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

T. Kawasaki^{a,b,*}, H. Hasumi^b

^a*National Institute of Polar Research, 10-3, Midoricho, Tachikawa, Tokyo, 190-8518, Japan*

^b*Atmosphere and Ocean Research Institute, the University of Tokyo, 5-1-5, Kashiwanoha, Kashiwa-shi, Chiba, 277-8568, Japan*

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.

Keywords: freshwater transport, Labrador Sea, winter deep convection,

*Corresponding author, Tel.: +81-4-7136-4417; Fax: +81-4-7136-4375
Email address: kawasaki@aori.u-tokyo.ac.jp (T. Kawasaki)

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

24 Current”, which has been used in many previous studies, since there is dy-
 25 namically one current (WGC) transporting two water masses (Fratantoni and
 26 Pickart, 2007). This layered structure is also seen in the central Labrador Sea
 27 and formed by the eddy-induced lateral transport from the shelf region (Lilly
 28 et al., 1999). High activity of mesoscale eddies is observed by surface drifters
 29 (Cuny et al., 2002), floats and satellites (Prater, 2002), and moorings (Lilly
 30 et al., 2003) in the Labrador Sea. Several observational studies have sug-
 31 gested the important role of such mesoscale eddies in the inhibition of winter
 32 deep convection and restratification of vertically homogenized water columns
 33 in the Labrador Sea interior by transporting buoyancy from the WGC (Lilly
 34 et al., 1999; Prater, 2002). Pickart et al. (2002) suggested that the eddy flux
 35 of buoyant water confines deep convection to the southwestern Labrador Sea.
 36 Coarse resolution models tend to simulate too wide or too deep convection in
 37 the Labrador Sea possibly because of the lack of such eddy-induced buoyancy
 38 transport (e.g., England, 1993; Komuro and Hasumi, 2005).

39 Chanut et al. (2008) classified the eddies into convective eddies (CEs),
 40 boundary current eddies (BCEs), and Irminger Rings (IRs). The CEs are
 41 generated by baroclinic instability of the rim current encompassing the con-
 42 vection area and more rapidly (less than two months) restratify the homoge-
 43 nized water columns than air-sea flux ($\sim$ half year) does in the shallow layer
 44 (Jones and Marshall, 1997).

45 The size of CEs is of the order of the Rossby deformation radius for
 46 the first baroclinic mode (~ 10 km). The BCEs have a comparable size
 47 to CEs and are formed by baroclinic instability of the boundary currents
 48 (WGC and LC) throughout a year. The IRs are generated by instability

49 of the WGC at the Cape Desolation (Katsman et al., 2004). There has
 50 not been an established agreement among several studies on whether the
 51 IRs are generated by barotropic, baroclinic, or mixed instability (Eden and
 52 Boning, 2002; Bracco and Pedlosky, 2003). The IRs are larger (diameter
 53 is ~ 40 -50 km) and more long-lived (several months to one year) than the
 54 BCEs and CEs (living time is less than one month). While the source of
 55 the IRs, which is detected by the maximum of sea surface height variability
 56 (Lilly et al., 2003), is confined to the northeastern Labrador Sea, their long
 57 lifetimes cause the significant buoyancy transport from the boundary current
 58 region to the Labrador Sea interior. The IRs contain cold/fresh water in the
 59 upper part and warm/saline core at depth and transport these properties to
 60 the Labrador Sea interior (Lilly and Rhines, 2002; Lilly et al., 2003; Hatun
 61 et al., 2007). Chanut et al. (2008) suggested that the buoyancy transported
 62 by the BCEs is also significant in the Labrador Sea, since the number of the
 63 BCEs is much larger than that of the IRs. On the other hand, Gelderloos
 64 et al. (2011) claimed that the BCEs are not essential for buoyancy transport
 65 toward the interior Labrador Sea.

66 High resolution modeling is necessary to examine the effect of eddies on
 67 the deep convection in the Labrador Sea, and such modeling became feasible
 68 only recently. Tréguier et al. (2005) reported the inter-comparison of four
 69 high-resolution models in the North Atlantic subpolar gyre (horizontal reso-
 70 lution is $1/6$ to $1/12$ °). They pointed out that the pattern of deep convection
 71 in the Labrador Sea depends on the transport of freshwater and heat by the
 72 boundary current (e.g., WGC and LC). The deep convection unrealistically
 73 developed (depth of convection is ~ 3000 -3500 m) in three models. Although

74 the depth of convection is about 1300 m in the remaining model, the location
75 of convection is the eastern Labrador Sea, which is unrealistic. They argued
76 that salinity drifts away from the observed state because of salinization of
77 the boundary current.

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

95 McGeehan and Maslowski (2011) quantitatively assessed the shelf-basin
96 freshwater transport in the western Labrador Sea by using an eddy-resolving
97 model (horizontal resolution is ~ 9 km). Unfortunately, the deepest con-
98 vection 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 $2.0 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$, 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).

123 The model domain is global. The model is formulated on the general
 124 curvilinear horizontal coordinates. Its poles are placed on the Labrador
 125 Peninsula and Greenland, and the horizontal resolution is eddy resolving
 126 (4-5 km) in the Labrador Sea, eddy permitting around the Cape Hatteras,
 127 Irminger and Greenland Seas (10-50 km), and coarse (> 50 km) in other
 128 regions (Figure 1a, b). The bathymetry is constructed from a 2-min to-
 129 pography dataset (ETOPO2, National Geophysical Data Center, National
 130 Oceanic and Atmospheric Administration). There are 45 vertical levels. The
 131 layer thickness varies from 5 m (top) to 500 m (bottom).

132 The model is initiated by climatological temperature and salinity (World
 133 Ocean Atlas 2009, Locarnini et al., 2010; Antonov et al., 2010) with no flow
 134 and sea ice. Temperature and salinity are restored to observed monthly
 135 climatology (World Ocean Atlas 2009, Locarnini et al., 2010; Antonov et al.,
 136 2010) with a damping time scale of 30 days at all depths over the whole
 137 domain during the first six months (January to June of the first year) to
 138 reproduce the mean circulation in the Labrador Sea. The sea surface heat,
 139 freshwater, and momentum fluxes are calculated using a daily climatology
 140 of surface air properties based on the ECMWF reanalysis (OMIP-forcing
 141 version 3; Roske, 2005). Restoring of sea surface salinity is not employed,
 142 since it may cause failure in reproducing the low salinity water in the WGC
 143 as in previous studies (Rattan et al., 2010; McGeehan and Maslowski, 2011).
 144 To avoid the drift of model, we employ nudging of temperature and salinity
 145 to monthly climatology in the northeastern Atlantic Ocean (shown as a box
 146 in Figure 1b), which is the source of warm/saline water of the WGC. Note
 147 that if we apply temperature and salinity nudging around the Denmark Strait

148 for the cold/fresh water from the Greenland Sea and the Arctic Ocean, the
149 low salinity water (the narrow East Greenland Current) is not reproduced
150 because of coarseness of climatology data as pointed out by Rattan et al.
151 (2010).

152 The 20-year time series of temperature, salinity and mean kinetic energy
153 averaged in the Labrador Sea show that the state of ocean is in a quasi-steady
154 state except for the first few years (Figure 2). This means that the boundary
155 conditions and modeled physics in the Labrador Sea are balanced with the
156 climatological initial state. So, it is meaningful to analyze the modeling
157 results for this period. On the other hand, when we run the model without
158 nudging of temperature/salinity, the modeled state continuously drifts away
159 from the initial state afterwards (dashed lines in Figure 2). This drift is
160 caused by biases in temperature and salinity in the upstream region (i.e.,
161 the North Atlantic Ocean) of the boundary current which comes into the
162 Labrador Sea, because the resolution is coarse and restoring for sea surface
163 salinity is not applied there. Although we succeeded in avoiding such a drastic
164 drift for basin-averaged kinetic energy, temperature, and salinity (Figure 2),
165 there still are drifts in localized features. For instance, gradual deepening of
166 winter convection in the interior Labrador Sea is induced after the sixth year
167 in our model. As discussed later, the eddy activity is still underestimated
168 due perhaps to insufficient resolution or lack of a proper parameterization,
169 and this causes the drift of deep convection. It is difficult to improve the
170 model further and remove this drift without nudging of sea surface salinity.
171 For the purpose of this study, therefore, it is desirable to analyze an early
172 period when the oceanic state is more realistic than in the later period. Thus,

173 the results presented hereafter are taken from the sixth and tenth years of
174 model integration.

175 To investigate the role of eddies, another case is conducted where mesoscale
176 eddies are artificially suppressed. As it is known that the IRs are generated
177 through instability of the WGC due to the sharp continental slope off the
178 Cape Desolation (Katsman et al., 2004; Bracco et al., 2008), a more gentle
179 continental slope is applied off the Cape Desolation in this case by smoothing
180 the bathymetry. This is hereafter referred to as the SMOOTH case (Figure
181 1d and e). Such bathymetry smoothing is also applied by a previous study
182 (Chanut et al., 2008). When the name of the case is not described explicitly,
183 the presented result is for the control case.

184 **3. Results**

185 *3.1. Mean state*

186 The cyclonic circulation including the WGC and LC, which are part of
187 the North Atlantic subpolar gyre, is qualitatively well reproduced in the
188 Labrador Sea (Figure 3a). The 5-year mean volume transports of the WGC
189 and LC are about 25 and 30 Sv, respectively, which are calculated by the local
190 minima of the barotropic stream function. The simulated mean WGC speed
191 is 35-40 cm s⁻¹ in the shallow layer, which is consistent with an estimate
192 from drifter data (35 cm s⁻¹; Cuny et al., 2002). The Labrador Current
193 speed around the Hamilton Bank is 25-35 cm s⁻¹ and in agreement with the
194 observed values of 20 cm s⁻¹ to the north and 30 cm s⁻¹ to the south of the
195 Hamilton Bank (Cuny et al., 2002). The weak counter currents adjacent to
196 the cyclonic boundary currents, which has been observed by Lavender et al.

Figure 1: (a) Geographical location of the grid points of the model. (b) The horizontal resolution of the model around the Labrador Sea (km). The rectangle indicates the domain of temperature and salinity nudging to climatology. The ocean depth around the northeastern Labrador Sea in (c) the control case and (d) the SMOOTH case. The contour interval is 500 m. (e) The cross section of the model bathymetry at right angle to the coast for control case (black) and SMOOTH case (gray). The location of this section is shown as a solid line in (c) and (d). The x and y labels indicate the distance from the coast of Greenland (km) and depth (m), respectively

Figure 2: 1-year running means of (a) potential temperature ($^{\circ}\text{C}$), (b) salinity (g kg^{-1}), and (c) mean kinetic energy ($10^{-4} \text{ m}^2 \text{ s}^{-2}$) averaged in the Labrador Sea for the integration period. The blue, red, and purple lines indicate the average in the boundary current region, the interior, and both of them, respectively, as shown in the upper panel. The dashed purple line indicates the average in whole of the Labrador Sea (boundary + interior regions) without temperature/salinity nudging in the upstream of the WGC (which is shown as a box in Figure 1 b).

Figure 3: 5-year mean horizontal velocity at (a) the sea surface and (b) 715 m depth. The color shade and vector indicate the scalar and direction of velocity, respectively. The vectors are plotted at every 0.5 degree in longitude and latitude.

197 (2000) and Fischer et al. (2010) and preconditions the deep convection in
 198 western Labrador Sea (Lavender et al., 2002) and enhances the export of
 199 LSW to the Irminger Sea (Myers and Deacu, 2004), are also reproduced at
 200 intermediate depths (Figure 3b).

201 Potential density, potential temperature, and salinity along the AR7W
 202 line in March (winter) and June (summer) in our model are shown in Fig-
 203 ure 4. Many observations have been conducted along the AR7W line, which
 204 is a hydrographic section across the Labrador Sea connecting Misery Point
 205 (Labrador Peninsula) and Cape Desolation (Greenland). The low density
 206 water along the both boundaries is transported from the Irminger and Green-
 207 land Seas by the cyclonic boundary currents (Figures 4). The low density

208 water is composed of the subsurface warm/salty water along the continental
 209 slope (IW) and the near-surface (above 150 m depth) cold/fresh water on
 210 the shelf in the WGC. These features of the WGC are consistent with ob-
 211 servations (Figure 5). The maximum of temperature and salinity of the IW
 212 along the AR7W in May are 5.5 °C and 35.01 g kg⁻¹, respectively, which
 213 are slightly warmer and saltier, but similar to the observed values of Cuny
 214 et al. (2002) (4.2-5.4 °C / 34.88-34.99 g kg⁻¹). The temperature and salinity
 215 of cold/fresh water near the surface is -1.0-0.0 °C and 31.5-33.0 g kg⁻¹, re-
 216 spectively, and these values are in the range of a high quality climatological
 217 data-set (Kulan and Myers, 2009).

218 Sea surface cooling vertically homogenizes density (potential temperature
 219 and salinity) on the Labrador side of the Labrador Sea interior in winter
 220 (Figures 4b, d). The location and width of the homogenized water column
 221 along AR7W are consistent with the observation in March 1997 (Figure 8
 222 in Pickart et al., 2002). The sea surface heating and eddy-induced lateral
 223 buoyancy transport from the WGC, as described later, stratify the Labrador
 224 Sea interior in summer (Figures 4a, c). The simulated salinity is vertically
 225 homogeneous even in June in the convection region as shown by the isolines
 226 of 34.89 and 34.91 g kg⁻¹ in salinity (Figure 4c). This qualitative feature of
 227 vertical uniformity of salinity is consistent with observations (Pickart et al.,
 228 2002; Yashayaev, 2007)

229 The potential density of LSW ranges between 27.8 and 27.83 kg m⁻³ in
 230 our model (Figure 4). The modeled LSW is denser than observed (27.77-
 231 27.80 in mid 1990s and 27.68-27.77 in 2000s; Figure 5; Yashayaev, 2007).
 232 Although the potential temperature of LSW is in the range of observed in-

233 terannual variability (Figure 4a and 5a, b), salinity ($\sim 34.9 \text{ g kg}^{-1}$; Figure 4c)
234 is larger than observed ($34.83\text{-}34.86 \text{ g kg}^{-1}$; Figure 5c, d) by $\sim 0.05 \text{ g kg}^{-1}$.
235 Since the salinity in Labrador Sea interior is vertically more uniform than
236 observed (Figure 4c, d), we understand that the vertical mixing induced by
237 deep convection is too strong in our model. In fact, the modeled winter time
238 convection is wider/deeper than observed, and this discrepancy is discussed
239 later (in Section 3.3).

240 3.2. Eddies

241 The three types of eddies (IR, BCE, CE) proposed by Chanut et al.
242 (2008) are well represented in our model. The IRs (their diameter is 40-50
243 km) are detected by cores of low relative vorticity, sea surface temperature
244 and sea surface salinity (Figures 6a-c). Such cold cores at the sea surface
245 are observed by high resolution satellite remote-sensing data (e.g., ATSR
246 and MODIS). These cores also have relatively high temperature and high
247 salinity in subsurface (Figure 6d). Although this layered structure of IRs is
248 revealed by many previous observational studies (e.g., Hatun et al., 2007;
249 Rykova et al., 2009), previous simulation studies by eddy-resolving models
250 (Chanut et al., 2008; McGeehan and Maslowski, 2011) represented only the
251 high temperature and high salinity feature and did not reproduce the low
252 temperature and low salinity near sea surface. Their time evolution in our
253 model shows that the IRs are first transported by the cyclonic mean flow
254 along isobaths and later propagate to the basin interior. The IRs are long-
255 lived (their typical lifetime is several months) and wander in the Labrador
256 Sea interior in our model. Such a feature of the trajectory of IRs is consistent
257 with observations (Prater, 2002; Cuny et al., 2002; Hatun et al., 2007). The

Figure 4: 5-year mean potential temperature (top; °C), and salinity (bottom; g kg⁻¹) along AR7W. Contours indicate isopycnal surfaces in potential density of 27.68, 27.74, 27.77, 27.80, 27.83, 27.86 kg m⁻³. The colors and contour values are so chosen as to coincide with Figure 5. The left and right panels are for June and March, respectively. The left/right ends are the Labrador Peninsula/Greenland. The horizontal labels indicate the distance from the coast of the Labrador Peninsula (km).

Figure 5: Observed potential temperature (top; °C) and salinity (bottom; g kg⁻¹) in 1994 (left) and 2005 (right) along AR7W (from Yashayaev, 2007).

258 moment of IRs carries the contained buoyancy (warm and fresh water) from
259 the WGC to the interior Labrador Sea.

260 The BCEs, whose diameter is smaller than that of the IRs (Chanut et al.,
261 2008), are found as the extremal relative vorticity in the vicinity of the whole
262 of the boundary currents (Figures 6a, e). Their time evolution indicates that
263 the BCEs also contain the low density boundary current water and transport
264 buoyancy from the region of the boundary currents to the interior Labrador
265 Sea. Although the number of BCEs is larger than that of IRs, the lifetime
266 of BCEs is shorter than that of IRs.

267 The CE (diameter ~ 10 km) are formed from winter to early spring in the
268 Labrador Sea interior and are recognized as frequent small cores of extremal
269 relative vorticity (Figure 6e). It is well known that the CEs are the result of
270 the convection induced by the sea surface cooling and directly and quickly
271 restratify the shallow layer (Jones and Marshall, 1997). As the horizontal grid
272 size of our model is not sufficiently small, eddies smaller than the mesoscale
273 (known as sub-mesoscale eddies) are not resolved. It is likely that mesoscale
274 eddies develop around the deep convection region instead of sub-mesoscale
275 eddies. The resolved mesoscale eddies seem to contribute to the rapid tilting
276 of the steeply sloping isopycnal surfaces in the deep convection area in our
277 model.

278 The (monthly mean) eddy kinetic energy (EKE) is defined as

$$\text{EKE} = \frac{1}{2} \left(\overline{u'^2} + \overline{v'^2} \right),$$

279 where overbar denotes the monthly mean, and (u', v') indicate the deviations
280 of horizontal velocities from the monthly mean velocities $(\bar{u}, \bar{v})$. The hor-
281 izontal map of annual-mean sea surface EKE is shown in Figure 7a. The

Figure 6: One-day mean of sea surface (a) ¹⁸relative vorticity, (b) temperature, (c) salinity, and (d) potential temperature at 635 m depth on 10th July of the eighth year. The units are 10^{-5} s^{-1} , $^{\circ}\text{C}$, and g kg^{-1} . The black circles in the interior Labrador Sea mark the IRs. The bottom two panels are the same as the top two but on 24th March of the eighth year. Enlarged views for the black boxes are also shown at the upper-right corners in the bottom two panels.

282 largest EKE is located at the northeastern region and extends toward the
 283 interior Labrador Sea. This tongue-like large EKE is also shown by the sea
 284 surface variability obtained from TOPEX/Poseidon altimeter (Prater, 2002;
 285 Lilly et al., 2003) and previous modeling studies (e.g., Eden and Boning,
 286 2002; Katsman et al., 2004; Chanut et al., 2008; Luo et al., 2011). As the IRs
 287 propagate and the CEs are also formed, the EKE in the interior Labrador Sea
 288 is largest from winter to spring (Figure not shown), and this seasonal vari-
 289 ability is consistent with the satellite observations (Lilly et al., 2003; Brandt
 290 et al., 2004). Note that the large EKE off Baffin Island and along the coast
 291 of Labrador Peninsula is caused by the sea-ice motion, which induces rapid
 292 changes of horizontal velocities near the sea surface.

293 The tongue of large EKE does not extend to the interior of the Labrador
 294 Sea in the SMOOTH case (Figure 7b). As we expected, the IRs cannot
 295 develop significantly and do not propagate toward the interior region when
 296 the continental slope off the Cape Desolation is smoothed. The effect of
 297 the IRs on the deep convection is described in the next section through
 298 comparison of the results of the SMOOTH and control cases.

299 *3.3. Convection*

300 Figure 7c shows the monthly mean depth of convection in March, when
 301 the convection develops most by sea surface cooling. Here, the depth of
 302 convection refers to the deepest level which convective adjustment reaches.
 303 While it is consistent with observations (Pickart et al., 2002; Lavender et al.,
 304 2002) that the deepest convection is found in the southwestern part of the
 305 Labrador Sea, both the depth and spatial extent are much larger than ob-
 306 served (Figure 7c). Previous observational studies suggested that the depth

Figure 7: (a) 5-year mean of sea surface eddy kinetic energy ($\text{m}^2 \text{s}^{-2}$). (b) Same as (a), but in the SMOOTH case. (c) Monthly mean of the depth of convection in March (m). (d) Sea surface heat flux (W m^{-2} ; positive means downward) averaged from January to March. (e) Same as (c), but in the SMOOTH case. (f) 3-month (from January to March) mean of sea ice concentration. The red line indicates the climatological sea-ice edge from satellite observations. The contour intervals are $0.02 \text{ m}^2 \text{s}^{-2}$, 500 m, 200 W m^{-2} , and 0.2 for EKE, convection depth, sea surface heat flux, and sea ice concentration, respectively.

307 of convection has large interannual variability (e.g., 2300 m in 1994-1995 and
 308 1000-1300 m in early and mid-2000s Avsic et al., 2006; Yashayaev, 2007).
 309 Pickart et al. (2002) argued that since the hydrographic state was pre-
 310 conditioned by severe winter conditions in early 1990s, even weak cooling
 311 caused the deep convection to ~ 1500 m depth in March 1997. The sim-
 312 ulated depth of convection (~ 2000 -2500 m) is maximum of the observed
 313 range (1000-2300m) and not representative as averaged conditions. Previous
 314 modeling studies often simulated too wide and too deep convection even in
 315 eddy-resolving cases (e.g., Chanut et al., 2008; McGeehan and Maslowski,
 316 2011). Although the convection of the present result is also too wide/deep
 317 compared with observations (Pickart et al., 2002), the present result is some-
 318 what improved compared with these studies.

319 The weak stratification in the central Labrador Sea accompanied by the
 320 wind-driven cyclonic circulation preconditions deep convection to take place
 321 in the whole of basin center. In addition, sea surface cooling in winter is
 322 strongest in the northwestern interior Labrador Sea (Figure 7d). On the
 323 other hand, the winter deep convection is furthest developed in southwestern
 324 interior Labrador Sea (Figure 7c). These facts imply that the deep convec-
 325 tion is suppressed over the central to northern and eastern interior Labrador
 326 Sea by lateral buoyancy transport from boundary current region. The role
 327 of eddy-induced buoyancy transport from the WGC on the restratification
 328 and suppression of deep convection has been suggested by the observational
 329 studies (Pickart et al., 2002; Lilly et al., 2003).

330 The convection is deeper and wider in the SMOOTH case (Figure 7e) than
 331 in the control case (Figure 7c). It reaches more than 2500 m depth, which is

332 larger than the result of control case and the above-mentioned observational
333 estimates. The winter deep convection is not sufficiently inhibited because
334 the IRs do not transport buoyancy toward the basin interior (Figure 7b).
335 The following analysis demonstrates that the suppression of deep convection
336 is caused by IRs.

337 Sea ice extends in the western and northern boundaries of the Labrador
338 Sea in winter (Figure 7f). The ice edge around the western and northern
339 Labrador Sea is similar to the climatological ice edge observed from satel-
340 lites. McGeehan and Maslowski (2011) suggested that the location of deep
341 convection is affected by the sea ice extent in winter. We conclude that the
342 well-represented sea ice extent is one of the reasons for the correct location
343 of deepest winter convection in our model.

344 Note that sea ice is also found along the southwestern coast of Green-
345 land, which is not in agreement with observation. This discrepancy does not
346 lead to a significant bias of freshwater transport from the eastern region to
347 the Labrador Sea interior. In addition, the lateral heat and freshwater flux
348 between the boundary current region and the interior ocean by sea ice is
349 significantly smaller than that by seawater in our model (result not shown).

350 3.4. *Heat/freshwater transport*

351 In order to investigate the role of eddies in the deep convection, the
352 model's buoyancy budget is analyzed here. Since the transported buoyancy
353 inhibits the deep convection in winter and restratifies the homogenized water
354 columns in summer, the year-round buoyancy transport should be analyzed.
355 Figure 8 shows the 5-year mean heat budget integrated vertically from the
356 sea surface to the bottom. Heat is lost over the whole of the Labrador Sea

357 through the sea surface (Figure 8a). This is essentially due to the sea surface
 358 cooling in winter (Figure 7d). In the boundary current region, heat is gained
 359 by the mean current (the WGC) and lost by the eddies (Figures 8b, c). On
 360 the other hand, the interior Labrador Sea gains heat by mesoscale eddies
 361 (Figure 8c). The small residual (Figure 8d) indicates that the the model is
 362 well thermally equilibrated during the analyzed period.

363 The freshwater content (FW) is defined here as

$$\text{FW} = \int_V \frac{S_{\text{ref}} - S}{S_{\text{ref}}} dV,$$

364 where V indicates the volume under consideration, S is salinity and S_{ref} is
 365 the reference salinity. The reference salinity is set to 34.8 g kg^{-1} , which is
 366 the average in the Labrador Sea (Figure 2). The budget of the vertically
 367 integrated freshwater content is shown in Figure 9. Because of the melting of
 368 sea ice in early spring, the net sea surface freshwater flux is downward around
 369 the WGC and LC (Figures 7f, 9a). Several patches of upward freshwater
 370 flux along the coast are due to the freezing at coastal polynyas. The large
 371 downward sea surface freshwater fluxes along the coasts of Greenland and the
 372 Labrador Peninsula correspond to river runoff. Since the surface fresh water
 373 flux caused by sea ice does not extend to the interior Labrador Sea, the role of
 374 freshwater transport in controlling the deep convection is not affected by the
 375 biased distribution of sea ice along the western coast of Greenland (Figure
 376 9a). The WGC transports low salinity water on the west Greenland shelf,
 377 and high salinity water (IW) from the Irminger Sea on the continental slope
 378 (Figure 9b). Mesoscale eddies transport high salinity water from the WGC
 379 toward the coast (Figure 9c). The eddy-induced transport of high or low
 380 salinity water does not extend to the interior Labrador Sea when vertically

Figure 8: Components in 5-year mean vertically integrated heat budget. (a) Sea surface flux, (b) mean advection, (c) eddy, and (d) residual. The positive and negative values mean the local heat gain and loss, respectively. The four boxes (WGC: West Greenland Current, NL: North Labrador Sea, INT: Labrador Sea interior, and LC: Labrador Current) are defined for quantitative analysis.

Figure 9: Same as in Figure 8 but for freshwater budget.

integrated over the whole depth, since the transport of fresher water in the surface layer and that of saltier water in the subsurface layer cancel each other (Figure 9c). The small residual (Figure 9d) means that freshwater content is well equilibrated during the analyzed period in the Labrador Sea.

The heat and freshwater budgets in four regions (the boxes labeled as WGC, NL, INT, and LC in Figure 8) are shown in Table 1. In the WGC (West Greenland Current) box, the WGC imports seawater from the Irminger

388 Sea. In the NL (North Labrador) box, the WGC and large EKE, which
 389 indicates active IRs, exist. The INT (interior) box corresponds to the area
 390 whose bottom is deeper than 3000 m in the central Labrador Sea. The LC
 391 (Labrador Current) box is the western boundary of the Labrador Sea, where
 392 EKE is not significantly large (less than $0.02 \text{ m}^2 \text{ s}^{-2}$) except for the noise
 393 accompanied by sea ice formation or melting near the Labrador coast.

394 Heat is injected to the WGC box by the WGC (mean advection; 374
 395 W m^{-2}) and exported by eddies (-343 W m^{-2}). In contrast, in the LC
 396 box the heat carried in by mean advection (the LC; 78 W m^{-2}) is removed
 397 not by eddies (-5 W m^{-2}) but by sea surface cooling (-74 W m^{-2}). This
 398 is consistent with the result that the eddy activity is small in the western
 399 boundary (LC box side) than in the eastern boundary (WGC box side). In
 400 the INT box, almost all heat injected by the eddies (152 W m^{-2}) is lost by
 401 winter sea surface cooling (-115 W m^{-2}). In the NL box, the warm water
 402 (IW) advected by the WGC (183 W m^{-2}) is mainly cooled at the sea surface
 403 (-154 W m^{-2}). At an observational point in the central Labrador Sea (53°W ,
 404 56°N), the annual mean surface heat loss (-100 W m^{-2}) is much larger than
 405 an estimate based on a reanalysis data (-28 W m^{-2} ; Straneo, 2006). This
 406 discrepancy is accompanied with the too wide/deep convection in the central
 407 Labrador Sea, and it is discussed later.

408 In the WGC and LC boxes, the freshwater supplied through the sea sur-
 409 face (389 and 232 cm year^{-1}) is exported by the mean advection (-434 and
 410 $-239 \text{ cm year}^{-1}$). In the WGC box, removal by eddies of high-salinity wa-
 411 ter originating from the Irminger Sea (86 cm year^{-1}) is also large. In the
 412 LC box, the eddy-induced freshwater transport (23 cm year^{-1}) is very small

Table 1: The area-averaged budgets of heat (W m^{-2}) and freshwater (cm year^{-1}) by sea surface flux, vertically integrated mean advection, eddies, and total (trend) in four boxes (WGC: West Greenland Current, NL: North Labrador, INT: Labrador Sea interior, and LC: Labrador Current) defined in Figure 8b. The values to the left and right of the slash indicate the heat and freshwater, respectively.

Heat / Freshwater	WGC	NL	INT	LC
Surface flux	-39/ 389	-154/ 111	-115/ 40	-74/ 232
Mean advection	374/-434	183/-119	-42/ 10	78/-239
Eddy advection	-343/ 86	-51/ 41	152/-61	-5/ 23
Total (Trend)	-9/ 42	-22/ 33	-5/-11	-2/ 16

413 because of the low activity of eddies, similarly to the heat transport. In the
414 NL box, the sea surface freshwater injection caused by the melting of sea ice
415 transported through the Davis Strait (111 cm year^{-1}) is exported by mean
416 advection ($-119 \text{ cm year}^{-1}$). In the INT box, precipitation supplies freshwa-
417 ter (40 cm year^{-1}) and high salinity water is supplied from the WGC box by
418 eddies (-61 cm year^{-1}).

419 The buoyancy budget over the whole water column does not necessar-
420 ily account for stabilization of the water column. Buoyancy injection near
421 the sea surface stabilizes a water column, while that at depth destabilizes
422 it. Therefore, it is more relevant to focus on the upper layer in order to
423 reveal the role of buoyancy transport in inhibition of deep convection and
424 restratification of homogenized water columns.

425 Figure 10 shows the components of the heat and freshwater budget due
426 to horizontal motion in the layer shallower than 120 m, where low salinity

427 water is contained along the WGC. The mean flow transports low salinity
 428 water toward the region bounded by 2000 and 3000 m isobaths in the north
 429 Labrador Sea (the NL box; Figures 10a, c). High temperature / low salin-
 430 ity water, which indicates the influence of WGC, is transported by eddies
 431 and extends toward the northern part of the interior (deeper than 3000 m)
 432 Labrador Sea (Figures 10b, d).

433 The heat originating from the atmosphere in summer is transported by
 434 eddies in the northeastern part of the interior Labrador Sea, where the winter
 435 convection is restricted by eddy activity (Figures 7a-c, e). The mean flow
 436 and eddies advect low salinity water from the WGC region to the north
 437 and northern part of the interior Labrador Sea, where consequently the deep
 438 convection is suppressed in spite of the strong sea surface cooling in winter
 439 (Figure 7c, d). As the surface low salinity water is colder than the buoyant
 440 water of the IW, it efficiently prevents the deep water from being cooled by
 441 the atmosphere. We find that both heat and freshwater transports work to
 442 stabilize the interior Labrador Sea. Next, we compare relative contributions
 443 from heat and freshwater.

444 3.5. Buoyancy transport

445 Buoyancy relative to a reference state, B , is decomposed into thermal and
 446 haline parts, B_θ and B_S , respectively, and they are defined here as

$$B_\theta = -g\alpha \Delta\theta,$$

$$B_S = g\beta \Delta S,$$

$$\alpha(\theta, S, P) = \frac{\partial \rho}{\partial \theta}, \quad \beta(\theta, S, P) = -\frac{\partial \rho}{\partial S},$$

Figure 10: Same as Figures 8b, c (upper panels) and 9b, c (lower panels), but vertically integrated only in the upper ocean (< 120 m).

449 where $\Delta\theta$ and ΔS are anomalies of potential temperature and salinity, re-
 450 spectively, from the reference state, g is acceleration of gravity, P is pressure,
 451 ρ is potential density. In the buoyancy budget of the whole water column,
 452 the total advection (mean and eddy) provides buoyancy by heat (Figure 11a)
 453 in almost all regions. Buoyancy is lost by high salinity water (IW) in the
 454 interior ocean, while it is gained by the low salinity water originated in the
 455 WGC, LC, and river runoff around the continental margin (Figure 11b).
 456 The boundary current (mean advection) injects buoyancy due to heat (Fig-
 457 ure 11c). The freshwater is injected by boundary current along the coasts,
 458 especially along the western coast of Greenland, and freshwater content is
 459 decreased along continental slope around the WGC and LC boxes where IW
 460 is transported (Figures 11d). Eddies transport heat and high salinity water
 461 from the boundary current region in the WGC box to the Labrador Sea in-
 462 terior (INT box) (Figure 11e, f). It is due to eddy activity that heat and
 463 salinity transport is larger in the northeastern Labrador Sea interior than in
 464 the southwestern Labrador Sea interior.

465 A quantitative comparison (Table 2) shows that the whole-column gain
 466 or loss of heat (temperature) contributes more to the buoyancy budget than
 467 that of freshwater (salinity) over the whole Labrador Sea. For instance, in
 468 the INT box, the buoyancy imported by heat ($661 \times 10^4 \text{ N s}^{-1}$) is about 17
 469 times larger than the buoyancy lost by high salinity water ($-38 \times 10^4 \text{ N s}^{-1}$).
 470 Even considering mean and eddy advection separately, freshwater does not
 471 contribute to the buoyancy budget in the Labrador Sea for the whole-column
 472 view.

473 The lower part of Table 2 shows the buoyancy budget in the near surface

Figure 11: Distribution of thermal contribution to the vertically integrated gain or loss of buoyancy over the whole column by (a) total of mean advection and eddy, (c) mean advection, and (e) eddy. (b), (d), and (f) are same as (a), (c), and (e), respectively, but for haline contribution.

Table 2: Budget of volume integrated buoyancy (10^4 N s^{-1}) by mean advection, and eddies in four boxes (WGC: West Greenland Current, NL: North Labrador, INT: Labrador Sea interior, and LC: Labrador Current) defined in Figure 8b. The values to the left and right of the slash indicate the thermal and haline buoyancy budget, respectively.

Heat / Freshwater	WGC	NL	INT	LC
Total advection	69/ 46	199/ 14	661/ -38	311/ 10
Mean advection	1009/ -39	316/ -8	-292/ 82	441/ -26
Eddy advection	-940/ 85	-116/ 22	952/-119	-100/ 36
Total adv. ($< 120 \text{ m}$)	57/ 46	69/100	199/ 73	42/ 98
Mean adv. ($< 120 \text{ m}$)	242/-332	42/ 91	7/ 14	22/-64
Eddy adv. ($< 120 \text{ m}$)	-185/ 378	27/ 9	192/ 58	20/162

layer (above 120 m depth). In WGC box, both heat and freshwater comparably (57 and $46 \times 10^4 \text{ N s}^{-1}$, respectively) contribute to the buoyancy injection in the total advection (mean plus eddy) (Figures 12a, b and Table 2). In the NL and LC boxes, the freshwater contribution to the buoyancy (100 and $98 \times 10^4 \text{ N s}^{-1}$, respectively) is larger than that induced by heat (69 and $42 \times 10^4 \text{ N s}^{-1}$, respectively). In the north Labrador Sea (NL box), the freshwater and heat by mean advection mainly causes the buoyancy gain in the surface layer (Figures 12d, e). Quantitative comparison (Table 2) shows that the buoyancy injection by mean advection of freshwater ($91 \times 10^4 \text{ N s}^{-1}$) is twice as large as that by mean advection of heat ($42 \times 10^4 \text{ N s}^{-1}$). In the LC box, the surface layer buoyancy gain is mainly contributed from the eddy-induced freshwater injection ($162 \times 10^4 \text{ N s}^{-1}$).

In the INT box, the thermal buoyancy gain ($199 \times 10^4 \text{ N s}^{-1}$) is about

three times larger than the haline buoyancy gain ($73 \times 10^4 \text{ N s}^{-1}$) for the to-
 tal advection. However, while the eddy-induced thermal buoyancy transport
 spreads over the whole Labrador Sea interior, the eddy-induced haline buoy-
 ancy transport is confined to the northern Labrador Sea interior (Figures
 12e, f). Thus, we calculate the buoyancy budget by separating the INT box
 into northern (INT_N), eastern (INT_E), and southwestern (INT_{SW}) parts
 (Table 3). In the northern interior (INT_N), although the total advection
 mainly transports buoyancy by heat ($67 \times 10^4 \text{ N s}^{-1}$) rather than freshwater
 ($28 \times 10^4 \text{ N s}^{-1}$), the eddy-transported buoyancy by freshwater ($44 \times 10^4 \text{ N}$
 s^{-1}) is comparable to the total heat contribution ($67 \times 10^4 \text{ N s}^{-1}$). In the
 eastern part of the interior Labrador Sea (INT_E), the injection of buoyancy
 by heat ($91 \times 10^4 \text{ N s}^{-1}$) is much larger than that by freshwater ($20 \times 10^4 \text{ N}$
 s^{-1}) even above 120 m depth. In the INT_{SW} box, where the deepest convec-
 tion develops in winter, the eddy-induced heat transport ($46 \times 10^4 \text{ N s}^{-1}$)
 mainly contributes to the buoyancy gain, and that caused by mean advection
 of freshwater is small ($17 \times 10^4 \text{ N s}^{-1}$). The result shows that the fraction of
 haline contribution is further significant in the northern part of the interior
 than that in the eastern and southwestern parts. Note that in the INT_N box,
 if we calculate the buoyancy above 70 m depth, eddy-induced contribution
 of freshwater ($46 \times 10^4 \text{ N s}^{-1}$) is larger than that of heat ($36 \times 10^4 \text{ N s}^{-1}$)
 (Table not shown).

McGeehan and Maslowski (2011) explained that unrealistic deepening of
 winter convection in the northern Labrador Sea in their model is due to
 shortage of the freshwater along the western coast of Greenland. Chanut
 et al. (2008) reported that the convection in the northern Labrador Sea is

Table 3: Budget of volume integrated buoyancy (10^4 N s^{-1}) by mean advection, and eddies above 120 m depth in three boxes in the INT box defined in Figure 12a. The values to the left and right of the slash indicate the thermal and haline buoyancy budget, respectively.

Heat / Freshwater	INT _N	INT _E	INT _{SW}
Total adv.	67/ 28	91/20	41/ 24
Mean adv.	-8/-15	20/13	-5/ 17
Eddy adv.	75/ 44	71/ 8	46/ 7

512 suppressed by the heat originating in the IW and transported by the IRs in
513 their model. However, as they did not represent the low salinity core in IRs
514 originating in the WGC, contribution of freshwater to the supply of buoy-
515 ancy was neglected. This study directly points out that the contribution of
516 freshwater to inhibiting the convection is important in the northern Labrador
517 Sea for the first time.

518 4. Discussion and conclusion

519 We investigated the relationship between the deep convection and buoy-
520 ancy budget in the Labrador Sea by using a realistically configured eddy-
521 resolving OGCM. In particular, the importance of near surface freshwater
522 transport from the vicinity of the WGC is revealed for the first time. The
523 mean currents and qualitative features of seasonal cycle of temperature and
524 salinity are well simulated by the model. The near surface water of relatively
525 low temperature and salinity in the WGC, which is not reproduced in previ-
526 ous models, is realistically represented. The three types of eddies, which are
527 specified in previous studies, are also well reproduced. The IRs and BCEs

Figure 12: Same as Figure 11, but vertically integrated above 120 m depth. Three boxes in INT box (INT_N : northern INT, INT_E : eastern INT, and INT_{SW} : southwestern INT) are defined for the qualitative analysis.

528 contain relatively low density water of the boundary currents and transport
 529 buoyancy by moving to the interior of the Labrador Sea. It is consistent
 530 with observations that the winter convection is confined to the southwestern
 531 Labrador Sea. As the deep convection unrealistically extends over the whole
 532 of Labrador Sea in the comparative experiment (the SMOOTH case) without
 533 propagation of the IRs, it is confirmed that the buoyancy transport by the
 534 IRs is crucial in inhibiting the winter deep convection. The lateral buoy-
 535 ancy transport caused by heat is significantly larger than that by freshwater
 536 for the whole water column in our model. However, it is demonstrated that
 537 freshwater mainly contributes to the buoyancy transport near the sea surface
 538 in the northern Labrador Sea.

539 A previous modeling study (McGeehan and Maslowski, 2011) failed to
 540 represent the low salinity water along the western coast of Greenland, and
 541 winter deep convection drastically and unrealistically developed in the north-
 542 ern Labrador Sea therein. Myers and Donnelly (2008) examined the roles of
 543 sea surface heat and freshwater fluxes in the interannual variability of forma-
 544 tion of LSW, and they concluded that the buoyancy induced by freshwater
 545 is smaller than heat. We studied the effect of both lateral transport and sea
 546 surface flux of heat and freshwater on the buoyancy in the central Labrador
 547 Sea, and the role of freshwater is not smaller than heat near the sea surface
 548 in the northern Labrador Sea. The freshwater provided in the upper 100 m
 549 of the interior Labrador Sea from March to September is ~ 30 cm in our re-
 550 sult, which is half of observational estimates (Lazier, 1980; Khatiwala et al.,
 551 2002). Existence of low salinity water near the sea surface would reduce
 552 oceanic heat loss under cold air because strong neat surface stratification

553 prevents heat uptake from depth and the sea surface is cooled immediately,
554 as suggested by Gelderloos et al. (2012). Then, it is naturally expected that
555 winter convection does not develop in the northern Labrador Sea even under
556 the influence of the strong cold winds from the continent. If the model's un-
557 derestimation of lateral freshwater flux is corrected, reproducibility of deep
558 convection might be drastically improved.

559 The modeled annual mean surface heat loss (-100 W m^{-2}) in the cen-
560 tral Labrador Sea (53°W , 56°N ; corresponding to the ocean weather station
561 Bravo) is much larger than an estimate based on reanalysis data (-28 W m^{-2} ;
562 Straneo, 2006). On the other hand, van Aken et al. (2011) estimated a larger
563 heat loss, -67 W m^{-2} , and discussed that such a magnitude is necessary for
564 explaining the interannual variability of LSW temperature. Thus, the actual
565 magnitude of annual mean heat flux is still controversial. The value -67 W
566 m^{-2} is closer to the result of our model. However, the modeled heat loss is
567 still significantly larger and would be associated with the simulated stronger
568 deep convection compared with observation. As the CEs, which actually are
569 sub-mesoscale and rapidly tilt the steeply sloping isopycnal surfaces around
570 deep convection, are not sufficiently resolved, the mesoscale eddies function
571 as their substitute in our model. If sub-mesoscale eddies were resolved, the
572 winter convection might be more efficiently suppressed. We should directly
573 resolve such small eddies or use parameterizations of sub-mesoscale eddies
574 (Fox-Kemper et al., 2008; Fox-Kemper and Ferrari, 2008) for better repre-
575 senting restratification of winter deep convection and heat loss in the central
576 Labrador Sea. It is likely that the too strong sea surface cooling and de-
577 velopment of convection enhance eddy activities through a large horizontal

578 density gradient and thus strong baroclinicity. It may be the case that the
579 lateral buoyancy (both heat and freshwater) flux is overestimated by such a
580 bias. However, our main conclusion that the role of freshwater is comparable
581 in magnitude to that of heat is unaffected.

582 McGeehan and Maslowski (2011) suggested the important role of eddy-
583 induced lateral freshwater flux from the western shelf of the Labrador Sea in
584 restratifying the homogenized water columns. They did not make a quanti-
585 tative comparison of the contribution of low salinity water from the western
586 and eastern shelves (WGC and LC) on the restratification or buoyancy in-
587 jection into the central Labrador Sea because of the poor reproduction of the
588 low salinity water along the western coast of Greenland. The eddy-induced
589 transport of freshwater from the western shelf (LC) is much smaller than that
590 from the eastern shelf (WGC) in our model. Furthermore, hydrographic and
591 mooring data demonstrate that the source of surface freshwater in the central
592 Labrador Sea is the WGC (Schmidt and Send, 2007).

593 The freshwater in the WGC is mainly originated from the Arctic Ocean
594 and propagated by the East Greenland Current (Aagaard and Carmack,
595 1989; Kwok et al., 2004). Kwok and Rothrock (1999) concluded that the
596 interannual and decadal variability of the sea ice export through the Fram
597 Strait has high correlation with the NAO or AO indices. This variability
598 propagates to the WGC by the East Greenland Current through the Den-
599 mark Strait and around Cape Farewell. Supply of low density water to the
600 Labrador Sea should have an associated interannual to decadal variability,
601 and it influences the climate through open-ocean deep convection and the
602 Atlantic thermohaline circulation. The relation between the interannual vari-

603 ability of freshwater transport to the Labrador Sea and climate variability is
604 worth investigating.

605 The Greenland Sea is also widely recognized as a region of open-ocean
606 convection and formation of deep water. Since the Greenland Sea has similar-
607 ities to the Labrador Sea with regard to the inflow of low density boundary
608 currents originating in the Arctic Ocean and the North Atlantic Current,
609 eddy-induced buoyancy transport from the boundary currents to the interior
610 may also be important there. Previous observational and modeling studies
611 suggested that deep convection events in the Labrador and Greenland Seas
612 are anti-correlated (Dickson et al., 1996; Oka et al., 2006). Oka et al. (2006)
613 concluded that the freshwater transport through the Denmark Strait con-
614 trols the variability of deep convection in the Labrador and Greenland Seas.
615 As the horizontal grid in their model is too coarse to resolve eddies, which
616 contribute to the transport of heat and freshwater between the boundary
617 currents and the Labrador Sea interior, it is worth reexamining this issue
618 from the viewpoint of the role of boundary current eddies by incorporating
619 an eddy-resolving ocean model.

620 **Acknowledgements**

621 This work is supported by the JST CREST and MEXT GRENE Arctic
622 Climate Change Research Project. Numerical calculations were performed
623 on SR16000 at Information Technology Center, University of Tokyo. Almost
624 all figures are drawn by using the Grid Analysis and Display System (GrADS)
625 developed by M. Doty and M. Fiorino.

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