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Effect of freshwater from the West Greenland Current on the winter deep convection in the Labrador Sea

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ABSTRACT

The effect of mesoscale eddies on the deep convection in the Labrador Sea is examined by using a realistically configured eddy-resolving ice-ocean model. The near-surface boundary current flowing into the Labrador Sea is realistically simulated, namely the West Greenland Current which carries upper/onshore fresh and lower/offshore warm water, and eddies separating from these boundary currents with cold/fresh water atop warm/salty water are also well reproduced. The modeled convection is confined to the southwestern Labrador Sea as observed, and its depth and width are reproduced better than in previous modeling studies. Although previous modeling studies demonstrated only the importance of eddy-induced heat transport in inhibition of deep convection over the central to northern Labrador Sea, our study found that the eddy-induced transport of near-surface fresh water also significantly contributes.

1. Introduction

The Labrador Sea is one of the most prominent regions in the World Ocean where open ocean convection takes place (e.g., Clarke and Gascard, 1983; Marshall and Schott, 1999). The open ocean convection reaches 1000–2300 m depth (Lazier et al., 2002) and forms Labrador Sea Water (LSW). Many studies have suggested that the LSW influences the global climate, as it is one of the source waters for the North Atlantic Deep Water (NADW) and thus affects the Atlantic meridional overturning circulation (Weaver et al., 1999; Bryden and Imawaki, 2001). The winter-time deep convection is preconditioned by weak stratification in the Labrador Sea interior associated with the cyclonic circulation which accompanies a system of strong boundary currents made up of the Labrador Current (LC) and the West Greenland Current (WGC). The smaller scale (~100–200 km) localized cyclonic circulation, which exposes water to wintertime heat loss for a longer period of time, also preconditions the deep convection in the western Labrador Sea (Lavender et al., 2002).

The large-scale cyclonic boundary current system (WGC and LC) transports buoyancy into the Labrador Sea (Cuny et al., 2002). The Irminger Water (IW), which originates in the Irminger Sea, is relatively warm and salty and transported by WGC and LC in the subsurface along the continental slope. On the other hand, cold/fresh water, which is originated in the Greenland and Arctic Seas, is transported by WGC along the vicinity of the Greenland coast in the surface layer. We use the "IW" instead of "Irminger Current", which has been used in many previous studies, since there is dynamically one current (WGC) transporting two water masses (Fratantoni and Pickart, 2007). This layered structure is also seen in the central Labrador Sea and formed by the eddy-induced lateral transport from the shelf region (Lilly et al., 1999). High activity of mesoscale eddies is observed by surface drifters (Cuny et al., 2002), floats and satellites (Prater, 2002), and moorings (Lilly et al., 2003) in the Labrador Sea. Several observational studies have suggested the important role of such mesoscale eddies in the inhibition of winter deep convection and restratification of vertically homogenized water columns in the Labrador Sea interior by transporting buoyancy from the WGC (Lilly et al., 1999; Prater, 2002). Pickart et al. (2002) suggested that the eddy flux of buoyant water confines deep convection to the southwestern Labrador Sea. Coarse resolution models tend to simulate too wide or too deep convection in the Labrador Sea possibly because of the lack of such eddy-induced buoyancy transport (e.g., England, 1993; Komuro and Hasumi, 2005).

Chanut et al. (2008) classified the eddies into convective eddies (CEs), boundary current eddies (BCEs), and Irminger Rings (IRs). The CEs are generated by baroclinic instability of the rim current encompassing the convection area and more rapidly (less than two months) restratify the homogenized water columns than air-sea flux (~half year) does in the shallow layer (Jones and Marshall, 1997).

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The size of CEs is of the order of the Rossby deformation radius for the first baroclinic mode (∼10 km). The BCEs have a comparable size to CEs and are formed by baroclinic instability of the boundary currents (WGC and LC) throughout a year. The IRs are generated by instability of the WGC at the Cape Desolation (Katsman et al., 2004). There has not been an established agreement among several studies on whether the IRs are generated by barotropic, baroclinic, or mixed instability (Eden and Boning, 2002; Bracco and Pedlosky, 2003). The IRs are larger (diameter is ∼40–50 km) and more long-lived (several months to one year) than the BCEs and CEs (living time is less than one month). While the source of the IRs, which is detected by the maximum of sea surface height variability (Lilly et al., 2003), is confined to the northeastern Labrador Sea, their long lifetimes cause the significant buoyancy transport from the boundary current region to the Labrador Sea interior. The IRs contain cold/fresh water in the upper part and warm/saline core at depth and transport these properties to the Labrador Sea interior (Lilly and Rhines, 2002; Lilly et al., 2003; Hatun et al., 2007). Chanut et al. (2008) suggested that the buoyancy transported by the BCEs is also significant in the Labrador Sea, since the number of the BCEs is much larger than that of the IRs. On the other hand, Gelderloos et al. (2011) claimed that the BCEs are not essential for buoyancy transport toward the interior Labrador Sea.

High resolution modeling is necessary to examine the effect of eddies on the deep convection in the Labrador Sea, and such modeling became feasible only recently. Tréguier et al. (2005) reported the inter-comparison of four high-resolution models in the North Atlantic subpolar gyre (horizontal resolution is 1/6° to 1/12°). They pointed out that the pattern of deep convection in the Labrador Sea depends on the transport of freshwater and heat by the boundary current (e.g., WGC and LC). The deep convection unrealistically developed (depth of convection is ∼3000–3500 m) in three models. Although the depth of convection is about 1300 m in the remaining model, the location of convection is the eastern Labrador Sea, which is unrealistic. They argued that salinity drifts away from the observed state because of salinization of the boundary current.

Chanut et al. (2008) investigated the effect of eddies on the open ocean convection and restratification in the Labrador Sea by using an eddy-resolving (horizontal resolution is 1/15°) ocean model. The water column homogenized by deep convection is restratified by the CEs in shallow layer during the early phase, and the BCEs and IRs transport the warm water from the WGC region to the Labrador Sea interior and restratify the water columns at depth during the later season in their model. These roles of eddies had previously been suggested by many studies (e.g., Jones and Marshall, 1997; Prater, 2002; Lilly et al., 2003). However, Chanut et al. (2008) demonstrated only the effect of the heat from the IW, since they failed to represent the cold/low-salinity surface water in the WGC perhaps due to coarse horizontal resolution in the data used for surface restoring of salinity. The observed trajectory of eddies originating from the WGC shows that these eddies move to the interior Labrador Sea with cold/fresh water in the upper layer (e.g., Pickart et al., 2002; Lilly et al., 2003), and their importance in inhibition of deep convection and restratification of vertically homogenized water columns has also been pointed out (Schmidt and Send, 2007).

McGeehan and Maslowski (2011) quantitatively assessed the shelf-basin freshwater transport in the western Labrador Sea by using an eddy-resolving model (horizontal resolution is ∼9 km). Unfortunately, the deepest convection takes place in the northeastern Labrador Sea in their model. They discussed that the discrepancy from the observed feature of deep convection is due to the absence of the IRs having low salinity water on their top. They argued that the lack of the surface low salinity is caused by the error in the surface salinity distribution used for nudging, shortage of runoff from Greenland, and insufficient resolution to reproduce the narrow coastal current along the Greenland coast in their model.

Thus, the importance of the eddy-induced freshwater transport for the restratification in the region of open ocean deep convection in the Labrador Sea has not been fully clarified from a quantitative point of view. We quantitatively investigate the roles of the eddies in the deep convection in the Labrador Sea through their buoyancy (heat and freshwater) transport by using an eddy-resolving ocean general circulation model.

The numerical model and experimental design are described in the next section. The results of experiments are shown in Section 3. Section 4 presents the discussion and conclusion.

2. Model description and experimental design

The ice-ocean general circulation model employed in this study is COCO version 4 (Hasumi, 2006). The model incorporates a second-order moments conserving scheme for tracer advection (Prather, 1986). The turbulence closure scheme of Noh and Kim (1999) is applied for diagnosing vertical viscosity and diffusivity. Background vertical viscosity and diffusivity are 1.0×10^{-4} and 2.0×10^{-5} m^2 s^{-1}, respectively. The horizontal bi-harmonic friction with Smagorinsky-like viscosity is utilized, where the value of controlling parameter is set to 3 (Griffies and Hallberg, 2000).

The model domain is global. The model is formulated on the general curvilinear horizontal coordinates. Its poles are placed on the Labrador Peninsula and Greenland, and the horizontal resolution is eddy resolving (4–5 km) in the Labrador Sea, eddy permitting around the Cape Hatteras, Irminger and Greenland Seas (10–50 km), and coarse (>50 km) in other regions (Fig. 1(a) and (b)). The bathymetry is constructed from a 2-min topography dataset (ETOPO2, National Geophysical Data Center, National Oceanic and Atmospheric Administration). There are 45 vertical levels. The layer thickness varies from 5 m (top) to 500 m (bottom).

The model is initiated by climatological temperature and salinity (World Ocean Atlas 2009, Locarnini et al., 2010; Antonov et al., 2010) with no flow and sea ice. Temperature and salinity are restored to observed monthly climatology (World Ocean Atlas 2009, Locarnini et al., 2010; Antonov et al., 2010) with a damping time scale of 30 days at all depths over the whole domain during the first six months (January to June of the first year) to reproduce the mean circulation in the Labrador Sea. The sea surface heat, freshwater, and momentum fluxes are calculated using a daily climatology of surface air properties based on the ECMWF reanalysis (OMIP-forcing version 3; Roske, 2005). Restoring of sea surface salinity is not employed, since it may cause failure in reproducing the low salinity water in the WGC as in previous studies (Rattan et al., 2010; McGeehan and Maslowski, 2011). To avoid the drift of model, we employ nudging of temperature and salinity to monthly climatology in the northeastern Atlantic Ocean (shown as a box in Fig. 1(b)), which is the source of warm/saline water of the WGC. Note that if we apply temperature and salinity nudging around the Denmark Strait for the cold/fresh water from the Greenland Sea and the Arctic Ocean, the low salinity water (the narrow East Greenland Current) is not reproduced because of coarseness of climatology data as pointed out by Rattan et al. (2010).

The 20-year time series of temperature, salinity and mean kinetic energy averaged in the Labrador Sea show that the state of ocean is in a quasi-steady state except for the first few years (Fig. 2). This means that the boundary conditions and modeled physics in the Labrador Sea are balanced with the climatological initial state. So, it is meaningful to analyze the modeling results for this period. On the other hand, when we run the model without nudging of temperature/salinity, the modeled state continuously