Buoyancy- and eddy-driven circulation in the Atlantic layer of the Canada Basin

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[1] The circulation driven by buoyancy and mesoscale eddy effects in the Atlantic layer of the Canada Basin is obtained diagnostically from a climatological density field. The buoyancy forcing is caused by the divergence of diapycnal flow. The diapycnal flow is estimated from a density balance between the diapycnal advection and diffusion based on the density field. On the other hand, the eddy forcing is formulated as the curl of vertical friction. The vertical friction is estimated from geostrophic shear, which is obtained from the density field. Both the flow fields caused by the buoyancy forcing and by the eddy forcing similarly suggest a new flow structure characterized by a transbasin westward flow, which appears from the eastern central basin to the southwestern basin. The transbasin westward flow is also shown in geostrophic circulation at 500 m relative to reference levels in a wide range (1000–3000 m). The significant part of the flow is consistent with a temperature distribution at 500 m. [INDEX TERMS: 4207 Oceanography: General: Arctic and Antarctic oceanography; 4203 Oceanography: General: Analytical modeling; 4568 Oceanography: Physical: Turbulence, diffusion, and mixing processes; 4520 Oceanography: Physical: Eddies and mesoscale processes; KEYWORDS: Arctic Ocean, Atlantic layer, Potential vorticity, Characteristic equation, Buoyancy-driven circulation, Eddy-driven circulation]

1. Introduction

[2] In the Arctic Ocean, warm water originating in the Atlantic Ocean circulates cyclonically along shelf slopes surrounding the Arctic and spreads even to the Canada Basin away from the Atlantic. Such circulation was inferred from analyses of water properties and current measurements [Aagaard, 1989; Rudels et al., 1994; McLaughlin et al., 1996]. Nazarenko et al. [1998] successfully reproduced the cycloic boundary currents along the basin peripheries by parameterizing a role of eddies interacting with bottom topographies, i.e., the neutron effect discussed by Holloway [1987, 1992, 1996]. On the other hand, different models where unresolved eddies were represented by eddy viscosity [e.g., Semtner, 1987; Zhang et al., 1998a, 1998b] indicated that the circulation at a depth of the Atlantic water is anticyclonic in the interior of the Canada Basin. Thus the driving mechanisms of the Atlantic water circulation would be different between the boundary and the interior of the Canada Basin.

[3] Furthermore, there is another possibility to cause motion at the depths of the Atlantic water. A diapycnal flow, which intersects a density surface, advects buoyancy balancing with diapycnal buoyancy diffusion to keep a steady density stratification [Munk, 1966]. The divergence (convergence) of diapycnal flow gives rise to a stretching (compression) of a water column in a layer defined by density surfaces, causing horizontal motion in the layer. This buoyancy forcing would compete against the eddy viscosity forcing in the interior of the Canada Basin to drive the Atlantic water.

[4] We do not have enough knowledge about the Atlantic water circulation and its dynamics, especially in the interior of the Canada Basin. In order to obtain the circulation there, a diagnostic analysis is proposed in the present study. The circulation there is assumed to be driven by the buoyancy and eddy viscosity forcings, which inject potential vorticity (PV) into a water column. These forcings are estimated from a climatological density field recently provided by Steele et al. [2001]. The buoyancy forcing is caused by the divergence of diapycnal flow. The diapycnal flow is estimated from the advective-diffusive equation describing a diapycnal density balance. On the other hand, the eddy viscosity forcing is formulated as the curl of vertical friction. The vertical friction is estimated from geostrophic shear, which is obtained from the density field. The motion driven by those forcings can be calculated from the PV equation in a water column assumed in the geostrophic balance. Details of the analysis are described in section 2, and the results of the analysis are depicted in section 3. Discussion is given in section 4, and conclusions are presented in section 5.

2. Analytical Method

[5] The Atlantic layer in the Canada Basin does not outcrop to the sea surface, and therefore waters in this layer are not directly forced by the wind and ice drag. The conceivable agencies, which cause water movement in the Atlantic layer, are buoyancy and mesoscale eddy effects. The former drives waters through the divergence of diapycnal flow. The latter
causes motion by means of various mechanisms like the Neptune effect [Holloway, 1987, 1992, 1996], the eddy viscosity, and other processes. In the present analysis, the eddy viscosity is assumed to be the predominant eddy effect in the interior of the Canada Basin as examined by, for example, Semtner [1987] and Zhang et al. [1998a, 1998b]. The motion driven by the buoyancy and eddy viscosity forcings is obtained by integrating the forcing terms in the PV equation as described below.

[a] At the eastern boundary of the Canada Basin, where PV contours strike the boundary, no normal flow condition in the Atlantic layer is applied to the present analysis, and therefore the obtained flow represents not a total flow but only part of the total flow crossing the PV contours driven by the buoyancy and eddy viscosity forcings. There can be a transport along the PV contours that the present analysis does not capture because of the non-normal-flow condition at the eastern boundary. In fact, an outflow of the Arctic deep water through the western Fram Strait [e.g., Aagaard et al., 1991] would accompany swift boundary currents along the eastern boundary of the Canada Basin, and the non-normal-flow condition at the eastern boundary might not be valid. However, from observations of the Atlantic water boundary current in the Lincoln Sea, Newton and Sotrin [1997] indicated that the velocity of the cross-stream boundary current is less than 1 cm s^{-1}, while in the interior of the Canada Basin the velocity of a significant flow reaches 1 cm s^{-1} in the resultant flow fields in the present analysis. Therefore at least the significant flow driven by the buoyancy and eddy viscosity forcings would capture one of the characteristic flows in the Atlantic layer of the Canada Basin.

[b] The motion in the interior of the Canada Basin caused by the buoyancy and eddy viscosity forcings is assumed to be a steady flow, approximately geostrophic (with the viscosity component) and hydrostatic, and the fluid satisfies continuity. In this case, the PV equation in the Atlantic layer is

\[
\frac{\partial}{\partial x} \left( \frac{f}{h} \right) + \frac{\partial}{\partial y} \left( \frac{f}{h} \right) = \frac{f}{h} \left( w_T^* - w_E^* \right) + \frac{\text{curl} \cdot \mathbf{F}}{h},
\]

where \( u \) and \( v \) represent zonal and meridional velocities, respectively; \( h \) is the layer thickness; \( f \) is the Coriolis parameter; \( w_T^* \) and \( w_E^* \) stand for diapycnal velocities at the top and bottom interfaces of the layer, respectively; and \( F \) is the vertical friction caused by the eddy viscosity. The horizontal components of vertical eddy friction are

\[
\begin{align*}
F_x &= \frac{\partial}{\partial x} \left( A \frac{\partial h}{\partial x} \right) \\
F_y &= -\frac{\partial}{\partial y} \left( A \frac{\partial h}{\partial y} \right),
\end{align*}
\]

where \( A \) is the coefficient of vertical eddy viscosity. Schematics of this diagnostic model are depicted in Figure 1.

[c] By using geostrophic relations for \( u \) and \( v \), the PV equation of (1) is written as a characteristic equation, the characteristics of which are given as PV contours (\( f / h = \) constant):

\[
-\frac{\partial P}{\partial y} \frac{\partial}{\partial x} \left( \frac{f}{h} \right) - \frac{\partial P}{\partial x} \frac{\partial}{\partial y} \left( \frac{f}{h} \right) = \nu_h \left( \frac{f}{h} \right)^2 (w_T^* - w_E^*) + \rho_0 \frac{\nu_h}{h} \text{curl} \cdot \mathbf{F},
\]

where \( P \) is the pressure and \( \rho_0 \) is the average density in the Atlantic layer. Then it is assumed that velocities across the eastern boundary of the Canada Basin vanish, i.e., \( P = 0 \) along the boundary. Integrating the buoyancy and eddy viscosity forcings, which are the first and the second terms, respectively, of the right-hand side of (3), along characteristics from the eastern boundary of the Canada Basin where \( P = 0 \), one can obtain flow (pressure) fields in the Atlantic layer of the basin.

[d] Both the forcing terms are estimated from a density field calculated from the Arctic climatological data of temperature and salinity recently provided by Steele et al. [2001]. The density field used in the present analysis is the one further smoothed by a median filter [Rabiner et al., 1975]. The buoyancy forcing due to the divergence of diapycnal flow is estimated from a density balance of diapycnal advection and diffusion. As shown by Tezgurman [1986], an equation of the diapycnal advective-diffusive density balance is expressed as a finite difference formula for a layer model, and the diapycnal velocity at the interface between the layers \( n \) and \( (n + 1) \), \( w_T^* \), can be written as follows:

\[
w_T^* = \frac{\nu_h \Delta_h \rho_P/h_{n+1} - \Delta_h \rho_P/h_n}{\rho_{n+1} - \rho_n},
\]

where \( \nu_h \) is the diffusion coefficient of diapycnal mixing; \( h_n \) and \( \rho_n \) are the thickness and average density in the \( n \)th layer, respectively; and \( \Delta \rho \) is the density difference between the top and bottom interfaces of the \( n \)th layer. A series of layers is defined with density surfaces as described later. The diffusion coefficient, \( \nu_h \), evaluated from a vertical 1°C distribution in the deep water of the Canada Basin is \( 0.39 \times 10^{-24} \text{ m}^4 \text{ s}^{-1} \) [Mackinnon et al., 1992]. This value might not apply to the Atlantic layer, because the actual diffusivity is variable in all space directions and time. Note that the circulation caused by the buoyancy forcing depends on the chosen value of the diffusivity. However, for simplicity, \( \nu_h \) is assumed to be constant as \( 0.5 \times 10^{-24} \text{ m}^4 \text{ s}^{-1} \), which is
almost as same as the evaluated value from the $^{14}$C distribution.

On the other hand, the vertical friction expressed in equation (2) can be estimated from geostrophic shears, $(\partial u/\partial z, \partial v/\partial z)$, which are obtained from the climatological density field. In this case, the curl of vertical friction is expressed as

$$ \text{curl} \cdot \mathbf{F} = -Ag \left( \frac{\rho_f - \rho_B}{\rho_f} \left[ \frac{\partial}{\partial x} \left( 1 \frac{1}{f} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( 1 \frac{1}{f} \frac{\partial h}{\partial y} \right) \right] \right) $$

where $g$ is the gravity; $\rho_f$ and $\rho_B$ are the densities at the top and bottom interfaces of the Atlantic layer, respectively. Therefore the eddy viscosity forcing is obtained from the density stratification, i.e., the Atlantic layer thickness $(h)$ defined with the density surfaces of $\rho_f$ and $\rho_B$. Similar to the diffusivity argument, the vertical eddy viscosity, $A$, is assumed to be constant as $1 \times 10^{-4}$ m$^2$ s$^{-1}$, although the actual viscosity is variable in all space directions and time and the circulation caused by the eddy viscosity forcing depends on the chosen value of the viscosity. The value of $A$ in the present analysis is much larger than that used in usual numerical models (1 to $5 \times 10^{-3}$ m$^2$ s$^{-1}$). In the usual numerical models the viscosity would be treated as the one induced by the Reynolds stress, which is caused by eddies accompanied by, for example, an internal wave breaking. On the other hand, in the present analysis, the viscosity is assumed to be induced by the stress resulted from mesoscale eddy perturbations. Such a stress was discussed by Rhines and Holland [1979] and Rhines and Young [1982b], who suggested that the momentum is vertically transferred by perturbations of density surfaces (vortex lengths). In this case, the order of viscosity, $A$, is equivalent to that expressed as $k f^2/\sqrt{N}$, where $k$ is the horizontal diffusion coefficient of PV and $N$ is the buoyancy frequency. Taking a typical value of $k = 10^3$ m$^2$ s$^{-1}$, and representative $f^2/\sqrt{N}$ of about $10^{-4}$, $A$ is estimated as $0.1$ m$^2$ s$^{-1}$, and this value is used in the present analysis.

The Atlantic layer is usually defined by 0°C isotherms sandwiching the temperature maximum originating in the warm Atlantic water [e.g., Aagaard, 1981; Carmack et al., 1997; Swift et al., 1997]. In the present analysis, the Atlantic layer is defined by density surfaces near the 0°C isolthersms, i.e., 27.5- and 28.0-$\sigma_0$ surfaces. To obtain the velocities of diapycnal flow at the top and bottom interfaces of the Atlantic layer, the upper and lower layers to the Atlantic layer are defined as follows. The upper layer at depths of cold halocline is defined by the density surfaces of 25.5 and 27.5-$\sigma_0$. The 25.5-$\sigma_0$ surface outcrops to the European side of the Lomonosov Ridge, and waters below the density surface do not outcrop to the sea surface in the western Arctic. The bottom interface of the lower layer is chosen as the density surface of 28.05-$\sigma_0$ lying near a depth of 1200 m.

3. Results

The thickness and PV distributions in the Atlantic layer of the Canada Basin are immediately obtained from the climatological density field based on the data sets of Swene et al. [2001]. Figure 2 shows the distribution of the Atlantic layer thickness defined with the density surfaces of 27.5 and 28.0-$\sigma_0$. The Atlantic layer is thin (thick) in the southern (northern) part of the Canada Basin. To explain the thickness distribution, a meridional section of density along 135°W is shown in Figure 3. The 28.0-$\sigma_0$ surface, the bottom interface of the Atlantic layer, has a ridge and a trough centered on 74°N and 82°N, respectively, resulting in the thin layer in the south and the thick layer in the north.

The PV in the Atlantic layer is calculated by dividing the Coriolis parameter by the layer thickness, and the distribution is shown in Figure 4. The PV distribution indicates an inverse pattern to the thickness distribution, i.e., PV is high in the south and low in the north, and PV contours from the eastern boundary extend into the interior of the basin.

Figures 5a, 5b, and 5c represent the distributions of diapycnal velocities at the top and bottom interfaces of the Atlantic layer and the difference of the velocities between the top and bottom interfaces multiplied by the square of
PV, i.e., the first term of the right-hand side of equation (3). In Figures 5a and 5b, the top (bottom) interface velocity increases (decreases) with latitude. As derived from equation (4), the distribution of the diapycnal velocity \( u'_z \) depends largely on the distribution of the upper layer thickness \( h_u \), because the stratification \( \Delta \sigma_0 / h_u \) is larger in the upper layer than the lower layer. Therefore the distribution of the top (bottom) interface velocity of the Atlantic layer is characterized by the thickness of the cold halocline (Atlantic) layer. In the southern part of the Canada Basin the diapycnal velocity is large at the bottom interface of the Atlantic layer (Figure 5b) compared with that at the top interface (Figure 5a), resulting in convergence of the diapycnal flow (Figure 5c), while in the northern part, the diapycnal flow diverges.

[15] Figure 6 shows the distribution of the curl of vertical friction multiplied by PV, i.e., the second term of the right-hand side of equation (3). The distribution is roughly determined by the laplacian of the inverse thickness of the Atlantic layer as expressed in equation (5). The thickness of the Atlantic layer has its minimum at 135°W and 74°N (Figure 2), and the minimum gives negative values for the laplacian of the inverse thickness there (Figure 6). On the other hand, northwest of the minimum thickness region,
Figure 5. Distributions of (a) diapycnal velocity at the top interface (27.5-\sigma_0 surface) and (b) at the bottom interface (28.0-\sigma_0 surface) of the Atlantic layer, and (c) the difference of the velocities between the top and bottom interfaces multiplied by the square of potential vorticity, i.e., the first term of the right-hand side of equation (3). In Figures 5a and 5b the units of the contour labels are $10^{-7}$ m s$^{-1}$, and the contour interval is $0.1 \times 10^{-7}$ m s$^{-1}$. In Figure 5c the units of the contour labels are $10^{-21}$ m$^{-1}$ s$^{-1}$, and the contour interval is $1 \times 10^{-21}$ m$^{-1}$ s$^{-1}$. 
the thickness has its maximum, resulting in positive laplacian of the inverse thickness there. Therefore negative vorticity is injected into a water column of the Atlantic layer in the minimum thickness region, and positive vorticity is injected northwest of the region.

[16] Flow fields are obtained by integrating the forcing terms of the right-hand side of equation (3) along PV contours. Figure 7a (8a) represents the pressure field obtained by integrating the buoyancy (eddy viscosity) forcing given by the diapycnal divergence (the curl of vertical friction), the first (second) term of the right-hand side of equation (3), and Figure 7b (8b) shows the geostrophic velocity field derived from the pressure field. The solutions of equation (3) are only calculated in regions where the PV contours from the eastern boundary extend into the interior of the basin, and in Figures 7a and 8a the points where the solutions are given are indicated by dots. In the flow field caused by the diapycnal divergence (Figures 7a and 7b), a southwestward flow appears from the northeastern to the southwestern Canada Basin. On the other hand, in the flow field resulting from the curl of vertical friction (Figures 8a and 8b), a southwestward flow appears only in the southwestern Canada Basin, and the flow returns to the east, resulting in a cyclonic gyre. Although details are different in both circulation patterns, a westward flow along 77°N between 130°W and 140°W (strictly speaking, the direction of the flow is toward the west-southwest for the buoyancy-driven circulation and west-northwest for the eddy-driven circulation) and the following southwestward flow in the southwestern Canada Basin are similarly prominent. The prominent westward and southwestward flows are hereinafter referred to as a transbasin westward flow.

[17] A cause of the transbasin westward flow is easily explained by the spread of characteristics of equation (3), i.e., PV contours and the distributions of the forcing terms. The transbasin westward flow appears between a PV minimum (147°W, 78°N) and a PV maximum (136°W, 75°N) in the PV field (Figure 4). Between the minimum and the maximum of PV, the PV contours are almost parallel to the contours of the buoyancy and eddy viscosity forcings (Figures 5c and 6, respectively). In that region the contour values of both the forcings increase as the PV decreases toward the northwest. As a result, the pressure values obtained by integrating the forcing terms along a PV contour increase toward the northwest, and the pressure gradient gives rise to the transbasin westward flow.

[18] The transbasin westward flow is also seen in geostrophic flow fields at 500 m relative to reference levels in a wide range (1000-3000 m). The depth of 500 m corresponds to the core depth of the Atlantic water in the Canada Basin. Therefore the geostrophic circulation at 500 m is a representative of the Atlantic water circulation. Figures 9a and 9c show the distributions of dynamic height at 500 m relative to 1000 m and 3000 m, respectively. The geostrophic velocities at 500 m derived from the dynamic height shown in Figures 9a and 9c are represented in Figures 9b and 9d, respectively. Geostrophic flow fields at 500 m relative to any other levels between 1000 and 3000 m were also calculated (not shown). Although detail circulation patterns depend on the reference levels, all the results of geostrophic calculation, as shown in Figures 9a to 9d, indicate that a westward flow appears between a cyclonic gyre around 75°N, 130°-145°W, and an anticyclonic gyre centered on 148°W, 77°N. The westward flow resembles the transbasin westward flow obtained in both the buoyancy-driven circulation and the eddy-driven circulation. In addition, the cyclonic gyre around 75°N, 130°-145°W in the geostrophic fields is also reproduced in the flow field caused by the eddy viscosity forcing.

[19] Figure 10 shows a temperature distribution at 500 m. The contours of temperature, e.g., 0.44°-0.46°C isotherms, orient northeast to southwest in the interior of the Canada Basin. West of 140°W a southwestward flow, which is the significant part of the transbasin westward flow shown in Figures 7, 8, and 9, is consistent with the isotherm orientation. East of 140°W, only the flow caused by the buoyancy forcing is toward the southwest, and therefore the
buoyancy-driven circulation explains well the whole isotherm orientation in the interior of the Canada Basin.

4. Discussion

[29] The total Atlantic water circulation in the interior of the Canada Basin would be obtained by summing the flows caused by the buoyancy forcing and by the eddy viscosity forcing and the flow originating in the eastern boundary. In the present analysis, the flow crossing the eastern boundary was assumed to be zero, and the buoyancy-driven circulation and the eddy-driven circulation excluding the flow from the eastern boundary were calculated. The summation of these two circulations is not so simple. This is because the diffusivity and viscosity, which were assumed to be constant in the present analysis, have spatial and time variability. Therefore, to obtain the Atlantic water circulation driven by the two combined forcings, the contribution rate of the two effects must be evaluated. Such an evaluation is beyond the scope of this study; however, it would be possible to infer which effect is predominant in a certain region by comparing each circulation with the geostrophic flow fields and the temperature distribution.

[31] In the geostrophic flow fields at 500 m, the southern branch of the cyclonic gyre around 75°N, 130°–145°W, i.e., the eastward flow south of 75°N, is only reproduced by the eddy-driven circulation. On the other hand, in the northeastern Canada Basin the isotherms at 500 m are along the southwestward flow, which is only caused by the buoyancy forcing. Therefore the eddy viscosity forcing might be predominant over the buoyancy forcing in the southern part of the Canada Basin, where the eastward flow appears as part of the cyclonic gyre. In contrast, the buoyancy forcing would prevail over the eddy...
Figure 8. Fields of (a) pressure and (b) velocity averaged in the Atlantic layer caused by the vertical friction resulting from perturbations of density surfaces due to mesoscale eddies. In Figure 8a, dots indicate positions where solutions of equation (3) are obtained. The contour interval is 20 Nm$^{-2}$. In Figure 8b the scale of velocity appears in the bottom right corner of the panel.

viscosity forcing in the northeastern Canada Basin, resulting in the buoyancy-driven circulation consistent with the temperature distribution there. The transbasin westward flow appears in both the buoyancy-driven circulation and the eddy-driven circulation and is consistent with the geostrophic flow fields and the temperature distribution west of 140°W. Although the contribution rate between the diffusivity and viscosity is not known, the transbasin westward flow would be caused by the combination of the buoyancy and eddy viscosity forcings. [22] In the northwestern Canada Basin, there is probably an anticyclonic gyre. The geostrophic flow fields (Figures 9a to 9d) indicate the anticyclonic gyre centered on 148°W and 77°N. The buoyancy driven circulation (Figures 7a and 7b) and the eddy-driven circulation (Figures 8a and 8b) would also capture the southern branch of the anticyclonic gyre around 150°W, south of 76°N. On the other hand, the existence of an eastward flow, which might be the northern branch of the anticyclonic gyre, is suggested by high concentration of radionuclide $^{129}$I, which is transported with the Atlantic water, appearing from the northern end of the Northwind Ridge (150°W, 78°N) to the interior of the Canada Basin [Smith et al., 1990]. Stuehle et al. [2000] also suggested a circulation pathway of the Atlantic water via the northern end of the Northwind Ridge to the interior of the Canada Basin by estimating a renewal time of the water from tritium/He and CFC distributions. [23] In the region where the anticyclonic gyre appears, PV is low and relatively uniform (Figure 4). Furthermore, the anticyclonic gyre shrinks toward the north compared with the surface Beaufort gyre. Therefore the theory of potential vorticity homogenization proposed by Rhines and Young [1982a, 1982b] is a candidate to explain the driving mechanism of the anticyclonic gyre. The anticyclonic gyre in the relatively uniform PV region is probably caused mainly by mesoscale eddy perturbations, which transfer momentum of the surface wind-driven circulation (Beaufort gyre) to deeper layers with shrinks of the circulation. The
Figure 9. Geostrophic flow fields at 500 m represented by (a) dynamic height and (b) velocity relative to 1000 m and (c) dynamic height and (d) velocity relative to 3000 m. In Figure 9a the contour interval is 0.002 m² s⁻², while in Figure 9c the contour interval is 0.005 m² s⁻². The scale of velocity appears in the bottom right corner in each panel of Figures 9b and 9d.
mesoscale eddies act to diffuse PV horizontally, resulting in homogenization of the PV in the region surrounded by a closed streamline [Rhines and Young, 1982b]. However, strictly speaking, PV is not uniform but has its minimum value in the anticyclonic gyre of the northwestern Canada Basin. Therefore the buoyancy forcing, which inputs PV through the diapycnal divergence preventing the homogenization, might also contribute to drive the anticyclonic gyre.

It is useful to compare the dynamics of the Atlantic water circulation in the Canada Basin with the dynamics of the mid-depth water circulation in the subtropical ocean, because the physics of the well-known subtropical ocean circulation may shed light on the circulation in the Canada Basin. Nishino and Minobe [2000] proposed a model combining the buoyancy effects with the PV homogenization and showed that a westward flow appears along the southern rim of the uniform PV region in the second layer, which is just below a layer directly ventilated by an anticyclonic wind field. The westward flow outside of the uniform PV region in their model was caused by the buoyancy effects. In the subtropics, mesoscale eddies play an important role in the uniform PV region, but the eddy activity decreases outside of the region. Consequently, the buoyancy effects are predominant over the mesoscale eddy effects outside of the uniform PV region. On the other hand, in the Canada Basin, mesoscale eddies are produced at the southern boundary of the basin because of higher horizontal shears of currents and lack of ice covers in late summer [Manley and Hunkins, 1985] and coastal upwellings of the Atlantic water [Nishino et al., 2001]. Therefore, in the present analysis, the mesoscale eddy effects were taken into account even outside of the uniform PV region, in addition to the buoyancy effects. As a result, in the southern part of the Canada Basin the circulation driven by the eddy viscosity forcing, which is derived from the mesoscale eddy effects, can reproduce the cyclonic gyre remarkably similar to that obtained in the geostrophic circulation at 500 m.

[25] Liu [1993a, 1993b] studied responses of density stratification against variations of wind stress in the subtropical ocean and indicated that the density stratification varies significantly just south of the wind-driven anticyclonic gyre. Therefore, in the Canada Basin, the stratification just south of the uniform PV region might be sensitive for wind variability. If a wind system changed as indicated by Walsh et al. [1996], the change would cause a change of density stratification and induce the accompanying motion, which would be significant just south of the uniform PV region. The decrease (increase) of wind stress might cause a northeastward (southwestward) flow just south of the wind-driven anticyclonic gyre within the uniform PV region, because of a stretching (compression) of a planetary vortex tube due to the upward (downward) anomalous Ekman pumping. Such a flow resulting from the varying wind would interact with the transbasin westward flow caused by the diapycnal divergence and the curl of vertical friction. In addition, the change of density stratification due to the varying wind induces changes of distributions of the buoyancy and eddy viscosity forcings. As a result, a change of wind system would play an important role in the circulation pattern outside of the uniform PV region in the Canada Basin.

5. Conclusions

[26] Flow fields in the Atlantic layer of the Canada Basin driven by buoyancy and eddy viscosity forcings were calculated by integrating the forcing terms of a PV (characteristic) equation. The buoyancy forcing is caused by the divergence of diapycnal flow. The eddy viscosity forcing is induced from perturbations of density surfaces, which transfer momentum from the sea surface to deeper layers acting as the vertical stress. Some problematic assumptions were used in the present analysis. No normal flow condition at the eastern boundary was applied, although there would be swift boundary currents. The diffusivity and viscosity were assumed to be constant, in spite of those variabilities in
space and time. The circulation pattern depends on these assumptions. However, the present analysis suggested a robust flow structure of the transbasin westward flow, which is consistent with the geostrophic flow fields at 500 m relative to reference levels in a wide range (1000–3000 m) and the temperature distribution at 500 m west of 140°W where the flow is significant. The transbasin westward flow in summer with both the flow fields caused by the buoyancy forcing and by the eddy viscosity forcing appears from the eastern central basin around 77°N and 130°–140°W to the southwestern basin roughly south of 75°N and west of 140°W. Taking the whole circulation pattern into account, the southern branch of the cyclonic gyre in the geostrophic circulation at 500 m, i.e., the eastward flow south of 75°N, is only reproduced by the eddy-driven circulation. On the other hand, in the northeastern Canada Basin the isotherms at 500 m are along the southwestern flow, which is only caused by the buoyancy forcing. The present study gives a new feature on the Atlantic water circulation in the Canada Basin. So far, it has been indicated that the Atlantic water circulates cyclonic along the shelf slope (boundary) of the basin; however, little attention has been paid to the basin interior circulation. The present study shed light on the basin interior circulation and suggested that there would be some characteristic flow structures such as the transbasin westward flow.

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