The Role of Oscillating Southern Hemisphere Westerly Winds: Global Ocean Circulation

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ABSTRACT

In the framework of a sea ice–ocean general circulation model coupled to an energy balance atmospheric model, an intensity oscillation of Southern Hemisphere (SH) westerly winds affects the global ocean circulation via not only the buoyancy-driven teleconnection (BDT) mode but also the Ekman-driven teleconnection (EDT) mode. The BDT mode is activated by the SH air–sea ice–ocean interactions such as polynyas and oceanic convection. The ensuing variation in the Antarctic meridional overturning circulation (MOC) that is indicative of the Antarctic Bottom Water (AABW) formation exerts a significant influence on the abyssal circulation of the globe, particularly the Pacific. This controls the bipolar seesaw balance between deep and bottom waters at the equator. The EDT mode controlled by northward Ekman transport under the oscillating SH westerly winds generates a signal that propagates northward along the upper ocean and passes through the equator. The variation in the western boundary current (WBC) is much stronger in the North Atlantic than in the North Pacific, which appears to be associated with the relatively strong and persistent Mindanao Current (i.e., the southward flowing WBC of the North Pacific tropical gyre). The North Atlantic Deep Water (NADW) formation is controlled by salt advected northward by the North Atlantic WBC.

1. Introduction

This paper extends the study of Cheon et al. (2018, hereinafter C18), which showed the following: 1) the formation rate of coastal polynyas as being in linear correlation with the oscillating intensity of Southern Hemisphere (SH) westerly winds, 2) the favorable conditions of the Weddell Sea in relation to open-ocean polynyas, and 3) the importance of recovery time of deep ocean heat content, which controls the occurrence frequency of open-ocean polynya. Coastal polynyas and open-ocean polynyas are closely linked to near-boundary convection and open-ocean deep convection, respectively; therefore, they play a major role in the Antarctic Bottom Water (AABW) formation. Moreover, it has been proposed that the SH westerly winds controlling northward Ekman transport in the Southern Ocean (SO) affect the surface western boundary currents (WBCs) in the North Pacific and Atlantic (McDermott 1996; Klinger and Cruz 2009). Herein, impacts of oscillating SH westerly winds on global ocean circulation are investigated from two perspectives: 1) the buoyancy-driven teleconnection (BDT) mode that results from SO sea ice/ocean interaction and directly controls AABW formation and 2) the Ekman-driven teleconnection (EDT) mode that results from northward Ekman transport and directly controls the upwelling of Circumpolar Deep Water (CDW) and outflow of North Atlantic Deep Water (NADW), as well as remotely influencing the North Pacific and Atlantic surface WBCs and the Atlantic meridional overturning circulation (MOC).

The SH westerly winds are closely linked to the southern annular mode (SAM) in association with their fluctuations in the latitudinal position and strength (Gong and Wang 1999; Thompson and Wallace 2000). At the surface, they are much stronger in the SH than in the Northern Hemisphere westerly winds (Gille 2005) owing to a unique zonal band that is not interrupted by the continent in the SO. The strongest surface westerly winds are located in the latitude band between 55° and 50°S.
(Gille 2005; Gent 2016) and are shifted poleward (equatorward) and intensified (weakened) in a positive (negative) phase of the SAM (Gong and Wang 1999; Thompson and Wallace 2000). According to analysis of satellite observation and atmospheric reanalysis data, the maximum zonal wind stress over the SO increased by at least 20% between 1980 and 2010 (Swart and Fyfe 2012; Bracegirdle et al. 2013; Farneti et al. 2015; Gent 2016) in association with SAM variation. The SAM changed from a prolonged negative phase beginning in the early 1960s to a positive phase in the 1980s and has maintained its positive phase until the present day (Gordon et al. 2007; Hartmann et al. 2013; Cheon and Gordon 2019). As illustrated in C18, the SH surface westerly wind variability is highly correlated with the SAM; the trend of this relationship is expected to be positive in the future climate (Zheng et al. 2013). Therefore, it is important to investigate the response of global ocean circulation to the SH westerly wind fluctuations associated with the SAM.

In the study by Cheon et al. (2014), who used a sea ice–ocean coupled general circulation model (GCM), SH westerly winds were proposed to trigger the formation of open-ocean polynya in the Weddell Sea. The intensified westerly winds induce spinup of the cyclonic Weddell Gyre, weakening and raising the pycnocline and helping the relatively warm Weddell Deep Water (WDW) to rise to the surface. The upwelled, warm WDW melts the overlying sea ice, or prevents it from forming, generating open-ocean polynya and deep convection. This hypothesis was clearly demonstrated by the analyses of Cheon and Gordon (2019) using diverse, long-term observation and reanalysis data. They showed that the relatively strong Maud Rise polynya observed from September to November of 2017 was generated by the combined effect of weak stratification near the Maud Rise and the wind-induced spinup of the cyclonic Weddell Gyre, which is in agreement with the observation data of Johnson (2008). It covers much of the ocean floor globally, except for in the Arctic Ocean and the northern North Atlantic Ocean, where the NADW overlies the ocean bottom (Mantyla and Reid 1983; Orsi et al. 2001). The NADW, whose upper water masses form in the Labrador Sea and lower water masses form in the Greenland–Iceland–Norway (GIN) Seas, flows southward as the deep WBC with the mean volume transport ranging from 12 to 17 Sv (1 Sv = 10⁶ m³ s⁻¹) (Schmitz and McCartney 1993; Ganachaud and Wunsch 2000; Lumpkin and Speer 2003; Schott et al. 2004). Analysis of a moored array in the western basin of the equatorial Atlantic (Hall et al. 1997) shows that the transition layer between the lower NADW and the AABW rises and falls synchronously with the quasi-annual cycle of AABW transport. This so-called “bipolar seesaw balance” between two water masses was also simulated on the multcentennial time scale by the sea ice–ocean coupled models with salt flux directly manipulated (Brix and Gerdes 2003) or indirectly manipulated by the dynamic air–sea ice interaction (Cheon and Stössel 2009) in the Weddell Sea. The inverse correlation between the deep WBCs of NADW and AABW associated with this bipolar seesaw balance was explained by Cheon and Stössel (2009) from the viewpoint of conservation of potential vorticity.

The SO has a unique zonal band that is not interrupted by the continent and passes through the Drake Passage of about 2000-m depth between the Antarctic Peninsula and South America. In this zonal band, zonal pressure gradients averaged around the globe are zero above this depth; therefore, there cannot be a net geostrophically balanced meridional flow. The overlying SH westerly winds push a large amount of very cold, fresh Antarctic Surface Water (AASW) away from the Antarctic continent via northward Ekman transport, which is geostrophically balanced and not affected by the above constraint. Since any subsurface return flow must be geostrophically balanced and therefore cannot exist, the net poleward geostrophic flow balancing this Ekman transport exists only below depths where meridional topographic barriers such as the East Scotia Ridge support the zonal pressure gradients (Toggweiler and Samuels 1995). This deep southward return flow is mixed with the CDW, which upwells south of the Antarctic Circumpolar Current (ACC). Numerous modeling studies have shown not only the impact of the SH westerly on the formation rate and physical properties of Antarctic Intermediate Water (AAIW) (Ribbe and Tomczak 1997; Sørensen

Regardless of where it forms, 60%–80% of its volume eventually ends up in the Pacific Ocean, 20%–30% in the Indian Ocean, and 5%–10% in the Atlantic Ocean, which is in agreement with the observation data of Johnson (2008). It covers much of the ocean floor globally, except for in the Arctic Ocean and the northern North Atlantic Ocean, where the NADW overlies the ocean bottom (Mantyla and Reid 1983; Orsi et al. 2001). The NADW, whose upper water masses form in the Labrador Sea and lower water masses form in the Greenland–Iceland–Norway (GIN) Seas, flows southward as the deep WBC with the mean volume transport ranging from 12 to 17 Sv (1 Sv = 10⁶ m³ s⁻¹) (Schmitz and McCartney 1993; Ganachaud and Wunsch 2000; Lumpkin and Speer 2003; Schott et al. 2004). Analysis of a moored array in the western basin of the equatorial Atlantic (Hall et al. 1997) shows that the transition layer between the lower NADW and the AABW rises and falls synchronously with the quasi-annual cycle of AABW transport. This so-called “bipolar seesaw balance” between two water masses was also simulated on the multcentennial time scale by the sea ice–ocean coupled models with salt flux directly manipulated (Brix and Gerdes 2003) or indirectly manipulated by the dynamic air–sea ice interaction (Cheon and Stössel 2009) in the Weddell Sea. The inverse correlation between the deep WBCs of NADW and AABW associated with this bipolar seesaw balance was explained by Cheon and Stössel (2009) from the viewpoint of conservation of potential vorticity.

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et al. 2001; Oke and England 2004; Santoso and England 2004) but also its remote influence on the Atlantic MOC (Toggweiler and Samuels 1993, 1995; McDermott 1996; Rahmstorf and England 1997; Gnanadesikan 1999; Gnanadesikan and Hallberg 2000; Gnanadesikan et al. 2007). However, the SO eddy compensation simulated in eddy-resolving models or in non-eddy-resolving models with the Gent–McWilliams (GM) scheme (Gent and McWilliams 1990), whose coefficient is variable in space and time, is known to weaken the influence of the SH westerly winds on the SO mean MOC (Farneti and Gent 2011; Hofmann and Morales-Maqueda 2011; Gent and Danabasoglu 2011; Gent 2016) and the ACC (Hallberg and Gnanadesikan 2006).

Regardless of the SO eddy compensation effect, the northward Ekman transport in the SO is linearly proportional to the strength of the overlying westerly winds (Gent 2016), which is a key factor controlling the EDT mode. According to previous studies using idealized ocean models with single- and two-basin geometries (McDermott 1996) and a coarse-resolution ocean GCM with realistic bottom topography (Klinger and Cruz 2009), the signal generated by the intensified SH westerly winds can propagate in the upper ocean to the North Atlantic and Pacific. This remotely affects each region’s WBCs, which are mainly controlled by local wind stress curl. In both modeling results, the increase rate of the North Atlantic WBC was much larger than that of the North Pacific when the SH westerly winds were uniformly intensified in the zonal direction. However, the cause of this different response has not been investigated, and neither has the detailed process by which the Atlantic MOC remotely responds to varying SH westerly winds. These aspects are investigated in the present paper from the perspective of the EDT mode. Whether or not the Antarctic MOC that is activated by the BDT mode can affect the Atlantic MOC is a further issue addressed in this study. This is considered in association with the bipolar seesaw balance between the AABW and NADW (Hall et al. 1997; Broecker 1998; Brix and Gerdes 2003; Cheon and Stössel 2009).

This paper aims to investigate the responses of the global ocean circulation to the oscillating SH westerly winds by use of the BDT and EDT modes. The model configuration and experimental setup are described in the following section, and the basic states of the global ocean’s properties and simulation of circulation in the absence of oscillating wind forcing in the control simulation (CTRL) are then described. In section 3 we investigate variations of the Antarctic MOC indicative of the formation rate of AABW, the abyssal flows, and the transition layer between the AABW and NADW in association with the BDT mode, and segues into the EDT mode, which remotely affects the WBCs of the North Pacific and North Atlantic and the Atlantic MOC. A summary and discussion are provided in section 4.

2. Model

a. Model and experimental design

This study employs the Modular Ocean Model version 4 (MOM4) of the Geophysical Fluid Dynamics Laboratory (GFDL), in which the primitive equation ocean model (Griffies et al. 2004) is coupled with the dynamic and thermodynamic sea ice model (Winton 2000), the two-dimensional global atmosphere energy balance model (Gerdes et al. 2006), and the land model (Anderson et al. 2004). The ocean model covers the global ocean from 80°S to 90°N with horizontal resolution of 2° × 1/3°–1° in longitude and latitude; it has 50 semivariable layers to 5500-m depth with the bottom layer following the actual topography data (Smith and Sandwell 1997). The thickness of the upper 22 layers is uniform at 10 m, and that of the 28 lower layers gradually increases to about 400 m at 5500-m depth. The mesoscale eddy transport is parameterized on isopycnal surfaces following the GM scheme (Gent and McWilliams 1990). The GM and Redi diffusivity coefficients are capped at 600 m² s⁻¹, which is generally used in the standard version of GFDL climate model version 2.1 (CM2.1), in order to simulate the realistic volume transport of the ACC under the observed wind stress strength. When the SH westerly winds are intensified, the GM coefficient is expected to increase, which causes the SO eddy MOC to turn in the direction opposite to the SO mean flow MOC (Gent 2016). Therefore, under intensifying SH westerly winds, capping the GM and Redi diffusion coefficients at 600 m² s⁻¹ can act to amplify the volume transport of the ACC owing to the underestimated eddy activity. This is assumed to decrease the meridional dissipation of potential energy reserved in the Antarctic Polar Front and thus enhance the meridional density gradient from the surface downward across the ACC. However, as previously stated, the northward Ekman transport, an important factor associated with the EDT mode, is in a linear correlation with the strength of SH surface westerly winds regardless of the effect of SO eddy compensation. Moreover, in this study, the Antarctic MOC that is indicative of the formation rate of the AABW is regulated by the SO air–sea ice–ocean interactions associated with the BDT mode. Therefore, the uniform GM and Redi diffusivity coefficients are judged to be suitable for this type of sensitivity experiment. The sea ice model consists of a dynamic component including the viscous-plastic constitutive law (Hibler 1979) and a thermodynamic component that follows the formulation.
of Winton (2000), that is, with one snow layer and two sea ice layers and run on the same grid as the ocean model. Whereas the atmosphere energy balance model provides thermodynamic forcing for the sea ice–ocean model, the wind field and precipitation are derived directly from the ERA-15 reanalysis data of the European Centre for Medium-Range Weather Forecasts (ECMWF).

As stated in C18, one standard deviation of the anomalous zonal wind stress derived from the monthly mean of the ECMWF ERA-40 and ERA-Interim data corresponds to about 49% of the climatological range. The current SO has experienced a prolonged positive SAM with its preceding negative mode beginning around 1965 (Gordon et al. 2007). This indicates that the most recent SAM period is longer than 50 years, whereas in the early twentieth century the periods were less than 30 years. In this study, we employ two experimental designs: SW050P02, in which the zonal wind stress oscillating within the range of ±50% over 50 years (two cycles per century) is added to its original value; and SW050P04, in which the zonal wind stress oscillating with the same amplitude over 25 years (four cycles per century) is added to its original value. These oscillations are applied only to positive values of the SH zonal wind stress that is indicative of the SH westerly winds (refer to Fig. 2 of C18). After 600 years of spinup, the model was run with each wind forcing for another 200 years. In C18, six experimental designs were employed to investigate the detailed responses of coastal and open-ocean polynyas in the Weddell and Ross Seas to diversely oscillating wind forcing consisting of two amplitudes and three periods. However, the present study employs only the two experimental designs (among the six designs used in C18) that are closest to the main characteristics of the SAM. Here we focus on the diverse responses of the global ocean circulation from the perspective of BDT and EDT. As presented in SW050P02 and SW050P04 in C18, the intensifying (weakening) SH westerly winds increase (decrease) the formation rate of coastal polynya and the ensuing ice-to-ocean salt flux in the western Weddell and Ross coasts, thereby enhancing (weakening) near-boundary convection. Open-ocean polynyas occur with the wind forcing being in a positive phase. This leads to open-ocean deep convection, which has much larger impact on the formation rate of AABW than near-boundary convection associated with coastal polynyas. The SO convection triggered by the oscillating SH westerly winds is a major forcing of the BDT mode.

b. Model evaluation

In this section the basic state of the global ocean’s properties and circulation simulated in CTRL are compared with the observation data derived from the World Ocean Atlas 2013 (WOA13; Locarnini et al. 2013; Zweng et al. 2013) and the reanalysis data derived from the Simple Ocean Data Assimilation version 3.3.1 (SODA3.3.1, downloaded from http://www.atmos.umd.edu/~ocean/index_files/soda3.3.1_mn_download.htm). Thus, we assess whether or not the simulated global ocean is sufficiently close to reality and is appropriate for this study. The potential temperature and salinity at the surface and 4500-m depth are averaged over 200 years in CTRL; these are compared with those derived from WOA13 (Fig. 1). The warm pools and cold tongues in the equatorial Pacific and Atlantic are well simulated, as are the high-salinity regions in the subtropics. In the midlatitudes the simulated surface temperature and salinity are lower overall than the observations, except for in the northwestern Pacific and Atlantic north of 40°N, which show positive surface temperature and salinity biases (Figs. 1c,f). As shown in Fig. 2, this results from WBCs such as the Kuroshio, Gulf Stream, and Brazil Current being simulated as slightly stronger than the reanalysis data. Since the surface WBCs play a key role in transporting the warm and saline water masses from subtropics to the high latitudes, the relatively strong simulation of WBCs in CTRL causes the surface water to be colder and fresher in most midlatitudes but to be warmer and saltier in the northwestern Pacific and Atlantic north of 40°N, as compared with the observations. Owing to the limitations of the coarse-resolution model, the simulated ACC in CTRL does not show the meandering feature that is captured in SODA 3.3.1 of the eddy-permitting model.

Compared with the observations, the simulated abyssal water is clearly warmer and saltier in the Atlantic and is colder and fresher in the Weddell Sea (Figs. 1i,l), which results from the enhanced formation of the NADW (characterized by the relatively warm and salty water mass) and of the AABW (characterized by the relatively cold and freshwater mass). As shown in Fig. 3, this is due to the Atlantic and Antarctic MOCs being simulated as stronger in CTRL than they are in the reanalysis data. Although persistent, larger-scale open-ocean polynya and deep convection (as confirmed in C18) never occurred in CTRL, some small episodes of convection associated with the convection scheme of Rahmstorf (1993), together with the GM and Redi coefficients being capped at 600 m²·s⁻¹, are the causes of the relatively strong Antarctic MOC. The simulated Atlantic MOC of about 25 Sv is 2 Sv stronger than the reanalysis product but is intermediate within the range of modeled historical MOC in phase 5 of the Coupled Model Intercomparison Project (CMIP5) (Cheng et al. 2013). Its maximum core is located farther north, which is more reasonable than the reanalysis data because the NADW predominantly forms in the Labrador Sea and Greenland Sea (Killworth 1983; Marshall and Schott 1999). The NADW outflow passing through 30°S reaches
FIG. 1. Horizontal distributions of annual mean (left) potential temperature and (right) salinity at (a)–(f) the surface and (g)–(l) 4500-m depth, derived from (a),(d),(g),(j) the World Ocean Atlas 2013 (WOA13) and (b),(e),(h),(k) CTRL. The variables derived from CTRL are averaged over the whole period (200 yr), and (c),(f),(i),(l) the differences between CTRL and WOA13 are presented. Note that even for the same variable the color bars for the surface and 4500-m depth are different.
18.7 Sv, which is in good agreement with the observed estimate [18 Sv according to Dong et al. (2009)]. Although not shown here, the simulated ACC of about 192.8 Sv is larger than the observed estimates, which span from 110 to 150 Sv (Whitworth et al. 1982; Orsi et al. 1995; Cunningham et al. 2003), but it is in line with the simulated estimates (Hallberg and Gnanadesikan 2006; Kuhlbrodt et al. 2012). According to the criteria of England (1993), namely a variation in global mean bottom water properties of less than 0.01°C (100 yr)^{-1} for temperature and 0.001 psu (100 yr)^{-1} for salinity, the bottom water mass has not completely reached an equilibrium state during the 600-yr spinup period; however, most of the main features of the global thermohaline circulation are within reasonable ranges. The SO sea ice also reached an equilibrium state and was within a reasonable range as described in C18. Therefore, the model used in this study is appropriate for this type of sensitivity experiment.

3. Results
a. BDT mode

The intensification of SH westerly winds plays an important role in generating open-ocean polynyas and deep convection in the SO (Cheon et al. 2014, 2015, 2018; Cheon and Gordon 2019). Figure 4 illustrates the seesaw balance between the Antarctic MOC and upwelling of the CDW, as well as the relationship between the occurrence of open-ocean polynyas in the Weddell and Ross Seas and the strength of the Antarctic MOC. The clockwise MOC of the SO is often described as a
Deacon cell (hereafter denoted as “Deacon MOC” in this paper) and consists of upwelling of the CDW (Figs. 3a,c) controlled by the SH westerly winds and subduction of the Antarctic Intermediate Water driven by the buoyancy loss (Speer et al. 2000). As the SH westerly winds intensify (weaken), the underlying northward Ekman transport increases (decreases), which intensifies (weakens) upwelling of the CDW in order to balance the Ekman transport. Since the northern limb of the Antarctic MOC and the southern limb of the Deacon MOC are in contact and rotate in opposite directions, there may exist a seesaw balance between these MOCs. Figure 4a shows the internal variations of 1) the Antarctic MOC calculated using the minimum value of the counterclockwise SO overturning cell and 2) the Deacon MOC calculated using the maximum value of clockwise SO overturning in CTRL. The Deacon MOC precedes the Antarctic MOC by about 2–3 years, and the two time series are positively correlated. However, since strengthening of the Deacon (Antarctic) MOC corresponds to a positive (negative) anomaly, such positive correlation supports the seesaw balance (i.e., anticorrelation) between the Antarctic MOC and upwelling of the CDW. The occurrence of open-ocean polynyas is estimated by the criterion that the sea ice concentration in the immediate area where open-ocean polynyas occur decreases below 40%. In SW050P02, the Weddell Sea open-ocean polynya (WSOP) occurs regularly around the peak of each cycle, whereas the Ross Sea open-ocean polynya (RSOP) occurs only once around the first peak. In SW050P04, the period is half that of SW050P02, and the WSOP occurs in every cycle except the second cycle; whereas the RSOP, as it does in SW050P02, occurs just once, in the first cycle. As discussed in C18, the unfavorable conditions for the RSOP are due to the relatively strong stratification of the Ross Sea and the Ross gyre being compressed by the southward shift of ACC fronts. Moreover, owing to intensified negative wind stress curl, the spinup of the Weddell Gyre activates the WDW eddies in the southwestern flank of the Maud Rise of the seamount rising up to 2000-m depth (Cheon and Gordon 2019), which subsequently weakens the pycnocline and provides the favorable conditions for the WSOP.

The response of the Antarctic MOC to the oscillating wind forcing consists of two impacts: 1) oceanic convection associated with coastal and open-ocean polynyas and 2) the seesaw balance driven by upwelling of the CDW. The coastal polynyas in the Weddell and Ross Seas are, as discussed in C18, in a linear correlation with
the oscillating wind forcing. Since their impact on the Antarctic MOC is smaller than the impact of open-ocean polynyas, an assessment of the coastal polynyas is not presented here. In the first cycle, as the SH westerly winds intensify, the Antarctic MOC weakens in both SW050P02 and SW050P04 in association with the seesaw balance driven by the intensifying upwelling. However, its weakening is insignificant because the oscillating wind forcing being in a linear correlation with the formation of coastal polynyas simultaneously intensifies near-boundary convection and thus the Antarctic MOC. The Antarctic MOC suddenly begins intensifying drastically in year 9 of SW050P02 and in year 7 of SW050P04 as a result of the occurrence of open-ocean polynyas and deep convection. It then again weakens with the intensifying winds until the WSOP reoccurs in the second cycle. In summary, when the SH westerly winds intensify, the Antarctic MOC is generally weakened by the seesaw balance driven by the intensifying upwelling; however, it is significantly intensified not only by the occurrence of open-ocean polynyas but also by the increased formation of coastal polynyas. When the SH westerly winds weaken, the Antarctic MOC is intensified by the combined effect of the seesaw balance driven by weakened upwelling with the open-ocean polynyas and the deep convection remaining. This overpowers the near-boundary convection, which is weakened by the decreased formation of coastal polynyas and acts in the opposite direction. This process is repeated throughout the entire period and provides the signal to be propagated northward through the abyssal ocean, which initiates the BDT mode.

Figure 5a shows the current and age of water (AOW) in CTRL at depths between 4400 and 5500 m where the AABW masses are present. The AOW indicates how old the water mass is after sinking from the ocean surface. Since the Pacific and Indian Oceans do not have...
FIG. 5. (a) Horizontal distributions of the annual-mean abyssal current and AOW of CTRL averaged over 4400–5500-m depths over 200 years, and the differences between CTRL and SW050P02 at years (b) 12, (c) 35, (d) 50, (e) 62, (f) 85, and (g) 100.
any predominant locations for deep water formation, very old bottom water masses occupy the North Pacific and the eastern Indian Oceans. Relatively young bottom water masses indicative of the AABW are observed in the southwestern Indian and Pacific Oceans as well as in the South Atlantic Ocean. The Mid-Atlantic Ridge plays an important role in determining the pathways of the AABW in the Atlantic (i.e., one in the western region of the ocean divided by the ridge and the other in the eastern region); this is in broad agreement with previous studies (Mantyla and Reid 1983; Stephens and Marshall 2000; Morozov et al. 2010). Orographic features also have a determining impact on the propagation of abyssal waters of the Pacific. The AABW flowing northward along the Kermadec-Tonga and Louisvile Ridges systems bifurcates at the southern tip of the Manihiki Plateau, with one portion passing through the Samoan Passage and the other passing anticlockwise around the eastern flank of the Manihiki Plateau. They join at the southern tip of the Gilbert Ridge and again bifurcate. The Pacific bottom current fields simulated in CTRL are in broad agreement with observations (Mantyla and Reid 1983; Roemmich et al. 1996; Kawabe et al. 2003).

The variations in AOW and current in SW050P02 (Figs. 5b–g) illustrate representative responses of the abyssal ocean to the BDT mode during the first 100 years. In year 12 (Fig. 5b), when the wind forcing reaches its first positive peak, although the Antarctic MOC is intensifying with the occurrence of WSOP and RSOP, the signal activated by the BDT mode does not reach the abyssal ocean north of 40°S. In year 35 (Fig. 5c), when the wind forcing approaches its first negative peak, most of the AABW masses that begin newly forming from year 10 enter the Pacific and flow along the bottom western boundary, affecting the abyssal water south of 20°N via advection. Although not shown here, the newly formed AABW with intensifying SH westerly winds in the Weddell and Ross Seas occupies most of the SO bottom south of the ACC fronts. Moreover, the relatively young (about 80–250 years old) water masses occupying the SO bottom before the application of the oscillating wind forcing come out through the ACC fronts, reaching the South Equatorial Pacific at about 10°S 25 years after activation of the BDT mode. The abyssal circulation in the Pacific is drastically intensified, e.g., the northward velocity of the abyssal currents averaged over the eastern flank of Manihiki Plateau (10°–5°S, 160°–155°W) quadruples from about 0.006 to 0.024 m s⁻¹. In contrast to the Pacific, the signals in the Atlantic and Indian Oceans are propagated only to 30°S. The northward AABW flows prevailing over the Pacific are consistent with those of previous modeling studies (van Sebille et al. 2013; Zanowski and Hallberg 2017). In year 50 (Fig. 5d), when the first cycle finishes, the enhanced bottom circulation substantially diminishes from the south, whereas the signal reaches farther north to 30°N in the Pacific. In the Atlantic and Indian Oceans the signals are also propagated farther north. In year 62 the oscillating wind forcing reaches the peak of its second cycle, and the Antarctic MOC begins intensifying with the reoccurrence of WSOP (Fig. 4b). Since the signal is just recently activated and thus remains south of the ACC fronts, the enhanced bottom circulation keeps diminishing (Fig. 5e). As shown in Figs. 5f and 5g, the bottom circulation is again intensified as the newly and substantially forming AABW around Antarctica sinks to the bottom and pushes the preexisting AABW northward. The change in bottom circulation is not as large as that in the first cycle because only the WSOP occurs and its scale is relatively small (Fig. 4b).

The abyssal WBCs passing through the equator show the BDT mode-induced oscillation with lags of about 30 years in the Pacific and of about 40 years in the Atlantic (Figs. 6a,c). Owing to the NADW occupying the deep Atlantic, the response of the AABW passing through the equator to the BDT mode is slower and smaller in the Atlantic than it is in the Pacific. The variations in MOC estimated at the equatorial Atlantic clearly illustrate the BDT mode-induced bipolar seesaw balance (Figs. 6b,d), which is in phase with the variation of the abyssal WBC. Therefore, under the oscillating SH westerly winds, the bipolar seesaw balance between deep and bottom waters at the equatorial Atlantic is controlled by the BDT mode-induced AABW formation.

Figure 7a shows the marked southward deep WBCs at 2000-m depth in the equatorial Atlantic. To show their representative responses to the oscillating SH westerly winds, the 16th and the 40th years in SW050P02 were selected for the positive phase and negative phase, respectively. As shown in Figs. 7b and 7c, the intensifying (weakening) SH westerly winds modulate the Ekman divergence-driven upwelling that increases (decreases) the NADW outflow. The temporal variations in the Atlantic deep WBC and the NADW outflow are presented in Figs. 7d and 7e, with the meridional velocity estimated at 2000-m depth near the northeastern coast of South America (5°N, 45°W) and the maximum Atlantic MOC estimated at 30°S. The positive (negative) anomaly of the meridional velocity indicates a weakening (intensifying) southward deep WBCs in the equatorial Atlantic, whereas the positive (negative) anomaly of the NADW outflow indicates an increase (decrease) in the NADW passing through 30°S. When the experimental setup employed in this study was designed, the BDT mode-induced change in the AABW formation was expected.
to affect not only the NADW outflow but also the Atlantic MOC. However, the NADW outflow is modulated by the oscillating wind forcing with the lag of about 2–3 years, rather than by the seesaw balance driven by the northward AABW flow varying with the BDT mode (cf. Figs. 7d and 7e with Figs. 6a and 6c). As will be discussed in the following subsection, the Atlantic MOC varies independently of not only the AABW-driven bipolar seesaw balance but also the NADW outflow.

b. EDT mode

In the North Atlantic, the South Equatorial Current (SEC) reaches the eastern coast of South America with the majority of its volume transport. It then changes direction, moving northward as the Guiana Current, most of which moves to the Gulf of Mexico and a part of which bifurcates to become the Gulf Stream together with the North Equatorial Current (NEC) (Fig. 8a). In the North Pacific, most of the SEC does not reach the eastern coast of Indonesia, instead turning eastward and joining the North Equatorial Countercurrent (NECC). The Pacific NEC, which is much stronger than the Atlantic NEC, reaches the eastern coast of Philippines and bifurcates, with one portion turning counterclockwise and joining the NECC, and the other portion moving northward as the Kuroshio (Fig. 8d).

Fig. 6. Variation in the meridional velocities averaged over 4400–5500-m depths in the Atlantic (5°S–5°N, 50°–30°W; red solid) and in the Pacific (5°S–5°N, 170°–180°E; blue solid), and time–depth cross sections of meridional overturning streamfunctions estimated at 30°S for (a),(b) SW050P02 and (c),(d) SW050P04. Time series of the maximum zonal-mean wind stress between 80° and 30°S (thin dashed black lines) are presented for reference.
when the oscillating wind forcing in SW050P02 passes through its first positive peak, the northward flowing surface WBCs such as the Guiana Current of the Atlantic and the Kuroshio of the Pacific intensify (Figs. 8b,e). However, in year 40, when the wind forcing is around its first negative peak, these WBCs weaken (Figs. 8c,f). As noted in Figs. 8a–f, the scale of variation is much smaller than that of the basic current field in CTRL. Moreover, the variation in WBC is smaller in the North Pacific than it is in the North Atlantic. A comparison of the deep (Figs. 7a–c) and bottom (Fig. 5) current variations indicates that the EDT mode-induced variation in the surface WBCs is much smaller than the BDT mode-induced variation in the deep and bottom WBCs. The Gulf Stream does not respond linearly to the oscillating SH westerly winds (in contrast to the Guiana Current, which shows a clear correlation near the northeastern coast of South America) because the Gulf Stream is significantly affected by the gyres that vary with oceanic deep convection and are associated with NADW formation.

Figures 8g and 8h illustrate the variations in surface WBCs: one, indicative of the Guiana Current in the Atlantic, is estimated by the meridional velocity near the northeastern coast of South America (8°–10°N, 60°–55°W); the other, indicative of the Kuroshio in the North Pacific, is estimated by the meridional velocity near the east coast of Taiwan (23°–25°N, 122°–130°E). Since both currents flow northward, the positive (negative) anomalies are indicative of strengthening (weakening). The North Atlantic surface WBC anomaly is highly correlated with the oscillating SH westerly winds, and their time lag is dependent on the oscillation period of each wind forcing. In SW050P02, the two factors are in good correlation without any time lag, whereas in SW050P04...
the response of the Guiana Current to the oscillating wind forcing is delayed by about 5–7 years. The Kuroshio anomaly is also highly correlated with the SH westerly winds, with a time lag of about 6–8 years in SW050P02 and of about 10–12 years in SW050P04. Without any atmospheric teleconnection, this EDT mode-generated signal is very quickly propagated to the North Atlantic and North Pacific. Using a coarse-resolution ocean-only GCM, Klinger and Cruz (2009) also showed that the signal generated by the 50%-intensified SH westerly winds was quickly propagated to the Northern Hemisphere in 3–10 years, which is in good agreement with our results. As such, it is important to elucidate how the change in the SH westerly winds remotely affects the WBCs in the Northern Hemisphere. According to the study of McDermott (1996), who used the single- and two-basin simple ocean...
models with a re-entrant channel indicative of the Drake Passage, the 50%-intensified SH westerly winds enhance the anticyclonic subtropical gyre, which can play a role in exciting baroclinic Kelvin-like signals that propagate northward to the equator along the western boundary. At the equator, these signals turn east and propagate to the eastern boundary like equatorial Kelvin modes. While some of the energy splits into two hemispheres at the eastern boundary, some is reflected to the west and propagates as Rossby modes. At the western boundary, the reflected Rossby modes act to enhance the northward WBC. This process, which is termed the “equatorial buffer” by Johnson and Marshall (2002), plays a role in restricting abrupt changes in the thermohaline overturning in the North Atlantic.

It should be noted that the WBC variations are generally larger in the Atlantic than in the Pacific, which is an important issue requiring our attention. The North Atlantic has major locations where deep water masses form and thus plays a crucial role as a “main engine” in driving the so-called Atlantic Ocean conveyor belt; this contrasts with the North Pacific, which exhibits no deep water formation. Since the Atlantic Ocean conveyor belt consists of 1) the northward Ekman transport associated with the EDT mode, 2) the NADW formation in the Labrador and GIN Seas, 3) its outflow toward the SO, and 4) upwelling of the CDW, the EDT mode-generated signal may be propagated on this conveyor belt to the Northern Hemisphere faster in the Atlantic than it is in the Pacific. However, this study considers another cause for the signal moving more rapidly in the Atlantic. The SH westerly winds generally give rise to northward Ekman transport uniformly over the underlying zonal band; therefore, the signal is propagated northward uniformly in all longitudes. Figure 9a illustrates the variations of the equatorial sea surface height (SSH) near the respective western and eastern boundaries of the North Atlantic and Pacific. The equatorial SSHs oscillate in the four different longitudes with nearly the same amplitude and phase, which implies that the EDT mode-generated signal passes through the equator uniformly in all longitudes.

As such, it is important to consider why the surface WBCs in the North Atlantic and Pacific respond differently to the EDT mode even under this zonally uniform propagation of the signal. Figure 9b shows the horizontal barotropic streamfunctions calculated in the North Atlantic and Pacific. Anticyclonic subtropical gyres controlled by the overlying wind stress curl dominate both oceans north of about 20°N, whereas the cyclonic gyre south of about 20°N is much stronger in the Pacific than it is in the Atlantic. The tropical cyclonic gyres in both oceans consist of the westward NEC in their northern limb and the eastward NECC in their southern limb. In the Pacific, the NEC reaching the western boundary along the Philippine coast near 125°E bifurcates with one portion turning to the north to become the Kuroshio and the other veering south to become the Mindanao Current (Qiu et al. 2015). The southward flowing Mindanao Current constitutes the western limb of the Pacific tropical gyre and separates into the Indonesian Throughflow and the eastward flowing NECC between 6° and 7°N (Lukas et al. 1991). On average, the NECC transports the warm pool water of more than 30 Sv to the eastern Pacific basin (Zhao et al. 2013), and the mean transport of the Mindanao Current measured over the thermocline is about 19 Sv. In the Atlantic, although there are negative barotropic streamfunctions (implying cyclonic circulation) between 0° and 20°N (Fig. 9c), the Guiana Current actually splits from the North Brazil Current and continues to flow northward along the western boundary between the NEC and NECC; this contrasts with the southward flowing Mindanao Current in the tropical Pacific (Condie 1991). Moreover, the NECC is robust in the boreal summer and fall but is very weak or even absent in the winter and spring (Yang and Joyce 2006). This difference between tropical gyres located in the North Atlantic and Pacific may be a crucial cause of the EDT-generated signal having a larger remote impact on the North Atlantic than it does on the North Pacific. As illustrated by the minimum barotropic streamfunction anomalies calculated in the north equatorial Atlantic and Pacific (Fig. 9d), the tropical gyre in the Pacific responds inversely to the oscillating wind forcing, whereas the response in the tropical Atlantic is negligible. That is, the distinct, cyclonic gyre in the equatorial Pacific weakens (intensifies) with the increasing (decreasing) northward Ekman transport under the intensifying (weakening) SH westerly winds and thus plays a role as a barrier to the EDT mode-generated signal. Therefore, in the Atlantic, the Ekman transport can keep moving northward without being blocked by the tropical gyre, which is a crucial cause of the response of the surface WBC to oscillating SH westerly winds to be larger in the Atlantic than it is in the Pacific. On the other hand, the lack of topographic waveguide in the upper ocean of the Indo-Pacific is also a potential cause of the North Pacific WBC responding belatedly to the SH wind forcing.

As such, we should consider how the transported mode affects the Atlantic MOC associated with NADW formation. Figure 10 shows the annual mean, surface horizontal circulation, and mixed layer depth in the northern North Atlantic. The mixed layer depth is indicative of the strength of oceanic deep convection. The relatively very deep mixed layer of the Labrador Sea and GIN Sea is in good agreement with previous studies.
The Gulf Stream, the Norwegian Current, the East/West Greenland Currents, and the Labrador Current are well reproduced in CTRL, as is the southward eastern boundary current connected to the Canary Current. To identify the factors controlling oceanic stratification in the vicinity of the Labrador Sea, a major location for NADW formation with the GIN Sea, variations in the upper ocean (0–100 m depths) salt levels and potential density at the surface ($\sigma_z = 0$) were estimated over 60°–50°W/55°–60°N (Figs. 11a,c). The Guiana Current (the WBC in the North Atlantic) transports heat and salt from subtropics northward and fluctuates in phase with the oscillating SH westerly winds (Figs. 8g,h), which causes upper ocean salt and potential density anomalies near the Labrador Sea. These are also generally in phase with the oscillating wind forcing, except for during the first cycles of both experiments. Although the upper ocean temperature anomaly near the Labrador Sea is also in positive correlation with the oscillating wind forcing (not shown here), the potential density anomaly is determined by the salt anomaly. These anomalies show overall decreases for the first 80 years, before increasing. It seems to be associated with the inherent dynamic features of the Labrador and GIN Sea cyclonic gyre circulations, which require an adjustment period for the EDT mode. Although not as extensive as these anomalies, the NADW outflow (Figs. 7d,e) and the Guiana Current (Figs. 8g,h) also show some decreases in the early period. The Atlantic MOC
formation. The green dashed-line box indicates the area of NADW formation in the vicinity of the Labrador Sea (55°–60°N, 60°–50°W).

anomaly estimated by the maximum values of meridional streamfunction over 45°–60°N/500–1500 m shows a clear correlation with the upper ocean salt and potential density anomalies near the Labrador Sea except for during the first cycles of both experiments and the third cycle of SW050P04 (Figs. 11b,d). This indicates that the oscillating SH westerly winds remotely affect the Atlantic MOC associated with the NADW formation via the EDT mode-generated signal, thereby modulating the North Atlantic WBC, northward salt transport from the tropics and thus stratification of the Labrador and GIN Seas. Note that the Atlantic MOC varies independently of the abyssal WBCs passing through the equatorial Atlantic (Figs. 6a,c), which implies that the BDT mode-generated signal does not affect the NADW formation.

4. Summary and discussion

Using a sea ice–ocean GCM coupled to an energy balance atmospheric model, the diverse responses of global ocean circulation to sinusoidally oscillating SH westerly winds are investigated in the context of the BDT mode (starting from the SO polynya events) and the EDT mode (starting from the northward Ekman transport). The Antarctic MOC that is indicative of the formation rate of AABW varies not only linearly with oceanic convection associated with coastal and open-ocean polynyas but also in an inverse correlation with upwelling of the CDW. Owing to the NADW occupying the deep North Atlantic, the AABW passing through the equator and flowing northward is highly predominant in the Pacific. In the Atlantic, the seesaw balance between deep and bottom waters at the equator appears to be controlled by BDT mode-induced AABW formation. The SH westerly winds play a further role in remotely affecting the North Atlantic and Pacific WBCs via the EDT mode-generated signal. Although the signal propagated northward passes through the equator uniformly at all longitudes, perturbation on the WBC is much stronger in the Atlantic than it is in the Pacific, which is due to the blocking effect of the relatively strong and persistent Mindanao Current, the southward flowing WBC of the North Pacific tropical gyre. Northward salt advection by the North Atlantic WBC appears to control the Atlantic MOC associated with the formation rate of NADW, which is independent of the BDT mode-generated signal.

Using the 640-day observation data estimated by a moored array at the equator in the western Atlantic, Hall et al. (1997) first showed that the rise and fall of the transition layer between the AABW and the lower NADW is associated with the bipolar seesaw balance between opposing changes in the AABW and NADW. The mechanism governing this seesaw balance was explained in relation to the conservation of potential vorticity by Cheon and Stössel (2009). For the climate community monitoring the Atlantic MOC, the variation in Arctic sea ice is an important factor because Arctic freshwater export directly determines the surface buoyancy and thus ocean stratification in the main NADW formation region in the Labrador and GIN Seas (Jungclaus et al. 2005; Jahn and Holland 2013; Sévellec et al. 2017). In this paper, we consider the effects of SAM on the Atlantic MOC. Since the SAM index has become more positive since the 1950s (Hartmann et al. 2013; Cheon and Gordon 2019) and showed a significant positive trend in the representative concentration pathway scenario 8.5 applied to fully coupled climate models by Zheng et al. (2013), the negative wind stress curl over the Weddell and Ross Seas may intensify in future climates, which implies an increase in the potential for activating the BDT mode. Therefore, the BDT mode-generated bipolar seesaw balance was expected to affect not only the NADW outflow but also the Atlantic MOC when the experimental setup was first designed for this study. However, the NADW outflow appeared to be directly controlled by the oscillating SH westerly winds and the ensuing Ekman divergence-driven upwelling of the CDW. Moreover, the Atlantic MOC appeared to be affected by the remote effect of the EDT mode-generated signal. That is, a positive trend in the SAM, which is expected in a warming climate, is not able to weaken the Atlantic MOC by the BDT mode.

The formation of the NADW is hypothesized to play a key role in driving the Atlantic Ocean conveyor belt. In this study, the North Atlantic surface WBC and the NADW outflow corresponding to the upper and lower parts of the Atlantic Ocean conveyor belt appear to be highly correlated with the oscillating SH westerly winds, which directly control the Ekman divergence-driven
upwelling of the CDW (i.e., the southern portion of the conveyor belt). With lags of about a decade the Atlantic MOC variation is correlated with the oscillating SH westerly winds, which modulate the transport of salinity from the subtropics to the northern North Atlantic. In this regard, the SH westerly wind anomalies associated with the SAM can drive the Atlantic Ocean conveyor belt variations.

Using an ocean-only GCM, Oke and England (2004) showed the impact of poleward shifting SH westerly winds on the ocean circulation in association with the positive trend in SAM (i.e., without sea ice–ocean interaction and AABW variation). Since the north–south migration of SH westerly winds is another important characteristic of the SAM that is not considered in this study, the oscillating position and its combination with oscillating strength are worthy of investigation in terms of the responses of sea ice/ocean interactions such as polynyas, oceanic deep convection, and global ocean circulation. In comparison with a coarse-resolution ocean GCM, a high-resolution ocean GCM that can explicitly resolve mesoscale eddies does not show large ACC responses to the intensifying SH westerly winds because of the role that eddies play in diffusing heat and salt meridionally (Hallberg and...
REFERENCES


