The Role of Oscillating Southern Hemisphere Westerly Winds: Southern Ocean Coastal and Open-Ocean Polynyas

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ABSTRACT

An oscillation in intensity of the Southern Hemisphere westerly winds is a major characteristic of the southern annular mode. Its impact upon the sea ice–ocean interactions in the Weddell and Ross Seas is investigated by a sea ice–ocean general circulation model coupled to an energy balance model for three temporal scales and two amplitudes of intensity. It is found that the oscillating wind forcing over the Southern Ocean plays a significant role both in regulating coastal polynyas along the Antarctic margins and in triggering open-ocean polynyas. The formation of coastal polynya in the western Weddell and Ross Seas is enhanced with the intensifying winds, resulting in an increase in the salt flux into the ocean via sea ice formation. Under intensifying winds, an instantaneous spinup within the Weddell and Ross Sea cyclonic gyres causes the warm deep water to upwell, triggering open-ocean polynyas with accompanying deep ocean convection. In contrast to coastal polynyas, open-ocean polynyas in the Weddell and Ross Seas respond differently to the wind forcing and are dependent on its period. That is, the Weddell Sea open-ocean polynya occurs earlier and more frequently than the Ross Sea open-ocean polynya and, more importantly, does not occur when the period of oscillation is sufficiently short. The strong stratification of the Ross Sea and the contraction of the Ross gyre due to the southward shift of Antarctic Circumpolar Current fronts provide unfavorable conditions for the Ross Sea open-ocean polynya. The recovery time of deep ocean heat controls the occurrence frequency of the Weddell Sea open-ocean polynya.

1. Introduction

The Southern Ocean (SO) surface water masses are known to sink to the bottom of the sea in two ways: 1) near-boundary convection, also called “continental shelf slope convection,” and 2) open-ocean deep convection (Killworth 1983; Gordon 2014). In the Southern Hemisphere (SH), near-boundary convection is closely linked to the coastal polynya, also called “latent heat polynya” (Curry and Webster 1999; Wadhams 2000), while open-ocean deep convection is closely linked to the open-ocean polynya, also called “sensible heat polynya” (Gordon 1978; 1982). Coastal polynyas occur mainly in the Weddell Sea (Kottmeier and Engelbart 1992; Fahrbach et al. 1995), Ross Sea (Jacobs and Comiso 1989), Adélie Depression (Gordon and Tchernia 1972; Williams et al. 2008), and Cape Darnley (65°–69°E; Ohshima et al. 2013). The accompanying descending plumes of dense water over the continental slope are key drivers in the Antarctic Bottom Water (AABW) formation (Orsi et al. 1999, 2013). The accompanying descending plumes of dense water over the continental slope are key drivers in the Antarctic Bottom Water (AABW) formation (Orsi et al. 1999; Jacobs 2004). In contrast to coastal polynyas, large-scale open-ocean polynya was observed only once in the Weddell Sea by the Electrically Scanning Microwave Radiometer carried on the Nimbus-5 satellite launched at December 1972. This so-called Weddell Polynya persisted for three

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consecutive winters, 1974–76, with a large average area of about $250 \times 10^3$ km$^2$ and was associated with deep reaching convection (Gordon 1978, 1982). Short-term, small-scale, open-ocean polynyas have been intermittently observed not only in the Weddell Sea but also in the Cosmonaut Sea (Comiso and Gordon 1987).

Coastal polynyas are generated and sustained by the wintertime offshore katabatic winds (Curry and Webster 1999; Wadhams 2000; Cappelletti et al. 2010; Rusciano et al. 2013). During the austral winter, the surface water (approximately $-1.8^\circ$C) around Antarctica is exposed to extremely cold air (approximately $-40^\circ$ to $-30^\circ$C), leading to the formation of sea ice. The local katabatic winds advect newly formed sea ice constantly away from coastline and thus give rise to coastal polynyas, allowing continuous new sea ice formation and associated vigorous brine rejection into the water column. This leads to formation of dense, high-salinity shelf water, which is a major source for the AABW (Wadhams 2000). Although polynyas occurring in the Adélie Land Coast are important sources for the AABW to the Australian–Antarctic basin (Williams et al. 2008), most bottom waters originate from the Weddell Sea due to coastal polynyas adjacent to the Ronne Ice Shelf, and from the Ross Sea adjacent to Ross Ice Shelf and Terra Nova Bay (Jacobs 2004; Drucker et al. 2011).

The process of open-ocean polynya formation shall be summarized in stages as follows: 1) preconditioning due to weak stratification, 2) destabilization of water column, 3) upwelling of the relatively warm deep water acting to prevent the overlying sea ice from forming or to melt it, 4) creation of an open-ocean polynya, 5) oceanic deep convection, and 6) sequential decay of the polynya and convection (Martinson et al. 1981; Gordon 1978, 1982; Killworth 1983). According to Gordon (1982), there was no evidence of open-ocean deep convection in 1973, which was one year before the Weddell Polynya occurred, and in 1977 the Weddell Deep Water (WDW) between 200- and 2700-m depths became colder and fresher than in 1973, which is a clear evidence of open-ocean deep convection. This confirms the order of the fourth and fifth stages, and the sixth stage is a natural process. However, the first, second, and third stages are still hypothetic reasoning.

In regard to the first and second stages, Gordon et al. (2007), statistically analyzing observational data and combining it with major climate modes, proposed that under the prolonged negative southern annular mode (SAM) a drier-than-normal air condition over the Weddell Sea could weaken the pycnocline and thus create preconditioning for open-ocean polynyas. The upper ocean can also be destabilized by the eddies spawned at the frontal boundary extending from Maud Rise to the northeast, where the relatively warm and salty WDW flowing along the eastern limb of the Weddell gyre encounters the already existing cold and fresh WDW (Gordon and Huber 1984, 1990, 1995). The WDW eddies have been studied and shown to originate from the interaction between the ocean currents temporally varying near the frontal boundary and the Maud Rise seamount via the Taylor column formation (Ou and Gordon 1986; Ou 1991; Alverson and Owens 1996; Holland 2001). The third stage, which triggers open-ocean polynyas, is ambiguous. The cyclonic eddies with horizontal scales of the Rossby radius (~10 km) might have thin surface water layer as the pycnocline domes up, resulting in small-scale convection mixing cold surface water with warm deep water within small eddy-scale features (Gordon 1978). Note that this small-scale convection is different from the fifth stage occurring with horizontal scales of a few hundreds of kilometers and reaching about 2700-m depth.

The SH westerly winds span the area between 70° and 30°S, while easterly winds dominate south of 70°S including the southern Weddell and Ross Seas. The corresponding negative wind stress curl over the Weddell and Ross Seas is a key driver for basin-scale cyclonic gyres, and its intensification plays an important role in spinup of gyres according to suggestion of Cheon et al. (2014). The SH westerly wind variation is closely linked to the SAM such that the westerly winds get stronger (weaker) and are shifted poleward (equatorward) in its positive (negative) mode (Gong and Wang 1999; Thompson and Wallace 2000). Analysis of the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis data, specifically the monthly-mean ERA-40 (1958–2001) and ERA-Interim (1979–2015) data, indicates that the SAM index is highly correlated with the zonal components of the wind speed at 10-m height and the wind stress, as illustrated in Fig. 1a.

Since the pioneering study of Toggweiler and Samuels (1995) proposing the link between the SH westerly winds and the global thermohaline circulation, numerous studies have been performed to investigate this link in depth with various viewpoints by using various ocean models such as an idealized single- or two-basin model (e.g., McDermott 1996; Tsujino and Suginohara 1999; Gnanadesikan and Hallberg 2000; Klinger et al. 2003), a coarse-resolution ocean-only model including realistic bottom topography (e.g., Toggweiler and Samuels 1995; Cai and Baines 1996; Rahmstorf and England 1997; Oke and England 2004; Klinger and Cruz 2009), a sea ice/ocean coupled model (Brix and Gerdies 2003), and an eddy-permitting isopycncal coordinate model (Hallberg and Gnanadesikan 2006). These studies have focused on the impact of the SH westerly winds upon the global
meridional overturning circulations (MOCs) and the Antarctic Circumpolar Current (ACC). Turning attention to the SO air–sea ice–ocean interaction, Cheon et al. (2014) first proposed a possibility for the intensified SH westerly winds to contribute to the formation of open-ocean polynya in the Weddell Sea by use of a coarse-resolution sea ice–ocean coupled general circulation model (GCM), which was the same model used in this study. Employing a more realistic wind forcing derived from the Coordinated Ocean-Ice Reference Experiments dataset version 2 (CORE2), Cheon et al. (2015) reproduced the 1970s’ Weddell Polynya and showed an importance of the deep ocean heat content as an essential precondition for open-ocean polynya formation. These two studies lead to four questions: 1) How does an open-ocean polynya form and dissipate under
the SH westerly winds oscillated by the SAM? 2) Is open-ocean polynya formation sensitive to the frequency of winds? 3) What is the unfavorable condition for open-ocean polynya formation in the Ross Sea? 4) Do synoptic-scale changes in the atmospheric circulation due to the oscillating winds affect coastal polynyas? These questions are addressed in this paper by use of the sinusoidally oscillating SH westerly winds with three temporal scales and two amplitudes of intensity, which are still artificial but closer to the main characteristics of SAM than the wind forcing used in Cheon et al. (2014). Although Brix and Gerdes (2003) simply showed the formation rate of AABW that was proportional to the strength of SH westerly winds, the change of AABW formation resulting from the wind-driven sea–ocean interactions and its impact upon the global ocean circulation have not been investigated yet. These will be addressed in a future publication.

In this study we employ the oscillation of only the wind strength in the main characteristics of SAM. This wind forcing is designed such that the intensity of SH westerly winds fluctuates within a range of ±25%–50% with one, two, and four periods per hundred years. Although precluding study of the mesoscale eddy effects, this coarse-resolution GCM enables us to simulate seven cases including a reference case over a global scale for 200 years. The model configuration, experimental setup, and model evaluation are described in detail in the following section. Sections 3 and 4 investigate responses of two types of polynya in the Weddell and Ross Seas, respectively, to the oscillating SH westerly winds. A summary and discussion follow in section 5.

2. Model

a. Model configuration

The model used in this study is the Modular Ocean Model version 4 (MOM4), a primitive equation ocean model (Griffies et al. 2004) with a dynamic and thermodynamic sea ice model (Winton 2000). A two-dimensional global atmosphere energy balance model (EBM; Gerdes et al. 2006) and a land model (Anderson et al. 2004) are also coupled to it via the Geophysical Fluid Dynamics Laboratory (GFDL) Flexible Modeling System (FMS). The ocean model spans from 80°S to 90°N with horizontal resolution of 2° × 1/3° to 1° in longitude and latitude, and reaches 5500-m depth with 50 semivariable layers (i.e., 22 upper layers with constant 10-m thickness and 28 lower layers with thickness gradually increasing to about 400 m). The bottom layer follows the actual topography data (Smith and Sandwell 1997), comprising of the National Oceanic and Atmospheric Administration (NOAA) 5-Minute Gridded Global Relief Data (ETOPO5), the International Bathymetric Chart of the Arctic Ocean (IBCAO), and satellite data spanning 72°S to 72°N. For simulation of the surface mixed layer, the K-profile parameterization (KPP) scheme (Large et al. 1994) is employed, and for the parameterization of mesoscale eddy transport on isopycnal surfaces the Gent–McWilliams (GM) scheme (Gent and McWilliams 1990) is employed. The GM diffusivity coefficient is fixed to be 600 m² s⁻¹, and so is the Redi diffusivity coefficient. The coefficients for vertical mixing change in the upper ocean from 10⁻² m² s⁻¹ in low latitudes to 3 × 10⁻² m² s⁻¹ in high latitudes and increase at depth to 1.2 × 10⁻⁴ m² s⁻¹ (Bryan and Lewis 1979). The convection scheme of Rahmstorf (1993) is also used for convective adjustment. The sea ice model consisting of a snow layer and two sea ice layers is run on the same grid as the ocean model. Its dynamic part includes the viscous-plastic constitutive law introduced by Hibler (1979), and its thermodynamic part follows the formulation of Winton (2000). The EBM extending globally has T42 horizontal resolution and provides thermodynamic forcing for the sea ice–ocean model. It should be noted that although the EBM is coupled to the sea ice–ocean GCM, the wind field and precipitation are directly derived from the reanalysis data (ERA-15) augmented by day-to-day variability of a selected year (1982). According to the SAM index [refer to Fig. 4 of Gordon et al. (2007)], 1982 is almost neutral. This allows the model to be adjusted to the state appropriate for the neutral SAM during spinup. These model configurations are same as those used in Cheon et al. (2014).

The model used in this study does not include ice sheet and shelf models and its horizontal resolution is too coarse to resolve coastal polynyas accurately. The wind field derived from the ERA-15 is also too coarse and is not accurate enough to resolve katabatic winds that are key drivers of coastal polynya. However, the offshore drift and formation of sea ice near the Antarctic coastline are closely linked to the prevailing winds, substantial parts of which are under an influence of the oscillating SH westerly winds. That is, since the western coasts of the Weddell and Ross Seas are under the influence of southwesterly winds, synoptic-scale changes in the atmospheric circulation due to the wind forcing should have a large influence on the nearshore sea ice field there. This is investigated in section 3.

b. Experimental setup

To determine the oscillation amplitude in the SH westerly winds, we calculate the monthly climatology of
zonal-mean zonal wind stress averaged over 65° and 40°S and its anomaly for the monthly-mean ERA-40 and ERA-Interim data. As shown in Fig. 1b, the maximum and minimum of the monthly climatology averaged between two sets of data are 0.0816 and 0.0305 N m⁻², respectively, and thus its range is 0.0511 N m⁻². It means that the zonal wind stress representative of the SH westerly winds between 1958 and 2015, on average, varies in this range. One standard deviation (σ) of the anomalous zonal wind stress is 0.0083 N m⁻² when calculated from the 12-month low-pass filtered data, and 3σ, explaining 99.7% of the anomalous zonal wind stress, is 0.0249 N m⁻², which corresponds to about 49% of the climatological range. Therefore, the wind forcing oscillates within the range of ±50%. The amplitude that varies within a range of ±25% is also employed for weakly oscillating wind forcing. Oscillations with these two amplitudes are added to the original zonal wind stress whose sign is only positive in the south of 20°S (i.e., the SH westerly winds), and their maximum and minimum distributions are presented with the reference wind stress in Fig. 2a.

The present SH sea ice–ocean system has experienced a prolonged positive SAM with a preceding negative SAM that began around 1965 (Gordon et al. 2007), an approximate oscillation period of about 50 years, whereas in the early twentieth century the period appears to be less than 30 years. Although the spectral analysis is in general a good method to identify the dominant period of variation accurately, because of the long period of SAM it cannot be applied to the relatively short ERA-40 (44 yr) and ERA-Interim (37 yr) data. Therefore, we selected the 50-yr period, the recent SAM period, and also selected half (25 yr) and double (100 yr) the 50-yr period in order to investigate the dependence of the sea ice–ocean interaction on the slowly to rapidly oscillating SH westerly winds.

Six sensitivity experiments carried out in this study are simply listed in Table 1. In SW025P01 the wind stress oscillating within the range of ±25% over 100 years (one cycle per century) is added to its original value; in SW025P02 the same amplitude of oscillation over 50 years (two cycles per century) is added, and in SW025P04 the same amplitude of oscillation over
25 years (four cycles per century) is added. In SW050P01 the wind stress oscillating within the range of ±50% over 100 years is added to its original value; in SW050P02 the same amplitude of oscillation over 50 years is added; and in SW050P04 the same amplitude of oscillation over 25 years is added. These oscillating wind stresses for the respective experiments are illustrated in Fig. 2b. The model was spun up from a “cold start” condition, which initialized the model by using climatological values of temperature and salinity derived from the World Ocean Atlas 2013 (WOA13) data and assuming the velocity field to be zero at the start. After 600 years of spinup, the model was then run with each wind forcing for another 200 years. For a control case, the model was run without any oscillation for 200 years (referred to as CTRL).

c. Model evaluation

Figure 3 shows meridional sections of potential temperature and salinity derived from CTRL and the WOA13 data for the South Atlantic and Pacific (i.e., the Weddell and Ross Sea sectors). Although differing in magnitude and pattern in comparison with the observed data, CTRL reproduces the low-salinity Antarctic Intermediate Water (AAIW; the water mass of purplish color originating at the surface between about 60° and 50°S in salinity distributions) spreading northward and the saline North Atlantic Deep Water (NADW; the water mass of light green occupying between about 60° and 3000-m depths in the Atlantic salinity distributions) spreading southward. However, south of 60°S, CTRL does not simulate the intrusion of warm deep water masses drawn from the Circumpolar Deep Water (CDW) between 200- and 2000-m depths in the Weddell Sea sector and between 150- and 1100-m depths in the Ross Sea sector. Although a slight evidence of oceanic convection is shown in the salinity distribution of the Weddell Sea sector, it should be noted that open-ocean deep convection of the scale enough to lead to open-ocean polynyas does not occur in CTRL (as described below with Fig. 5c).

To evaluate relative stratification within the open-ocean core of the Weddell and Ross Sea cyclonic gyres, vertical profiles of their potential temperature and density referenced to the surface and salinity are compared with the observed data in Fig. 4. Although there is some difference in potential temperature, the model mimics water properties of both seas well, in particular for the potential density and salinity. In comparison with the Weddell Sea, in the Ross Sea the deep water below about 150-m depth is simulated to be warmer at maximum by 1.26°C, and the surface water is simulated to be fresher by 0.26 psu. Although differences between two seas in the observation are smaller than in the simulation (warmer by 0.93°C and fresher by 0.16 psu), the characteristics of two datasets are consistent in that both reveal a stronger barrier to vertical mixing across the pycnocline of the Ross Sea gyre relative to the Weddell Sea gyre. In addition, from the WOCE Southern Ocean atlas (Orsi and Whitworth 2005) we find that the salinity difference between the surface layer at 50 m and the upper deep water at 200 m in the core of the Ross Sea gyre is approximately 0.5 psu, whereas that in the core of the Weddell Sea gyre is only 0.25 psu. The relatively low salinity of the Ross Sea surface layer is likely derived from the inflow from the Bellingshausen Sea and the Amundsen Sea, both of which have the lowest salinity surface waters of any coastal segment of Antarctica (Orsi and Whitworth 2005).

The modeled winter-mean (June–August) sea ice concentration is also compared with the observed one derived from the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) data (Figs. 5a,b) and appears to extend farther northward. Low sea ice concentrations are observed along the coastline over which the offshore katabatic winds are predominant, particularly in the southeastern Weddell Sea (30°–15°W), the southern Ross Sea (170°–160°W), and near the Mac. Robertson Land (62°–75°E). These correspond to coastal polynyas. Although most of the sea ice cover consists of highly concentrated sea ice (>92%), there is small area of the central Weddell Sea whose sea ice concentration drops to 82%. It is not associated with open-ocean polynyas. As shown in Fig. 5c, in both the Weddell and Ross Seas the very young age of water (AOW) at the second layer remains unchanged during 200 years of CTRL. Since the AOW indicates how old the water mass is after sinking from the ocean surface, it indicates that the relatively warm and salty deep water does not upwell. Therefore, the sea surface there keeps it at near-freezing temperatures during winter, and the overlying sea ice has a high concentration, which implies that open-ocean polynyas and deep convection never occur in CTRL. However, in

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**Table 1. List of experiments.**

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Range of fluctuation</th>
<th>Number of cycle per 100 years</th>
<th>Simulation time (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>—</td>
<td>—</td>
<td>200</td>
</tr>
<tr>
<td>SW025P01</td>
<td>±25%</td>
<td>1</td>
<td>200</td>
</tr>
<tr>
<td>SW025P02</td>
<td>±25%</td>
<td>2</td>
<td>200</td>
</tr>
<tr>
<td>SW025P04</td>
<td>±25%</td>
<td>4</td>
<td>200</td>
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<td>SW050P02</td>
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<tr>
<td>SW050P04</td>
<td>±50%</td>
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</table>
comparison with the Ross Sea, the Weddell Sea shows very small fluctuations in the AOW, sea surface temperature (SST), and sea ice concentration, which indicate some small episodes of convection that do not lead to open-ocean polynyas and deep-reaching convection and are associated with the difference between the model and the observation for the potential temperature in the Weddell Sea (Fig. 4). Although there is no evidence of oceanic convection, the fact that the simulated Ross Sea is less stratified than the observation may be reason why open-ocean polynyas that have never been observed in the real world occur under the oscillating wind forcing.

3. Coastal polynyas and near-boundary convection

The wind stress pattern of CTRL in Fig. 5b illustrates that the western and southern coasts of the Weddell and Ross Seas are under the influence of southwesterly and southeasterly winds, respectively. The responses of the western coasts of the Weddell and Ross Seas to the intensifying westerly winds at year 6 are investigated for
SW025P04 and SW050P04 in Fig. 6. Although there are small areas where sea ice concentration increases, the sea ice concentrations along most western coastlines of the Weddell and Ross Seas decrease as the offshore sea ice drifts intensify, with the corresponding magnitude larger in SW050P04 than in SW025P04, suggesting an increase in coastal polynya formation. As the coastal polynyas expand, more sea ice forms, as indicated by increases in energy flux of frazil formation, resulting in a larger amount of ice-to-ocean salt flux than in CTRL. These features are observed in other experiments as well. The aforementioned small areas, whose sea ice concentration, energy flux of frazil formation, and ice-to-ocean salt flux all increase, seem to be associated with the winds that are not powerful enough to blow newly formed sea ice offshore. This could be partly due to katabatic winds not being resolved in the forcing.

The process described above is investigated in further detail with variations in sea ice concentration and ice-to-ocean salt flux near the western coasts of Weddell and Ross Seas for the full 200-yr model run. As shown in Fig. 7a, the sea ice concentration within both areas remains unchanged in CTRL, and so does the salt flux. Although the sea ice concentrations are higher near the western Ross Sea coast than near the western Weddell Sea coast, implying smaller-scale coastal polynyas, the ensuing salt fluxes are, on the contrary, larger. This is due to the offshore winds of the western Ross Sea coast that are relatively strong and are located farther south (Fig. 5b). That is, the strong offshore winds tend to be colder and cause a large amount of new sea ice to form within relatively small-scale coastal polynyas, leading to large ice-to-ocean salt flux (Ou 1988).

As shown in Fig. 7b, in SW025P01 the sea ice concentration along the western Weddell Sea coast begins decreasing simultaneously as the westerly winds intensify, indicating an increase in the formation of coastal polynya. A general expectation is a simple negative correlation between the oscillating wind forcing and anomalous formation of coastal polynyas. However, the sea ice concentration increases drastically between years 14 and 20, maintaining its peak, and suddenly decreases between years 29 and 32. Shortly afterward, it is merged into the general negative correlation. This event occurs.

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**Fig. 4.** Vertical profiles of potential temperature (red lines) and density (black lines) referenced to the surface, and salinity (blue) averaged over the (a) Weddell Sea and (b) Ross Sea, from model results (solid lines) and WOA13 (http://www.nodc.noaa.gov/OC5/WOA13) observations (dashed lines). Since the stratification of the upper ocean ($\leq 500$-m depth) is important for preconditioning the formation of open-ocean polynya, potential density referenced at the surface is used. The potential temperature, density, and salinity in the Weddell Sea are averaged over $80^\circ-60^\circ S$, $60^\circ W-20^\circ E$, and those in the Ross Sea over $80^\circ-60^\circ S$, $160^\circ W-140^\circ W$. The model results shown here are averages of CTRL over the whole period (200 yr).
again in the second period and is associated with the occurrence of an open-ocean polynya that is investigated further in the following section. When open-ocean polynyas and accompanying oceanic deep convection occur, the Weddell gyre is drastically intensified (see Fig. 11a), causing a large amount of sea ice to accumulate along the western coast via the westward southern limb of the gyre. The piling up of sea ice is larger than the decrement due to the formation of coastal polynya and thus dominates the western margin sea ice cover. The anomalous ice-to-ocean salt flux is generally in positive correlation with the oscillating winds. This process is also observed in the western Ross Sea coast. It is noted that the range of fluctuation in the sea ice concentration is much smaller in the western Ross coast than in the western Weddell coast, while that in the salt flux is even slightly larger. As stated above, since the offshore winds in the western Ross coast are relatively strong, the variation is large under the same wind forcing. Therefore, large (small) variations in the offshore winds along the western Ross (Weddell) Sea coast lead to large (small) variations in the ice-to-salt flux, which is confirmed in Fig. 6. In the SW050P01, with twice the amplitude of the SW025P01, the variation in coastal polynya formation becomes larger, and so does that in the ice-to-salt flux (Fig. 7c). Even in the experiments whose amplitude is identical but period is shorter, the aforementioned phenomena occur every period, and the ranges of fluctuation in these are similar.

4. Open-ocean polynyas and deep convection

a. Creation and destruction of open-ocean polynyas

Figure 8 illustrates the creation and destruction of open-ocean polynyas in the Weddell and Ross Seas as represented by the variations in the winter-mean sea ice concentration and AOW at 4000 m depth. As shown in Fig. 8a, in SW025P01 an open-ocean polynya occurs first in the Weddell Sea, while the wind forcing is intensifying, and occurs five years later in the Ross Sea.
Areal-mean sea ice concentration appears to drop to about 40%, but in the immediate area, where the open-ocean polynya occurs, the sea ice concentration drops below 20% (see Fig. 12). Simultaneously with this, the AOW at 4000-m depth sharply drops to 1–2 years old, which means that it only takes 1–2 years for the surface water mass to spread along this depth and is indicative of spatially focused open-ocean deep convection. The first occurring open-ocean polynyas are followed by small-scale ones with a resting period of 2–3 years in both seas that disappear once the wind forcing weakens. Oceanic deep convection also disappears, and the underlying

**Fig. 6.** Differences of (left) the winter-mean sea ice concentration/drift, (middle) the energy flux of frazil formation, and (right) ice-to-ocean salt flux between SW025P04/SW050P04 and CTRL at year 6 near the western (top) Weddell and (bottom) Ross Sea coasts. The arrows correspond to the sea ice drift, and the colors to the sea ice concentration, energy flux of frazil formation, and ice-to-ocean salt flux.
deep ocean is then replaced with the “older” water drawn from the CDW. Just before the wind forcing reaches its second peak, the Weddell Sea open-ocean polynya (WSOP) reappears and is followed by oceanic deep convection. However, the Ross Sea open-ocean polynya (RSOP) does not occur during the second cycle, although there are slight reductions in the sea ice concentration and the 4000-m depth AOW.

As shown in Fig. 8b, in SW050P01, whose amplitude in the oscillating winds is twice the SW025P01, the first WSOP and RSOP occur three years earlier than those in SW025P01. When the wind forcing reaches the second
peak, the WSOP occurs and persists longer by two years than the first peak, and the RSOP, which does not occur in the second cycle of SW025P01, occurs, although its scale is relatively small. In SW025P02, whose period is half the SW025P01, the WSOP occurs regularly around a peak of every cycle, whereas the RSOP occurs only once around the first peak and does not occur again (Fig. 8a). Although larger in amplitude than SW025P02, the responses of the Weddell and Ross Seas in SW050P02 are very similar to those in SW025P02.

In low- and midfrequency experiments (i.e., SW025P01, SW050P01, SW025P02, and SW050P02), sea ice concentrations in the immediate area, where open-ocean polynyas occur, consistently drop below 20% (see Fig. 12), which means almost ice-free ocean. In high-frequency experiments (i.e., SW025P04 and SW050P04), decreases of sea ice concentrations in the area are smaller than those in the former experiments but are still judged to be open-ocean polynyas. Therefore, the criterion for occurrence of open-ocean polynyas is slightly loosened and is increased up to 40%. According to this criterion, in SW025P04 the WSOP occurs in the first, third, fifth, and eighth cycles, and the RSOP never occurs (Fig. 12a). In the SW050P04 the WSOP occurs in every cycle but the
second cycle, and the RSOP occurs just once in the first cycle and never occurs again after then (Fig. 12b).

Every open-ocean polynya appeared to occur during the positive phase of the oscillating wind forcing. In designing these sensitivity experiments, open-ocean polynyas were expected to occur when the wind forcing began to regain its strength immediately after reaching the lowest limit of oscillation, because the observed 1970s’ Weddell Polynya occurred right after the prolonged negative SAM reached its peak. However, our model never reproduces open-ocean polynyas in that moment, which implies an important point. The effect of SH westerly winds alone is not enough to trigger the formation of open-ocean polynya and must be combined with preconditioning factors such as drier-than-normal and/or colder-than-normal air conditions (Gordon et al. 2007), and warm and salty WDW eddies, the so-called Maud Rise effect (Ou and Gordon 1986; Ou 1991; Alverson and Owens 1996), which are not resolved in the model used in this study. Therefore, if those preconditioning factors are individually strong and are synchronously combined, an instantaneous and drastic intensification of SH westerly winds may be able to trigger an open-ocean polynya, which seems to be what happened in the observed 1970s’ Weddell Polynya.

Although the WSO and RSOP respond differently to the respective oscillation periods and amplitudes of the wind forcing, the occurrence mechanism that starts from the spinup of gyre and leads to upwelling of warm deep water is applied identically to both seas. The SW050P01 is selected to show the moment when open-ocean polynyas occur at the respective seas (Fig. 9). To investigate the nature of the year-to-year variation, the spinup of gyre, changes of barotropic streamfunction, and surface current for a given year are calculated by subtracting the previous year’s value from the given year’s value. Values of −1.1°C and −0.3°C are potential temperatures estimated at the respective thermocline depths of the Weddell and Ross Seas in CTRL (see Fig. 4) and represent a barrier between the cold, fresh surface water and the warm, salty deep water. The areas masked along the southwestern coasts of the Weddell and Ross Seas (Figs. 9b,d) mean that the whole water column in those areas is colder than the potential temperatures at the respective thermocline depths, which indicates that the surface water colder than the thermocline temperature sinks to the bottom of the sea via near-boundary convection due to the enhanced formation of coastal polynyas, as previously described.

As described in Figs. 9a and 9b, in the Weddell Sea a small-scale cyclonic circulation anomaly occurs in year 9 with doming of the 1.1°C isotherm and is slightly intensified in the following year, pulling the isotherm to the surface and consequently bringing the relatively warm deep water into direct contact with sea ice. Upwelled warm deep water acts to prevent sea ice from forming and to give rise to the open-ocean polynya. Once the open-ocean polynya occurs, a large amount of oceanic heat is released to the atmosphere within the ice-free area, and new sea ice forms, releasing salt [see Fig. 11 of Cheon et al. (2014)]. The cold surface water mixes with this ice-to-ocean released salt, thereby becoming denser than the underlying deep water and generating open-ocean deep convection. Although not shown here, the water column periodically restratifies as general convective ceases. Such an oscillatory mode is inherent in the formation process of open-ocean polynyas (Gordon 1991; Goosse and Fichefet 2001). This process is also observed in the Ross Sea, as shown in Figs. 9c and 9d.

Different responses between the WSO and RSOP to the oscillating wind forcing are summarized as follows. First, the WSO occurs earlier than the RSOP when both occur within one cycle. Second, the WSO occurs much more frequently than the RSOP under the same wind forcing, which implies that the Weddell Sea has stronger potential for the formation of open-ocean polynyas than the Ross Sea. Third, even the WSO does not occur in every oscillation cycle when the oscillation period is too short. Causes for these may be clues for why the RSOP has been never observed and why the long-lived, large-scale WSO has never observed since the mid-1970s. We investigate associated mechanisms below.

b. Unfavorable conditions of the Ross Sea for open-ocean polynya

For the first and second issues associated with the unfavorable condition of the Ross Sea for open-ocean polynya, the most basic factor is much stronger stratification in the Ross Sea than in the Weddell Sea, as previously stated with Fig. 4. However, since in this study the oscillating SH westerly winds are the only driver in the occurrence of open-ocean polynya, hereafter its associated factors are investigated. Figures 10a and 10c, illustrating barotropic streamfunctions of the South Pacific and Atlantic in CTRL, show that the gyre simulated in the Weddell Sea is basically stronger than that in the Ross Sea. This is supported by satellite altimetry data [Archiving, Validation and Interpretation of Satellite Oceanographic Data Project (AVISO); http://www.aviso.altimetry.fr/; figure not shown]. The curl of oscillating wind stress playing a crucial role in driving these gyres is also stronger in the Weddell Sea than in the Ross Sea (Fig. 10c). In keeping with the theory of Ekman pumping, the Weddell Sea has more potential for its relatively warm deep water to upwell and thus to
give rise to open-ocean polynyas, which is another factor for the Ross Sea to be unfavorable for the formation of open-ocean polynyas.

When the SH westerly winds intensify, not just the Weddell and Ross gyres but also the South Atlantic and Pacific subtropical gyres intensify in association with the enhanced positive wind stress curls over these two basins. According to comparison of the steric height observed from Argo program between 2004 and 2007 with that derived from the World Ocean Atlas 2001 (WOA01) including observation up to August 2001, the subtropical gyres in both basins intensify, and more importantly the

Fig. 9. (left) Annual variation of the winter-mean barotropic streamfunctions and surface currents, and (right) the winter-mean depths of the −1.1°C and −0.3°C isotherms in the (a),(b) Weddell and (c),(d) Ross Seas at the respective years for SW050P01. The depth of −1.1°C (−0.3°C) is the thermocline depth of the Weddell (Ross) Sea calculated from CTRL. To shed light on the year-to-year variation such as spinup of gyre, annual variation for a given year is used in this figure and is calculated by subtracting the previous year’s value from the given year’s value. The arrows correspond to the surface currents, and the colors to the barotropic streamfunctions and depths of the −1.1°C and −0.3°C isotherms.
intensification in the South Pacific subtropical gyre is larger than that in the South Atlantic subtropical gyre (Roemmich and Gilson 2009). Using the altimetric height from AVISO, Roemmich et al. (2007) also showed that the South Pacific subtropical gyre was spinning up from the mid-1990s to the mid-2000s, which was due to the enhanced wind stress curl reflecting a decadal or longer increase in the SAM. The 10th year of SW050P01 is selected to show the moment when the Weddell Sea gyre begins to spin up just before the occurrence of WSOP. In SW050P01 the oscillating wind forcing reaches its first peak at year 25, and therefore year 10 is located in the middle of increasing trend. As illustrated in Figs. 10b and 10d, both of the subtropical gyres spin up with the SH westerly winds intensifying at this time, and so does the Weddell Sea gyre. However, the Ross Sea gyre appears to weaken slightly and shrinks to the east of 130°W. This is supported by the analysis of the observed sea surface height anomaly (Sokolov and Rintoul 2009) showing that the ACC fronts in the Australian–Antarctic basin (i.e., in the South Pacific) clearly shift poleward over 15 years from 1992 to 2007, which corresponds to the period of prolonged positive SAM, whereas those to the east of the Drake Passage (i.e., in the South Atlantic) remain unchanged.

Figure 11 shows variation in the Weddell and Ross Sea gyres and in the South Atlantic and Pacific subtropical gyres during the whole period. In the polar gyres the positive (negative) anomalies indicate the spinup (spindown) of gyres, and the opposite is true in the

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subtropical gyres. Note that drastic intensification in the polar gyres is generated by oceanic deep convection due to the occurrence of open-ocean polynyas, not just by the enhanced curl of the overlying wind stress, in contrast to that in the subtropical gyres. Consistent with the analysis of observed data (Roemmich and Gilson 2009), the increment in the South Pacific subtropical gyre is larger than that in the South Atlantic subtropical gyre. Both the subtropical gyres vary in a positive correlation with the oscillating winds, intensify at most by a factor of about 1.4 in the positive phase, and return to their initial states in the negative phase. The subtropical gyres in SW02P04 and SW050P04 have small variations in comparison with experiments whose periods are longer. As discussed in conjunction with Figs. 10b and 10d, the Ross gyre is clearly weakened by the poleward-shifted

FIG. 11. Variation in strength of the Weddell (solid red) and Ross (solid blue) gyres and of the South Atlantic (dashed red) and South Pacific (dashed blue) subtropical gyres for the respective cases amplified by (a) 25% or (b) 50%. The strength is estimated by finding the maximum (minimum) value in the barotropic streamfunctions of the Weddell and Ross gyres (the South Atlantic and Pacific subtropical gyres), shown by thin dashed black lines. Time series of the maximum zonal-mean zonal wind stress between 80° and 30°S are respectively presented for reference.
and intensified South Pacific subtropical gyre in the positive phase unless the RSOP occurs. When the oscillation period is sufficiently long as in low-frequency experiments, the Ross gyre grabs a chance to spin up under the prolonged positive phase. However, when the oscillation period is not sufficiently long, the transition from spin-down to spinup does not occur in the Ross gyre during the relatively short positive phase. In this study, the competitive relation between the South Pacific subtropical gyre and Ross gyre under the intensifying SH westerly winds is the cause for the RSOP to occur later than the WSOP and not to occur under the relatively short oscillation periods. This hypothesis is supported by the aforementioned observation studies (Roemmich and Gilson 2009; Sokolov and Rintoul 2009).

Another cause for this is illustrated in Fig. 12 with the deep ocean heat (from 150-m to 4000-m depth) and minimum sea ice concentration in areas where the WSOP and RSOP are mainly simulated. The minimum sea ice concentration is analyzed to show the aforementioned criterion for open-ocean polynyas, which is not identified by the areal-mean sea ice concentration presented in Fig. 8. In most experiments except SW025P04, the RSOP occurs in the first cycle and persists longer than the WSOP occurring in the same cycle, and its minimum sea ice concentration is lower. In low- and midfrequency experiments where the WSOP occurs in every cycle, the deep ocean heat of the Weddell Sea that is depleted by the WSOP and oceanic deep convection is always fully recovered before the wind forcing reaches the following peak. A larger amount of deep ocean heat is released to air or is used to prevent new sea ice formation in the Ross Sea than in the Weddell Sea, which is due to the RSOP lasting longer than the WSOP in the first cycle. Together with this, the relatively slow recovery speed of the Ross Sea, which is due to the relatively weak Ross gyre bringing the warm CDW into the Ross Sea, causes the RSOP to rarely occur. Only in SW050P01, whose oscillating period and amplitude are longest and largest, is the depleted deep ocean heat of the Ross Sea almost recovered, and the RSOP occurs again shortly.

c. Dependence of the Weddell Polynya on the period of the SAM

Given that the Weddell Sea is much more favorable for the formation of open-ocean polynyas than the Ross Sea, why does the WSOP not occur in every cycle of the high-frequency experiments? In low- and midfrequency experiments, the drastically intensified Weddell Sea gyre always returns to its initial state before the wind forcing reaches its next peak (Fig. 11), and so does the underlying deep ocean heat that is depleted (Fig. 12). However, the Weddell Sea gyre intensified in the first cycle of the high-frequency experiments is not stabilized until the wind forcing reaches the second peak, which implies that open-ocean deep convection does not disappear completely until then. That is, on returning to the initial state, the Weddell gyre spins down, preventing the warm deep water mass from upwelling. Moreover, the depleted deep ocean heat is also not fully recovered before the second peak of the oscillating winds. Therefore, even if the WDW upwells, it is not warm enough to prevent sea ice from forming or to melt it, which is consistent with observation (Robertson et al. 2002; Smelstrud 2005) and modeling (Cheon et al. 2015) studies. These two factors make an adverse condition for the occurrence of WSOP in the second cycle. As previously stated, according to the slightly loose criterion for open-ocean polynyas, the WSOP does not occur in the second, fourth, sixth, and seventh cycles of SW025P04 and the second cycle of SW050P04 (Fig. 12), to which both or either factor is applied. Each cycle of oscillation is too short for the Weddell Sea to be sufficiently stabilized before the following cycle begins, providing an unfavorable condition for the WSOP. It implies that the SAM period (i.e., the durations of the positive and negative SAM phases) can be an important factor to control occurrence of open-ocean polynyas.

5. Summary and discussion

A sea ice–ocean GCM coupled to an energy balance model is used to investigate diverse responses of sea ice–ocean interaction to the sinusoidally oscillating SH westerly winds with three periods and two amplitudes. With the intensifying (weakening) westerly winds, the formation of coastal polynyas in the western Weddell and Ross Seas increases (decreases), resulting in an increase (decrease) in the ice-to-ocean salt flux within the polynyas. Coastal polynyas respond linearly to the oscillating wind forcing regardless of the seas and the oscillation periods, which, however, have a large influence on the formation of open-ocean polynyas. In the low- and midfrequency experiments the WSOP occurs near the peak of oscillating winds, whereas in the high-frequency experiment it happens not to occur. Although smaller in size and incidence than the WSOP, the RSOP responds to the same triggering mechanism that starts from spinup of the gyre and leads to upwelling of warm deep water. The cyclonic gyre is stronger in the Weddell Sea than in the Ross Sea, and so are the negative curl of overlying wind stress and the instability of water column. The South Pacific subtropical gyre strengthens more than the South Atlantic subtropical gyre, and only the ACC fronts adjacent to the northern
limb of Ross gyre shift slightly poleward. Once the deep ocean heat is depleted by the RSOP and oceanic deep convection, it is unlikely to recover until the oscillating winds reach the following peak. These cause the RSOP to occur later and less than the WSOP. In high-frequency experiments even the Weddell Sea does not return to the initial state at every oscillation cycle, revealing dependence of the WSOP on the period of the SAM.

The oscillation in intensity of the SH westerly winds, a focus of this study, is just one of several factors required for occurrence of open-ocean polynyas. In nature the

Fig. 12. Variation in the heat content from 150- to 4000-m depth of the Weddell Sea (solid red lines) and the Ross Sea (solid blue lines) and the minimum sea ice concentration of areas where open-ocean polynyas mainly occur in the Weddell Sea (dotted red lines) and the Ross Sea (dotted blue lines) for the respective cases amplified by (a) 25% or (b) 50%. Variation in the heat content is calculated by subtracting the heat content averaged over the last 100 years of spinup from that at the respective years, and 0 joules (thin solid black line) is plotted as an initial state. The centers of simulated WSOP and RSOP are mainly located in the areas of 75°–65°S, 45°–30°W and 75°–70°S, 160°–140°W, respectively. Time series of the maximum zonal-mean zonal wind stress between 80° and 30°S are respectively presented for reference (thin dashed black lines)
surface wind is highly variable in its temporal pattern (i.e., it is more complex than just a simple oscillation due to the SAM). In the mid-1970s when the Weddell Polynya occurred, the actual wind forcing would have been weaker than the oscillating wind forcing used in this study and might have been accompanied by other preconditioning factors, such as the drier-than-normal air conditions over the Weddell Sea during the prolonged negative SAM, the colder-than-normal air conditions during the La Niña period (Gordon et al. 2007), and enhanced upwelling within warm and salty eddies associated with the Maud Rise (Gordon and Huber 1984, 1990, 1995; Ou and Gordon 1986; Ou 1991; Alverson and Owens 1996). Although the debate on these factors is still open, the complex combination among these can be the cause for a rare occurrence of the WSOP.

According to the Fifth Assessment Report (AR5) of the Intergovernmental Panel on Climate Change (IPCC), the SAM index has become more positive since the 1950s (Hartmann et al. 2013). Although we showed a high correlation between the SAM index and zonal components of the wind speed at 10-m height and the wind stress derived from ERA-40 and ERA-Interim (Fig. 1a), the SAM index was recently proposed not to be a direct proxy for changes in the SH westerly winds (Swart et al. 2015; Thomas et al. 2015). However, Thomas et al. (2015), analyzing 30 climate models included in phase 5 of the Coupled Model Intercomparison Project (CMIP5) and 6 reanalysis datasets, showed that the SAM index and SH westerly jet strength had a positive trend over 1951–2011. The jet position did not show a significant trend. The representative concentration pathway scenario 8.5 applied to CMIP5 models also provided a significant positive trend in the SAM (Zheng et al. 2013). If this scenario comes true, the negative wind stress curl over the Weddell and Ross Seas is expected to intensify in the future climate, and so are the offshore winds in the western coasts of both the seas. These can increase both the formation of coastal polynya and the occurrence possibility of open-ocean polynya. However, de Lavergne et al. (2014), analyzing historical observations and CMIP5 model simulations, proposed that the SO stratification increased by the surface ocean freshening would prevent open-ocean polynya and deep convection under anthropogenic climate change. It may reduce the occurrence possibility of open-ocean polynya proposed in this paper.

Among many factors required for preconditioning or triggering the open-ocean polynya and deep convection in the SO, only the dynamic effect of the SAM was examined in this study. Since the Maud Rise was barely resolved in the bottom topography used in the model, an assessment its effect of a warm and salty WDW cell was not possible. Therefore, further studies are necessary to understand completely the whole process of the Weddell Polynya event and its impact on climate. For example, if a high-resolution regional sea ice–ocean coupled model able to permit effects of mesoscale eddies is used with the same wind forcing used in this paper, the combined effect of instantaneous spinup with warm and salty WDW eddies can be investigated. An application of the SAM component extracted from the atmosphere reanalysis data can be a more realistic wind forcing. Last, the application of the wind forcing used in this study to ocean models belonging to the “convecting model” group and the “non-convecting model” group classified in de Lavergne et al. (2014) and their comparison will lead to in-depth understanding for climate models.

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