM can be induced by the internal variability of the climate system (e.g., Cai and Watterson 2002; Deser et al. 2012). The physical process involves the equator-to-pole temperature gradient in the upper troposphere, which has been considered to be the dominant factor controlling the atmospheric circulation (e.g., Barron and Washington 1982; Toggweiler and Russell 2008). A higher gradient generates a strengthening of the SWWs and causes them to shift poleward (Toggweiler and Russell 2008). Variability of this temperature gradient can be induced by the internal atmospheric dynamics (Cai and Watterson 2002).
maintained by the embedded synoptic eddies (Yu and Hartmann 1993). Many previous studies (e.g., Toggweiler and Samuels 1993; Gnanadesikan and Hallberg 2000; Hirabara et al. 2007; Sijp and England 2009) have examined the relationship between the SWW-induced SO overturning circulation change and AMOC variation. The basic mechanism can be explained as follows: The wind-induced increased Ekman transport, on one hand, leads to enhanced ventilation in the SO (Russell et al. 2006; Delworth and Zeng 2008). Stronger upwelling removes the deep water and increases the outflow from north to the SO, which needs to be compensated by more deep-water formation in the north. The North Atlantic is considered the only deep-water formation place north of the latitude band of SWW (Toggweiler and Samuels 1995). Thus, the NADW formation is expected to strengthen, which has been captured by the NH mode in our simulation. On the other hand, the increased Ekman transport pushes more water to move northward and to be injected into the AAIW and results in a deepening of the thermocline (Gnanadesikan and Hallberg 2000). As a consequence, the NADW formation increases, which in turn increases the NADW outflow and ventilation in the SO. Santoso et al. (2006) analyze the natural variability of the circumpolar deep water using their long integration of a coupled GCM and find that the NADW formation is mainly buoyancy driven, but with a weak connection with the SWW on multicentennial time scales. We show that the NH mode, mainly projected onto the AMOC, is driven by a combination of wind forcing and buoyancy forcing, with the last one dominating. Furthermore, we highlight that the impact of the SWW on the AMOC needs 3 centuries to reach the full adjustment, which is attributed to the meridional advection of the anomalies from south to north in the Atlantic. Results from our wind perturbation experiments are also consistent with their argument that the SWW has a more direct influence on the SWW outflow (Santoso et al. 2006).

A number of factors for the wind-induced GMOC need to be taken into account. One is linked to the uncertainty in the parameterization of mesoscale eddies in the SO. Coarse-resolution models without resolving these eddies can exaggerate the effect on the wind-induced Ekman transport (e.g., Straub 1993; Hallberg and Gnanadesikan 2006; Böning et al. 2008; Farneti et al. 2010; Farneti and Delworth 2010). The eddy-induced contribution is important for the water mass transport in the vicinity of the ACC, because it balances the horizontal northward Ekman transport as well as the vertical Ekman pumping (Sallée et al. 2008). The heat flux-induced temperature change can also be reversed in an eddy-saturated ocean state compared to the direct influence from the SWW change (Hogg et al. 2008; Screen et al. 2009; Spence et al. 2010), because the total heat flux response is more in an eddy-permitted model setup than that in a coarse resolution one by a factor of 2 (Hogg et al. 2008). However, Treguier et al. (2010) report that an increase of overturning circulation as a response to SH wind is due to the time-mean component and is compensated by an increased buoyancy gain at the surface. We suggest that the internal variability of the GMOC associated with the SWW change on longer time scales needs to be considered, including the role of mesoscale eddies, in order to understand GMOC changes under past, present, and future climate conditions.

The cold temperature response in the tropical and subtropical Pacific Ocean related to the SO mode (Fig. 10a) is due to intensified upwelling in the Pacific and Indian Oceans (Figs. 3e, 7a). SST shows a large-scale zonally asymmetric response between the Atlantic and Pacific–Indian Oceans (Fig. 10a) caused by wind-induced changes (Figs. 4a,b) at the mixed layer depth. Enhanced upwelling over the latitude band of the SWW pumps more cold recirculated deep water to the surface and causes negative SST anomalies west of the Drake Passage (Fig. 10a). Similar asymmetric heat budgets were observed recently by Sallée et al. (2010) in relation to the southern annular mode (projecting strongly onto the patterns in Figs. 4a,b). The stratification exhibits opposite anomalies over the South Atlantic Ocean. Positive SST anomalies in the South Atlantic Ocean in the NH mode can be furthermore potentially linked to increased AMOC transports of warm and salty Indian Ocean waters into the cold and fresh South Atlantic. There is evidence from modeling studies (De Ruijter 1982; Biastoch et al. 2009) manifesting that this interbasin transport can be modulated by change in the SWW: for example, a poleward shift can result in a warming in the South Atlantic, which affects the AMOC on decadal time scales (Biastoch et al. 2008). Such effect may be also relevant for an onset of a strong present AMOC from the weaker glacial mode during deglaciation (Knorr and Lohmann 2003, 2007).

The wind perturbation experiments demonstrate that the GMOC and AMOC are sensitive to the SWW change in our model and capture features of the observed system. A 1.5-times-amplified SWW in $W_{x1.5}$ and 3.5° southward shift of the SWW in $W_{PLWD}$ result in an intensifying of the divergence over the latitude band of the SWW (Figs. 6a,c), as can be seen in the SO mode. The corresponding AMOC in both experiments acts to reinforce (Figs. 6b,d). In the recent Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4), the models (as a group) predict a 25% intensification and 3.5° poleward shift of the SWW (Fyfe
et al. 2007), which is comparable to the SWW change in WPLWD×1.5. Here, WPLWD×1.5 shows an intensified northward Ekman transport across the ACC and increased upwelling over the SO, consistent with the SO overturning circulation response to the SWW change in IPCC AR4 runs (Fyfe and Saenko 2006). However, the AMOC in the WPLWD×1.5 experiment strengthens by more than 3 Sv, which is contrary to the predicted weakening of the AMOC in most IPCC runs that results from the increase of anthropogenic CO2 release (e.g., Schmittner et al. 2005).

Paleostudy based on model–data comparison (Varma et al. 2011) indicates that the centennial-scale variability in the SWW during the Holocene can be caused by the change in solar forcing, which implies possible application of the modes found in this study to understand the past GMOC change. From the future perspectives, CO2-induced warming in the North Atlantic might overlay the AMOC response to the poleward-intensifying SWW. However, it is unclear whether the response of the overturning circulation in the NH takes longer or is more pronounced than changes induced in the south (Klinger and Cruz 2009). The observed SAM undergoes a pronounced upward trend in the last 50 yr (Marshall 2003), which has been attributed to stratosphere ozone depletion (e.g., Thompson and Solomon 2002) and greenhouse gas increases (Hartmann et al. 2000; Marshall et al. 2004). This trend is also related to the observed increase in the SWW. In the same period, the Antarctic Peninsula in general experiences a warming trend, but with distinct regional difference (Turner et al. 2005). Driven by surface warming and freshening, a decrease in the projected subduction of SO upper-water masses implies a slowdown in the SO circulation in the future (Downes et al. 2010). A reduction in the subduction of intermediate waters affects the ocean’s capacity to sequester heat and carbon and would work against changes in the wind-induced variations in the Southern Ocean.

5. Conclusions

Identification of the link between the SWW and the two leading modes in the internal variability of the GMOC on multidecadal and centennial time scales enables us to confirm that the winds play an essential role in modulating the GMOC. Strengthening and poleward shift of the SWW generate a direct response in the SO, with Ekman transport enhanced and the Deacon cell strengthened. Such a change in the SO circulation produces an increase of the NADW outflow in one decade, eventually propagating this anomaly to the North Atlantic Ocean after centuries.

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