Weddell Polynya Transport Mechanisms in the Abyssal Ocean

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ABSTRACT

Weddell Polynya transport mechanisms in the deep and abyssal oceans are examined in the NOAA Geophysical Fluid Dynamics Laboratory's (GFDL) coupled climate model CM2G. During an 1820-yr-long integration of the model, polynyas are forced every 29 years in the Weddell Sea via an increase in the diapycnal diffusivity. Composites of the events are used to examine the mechanisms responsible for transporting polynya signals away from the Weddell Sea. Polynya signal transport is governed by two dynamical mechanisms that act on different time scales and spread at different rates. Large-scale waves, such as Kelvin and planetary and topographic Rossby waves, propagate the polynya signal rapidly, on interannual-to-decadal time scales, while advection transports the signal more slowly, on decadal-to-centennial time scales. Despite their different spreading rates, these mechanisms can act contemporaneously, and it is often their combined effect that governs the property changes in the global deep and abyssal oceans. Both waves and advection cause temperature changes on isobaths. In the deep Atlantic, advection accounts for <15% of the total temperature change in the model, indicating that waves are strongly dominant there. Elsewhere, waves are still the stronger contributor, but advection accounts for 20%–40% of the total temperature change.

1. Introduction

The global deep and abyssal oceans have experienced marked changes in temperature, salinity, and heat content recently. Antarctic Bottom Water (AABW) production is declining (Purkey and Johnson 2012; Azaneu et al. 2013), and much of the deep and abyssal ocean is warming and freshening (e.g., Purkey and Johnson 2010, 2013; Johnson et al. 2014; Schmidtko et al. 2014; Anilkumar et al. 2015). Although the majority of these

changes are attributable to anthropogenic climate change, Zanowski et al. (2015) have shown that abyssal warming after the 1970s Weddell Polynya could account for up to 10% of that observed in the abyssal Southern Ocean. This enormous open-ocean polynya ($2\text{--}3 \times 10^5 \text{ km}^2$; Zwally and Gloersen 1977; Zwally et al. 1983; Carsey 1980) occurred during the 1974–76 austral winters. No polynya of similar size or duration has been observed in the Weddell Sea since that time. Within the polynya, surface cooling and brine rejection destabilized the weakly stratified water column and initiated intense convection that reached to nearly 3000 m (Martinson et al. 1981; Gordon 1982). As a result, the rate of AABW formation increased and the Weddell Sea cooled and freshened for several years after the event (Gordon 1978, 1982; Martinson 1991; Robertson et al. 2002). Since the Weddell Polynya's cessation, the deep Weddell Sea warmed (Robertson et al. 2002; Fahrbach et al. 2004; Smedsrud 2005; Fahrbach et al. 2011), albeit with significant temporal variation that has also been linked to Weddell Gyre variability and changes in the properties of Weddell seawater mass constituents [e.g., Circumpolar Deep Water (CDW);

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Fahrback et al. 2004, 2011; Couldrey et al. 2013; Ryan et al. 2016].

Modeling studies have attempted to elucidate the impacts of Weddell Sea polynyas on the global oceans (e.g., Goosse and Fichefet 2001; Hirabara et al. 2012; Latif et al. 2013; Martin et al. 2013, 2015; Wang et al. 2016). During active convection, large open-ocean polynyas cause cooling and freshening that spreads into the deep and abyssal oceans. After convection ceases, the cooling and freshening is followed by warming and salinification as the deep and abyssal oceans relax to their mean state (Zanowski et al. 2015). Recent studies suggest that polynya impacts are neither confined to the deep and abyssal oceans nor do they solely manifest themselves in water mass property changes (e.g., Hirabara et al. 2012; Latif et al. 2013; Martin et al. 2013; Wang et al. 2016). Evidence of strong dynamical and circulation impacts exists as well. Martin et al. (2015) find that Weddell Sea convection can drive variations in the Atlantic meridional overturning circulation (AMOC) on multicentennial time scales, while Behrens et al. (2016) find that deep convection throughout the Southern Ocean is responsible for increased Drake Passage transport and increased subpolar gyre variability on decadal time scales. Using a high-resolution climate model, Wang et al. (2016) find that Weddell Sea polynyas can affect ice shelf basal melt via their impact on Weddell Gyre circulation. Additionally, polynya impacts need not be oceanic at all, as evidenced by Pedro et al. (2016) and Cabré et al. (2017), who find that Southern Ocean deep convection can drive Southern Hemisphere atmospheric circulation and temperature variability.

Despite advances in our understanding of polynya impacts on the climate system, it is still unclear how polynya signals are transported in the deep and abyssal oceans and on what time scales these transport mechanisms act. Previous studies of deep and abyssal ocean dynamics indicate that large-scale waves (e.g., Kelvin and Rossby waves) also propagate oceanic signals but on time scales faster than advection (e.g., Kawase 1987; McDermott 1996; Stössel and Kim 2001; Nakano and Sugimoto 2002; Johnson et al. 2008; Santoso and England 2008; Masuda et al. 2010; Meredith et al. 2011; Hirabara et al. 2012). Coles et al. (1996) found that via wave propagation, Weddell Sea cooling signals, possibly because of the 1970s Weddell Polynya, can reach as far north as 18°S in the Atlantic Ocean in as little as 5–11 years. Nakano and Sugimoto (2002), Fukasawa et al. (2004), and Masuda et al. (2010), among others, found similar results in the Pacific Ocean. Using an adjoint model, Masuda et al. (2010) found that warming signals originating near the Adélie Coast, Antarctica, can propagate to the North Pacific in approximately 40 years.

In this study, we regularly force Weddell polynyas in a coupled climate model to examine polynya signal transport dynamics in the deep and abyssal oceans. We find that polynya signals are transported by two dynamical mechanisms that act on different time scales and spread at different rates. Large-scale waves, such as topographic and planetary Rossby and Kelvin waves, propagate polynya signals on interannual-to-decadal time scales, while advection acts on decadal-to-centennial time scales. Both mechanisms give rise to temperature changes on isobaths: waves via the vertical displacement of isopycnals and advection via transport of water from the source region. We find that waves are the dominant driver of temperature anomalies in most regions, as advection accounts for less than half of the total polynya-induced temperature change.

2. Methods

a. Model

The model used for this study is the NOAA Geophysical Fluid Dynamics Laboratory's (GFDL) coupled climate model CM2G. CM2G is a coupled land–atmosphere–ocean–sea ice model. It is the physical component of one of two GFDL Earth System Models (ESM2G; Dunne et al. 2012) that participated in phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012). Details of the atmosphere, sea ice, and land components of CM2G can be found in Anderson et al. (2004), Winton (2000), and Milly et al. (2014), respectively. Because waves propagate along density surfaces in the real ocean, we have chosen to use an ocean model that explicitly represents these surfaces. The ocean component of CM2G, Generalized Ocean Layer Dynamics (GOLD; Adcroft and Hallberg 2006; Hallberg and Adcroft 2009), is an isopycnal coordinate model with 1° horizontal resolution. The latitudinal resolution refines to 1/3° near the equator. GOLD is configured with 63 vertical density layers referenced to 2000 dbar.

b. Simulations

Two simulations, one with forced polynyas (FP) and a control without forced polynyas (CTL), are branched from the end of a 5000-yr spinup of CM2G that was forced with 1860 preindustrial conditions. CTL is run for 1500 years, and FP is run for 1820 years.

1) POLYNIA FORCING

In FP, polynyas are forced every 29 years in the Weddell Sea. We chose a 29-yr interval for the polynyas

FIG. 1. (a) Freshwater forcing (60° – 68° S, 50° W– 130° E; blue region), polynya forcing (increased diapycnal diffusivity; gray circle), and polynya averaging regions (61° – 68° S, 15° – 25° E; orange box). (b) Composite winter (June–September) SIC (color) and 1000- (dashed) and 2000-m (solid) MLDs averaged over the 10-yr polynya on period.

because it is long enough to ensure heat and salt reservoir replenishment in the Weddell Sea after each polynya, but short enough to prevent unforced convection between the forced events. The polynyas are forced via a diapycnal diffusivity increase in a circular region of 500-km radius ($7.8 \times 10^5 \text{ km}^2$ surface area), centered at 64° S, 20° E (Fig. 1a). The 1970s Weddell Polynya developed about 800 km farther west, near Maud Rise ($\sim 65^{\circ}$ S, 3° E). However, we find that in order to ensure strong oceanic heat loss during the forced events, we are required to force them over the same region where the model has its strongest heat and salt reservoirs at depth, hence the forcing centered at 20° E.

The diffusivity increase acts over the top 47 layers, which is approximately equivalent to 3000-m depth in the forcing region. The diapycnal diffusivity profile is Gaussian in the horizontal, and its value reduces to that of the background diffusivity ($\sim 10^{-4}$ – $10^{-5} \text{ m}^2 \text{ s}^{-1}$) at the edge of the polynya. Its maximum value is $0.5 \text{ m}^2 \text{ s}^{-1}$ at the polynya center. The polynyas are forced between June and October for 10 years of their 29-yr period. We refer to these 10 years as the polynya “on” period and the following 19 years as the polynya “off” period. During the unforced months (November–May) and unforced years (years 11–29 of each event), the diapycnal diffusivity remains at its background value. The forced events are not intended to be comparable to the 1970s Weddell Polynya in terms of their size, strength, or duration. They are deliberately designed to be extremely large, powerful events so that their perturbations in the deep and abyssal ocean are more easily

distinguishable from ambient variability, and thus their transport mechanisms can be cleanly studied.

2) SUPPRESSING NATURALLY OCCURRING CONVECTION

To suppress the naturally occurring open-ocean convection in the model’s Southern Ocean, FP and CTL are forced with 40 cm yr^{-1} of freshwater in a band covering the most active regions of open-ocean convection. The added freshwater is balanced by freshwater removal everywhere north of 60° S. In CM2G, open-ocean convection typically occurs in the area spanning 60° – 68° S, 50° W– 130° E (Fig. 1a; blue region). The 40 cm yr^{-1} of freshwater forcing in this band is equivalent to $\sim 15 \text{ cm yr}^{-1}$ ($\sim 3000 \text{ Gt yr}^{-1}$) of freshwater spread evenly south of 60° S. This is a substantial addition of freshwater. Current estimates of Southern Ocean freshwater input from Antarctic ice shelf and iceberg melt are ~ 2300 – 2400 Gt yr^{-1} (e.g., Rignot et al. 2013; Liu et al. 2015). The freshwater forcing causes a maximum 0.15 kg m^{-3} density change in the densest waters over the length of the simulations. Although this drift is large, we are still able to examine the polynya signals as long as the drift is removed via detrending or filtering. For a more complete diagnosis of the impact of the freshwater forcing on the model’s circulation and deep-water properties, refer to the supplemental information.

Despite the freshwater forcing, Weddell Sea convection still occurs in CTL but at a rate reduced from one event per 50 years to one event per 200 years. Both Stössel et al. (2015) and Kjellsson et al. (2015) found similar results using freshwater forcing in their models.

Additionally, Kjellsson et al. (2015) and Heuzé et al. (2015) indicate that tuning the vertical diffusivity can help reduce convection. We did not attempt this modification because it is unclear from these studies whether or not vertical diffusivity tuning alone is sufficient to remove convection in long-running simulations. In FP, the regularity of the forced polynyas prevents unforced Weddell Sea convection from occurring. In both CTL and FP, open-ocean convection outside the Weddell Sea disappears completely.

c. Composites and polynya metrics

A total of 63 polynyas are forced during FP. Composites of these events are used to study the polynya signals in the deep and abyssal oceans. For most of the fields analyzed, the number of polynyas used in the composite depends on the temporal length of the composite. All 63 polynyas are used when the composite length is 29 years, as the final year of the simulation occurs just before the sixty-fourth polynya would be forced. Longer composites (100–250 years) contain 50 polynyas because of the extended length of time utilized for the averaging periods. Variables analyzed on density surfaces are subject to further restrictions because the freshwater forcing causes the deepest isopycnal layers to disappear over time. When this is the case, only polynyas that occurred before the isopycnal layer disappeared are used for the composite. For the deepest isopycnal layer, which disappears the earliest, the composite consists of three polynyas. This is the smallest number of polynyas used in any composite. The only variable we analyze on density surfaces is the interface height, which is the depth of an isopycnal relative to the mean sea level. When discussing interface height, we will note how many polynyas are used in each composite.

Composite polynya winter sea ice concentration (SIC) and mixed layer depth (MLD) are calculated by averaging June–September values over the polynya “on” period (Fig. 1b). All other variables used to diagnose the composite polynya state are annual-mean values averaged in the polynya region (61°–68°S, 15°–25°E; Fig. 1a; orange box). Dense-water formation in the polynya region is estimated by calculating the change in thickness of all deep layers with a positive thickness perturbation during the polynya “on” period. This value is then multiplied by the polynya area and divided by the time (10 years).

d. Advective signals

Dye tracer is used to study the polynya advective signal. During each event, a separate dye tracer is added to the two uppermost layers of the polynya region to tag the polynya water. Tracer values are reported as a mass

fraction (kg kg^{-1}) and so are unitless. The initial value of the tracer is 1.0. A 0.001 threshold is used to determine the tracer arrival time at specific locations. The dye tracer’s dynamic range (1000) is similar to that of other passive tracers such as chlorofluorocarbons (CFCs; 350 for CFC-11 and 700 for CFC-12; Bullister and Weiss 1988).

e. Wave signals

Interface height anomalies (variations in the depth of isopycnal layers) are used to study the polynya wave signal. To distinguish the polynya signal from other internal variability, all interface heights are bandpass filtered with a Butterworth filter in the frequency range $1/30$ to $1/10 \text{ yr}^{-1}$ and then composited. Filtering also removes the effect of the freshwater forcing on the interface heights because it treats the long-term isopycnal heave as a very low-frequency signal that is removed. All subsequent references to “low frequency” signify the bandpassed 10–30-yr time scales.

Wave arrival times are calculated by convolving a square wave kernel with the unfiltered, normalized interface height time series at a particular location. The time of the first extremum of the convolution gives the arrival time of the waves relative to the beginning of the first forced polynya. In this case, the CTL interface heights are subtracted before convolving because no filtering is applied.

Composite potential temperature and salinity contributions to density changes at 3900 m are calculated via the following formula, where θ is potential temperature, S is salinity, and ρ is density:

$$\Delta\rho = \frac{\partial\rho}{\partial S}\Delta S + \frac{\partial\rho}{\partial\theta}\Delta\theta. \quad (1)$$

Partial derivatives are calculated using the 1980 International Equation of State of Seawater (UNESCO 1981). Density changes are calculated over the polynya “on” and “off” periods. The control simulation density changes are subtracted before compositing.

f. Temperature anomaly calculations

Composite bandpass filtered potential temperature anomalies are examined in several regions where both wave and advective signals from the polynyas are likely to occur. Potential temperature anomalies are calculated as a function of distance along the deep (3900 m) western boundaries of the Atlantic, Pacific, and Indian Oceans, at 860 m across the North Atlantic (averaged 13°–15°N), and at 530 m across the South Pacific (averaged 8°–10°S). Regions whose anomalies are statistically

different from the mean are determined via bootstrapping. The underlying sampling distribution for the control simulation variability is determined by taking the mean of 50 random samples 10^5 times. Two-sided 90% confidence intervals are determined from the sample distribution. Temperature anomaly mean speeds are estimated by fitting a line to the locations of the potential temperature maxima or minima in time. The temperature anomaly mean speeds are then compared to the mean speeds of the dye tracer in the same locations to determine whether the temperature anomalies are the result of waves or advection.

A comparison of the 3900-m total minimum temperature anomaly and a dye tracer–derived advective cooling field is also used to determine whether the temperature anomalies are the result of waves or advection. The constructed field is not exactly the same as the true advective temperature field mainly because heat and dye tracer diffusion are subject to different background gradients. Nonlinear processes such as diffusion are likely small over the time scales we examine, so we do not expect them to significantly contribute to the temperature anomalies. The total minimum temperature anomaly is calculated by taking the composite minimum temperature anomaly at each point over the 29-yr polynya period. The advective cooling field is constructed by multiplying the average cooling in the polynya region (-0.2°C) by the composite maximum tracer field. The composite maximum tracer field is calculated by taking the temporal maximum value of the tracer at each point and then masked to ignore tracer in regions where the magnitude of the minimum temperature field is $<0.002^\circ\text{C}$. The ratio of the constructed advective cooling signal and the total minimum temperature signal yields the fraction of the total temperature signal that is the result of advection. Temperature changes that are not fully explained by the constructed advective temperature field are attributable to large-scale waves.

3. Results

a. Polynya characteristics

The composite polynya is characterized by a mean winter sea ice concentration below 10% and MLDs that exceed 1000–2000 m (Fig. 1b). The areas of MLD >1000 m (Fig. 1b; dashed line) and >2000 m (Fig. 1b; solid line) are 5.1×10^5 and $2.2 \times 10^5 \text{ km}^2$, respectively. Annual-mean dense-water formation in the polynya is $\sim 1.5 \text{ Sv}$ ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). This is a significant input of dense water to the abyssal oceans. In FP, the total mean deep overturning across 32°S is 25.6 Sv and in the Atlantic sector, 0.79 Sv. Averaged

over the final 2000 years of the spinup, these values are 28.7 and 2.3 Sv, respectively. In both simulations, Atlantic sector values are low compared to observations (~ 4 – 6 Sv ; Ganachaud and Wunsch 2000; Talley 2008, 2013; Naveira Garabato et al. 2014). By the end of FP, the forced polynyas create the majority of the dense water in the Atlantic sector.

During active convection, heat loss to the atmosphere leads to surface warming and subsurface cooling, and the sea ice concentration reduces to nearly zero (Fig. 2). An average cooling of 0.2°C occurs between 1500 and 3000 m. After convection ceases, the surface cools and the subsurface warms. The sea ice concentration recovers rapidly, typically reaching prepolynea values (0.5–0.6) in the year following each event's cessation.

Despite constraining the depth of the diapycnal diffusivity increase, convective mixing reaches the seafloor, as demonstrated by the warming of the deepest waters (>4000 m) after each event's commencement. This effect is unrealistic (these waters should either cool initially or remain unperturbed) and is most likely due to the strength of the imposed mixing. However, the waters below 4000 m are still deep enough and dense enough that they do not escape the model's Argentine–Weddell–Enderby Basin complex, and so their impact on external abyssal regions is negligible.

b. Advective signals

The dye tracer identifies the advective pathways of the polynya signal in the deep and abyssal oceans. These pathways generally match those of the large-scale deep and abyssal ocean circulation (e.g., Reid 1986, 1989, 1994, 1997, 2003) and those of southern-sourced dense water (e.g., Johnson 2008). At 3900 m, the most northward extent of the tracer is about 30°S after 25 years (Fig. 3a). The largest mass fractions occur in the Atlantic and Indian sectors of the Southern Ocean. Somewhat weaker mass fractions exist along the Antarctic continental boundary in the eastern Pacific sector as well. Tracer in this region is advected westward through Drake Passage despite the prevailing eastward flow of the Antarctic Circumpolar Current (ACC). Boundary currents that develop in the wake of topographic Rossby waves most likely advect this signal.

In the Atlantic, Pacific, and Indian Oceans, the strongest tracer mass fraction and farthest northward extent occur in the west where deep western boundary currents (DWBCs) rapidly advect the polynya water northward. Via the DWBCs, tracer reaches the equatorial Pacific and Indian Oceans after 50 years (Fig. 3b). After 100 years (Fig. 3d), tracer is found in the equatorial Atlantic, well into the North Pacific, and throughout the entire Indian Ocean. Small values (~ 0.001) of the

FIG. 2. (top) Annual mean SIC and (bottom) potential temperature ($^{\circ}\text{C}$) profile over time, averaged in the polynya region (Fig. 1a; orange box). The contour interval (gray lines) is 0.2°C . Vertical black lines indicate the beginning of each forced polynya.

tracer also exist in the North Atlantic. The southward flow of the DWBC in the North Atlantic precludes the possibility of the tracer arriving there via this deep pathway. Instead the tracer is advected northward above 1000 m (not shown). It then sinks in the dense-water formation regions in the North Atlantic and is advected southward in the DWBC.

Tracer mass fractions are not vertically coherent, consistent with the three-dimensional structure of the large-scale overturning. Composite tracer profiles at selected locations (Fig. 4) suggest that the tracer's vertical structure is dependent on the locally dominant water masses. Although tracer is mixed throughout the water column in the polynya region (not shown), the majority of the polynya water is dense enough that it sits in the bottom of basins outside of the Weddell Sea, similar to AABW and Lower Circumpolar Deep Water (LCDW) (e.g., Orsi et al. 1999; Sloyan and Rintoul 2001; Talley 2013). At all four profile locations, the highest tracer mass fractions occur below 3000 m, indicating that the polynya signal most strongly influences the bottom waters in these regions. Somewhat lower tracer mass fractions (0.005–0.01) occur above 1000 m, while the lowest values occur between 1000 and 3000 m, especially in Vema Channel (VC; Fig. 4c) and the entrance to the Madagascar–Mascarene Basin (MMB; Fig. 4d). In both locations, the local deep-water component, that is, North Atlantic Deep Water (NADW) and/or Indian Deep Water (IDW)/Pacific Deep Water (PDW), has a

prevailing influence from ~ 1000 to 3000 m (e.g., Talley 2013). Tracer mass fractions remain lower for longer at these depths because the tracer, having not originated in these water masses, must enter them via mixing. This is also why the tracer arrival times at these depths (0.001 threshold; Fig. 4; navy lines) tend to be much later than in other parts of the water column.

At the more southerly points in the southwest Pacific (SWP) and Amundsen–Bellingshausen Basin (ABB; Figs. 4a,b), the range of tracer arrival times over all depths is smaller (5–40 yr in Figs. 4a,b vs 20–160 yr in Figs. 4c,d), likely because of a combination of the local circulation and the stronger influence of southern-sourced waters throughout the water column in these regions. At SWP (Fig. 4a), where the circulation is predominantly governed by the ACC, the tracer arrival time increases with depth, owing to the decrease in current speed with depth. In all four regions, the wind-driven circulation dominates the upper water column, leading to relatively fast (~ 5 –40 yr) arrival times above 1000 m. Similarly rapid arrival times (~ 10 –30 yr) occur below 4000 m. At these depths, the tracer arrival time is governed by the speed of the abyssal flow, and the comparably fast arrival times are mostly a consequence of regional boundary currents that advect the tracer quickly.

c. Wave signals

Polynya wave signals manifest themselves as deviations in the depth of the model's isopycnal surfaces.

FIG. 3. Composite tracer mass fraction (kg kg^{-1}) at 3900 m at (a) 25, (b) 50, (c) 75, and (d) 100 years relative to the start year of the polynya. Black boxes indicate the polynya region. Bathymetric contours (gray lines) are at 2500 and 3500 m.

We refer to these deviations as interface height anomalies. Composite interface height anomalies along $\sigma_2 = 36.99 \text{ kg m}^{-3}$ (Fig. 5; composite of 15 polynyas) and $\sigma_2 = 37.09 \text{ kg m}^{-3}$ (Fig. 6; composite of 3 polynyas) illustrate the horizontal pattern of the wave signals (σ_2 is defined as $\rho_2 - 1000 \text{ kg m}^{-3}$, where ρ_2 is the in situ density at 2000 dbar). In the polynya region the mean depths of these isopycnals are 500 and 3200 m, respectively. Polynya wave signals are present globally. Large, 100–500-m interface height anomalies occur in the Weddell Sea and South Atlantic (Fig. 5). As distance from the polynya region increases, smaller, 10–50-m anomalies occur elsewhere in the Southern

Ocean and in the North Atlantic. Very small ($<10 \text{ m}$) anomalies occur in the North Pacific. Although these anomalies appear in the composite, they are too small to significantly impact local temperature, which will be discussed later.

The patterns of isopycnal displacement evolve over the polynya's 29-yr period. These temporal variations are evident when comparing Figs. 5a–d, which show the quarterly (~ 7 years) signal evolution. Figures 5a and 5c (Figs. 5b,d) differ by 15 years or approximately half the period of the forced polynyas. Therefore, the anomalies in Fig. 5a (Fig. 5b) should be reversed in sign relative to those in Fig. 5c (Fig. 5d), which is the case. An equivalent relationship holds when comparing Figs. 6a–d. Because it takes many years for the waves to propagate to remote regions, not all of the signals in Figs. 5 and 6 are due to the same polynya. This is why at the start year of the composite polynya (year 0; Figs. 5a, 6a) wave signals exist throughout the ocean even though no time has passed since the beginning of the event. Figures 5 and 6 demonstrate the evolution of the polynya signal once it reaches a certain location rather than its transit from place to place.

The group velocities of the waves propagating the polynya signal determine the wave arrival times. It is unlikely that dispersive waves, because of their wavenumber-dependent group velocities, will give rise to a coherent signal with a well-defined arrival time. Instead, it is more likely that fast, nondispersive waves, such as Kelvin and long Rossby waves, determine the arrival times. At SWP, ABB, VC, and MMB the waves arrive in less than 15 years at 4000 m (Table 1). Compared to the tracer arrival times, wave arrival times are 1.5–2.6 times faster.

The waves impact the local velocity field as they propagate from the polynya region. When low-frequency topographic Rossby waves are present, the local along-slope velocity component experiences the largest perturbations, while the across-slope velocity perturbations are typically much weaker (e.g., Rhines 1970). This is the case in Fig. 7, which shows composite profiles of along- and across-slope velocity perturbations at SWP, ABB, VC, and MMB. The along- and across-slope velocity components are aligned with the zonal and meridional velocity components at these locations. At each location, the deep ($>3000 \text{ m}$) along-slope velocity perturbations are large in comparison to the corresponding across-slope velocity perturbations, consistent with the presence of topographic Rossby waves. Large velocity perturbations are also present in the upper 1000 m. The zonal velocity u (u is along slope in Figs. 7a,b and across slope in Figs. 7c,d) experiences the largest perturbations, indicating the presence of planetary Rossby waves.

FIG. 4. Vertical profiles of composite tracer mass fraction (kg kg^{-1}) over time in the (a) southwest Pacific, (b) Amundsen–Bellingshausen Basin at 100°W , (c) Vema Channel, and (d) entrance to the Madagascar–Mascarene Basin. Times are relative to the start year of the polynya. Thick navy contours delineate the 0.001 threshold used to determine tracer arrival times. (e) A map of the profile locations (stars). Bathymetric contours (gray lines) are at 2000, 3000, and 4000 m.

Under certain circumstances, the presence of the planetary vorticity gradient and the topography allows for a dynamical coupling between the deep and surface ocean. Provided that the topographic and planetary vorticity gradients are parallel or antiparallel, deep perturbations can excite perturbations in the surface (e.g., Hallberg and Rhines 1996; Hallberg 1997). These surface perturbations can then propagate as planetary Rossby waves. Despite the strong zonal velocity perturbations at each location, only SWP and ABB (Figs. 7a,b) are oriented such that the planetary and topographic vorticity gradients are appropriately aligned for such a coupling to occur for low-frequency waves (such as those picked out by the compositing). Elsewhere, deep and surface velocity perturbations exist, but they are not necessarily coupled.

d. Temperature anomalies

Both advection and waves cause temperature changes on isobaths. However, determining whether temperature signals are most likely due to waves or advection requires an examination of the transit speeds of the temperature anomalies and the dye tracer. In most of

the regions where the polynya signal exists, complete density compensation caused by temperature and salinity changes does not occur (Fig. 8). This incomplete compensation is consistent with the presence of large-scale waves, but it does not confirm that waves are the sole driver of the changes. Advective signals also need not be density compensated. However, complete density compensation in a particular region does suggest that temperature and salinity signals there are moving advectively.

At 3900 m, statistically significant warming and cooling occurs along the western boundaries of the Atlantic, Indian, and Pacific Oceans (Figs. 9–11). Temperature anomaly magnitudes range from 0.001° to 0.2°C . Maximum temperature anomalies are smaller in the Pacific ($\pm 0.05^\circ\text{C}$) than in the Atlantic ($\pm 0.2^\circ\text{C}$) or Indian ($\pm 0.1^\circ\text{C}$) Oceans because the distance from the polynya forcing region is larger. In each basin, the temperature anomalies are largest in the south and decay northward.

In the Atlantic (Fig. 9), large temperature anomalies ($\sim \pm 0.02^\circ\text{C}$ – 0.1°C) also exist north of $\sim 40^\circ\text{N}$. However, most of these anomalies are indistinguishable from

FIG. 5. Bandpassed composite interface height anomalies (m) along $\sigma_2 = 36.99 \text{ kg m}^{-3}$ at (a) 0, (b) 7, (c) 15, and (d) 22 years relative to the polynya start year. Black boxes mark the polynya region. Because of the drift caused by the freshwater forcing, only 15 polynyas are used for this composite (refer to [section 2c](#)).

other variability at 80% confidence. None of the signal in the North Atlantic is statistically different from the mean at 90% confidence. Observations indicate that the North Atlantic experiences strong, low-frequency variability such as that forced by the Atlantic multidecadal oscillation (AMO; [Kerr 2000](#)). Multidecadal, AMO-like variability is also present in many coupled models (e.g., [Delworth and Mann 2000](#); [Ting et al. 2009](#)) and is linked

to a wide range of climate phenomena, including variations in the AMOC, Atlantic hurricane activity, and India/Sahel rainfall (e.g., [Zhang and Delworth 2006, 2009](#); [Zhang et al. 2011](#)). Given that CM2G experiences North Atlantic variability on time scales similar to that of the polynya forcing, it is inappropriate to attribute all of the temperature anomalies in this region to the polynyas. Possible connections between the polynya

FIG. 6. As in [Fig. 5](#), but along $\sigma_2 = 37.09 \text{ kg m}^{-3}$. Because of the drift caused by the freshwater forcing, only three polynyas are used for this composite (refer to [section 2c](#)).

TABLE 1. Tracer and wave signal arrival times at 4000 m at SWP, ABB at 100°W, VC, and the entrance to the MMB. Arrival times are relative to the initial year of the first forced polynya. Refer to Fig. 4e for locations.

Location	Wave arrival time (yr)	Tracer arrival time (yr)	Ratio ($t_{\text{tracer}}/t_{\text{wave}}$)
ABB	6	11	1.8
MMB	12	18	1.5
SWP	10	26	2.6
VC	11	24	2.2

forcing and North Atlantic variability are further discussed in section 4c.

The 3900-m western boundary temperature anomalies (Figs. 9–11) move faster than the dye tracer in the same locations (not shown). This suggests that the temperature signals in these regions cannot be entirely

advective. Mean temperature anomaly propagation speeds fall between 0.008 and 0.022 m s^{-1} , while tracer speeds fall between 0.005 and 0.01 m s^{-1} (Table 2). In the Atlantic Ocean, the temperature anomalies travel ~ 2.5 times faster than the tracer, while in the Indian Ocean the temperature anomalies are 1.6 times faster. In these regions, large-scale waves are most likely propagating the temperature anomalies because advection is too slow to explain their rapid movement.

In the Pacific Ocean, the temperature anomalies move 1.1 times faster than the tracer. Because the temperature anomalies are moving faster than the tracer, their behavior is consistent with the presence of large-scale waves. Additionally, at 3900 m in the southwest Pacific, complete density compensation does not occur everywhere (Fig. 8). However, the total density change is typically very small ($<0.005 \text{ kg m}^{-3}$), suggesting that

FIG. 7. Mean horizontal velocity and composite bandpassed along- and across-slope velocity (cm s^{-1}) anomalies at (a) SWP, (b) ABB, (c) VC, and (d) MMB. Refer to Fig. 4e for locations. Mean velocity profiles are averaged over the length of FP, while the along- and across-slope velocity anomalies show the evolution of the perturbations over one polynya period (29 yr). Dashed gray lines denote zero velocity. At SWP and ABB, the topography is oriented such that the along-slope velocity is in the direction of the zonal velocity u , and the across-slope velocity is in the direction of the meridional velocity v . At VC and MMB, the topography is oriented such that the along-slope velocity is in the direction of the meridional velocity v , and the across-slope velocity is in the direction of the zonal velocity u .

FIG. 8. Composite density change at 3900 m caused by (a),(d) temperature change, (b),(e) salinity change, and (c),(f) their sum during the (top) polynya “on” and (bottom) polynya “off” periods. Black boxes mark the polynya region. During both periods, temperature changes dominate the density changes and are not fully compensated by opposing salinity changes. Density changes are not exactly equal and opposite during the “on” and “off” periods because of their different temporal lengths.

advection contributes to temperature anomaly transport in this region. Given that the temperature anomaly and tracer speeds differ by only 10%, we cannot use this speed comparison method to unambiguously determine the dominant transport mechanism in this region.

Instead, we employ an additional method to determine wave and advective temperature change contributions: the ratio of the polynya’s advective cooling to its total minimum cooling (refer to [section 2f](#) for further details). The 3900-m total minimum cooling, the advective cooling, and their ratio are shown in [Fig. 12](#). Ratios above 0.5 indicate that advection is mostly responsible for the temperature changes, while ratios below 0.5 indicate that large-scale waves are mostly responsible. Most of the temperature anomalies are a result of a combination of waves and advection, but waves often dominate ([Fig. 12c](#)). Consistent with the results of the temperature anomaly and tracer speed comparison, very low ratios (<0.15) occur over much of the Atlantic Ocean, suggesting that large-scale waves are mostly responsible for propagating the temperature anomalies there. Similarly, in the Indian Ocean, the lowest ratios occur along western boundaries, again suggesting the influence of large-scale waves. However, the ratios are typically larger (~ 0.2 – 0.4) than in the Atlantic, and they approach 1.0 (all advective) toward the north. This suggests that advection is also important for polynya-induced temperature changes in the Indian Ocean and perhaps explains why the temperature anomaly to tracer speed ratio is smaller here than in the Atlantic (1.6 vs ~ 2.5).

Ratios below 0.5 also occur along the western boundary of the southwest Pacific. As in the Indian

Ocean, these ratios are >0.5 and approach 1.0 toward the north. In minor places throughout the oceans, the ratio exceeds 1.0. A ratio >1.0 means that the values of the advective cooling field are larger than the total cooling, which is unphysical. This could indicate that other processes, such as diffusion, are important in those locations. Another possibility is that the advective and wave signals destructively interfere in these regions. If this is the case, then the total cooling will be smaller than the temperature perturbations that arise from the individual mechanisms, and thus the ratio can exceed 1.0. We will further discuss our wave versus advection attribution method and its caveats in [section 4b](#).

At shallower depths, planetary Rossby waves propagate the polynya signal westward across the Atlantic and Pacific Oceans ([Fig. 13](#)). At 860 m in the North Atlantic ([Fig. 13a](#)), temperature anomalies propagate at a mean speed of 0.073 m s^{-1} . At 530 m in the South Pacific ([Fig. 13b](#)), temperature anomalies propagate at a mean speed of 0.23 m s^{-1} . In both locations, the speed of the temperature anomalies is at least one order of magnitude larger than the advective speed of the tracer (not shown). Anomalies with magnitude $>0.02^\circ\text{C}$ in the Atlantic and $>0.005^\circ\text{C}$ in the Pacific are statistically significant at 90% confidence.

4. Discussion

a. Polynya transport mechanisms

Polynya signal transport is governed by two dynamical processes that act on different time scales and spread at

FIG. 9. Bandpassed composite potential temperature anomalies ($^{\circ}\text{C}$) at 3900 m as a function of distance along the Atlantic western boundary. Hatching (diagonal lines) and stippling (dots) denote anomalies statistically different from the mean at the 90% and 80% significance levels, respectively. In the left panel, black dots and red stars mark the path of the signal. Red stars occur at 5000-km intervals and correspond to the y-axis distances in the right panel.

different rates. Advection is the slower of these two mechanisms. In the deep ocean, polynya advective signals spread according to the large-scale abyssal circulation (e.g., Reid 1986, 1989, 1994, 1997, 2003). Polynya water advection causes deep and abyssal ocean property changes on decadal-to-centennial time scales. At 3900 m, tracer from the polynya arrives in the equatorial Pacific after 50 years and in the North Atlantic after 100 years (Fig. 3). The large-scale circulation and the dominant water masses determine the tracer's vertical structure at a particular location. Tracer arrival times at depths dominated by southern-sourced waters are typically faster than at depths dominated by other water masses, such as NADW (e.g., Fig. 4c).

Kelvin and topographic and planetary Rossby waves propagate the polynya signal on interannual-to-decadal time scales. Wave signals exist globally, with the largest density perturbations occurring closest to the polynya region. At 4000 m, wave signals arrive 1.5–2.6 times faster than advective signals (Table 1) and reach locations in the southern Atlantic, Pacific, and Indian

Oceans in less than 15 years. This is consistent with estimates from Coles et al. (1996) and Hirabara et al. (2012), who find that wave signals from the Weddell Sea can reach the South Atlantic and southwest Pacific in a decade or less.

Both waves and advection cause temperature changes on isobaths. At 3900 m along the western boundaries of the Atlantic, Pacific, and Indian Oceans, temperature anomaly magnitudes fall between 0.001° and 0.2°C and decrease with distance from the polynya region. However, only a small subset of the anomalies is statistically significant at 90% confidence (Figs. 9–11). In the Atlantic and Indian Oceans, these deep temperature anomalies travel 1.6–2.75 times faster than the tracer along the same paths, suggesting that large-scale waves are propagating the anomalies there.

In the deep southwest Pacific, the temperature anomalies travel only 1.1 times faster than the tracer. Although the temperature anomaly speed is faster than the advective speed, the small difference between the two suggests that advection is also an important transport mechanism for polynya signals in this region.

FIG. 10. As in Fig. 9, but for the Indian western boundary. Red stars occur at 2000-km intervals and correspond to the y-axis distances in the right panel.

Perhaps the most interesting aspect of the deep Pacific signal is its rapid advective speed; the tracer can travel from $\sim 55^\circ$ to $\sim 10^\circ\text{S}$ ($\sim 10^5$ km) in ~ 30 years (Fig. 3; Table 2). While wave signals travel faster, the tracer's advective speed in this region is such that the advective time scale is not as drastically different from the wave time scale, as might be expected based on ventilation time scales inferred from the meridional overturning circulation. Abyssal waters in the North Pacific are ventilated on an ~ 800 – 1000 -yr time scale (e.g., Matsumoto 2007), but this time scale includes the effects of mixing and diffusion and thus represents the average time it takes for the water to be ventilated. However, as the dye tracer advection along the deep western boundary of the Pacific indicates, the amount of time it takes for an advective signal to reach the abyssal North Pacific may be considerably less than ventilation time scales suggest.

b. Separating wave and advective effects

Although the polynya wave and advective signals typically act on different time scales, there are still regions where both mechanisms act contemporaneously. Signal overlap regions are a function of a region's distance from the Weddell Sea and the length (29 years) of the polynya

signal. When both wave and advective signals are present at a particular location, their relative magnitude and difference in phase with respect to each other ultimately determine the sign of the property changes. The <29 -yr difference between the tracer and wave arrival times at 4000 m (Table 1) is suggestive of such an overlap occurring in those locations.

In most of the deep ocean, large-scale waves appear to account for more than half of the total polynya cooling, but the magnitude of their dominance varies regionally. Throughout the Atlantic Ocean, waves appear to drive almost all of the cooling, with advection accounting for $<15\%$ of the temperature decrease in most regions. In the Pacific and Indian Oceans, waves still account for more than half of the cooling (except at the northern boundaries of the signal), but advection often accounts for a larger percentage ($\sim 20\%$ – 40%) of the cooling than in the Atlantic.

The more strongly wave-driven cooling in the Atlantic arises from a combination of factors. In the South Atlantic, it is likely due to dynamical changes in the ACC brought about by polynya-induced meridional density gradient modification. In the North Atlantic, the polynya advective signal (Figs. 4, 12b) is too weak to explain the relatively strong total minimum cooling

FIG. 11. As in Fig. 9, but for the Pacific western boundary. Red stars occur at 5000-km intervals and correspond to the y-axis distances in the right panel.

(Fig. 12a), so polynya signals in this region are most likely driven by waves.

At the northern edges of the signal in the Pacific and Indian Oceans, the ratio approaches, and occasionally exceeds, unity. This suggests that the signal in these locations is either entirely advective or that the advective signal is larger than the total cooling, which is unphysical. However, at these northern boundaries, the total cooling signal is typically very small ($\sim 0.002^{\circ}\text{C}$), and although the composite consists of many polynyas, these far-field signals are probably spurious. Therefore, the ratios in these northern regions may not reliably estimate the relative wave and advective contributions to the polynya-induced temperature changes. In addition, when we compare the

advective cooling to the total cooling, we assume that both wave and advective cooling occurs at exactly the same time, but it is unlikely that this is the case. It is possible that in some locations, wave and advective signals of opposite sign can lead to significant cancellation, ultimately reducing the magnitude of the total cooling in Fig. 12a. Because of this, the relative proportion of each signal to the total may be inflated or deflated depending on the amount of cancellation.

c. North Atlantic low-frequency variability

Because of intrinsic, low-frequency variability in the model North Atlantic, it is unlikely that the forced polynyas cause all of the temperature changes in this

TABLE 2. Potential temperature anomaly and tracer transport speeds (m s^{-1}) and their ratio at 3900 m along the western boundaries of the Atlantic, Pacific, and Indian Oceans. Negative signs indicate southward flow. The distances along the boundaries over which the speeds are calculated are noted parenthetically. Refer to Figs. 9–11 for locations.

Region	Tracer mean speed (m s^{-1})	θ Anomaly mean speed (m s^{-1})	Ratio (θ :tracer)
South Atlantic (0–7000 km)	0.008	0.022	2.75
North Atlantic (10 500–22 000 km)	−0.006	−0.015	2.50
Indian (0–6000 km)	0.005	0.008	1.60
Pacific (0–12 000 km)	0.010	0.011	1.10

FIG. 12. Composite 3900 m (a) total minimum cooling ($^{\circ}\text{C}$), (b) constructed advective cooling ($^{\circ}\text{C}$), and (c) the ratio of (b) to (a). The total minimum cooling is calculated by taking the composite minimum temperature anomaly at each point over the 29-yr polynya period. The advective cooling field is constructed by multiplying the average cooling in the polynya region (-0.2°C) by the composite maximum tracer field. Refer to [section 2f](#) for further details. Ratios above 0.5 indicate that advection likely causes the majority of the cooling, while ratios below 0.5 indicate that large-scale waves likely cause the majority of the cooling. Black boxes mark the polynya region.

region. Small areas of statistically significant temperature anomalies occur at multiple depths (Figs. 9, 13a), but the majority of the anomalies are not significant. None of the 3900-m North Atlantic temperature

anomalies are statistically different from the mean at 90% confidence.

However, given the strong forcing imposed in the Weddell Sea, it is plausible that, in a mechanism akin

FIG. 13. Hovmöller diagrams of composite potential temperature anomalies ($^{\circ}\text{C}$) showing planetary Rossby waves in (a) the Atlantic at 860 m and (b) the Pacific at 530 m. Anomalies are averaged over (a) $13^{\circ}\text{--}15^{\circ}\text{N}$, and (b) $8^{\circ}\text{--}10^{\circ}\text{S}$. Hatching denotes anomalies statistically different from the mean at the 90% significance level.

to the bipolar ocean seesaw (e.g., [Broecker 1998](#); [Swingedouw et al. 2009](#)), the increased dense-water formation in the polynyas, and the subsequent decrease in dense-water formation after their cessation, could lead to changes in the North Atlantic via AMOC adjustment. Anomalies in the North Atlantic could then propagate south along the DWBC, much like the southward temperature anomaly propagation in the deep North Atlantic that occurs in our model ([Fig. 9](#)). Composite bandpassed AMOC anomalies at 26°N in FP (not shown) are ~ 0.1 Sv larger relative to CTL. This suggests that the forced polynyas do impact the AMOC, but these changes are small compared to the 21.8-Sv mean AMOC in FP and fall within the simulation's natural AMOC variability range. The AMOC could also be impacted via atmospheric teleconnections, but preliminary examination of the forced polynya's atmospheric impacts does not suggest that this is the case.

In their modeling studies, [Martin et al. \(2013, 2015\)](#) found that convective events in the Weddell Sea impact the AMOC and drive changes in the North Atlantic on multicentennial time scales. [Patara and Böning \(2014\)](#) found that AABW warming, and the subsequent reduction in its northward extent, lead to modest increases in the strength of the upper limb of the AMOC in their model. Furthermore, [Johnson et al. \(2008\)](#) presented observational evidence indicating that, via Kelvin and Rossby waves, AABW anomalies can propagate to the North Atlantic in a few decades and perturb the water

mass properties there. While it is possible that similar behavior occurs in our model, an in-depth analysis of the potential connection between the forced polynyas and AMOC variations is beyond the scope of our study.

5. Conclusions

We explored open-ocean polynya signal transport mechanisms in a coupled climate model by forcing large polynyas in the Weddell Sea every 29 years. We determined that large-scale waves and advection transport the polynya signal and that waves dominate this transport. The polynya advective signal acts on decadal-to-centennial time scales and can reach remote locations such as the abyssal equatorial Pacific and North Atlantic in 50–100 years. Below 4000 m, the abyssal circulation controls the polynya signal advective pathways. Large-scale waves transport the polynya signal much faster than advection. Long, nondispersive waves such as Kelvin and topographic and planetary Rossby waves propagate the polynya signal on interannual-to-decadal time scales. Polynya wave signals can reach locations in the South Atlantic, southwest Pacific, and south Indian Oceans in 15 years or less.

Both waves and advection cause temperature anomalies on isobaths. In the deep Atlantic, waves are responsible for the majority of the temperature anomalies. This is also true in the deep Indian and Pacific Oceans, although to a lesser extent. In these oceans, advection may also be responsible for a significant fraction of the

total polynya-induced temperature variations. In the deep South Pacific, it is difficult to distinguish between the wave and advective components of the signal because advective signals also travel rapidly ($\sim 10^5$ km in ~ 30 yr or about 1 cm s^{-1}) along the DWBC pathway. This result indicates that, at least for polynya-induced property changes, advection in the deep southwest Pacific occurs on time scales similar to that of the waves.

Although we have studied polynya signal spreading in the deep and abyssal oceans, the wave and advective transport mechanisms discussed here are also relevant to other climatic signals that originate in Antarctic marginal seas. Modeling studies indicate that bottom water anomalies arising from natural and forced Weddell Gyre variability reach the interior Atlantic via the same mechanisms (Santoso and England 2008; Zhang and Delworth 2016), and recently observed AABW warming and freshening is likely due in part to the propagation of wave signals from Antarctica (Purkey and Johnson 2010, 2013). In our study, the combination of fast wave signals and slower advective signals made it difficult to distinguish one mechanism from the other in many regions. Despite this, it is clear that both mechanisms are integral to polynya signal transport in the deep and abyssal oceans. It is important that we fully understand the time scales on which these mechanisms act and to what extent the superposition of their signals—be they polynya generated or otherwise—determine trends in deep and abyssal ocean properties.

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