Dynamics of the Coastal Oyashio and Its Seasonal Variation in a High-Resolution Western North Pacific Ocean Model

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ABSTRACT

The Coastal Oyashio (CO) carries the cold, fresh, and relatively light water mass called the Coastal Oyashio Water (COW) westward along the southeastern coast of Hokkaido in winter and spring. To investigate dynamics of the CO and its seasonal variation, model experiments are executed using a western North Pacific general circulation model with horizontal resolutions of approximately 2 and 6 km. The 2-km resolution model reproduces the properties of COW with temperature of 0–2°C and salinity of 32.2–32.6 and reproduces its distribution. COW is less dense than offshore water by 0.2 kg m⁻³, and it forms a surface-to-bottom density front with a width of 10 km near the shelf break. The CO appears as a baroclinic jet current along the front with a maximum velocity of approximately 40 cm s⁻¹. The velocity and density structures and the front location relative to bathymetry indicate that the CO can be understood in terms of a simplified dynamical model developed for the shelfbreak front in the Middle Atlantic Bight. In contrast to the 2-km resolution model, the 6-km model cannot realistically reproduce the COW distribution. This is because only the 2-km model can represent the sharp density structure of the shelfbreak front and the accompanying CO. The CO exists during the limited period from January to April. This is directly connected with seasonal variation of the COW inflow from the Okhotsk Sea to the North Pacific Ocean through the Nemuro and Kunashiri Straits, indicating that the seasonal variation of the CO is ultimately controlled by the variation of the circulation in the Okhotsk Sea induced by the monsoon.

1. Introduction

Seawater in the southeastern coastal region of Hokkaido, the northern island of Japan (Fig. 1), becomes remarkably cold and fresh compared to offshore water in winter and spring (Ohtani 1989). In observations executed across the southeastern region of Hokkaido (the Doto area), water with temperature less than 2°C and salinity less than 33.0 has been detected over the shelf almost always from January to April (Kono et al. 2004; Oguma et al. 2008; Kusaka et al. 2009). This water mass is called the Coastal Oyashio Water (COW), and its density of \( \approx 26.0 \rho_0 \) is lower than offshore surface water \( (> 26.2 \rho_0) \) because of its low salinity. Figure 2a shows sea surface temperature (SST) observed by a satellite on 22 February 2007. SST over the southeastern shelf is low (0°C–2°C) compared to offshore (1°C–6°C), indicating that COW is widespread over the shelf from the Kunashiri Strait to Cape Erimo.

It has been thought that most of COW originates in low-density surface water formed by ice melting and
river runoff in the Okhotsk Sea (e.g., Ohtani 1989). Previous studies unanimously reported that COW intruding into the Pacific Ocean from the Okhotsk Sea is transported westward along the southeastern coast of Hokkaido (the Doto coast), and the current carrying COW has been conventionally called the Coastal Oyashio (CO). However, a unified view has not been achieved with respect to the driving mechanism of the CO. Nakamura et al. (2003), who analyzed results of a general circulation model (GCM) with a horizontal resolution of $12 \text{ km}$, concluded that the CO is a coastally trapped current driven by horizontal density gradient and that it is distinguished from the Oyashio, the western boundary current of the subpolar gyre flowing southwestward above the shelf slope. In contrast, Kono et al. (2004), who diagnosed a velocity field based on a hydrographic observation with a horizontal interval of $10 \text{ km}$, argued that the CO is not a density current but a part of the Oyashio. More detailed analysis is necessary to reach a solid conclusion.

An important feature of COW is that COW occupies the whole water column on the shelf. As a result, near the shelf break, COW forms a surface-to-bottom temperature and salinity front with offshore warm and saline water, as seen in vertical profiles obtained across the shelf slope (Oguma et al. 2008). Because COW is less dense than the offshore water, this surface-to-bottom front is a density front. Chlorophyll-a concentration is observed to vary abruptly near the shelf break (Fig. 2b), also suggesting that a density front separates COW on the shelf from the offshore water. This front is expected to play an important role in formation of the CO.

A density front accompanied with a current is common in coastal regions of the World Ocean (Hetland and

Fig. 1. (a) Topography of the North Pacific Ocean. The domain of the figure indicates the outer model region, whereas the dashed rectangle is the nested model region. (b) Oceanic bathymetry near the southeast coast of Hokkaido. The values on the thin lines indicate depth (m). The thick line is a 150-m isobath, corresponding to the shelf break.

Fig. 2. (a) SST and (b) chlorophyll-a concentration observed by the satellite Aqua on 22 Feb 2007. The thick line indicates a 150-m isobath, corresponding to the shelf break. The arrows in (a) indicate characteristic perturbations of the shelfbreak front. The dataset is downloaded from the Web site of the Earth Observation Research Center (EORC), the Japan Aerospace Exploration Agency (JAXA), courtesy of JAXA.
Signell 2005; Brink et al. 2007; Spall et al. 2008). Among them, the situation of the Middle Atlantic Bight (MAB), a continental shelf from Massachusetts to North Carolina, is similar to that of the Doto area in that cold and freshwater is carried by a coastal current from high latitudes to form a surface-to-bottom density front near the shelf break (Linder and Gawarkiewicz 1998). Numerous high-resolution observational and model studies have been executed in the MAB to reveal that the front with a width of approximately 10 km controls the coastal current. Among the studies, the work of Chapman and Lentz (1994) is a milestone. They proposed a dynamical model as follows, by focusing on a role of downslope transport within the bottom boundary layer (BBL) in the adjustment process of a localized freshwater inflow from the coast.

First, the inflow of less dense water on the shelf forms an along-shelf current through the Coriolis effect. The along-shelf current forces offshore advection of buoyancy in the BBL, and then the density front is established. The front continues moving seaward until the along-shelf velocity within the front vanishes at the bottom because of thermal wind relation. At this point, the BBL cannot move the front farther seaward, so the steady front is achieved. The along-shelf current remains in the upper layer of the front, which is called a shelfbreak frontal jet. This mechanism (referred to here as the shelfbreak front model) has been verified through several observations (Barth et al. 1998; Houghton and Visbeck 1998; Pickart 2000; Fratantoni et al. 2001) and used as a theoretical base for studies of coastal currents (e.g., Yankovsky and Chapman 1997; Avicola and Huq 2002). Considering common characteristics between the Doto area and the MAB, it is expected that the CO could be understood as a shelfbreak frontal jet. If this speculation is correct, it is likely that previous observational or modeling studies with the horizontal resolutions of ~10 km could not capture the dynamical structure within the CO.

Meanwhile, there is a difference between the Doto area and the MAB. The fresh coastal water in the MAB is supplied from high latitudes throughout the year (Linder et al. 2004), whereas COW is observed only from January to April (Kono et al. 2004; Oguma et al. 2008). Although seasonal variations of the COW distribution and CO have been reproduced in a western North Pacific GCM to some extent, mechanisms controlling the variations have not been discussed thoroughly (Nakamura et al. 2003).

Our approach to the problems about dynamics of the Coastal Oyashio and its seasonal variation is to analyze results of a realistic GCM where the CO is reproduced. Because of a nesting method, the horizontal resolution of the GCM can be set so fine that the grid interval is approximately 2 km around Japan. This resolution turns out to be sufficient to represent some key dynamical features of the CO with a horizontal scale of 10 km, which has not been possible so far. The contribution of the high horizontal resolution to the model ability to reproduce the CO is examined through comparison with a lower-resolution model with a grid interval of ~6 km. We also have reproduced seasonal variation of the COW distribution using climatological forcings of a repeated annual cycle to investigate mechanisms controlling seasonal variation of the CO.

2. Model and experimental design

The GCM used is the Meteorological Research Institute Community Ocean Model (MRI.COM; Ishikawa et al. 2005). The MRI.COM is a z-coordinate multilevel model that solves the primitive equations under the hydrostatic and Boussinesq approximations. It adopts an Arakawa B-grid arrangement (Arakawa 1972), and coastlines are created by connecting tracer points instead of velocity points. The vertical coordinate near the surface follows the surface topography like sigma coordinate models.

The model domain is the western North Pacific, spanning from 117°E to 125°W and from 15° to 65°N (Fig. 1a). The horizontal resolution is \( \frac{1}{36} \) \( ^\circ \) (\( \frac{1}{60} \) \( ^\circ \)) in the zonal (meridional) direction, corresponding to approximately 2 km. The model has 50 levels in the vertical direction, with thickness increasing from 4 m at the surface to 600 m at 6300-m depth.

The quadratic upstream interpolation for convective kinematics with estimated streaming terms (QUICKEST; Leonard 1979) and the uniformly third-order polynomial interpolation algorithm (UTOPIA; Leonard et al. 1993) are used as the vertical and horizontal tracer advection schemes, respectively, whereas the Arakawa scheme (Arakawa 1972) is used for the momentum advection. A biharmonic operator is used for horizontal turbulent mixing of tracers with a diffusivity coefficient of \( 1.5 \times 10^8 \) m\(^4\) s\(^{-1}\). A biharmonic friction with a Smagorinsky-like viscosity (Griffies and Hallberg 2000) is used for momentum with a scaling constant \( C = 2.5 \) in their notation. The vertical viscosity and diffusivity are determined by the turbulence closure level 2.5 scheme (Mellor and Yamada 1982). The parameterization of St. Laurent et al. (2002) is used as background diffusion to include the effect of tidally driven mixing over rough topography. When the static instability occurs, the vertical diffusivity is set up to 1.0 m\(^2\) s\(^{-1}\) to eliminate the local inversion. Sea ice is also incorporated. Thermodynamics are based on the work by Mellor and Kantha (1989), and dynamics are based on an elastic–viscous–plastic rheology of Hunke and Ducowicz (1997, 2002).
The model is driven using the normal year forcing of the Coordinated Ocean-Ice Reference Experiments (COREs; Large and Yeager 2004; Griffies et al. 2009). This dataset is designed for running ocean-ice models by merging and correcting various reanalysis and remote sensing data products. It consists of climatological annual-mean continental runoff, monthly varying precipitation, daily varying shortwave and longwave radiative fluxes, and 6-hourly varying meteorological fields. They are linearly interpolated in time within the model. Wind stress, sensible heat flux, evaporation, and accompanying latent heat flux are calculated using the bulk formulas of Large and Yeager (2004). Freshwater fluxes are converted to equivalent salinity fluxes. Sea surface salinity (SSS) is also restored toward the monthly climatology from the World Ocean Atlas 2001 (WOA01; Conkright et al. 2002) with a piston velocity of 4 m per 4 days to prevent model drift. No tracer flux is imposed at the bottom boundary. The bottom friction is represented by the quadratic friction with a drag coefficient of $1.225 \times 10^{-3}$.

The model is nested within the outer model using a one-way nesting method presented in Tsujino et al. (2006). The outer model domain is the Pacific Ocean north of 15°S from 100°E to 75°W (Fig. 1a), and the horizontal resolution is $1/6^\circ (1/18^\circ)$ in the zonal (meridional) direction. At the southern edge, temperature and salinity are restored to the climatology of WOA01 with a restoring time of 30 days. To save computational resources, the horizontal resolution of the inner model is set coarse near the lateral boundary (three times at most), whereas the horizontal diffusivity is increased (1.0 \times 10^8 m^4 s^{-1}). The boundaries are so far from Hokkaido (Fig. 1a) that they could hardly affect the Coastal Oyashio.

The outer model is integrated for 35 yr from a state of rest with an initial stratification derived from WOA01 with a restoring time of 30 days. To save computational resources, the horizontal resolution of the inner model is set coarse near the lateral boundary (three times at most), whereas the horizontal diffusivity is increased (1.0 \times 10^8 m^4 s^{-1}). The boundaries are so far from Hokkaido (Fig. 1a) that they could hardly affect the Coastal Oyashio.

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To illustrate a typical ocean state in the Doto area of the 2-km model, temperature $T$, salinity $S$, and potential density $\sigma_T$ at the sea surface on 10 February are shown in Fig. 3. In the shelf region, there is a cold and freshwater mass (0°C $< T < 2°C$, 32.2 $< S < 32.6$). In the offshore region, there are two types of water, a warm and saline water (8°C $< T$, 33.5 $< S$) to the west of 146°E and a relatively cold and saline water (1°C $< T < 5°C$, 32.6 $< S < 33.3$) to the east. The former is a warm eddy existing since autumn, whereas the latter is the Oyashio water transported from high latitudes by the Oyashio. Density of the shelf water is low (25.8 $< \sigma_T < 26.0$) compared to the offshore Oyashio water (26.2 $< \sigma_T < 26.2$) and the warm eddy (26.0 $< \sigma_T < 26.2$).

The shelf and offshore waters correspond well to previous observations with respect to both the properties and the distributions (Kono et al. 2004; Oguma et al. 2008). The model SST is also consistent with the satellite observation shown in Fig. 2a, except for relatively high $T$ of the offshore warm eddy ($\sim 10°C$). The shelf water moves westward along the coast beyond Cape Erimo to intrude into the Hidaka Bay, as observed by Kuroda et al. (2006). Sea ice flows out from the Okhotsk Sea into the Pacific Ocean through southern straits of the Kuril Islands during several days in February, which is also consistent with observation (JMA 1998).

In the 2-km model, the cold and fresh shelf water remains on the shelf; as a result, the sharp front between the shelf and offshore waters is formed near the shelf break. For the first time by a GCM, the sharp front is reproduced as inferred from satellite images of SST and chlorophyll-a concentration (Fig. 2). The front is a density front with a density difference across the front of approximately 0.2 kg m$^{-3}$ (Fig. 3c), which is consistent with Oguma et al. (2008). The front varies actively with a period of several days and a horizontal scale of several tens of kilometers. For example, Fig. 3b shows that the front meanders with a wavelength of 30–50 km near 144°E and this meander develops into an isolated small eddy near 146°E. Similar undulating fluctuations of the front are found in the satellite SST (the arrows in Fig. 2a), also suggesting that the 2-km model reproduces small-scale phenomena characteristic to coastal regions.

COW is defined as a water with $T < 2°C$ and $S < 32.6$ in the present paper. These thresholds were chosen so that the boundary of COW coincides with the front location (Fig. 3). Potential density of COW defined here corresponds roughly to $\sigma_T < 26.05$. Our definition is not far from a conventional one of $T < 2°C$ and $S < 33.0$ (e.g., Oguma et al. 2008). Note that SSS in the Doto area

3. Simulated features of the Coastal Oyashio Water

In addition to the main experiment, we execute an experiment using a model with a horizontal resolution of three times as coarse, that is, $1/12^\circ (1/36^\circ)$ in the zonal (meridional) direction, corresponding to approximately 6 km. The settings of the low-resolution model are same as the original model, except for increasing the coefficient of the biharmonic horizontal diffusivity to $1.0 \times 10^7 m^4 s^{-1}$ in order to suppress grid-scale numerical instabilities. Hereafter, we refer to the high-resolution (low-resolution) model as the 2-km (6 km) model.
Fig. 3. Instantaneous fields of (a) temperature, (b) salinity, and (c) density at the sea surface on 10 Feb (2-km model). The thick solid and dashed lines are the shelf break (150-m isobath) and boundary of COW, respectively. The arrows in (b) indicate perturbations of the shelfbreak front. The 146.5°E transect for Fig. 9 is also shown in white in (b).
is restored toward the value larger than 32.6 because the Levitus data do not resolve this low salinity signal, indicating that COW there is not generated by the local salinity restoring of the model.

Figure 4 shows SST and SSS on the same day as Fig. 3 in the 6-km model. For the 6-km model, the water-mass distributions in the Doto area are basically same as for the 2-km model. COW with $T < 2^\circ \text{C}$ and $S < 32.6$ exists on the shelf, whereas relatively warmer and more saline water originating in the Oyashio is found offshore (there is not a warm eddy at this time in the 6-km model). However, the 6-km model (Fig. 4) is rather different from the 2-km model (Fig. 3) with respect to the COW distribution. COW does not remain on the shelf, but flows offshore extensively beyond the shelf break in the west of 146$^\circ \text{E}$. As a result, a sharp front is not found near the shelf break, in contrast to the 2-km model. In addition, undulating perturbations with a small scale do not appear. These differences indicate that COW and the accompanied front are not reproduced as realistically in the 6-km model as they are in the 2-km model. Moreover, the westward migration of COW stops at Cape Erimo, so COW is hardly found in the Hidaka Bay. This is also different from the 2-km model, which shows realistic intrusion of COW into the Hidaka Bay.

The results of the 2-km model indicate that COW and its sharp front are well reproduced in a realistic model for the first time, to the best of our knowledge. In the next section, we analyze the physical field to investigate dynamics of the CO. Meanwhile, reproducibility of the 6-km model is not satisfactory with respect to the COW distribution. Through comparison between the two models, we also elucidate what kind of dynamical processes need to be represented to reproduce realistically oceanographic features in the Doto area.

4. Dynamics of the Coastal Oyashio

a. Application of a dynamical model for the shelfbreak frontal jet

A surface velocity field around Hokkaido for the 2-km model is shown in Fig. 5a, where contours of sea surface height (SSH) are superposed to show roughly the direction of surface currents (geostrophic currents). The Oyashio, flowing southwestward with a width of $\sim 100 \text{ km}$ south of the Kuril Islands, is most dominant in this region. The Oyashio turns to the south near 43$^\circ \text{N}$, combining a current into the Pacific through the Etorofu Strait. Meanwhile, a contour line of SSH extends along the Doto coast from the Kunashiri Strait to Cape Erimo, not coalescing with the Oyashio, and the speed field (shades) shows a narrow band of local maximum of $\sim 30 \text{ cm s}^{-1}$ near the shelf break. That is, the CO is trapped along the shelf break and distinguished from the Oyashio.

It should be noted that the latitude of the Oyashio’s veering changes seasonally and interannually, in similar fashion to the real ocean (Isoguchi and Kawamura 2006). Actually, in some other years of the experiment, the Oyashio flows closer to the Doto coast and partly overlaps the CO (even then, the Oyashio and CO can be distinguished in the 2-km model because of the high horizontal resolution). The final year of the experiment, which is used in this study, might be suited for analysis of the CO, because the Oyashio is relatively far from the Doto coast.

Velocity and density fields are expanded near the shelf break to focus on the CO structure (Fig. 5b). A density front exists near the shelf break as noted in section 3, and the CO flows westward following the front. The CO has a jet structure with a maximum velocity of approximately 40 cm s$^{-1}$ and a width of 8–10 km (based on a half of the core velocity; Linder and Gawarkiewicz 1998). Apart
FIG. 5. Velocity fields at the sea surface averaged over 10–14 Feb (2-km model). (a) Shades, contours, and the arrows on the contours indicate speed, SSH (the contour interval is 10 cm), and direction of geostrophic currents, respectively. (b) The box region in (a) is expanded. Shades, contours, and vectors indicate speed, sea surface density (the contour interval is 0.05), and velocity, respectively. The velocity vectors are shown at every grid of the model, and the meridional dashed line is the transect for Fig. 6. In both (a) and (b), the thick line is the shelf break (150-m isobath).
from the CO, a westward current with a relatively small speed of 10–20 cm $s^{-1}$ is found on the shelf. This current might be generated in the formation process of the shelfbreak front as noted later, or it might be relevant to a barotropic current formed by confluence of COW from the Nemuro Strait to the Doto area. In fact, the surface speed on the shelf is more than 10 cm $s^{-1}$ in the confluent zone of 144°–146°E but less than 10 cm $s^{-1}$ in the other shelf regions (Fig. 5a). In each case, the CO concentrating in the front is dominant in the coastal region. To focus on the basic dynamics of the CO, 5-daily averages are used in this section.

The CO is represented as a shelfbreak-trapped current accompanied with a density front in the 2-km model; however, it was recognized as a barotropic current on the shelf in some observational studies. For example, Kusaka et al. (2009) stated that the CO is a barotropic current with a speed of ~20 cm $s^{-1}$, based on an observation with a horizontal interval of ~10 km. On the other hand, Kuroda et al. (2006) carried out a surface velocity observation with a close horizontal interval of ~2 km and detected a shelfbreak-trapped current with a width of ~10 km and a maximum velocity of 40 cm $s^{-1}$. Comparing the present results with previous observations, it is suggested that a resolution of a few kilometers is necessary to capture the jet structure of the CO.

Figure 6 shows typical velocity and density structures of the CO in a vertical section. As suggested by previous observations (e.g., Oguma et al. 2008), the density front found at the sea surface intersects the sea floor near the shelf break. The CO is a westward baroclinic jet and is concentrated in the upper layer of the surface-to-bottom front. Because of the horizontal density gradient in the front (the thermal wind relation), the westward velocity weakens with depth to almost vanish at the bottom of the front (<10 cm $s^{-1}$). Temperature and salinity shoreward of the front are homogenized vertically and horizontally so that the current is barotropic. As an exception, though ambiguous, a weak current carries the shelf water downslope near the bottom, which might be attributed to the effect of a bottom friction. The downslope current near the bottom stops at the front (42.75°N), where the alongshore current (and the bottom friction) vanishes. These features in the velocity and density structures coincide with the shelfbreak front model proposed by Chapman and Lentz (1994), indicating that the CO can be recognized as a shelfbreak frontal jet. The upwelling from the bottom to the sea surface in the onshore side of the front (Fig. 6) indicates detachment of the BBL, which has been often reported as an evidence of the shelfbreak front model (e.g., Barth et al. 1998).

Chapman (2000) proposed a theoretical range of the water depth $h_{\text{front}}$ at which the shelfbreak front is trapped, $c h_b < h_{\text{front}} < h_b$, based on numerical experiments under ideal settings of buoyant inflows on the shelf (corresponding to COW in the present situation), where the coefficient $c$ is the lower limit. The depth $h_b$ is given by

$$h_b = \left(\frac{2Vf\rho_0}{g\Delta\rho}\right)^{1/2},$$

where $V$ is the volume transport rate of the buoyant inflow, $\Delta\rho$ is the density difference across the front, $f$ is the Coriolis parameter, $\rho_0 = 1.0 \times 10^3$ kg m$^{-3}$ is the standard density, and $g = 9.8$ m $s^{-2}$ is the gravitational acceleration. The coefficient $c$ depends on offshore stratification [the dependency is shown in Fig. 12 of Chapman (2000)], and $c \approx 0.9$ for the Doto area (buoyancy frequency of $\sim 2 \times 10^{-3}$ s$^{-1}$). Considering the situation of Fig. 6 and regarding $V$ as the transport rate of COW across the transect, $h_{\text{front}} = 180$ m [$V = 0.32$ Sv (1 Sv = $10^6$ m$^3$ s$^{-1}$) = $0.32 \times 10^6$ m$^3$ $s^{-1}$, $f = 1.0 \times 10^{-4}$ s$^{-1}$, and $\Delta\rho = 0.2$ kg m$^{-3}$], which indicates that the front would be trapped at the water depth of 160–180 m. The front in the model is found to be located at the water depth of 140–150 m, which is close to the theoretical range though a little shallower.

Equation (1) indicates that the water depth of the front location becomes larger (i.e., the front is located more offshore) as $V$ increases. To verify it, we examine the velocity and density profiles of the CO at upstream and downstream transects, where $V$ is different (Fig. 7). Note that, instead of the zonal and meridional velocity components ($u$, $v$), Fig. 7 shows the along- and cross-shelf components ($u_s$, $v_s$) defined by

$$(u_s, v_s) = (-un_y + un_x, un_x + vn_y),$$

where $n = (n_x, n_y)$ is the unit vector in the downslope direction (positive $u_s$ to the direction with the coast on the right while positive $v_s$ offshore). At the upstream transect (145.5°E; Fig. 7a), where $V = 0.46$ Sv is larger than 0.32 Sv at 144.7°E (Fig. 6), the front is located more offshore at the water depth of 180 m. At the downstream transect with $V = 0.11$ Sv (42.0°N; Fig. 7b), the front moves onshore at the water depth of 100 m. This displacement of the front is consistent with Eq. (1). The front tends to be located a little onshore (shallower by 20 m at most) compared to the theoretical range (the depths of 180–200 and 90–100 m at the upstream and downstream transects, respectively), as also seen at the 144.7°E transect (Fig. 6). This may be because the front does not achieve the final equilibrium state, as discussed later. Nevertheless, it can be concluded that the location of the front is in good agreement with the shelfbreak
front theory, considering that the deviations from the theoretical range are less than 10% of the total depth (10–20 m). The vertical circulations on the shelf shown in Fig. 7 have basically the same structure as in Fig. 6, except that downslope transport in the BBL is found more clearly in Fig. 7a.

A key process in the shelfbreak front model is that the low-density water on the shelf is carried downslope within the BBL to the front location. In the present results, downslope currents near the bottom are found on the shelf at most of transects; additionally, upwelling on the shore side of the front suggests the downslope transport in the lower layer. However, the BBL current is often ambiguous in velocity profiles (Figs. 6, 7b), though appeared clearly sometimes (Fig. 7a). To investigate the effect of the bottom friction on downslope transport over the entire CO region, Fig. 8a shows downslope transport within the bottom layer with a height of 20 m (\(V_{BBL}\)), where 20 m is the Ekman layer thickness for the vertical viscosity of \(\nu = 0.02 \text{ m}^2 \text{s}^{-1}\) predicted in this region. The transport is downslope almost everywhere on the shelf but rarely in the offshore of the shelf break. That is, seawater on the shelf moves downslope to the front location within the bottom layer.

For comparison, the bottom Ekman transport \(V_{Ek}\) is estimated following Eq. (4.3.32) in Pedlosky (1987),

\[
V_{Ek} = \left( \frac{\tau_b}{\rho_0 f} \right) \cdot n,
\]

where the lateral stress on the stepwise slope is included in the stress \(\tau_b\), in addition to the bottom stress. Figure 8b shows that \(V_{Ek}\) is large on the shelf but vanishes offshore of the shelf break. Thus, the distribution of \(V_{BBL}\) is similar to \(V_{Ek}\), though \(V_{BBL}\) is somewhat smaller than \(V_{Ek}\) possibly because of the rough estimate of 20 m for the BBL thickness. This similarity implies that the present model reproduces to some extent the downslope transport forced by the bottom friction, which is essential to formation of the shelfbreak front. Because the downward transport is considerably small in terms of velocity (1–4 cm s\(^{-1}\)), it seems ambiguous at each transects, as Houghton (1997) noted.

Because density is almost homogenized horizontally and vertically on the shelf, the downslope \(V_{BBL}\) corresponds to the barotropic current flowing with the coast on the right as shown in Fig. 6. After numerical studies about the shelfbreak front (e.g., Chapman and Lentz...
1994), most of the inflow water is transported by a baroclinic jet within the front after the shelfbreak front reaches the final equilibrium state, whereas a weak barotropic current is maintained on the shelf before that. Considering that a time scale of 100 days was necessary to reach the equilibrium state in the numerical experiments, the CO is possibly in the adjustment process because the COW input starts in January (section 5). Therefore, the barotropic current and $V_{BBL}$ on the shelf are interpreted as a part of the formation process of the shelfbreak front. This interpretation is also supported by the result that the water depth of the front location is a little shallower than the theoretical range ($ch_b-h_b$) resulting in $V_{BBL}$ gradually decreasing with time (not shown). Further discussion of the adjustment process might be needed for the CO because of its seasonality, though this has been hardly referred to because the shelf water inflow occurs throughout the year in MAB.

The results of Avicola and Huq (2002) are instructive to elucidate the reason why the shelfbreak front is formed along the Doto coast in winter. Based on water tank experiments, they derived a condition for formation of the shelfbreak front by using $h_b$ given by Eq. (1) and the water depth at the distance $R$ offshore from the coast $H$, where $R$ is the baroclinic deformation radius for the depth $h_b$:

$$ R = \frac{1}{f} \left( \frac{g \Delta \rho h_b}{\rho_0} \right)^{1/2}. $$

That is, the shelfbreak front is formed when the depth scale of the buoyant coastal current $h_b$ exceeds the ambient ocean depth $H(h_b/H > 1)$. Conversely, when $h_b/H < 1$, a surface-to-bottom front is not formed, because the coastal current is not in contact with the bottom.

![Fig. 7. Velocity fields as in Fig. 6 at (a) upstream (145.5°E) and (b) downstream (42.0°N). Note that the velocity components $u_0$ and $v_0$ are shown instead of $u$ and $v$.](image)

![Fig. 8. (a) Downslope transport below 20 m above bottom $T_{BBL}$ and (b) downslope Ekman transport $T_{Ek}$, averaged over 10–14 Feb (2-km model).](image)
Applying this condition to the 144.7°E transect of the CO in the model (Fig. 6; $\Delta \rho = 0.2$ kg m$^{-3}$, $f = 1.0 \times 10^{-4}$ s$^{-1}$, and $h_b = 180$ m), we obtain $R = 6.0$ km, $H = 60$ m, and $h_b/H = 3.0$. For the 145.5°E transect, $R = 6.3$ km, $H = 60$ m, and $h_b/H = 3.3$ (Fig. 7a), whereas $R = 4.5$ km, $H = 50$ m and $h_b/H = 2.0$ for 42.0°F (Fig. 7b). Thus, $h_b/H$ is larger than unity at each transect, indicating that the transport and the density anomaly of COW satisfy the condition for formation of the shelfbreak front. The application of Avicola and Huq (2002) gives the reason why the CO appears as a shelfbreak frontal jet.

Undulating perturbations of the front are found in the 2-km model (Fig. 3b). Such perturbations are expected to be induced by instabilities of the front, because the wavelength of 30–50 km and the phase speed of ~20 cm s$^{-1}$ are close to observational values of 40 km and 11 cm s$^{-1}$, respectively, for instabilities of the MAB shelfbreak front (Gawarkiewicz et al. 2004). Perturbations often develop into detached eddies and make contribution to outflow of COW over the shelf break. Dynamics and influence of perturbations of the front should be investigated in future, preferably using a higher-resolution model.

b. Dependence on the horizontal resolution

As compared to the 2-km model, the 6-km model cannot reproduce the COW distribution realistically. This is expected to be attributed to difference in representations of the CO, because COW on the shelf is transported by the CO. Figure 9 shows the representations of the CO and the shelfbreak front at 146.5°E for the two models on 10 February. Because this transect is in the upstream region of the CO (Figs. 3c, 4), the COW distributions there seem still similar. In both models, COW on the shelf forms a density front with offshore water near the shelf break, and isopycnals within the front intersects both the sea surface and the bottom. However, sharpness of the front is quite different, because the front width is approximately 10 km for the 2-km model but 30 km for the 6-km model (about five grids for both of the models). Furthermore, the CO is found in the upper layer of the front for the 2-km model, whereas a coastal current distinguished from the Oyashio is not found for the 6-km model. As a result, COW in the 6-km model is transported westward by the Oyashio, not by the CO.

The differences in representing the shelfbreak front and the CO can be attributed to the relative length of the horizontal grid interval against the internal deformation radius. In the shelfbreak front, the density profile (tilting of isopycnals) corresponds to that formed by geostrophic adjustment, and the width scale of the front is given by the internal deformation radius $R$ of Eq. (4) (Chapman and Lentz 1994; Avicola and Huq 2002). This implies that a model resolution needs to be sufficient to resolve $R$ to represent the shelfbreak front and the CO. Because $R$ is 5–6 km for the CO, the sharp shelfbreak front can be represented in the 2-km model but not in the 6-km model. In fact, the isopycnals in the front are tilting for the 2-km model but almost vertical for the 6-km model (Fig. 9).

To illustrate the difference in representing the shelfbreak front in the Doto area, magnitude of horizontal density gradient $\rho_h$ is shown in Fig. 10, normalized by the density gradient scale of the shelfbreak front $\Delta \rho/R$. The density gradient $\rho_h$ is defined by

$$
\rho_h = \sqrt{\left(\frac{\partial \rho}{\partial x}\right)^2 + \left(\frac{\partial \rho}{\partial y}\right)^2},
$$

and typical values of $R = 5$ km and $\Delta \rho = 0.2$ kg m$^{-3}$ are used. The normalized value $\rho_h/(\Delta \rho/R)$ reaches unity along the shelfbreak front for the 2-km model, indicating that the front can be represented with inherent sharpness. In contrast, the value is less than 0.5 for the 6-km model. Thus, although basic characteristics of COW and the offshore water are the same in the two models (Figs. 3, 4), the boundary between them is represented differently depending on the horizontal resolution. This difference in model ability leads to the difference in representing the CO. It should be noted that, also in the offshore region, the 2-km model shows sharp density fronts, which are not seen in the 6-km model. This may be because the horizontal resolution of 2 km makes it possible to reproduce frontogenesis processes in the mixed layer (Capet et al. 2008a,b).

5. Seasonal variation

The 2-km model results indicate that the CO can be understood as a shelfbreak frontal jet, as the coastal current in the MAB is. Previous observations in the Doto area (e.g., Oguma et al. 2008) suggest that the shelfbreak front exists only when COW appears on the shelf from winter to spring. This is in contrast to the case of MAB, where the front has been observed throughout the year (Linder et al. 2004).

Seasonal variation of the COW distribution is illustrated in Fig. 11, showing monthly averaged fields of SST and SSS from November to May. In November, COW is absent so that SST and SSS over the shelf is hardly different from offshore. COW of $T < 2.0°C$ and $S < 32.6$ appears south of the Nemuro and Kunashiri Straits in January and spreads over the shelf of the Doto area to reach Cape Erimo in March. Subsequently, COW disappears by May, except for the narrow band near the shelf break. The shelfbreak front and the associated CO also appear in January and disappear in May.
The satellite SST observation is shown in Figs. 11i–l to verify the seasonal evolution of the 2-km model, although it is difficult to distinguish clearly between COW and the offshore Oyashio water using SST (Murakami 1984). As a whole, the seasonal evolution of the SST observation is quite similar to that in the 2-km model, except that a warm eddy is found offshore in May in the observation but from January to March in the model. With respect to the COW distribution, COW appears from the north in January, spreads over the Doto shelf region in March, and disappears in May, as seen in the model. It is also similar that a relatively cold water band remains near the shelf break in May. The seasonal evolution of the model is consistent with Rosa et al. (2007) and Kusaka et al. (2009), both of whom compiled long-term observations.

As a first step to investigate mechanisms controlling the seasonal variations of the COW distribution and the front, we identify the origin of COW, using the 2-km model results. The SST and SSS variations shown in Fig. 11 imply that COW intrudes from the Okhotsk Sea into the North Pacific Ocean through the Nemuro and Kunashiri Straits. To confirm this hypothesis, we inject a passive tracer with concentration $C$ of 1.0 into COW passing through the two straits from 1 January and calculate its time evolution using the advection–diffusion equation of the model (see appendix for settings of the tracer experiment). Figure 12 shows $C$ at the sea surface on 14 March. The concentration $C$ is more than 0.9 over the southeastern shelf, indicating that most of COW there originates in the two straits. This means that the CO is driven by the COW inflow through the two straits, because the shelfbreak frontal jet is essentially driven by the buoyant inflow on the shelf (Chapman and Lentz 1994; Avicola and Huq 2002). Though some COW flows into the Pacific Ocean through the Etorofu Strait in
March, the water is mixed with ambient water too soon to reach the shelf.

The COW origin is also confirmed in terms of spatial variation of the CO transport rate $V$. We estimate $V$ across the two straits and the four sections in the Doto area (Fig. 12). The total inflow rate from the Okhotsk Sea is 0.51 Sv on average from January to April, with more than 90% passing through the Kunashiri Strait (0.04 Sv through the Nemuro Strait and 0.47 Sv through the Kunashiri Strait). Meanwhile, $V$ decreases with distance from the straits from 0.33 to 0.12 Sv along the Doto coast [the value of 0.12 Sv at Cape Erimo is small compared to previous estimates of 0.2–0.5 Sv by Murakami (1984) and Shimizu and Isoda (1999), possibly because of our strict definition of COW]. This result confirms the implication of the tracer experiment that there is little inflow from offshore to the shelf and that most of COW originates in the two straits. For the decrease of $V$, two
reasons are cited. One reason is that COW spills out from the shelf as a result of the development of small-scale perturbations of the front (Fig. 3b). Another reason is the seasonal variation of the CO; that is, COW is used to replace water on the shelf in the developing stage of the shelfbreak front so that $V$ decreases at the head (downstream) of the CO.

Seasonal variation of the COW inflow through the two straits is shown by the thick line in Fig. 13. The COW inflow begins in early January. The inflow rate stays at 0.5 Sv until April and vanishes in May. Though somewhat smaller, $V$ at 145.5°E (the thin line) follows the COW inflow rate well, indicating that the seasonal variation of the CO is controlled by the COW inflow through the two straits. In addition, the COW inflow rate places an upper limit on $V$ because of little water inflow from offshore. This result implies that the COW inflow through the two straits, which is regarded an external condition for the CO, determines the location of the CO as well as its existence, because the front location is given by the transport $V$ and the density anomaly $\Delta \rho$ of COW as explained in section 4a.

The seasonal variation of the COW inflow can be explained qualitatively in terms of circulation in the Okhotsk Sea, which has been revealed by previous studies. In the model, water inflow from the Okhotsk Sea into the Pacific Ocean through the two straits occurs also in summer and autumn, as shown by the total water inflow rate including COW in Fig. 13. However, in these seasons, the inflow water is the warm and saline water mass originating in the Soya Warm Current as reported by Nakamura et al. (2003). In early January, the water switches abruptly from the warm and saline water to the cold and freshwater originating in the East Sakhalin Current (ESC; i.e., COW), leading to formation of the surface-to-bottom front along the Doto coast. This switch of the water masses near the straits has been observed by Itoh and Ohshima (2000) and attributed to seasonal variation of the monsoon over the Okhotsk Sea (Simizu and Ohshima 2002, 2006). Therefore, the present results mean that the Doto area responds to the monsoon over the Okhotsk Sea through the mechanism of the shelfbreak front. This conclusion is consistent with Tsujino et al. (2008), who analyzed sea surface elevation around Hokkaido to show that the monsoon controls coastal currents around Hokkaido.

Although the CO is found to be related to circulation in the Okhotsk Sea, a question about formation of COW remains. It has been already revealed based on water properties that COW originates in the East Sakhalin Current Water (ESCW), but the potential vorticity $q$ is different between the two water masses. ESCW is strongly stratified, which leads to large $q$, whereas COW at the Doto coast is nearly homogenized so that $q$ is small (Kusaka et al. 2009). In the 2-km model, COW on the shelf is already vertically homogenized and accompanied
with small \( q \) at the Kunashiri Strait, through which most of COW flows into the Pacific Ocean. The modification process from ESCW to COW in the Okhotsk Sea is investigated using the 2-km model as a final analysis in this section.

Figures 14a–c show SSH and vertical sections of \( T \) and \( S \) in the southern Okhotsk Sea. A part of ESC advances southward near 144\(^\circ\)E, turns to the east at latitudes lower than 45\(^\circ\)N, and flows out through the Kunashiri Strait to subsequently become the CO (Fig. 14a). Near 144\(^\circ\)E, ESCW is located in the surface layer with a thickness of 40 m and is very cold and fresh with \( T, 0.8^\circ\)C and \( S, 32.1 \) (Figs. 14b,c). The ESCW layer gradually gains thickness in moving eastward, as the mixed layer depth, which is defined by the density difference of 0.05 kg m\(^{-3}\) from the sea surface, increases from approximately 50 m at 144\(^\circ\)E to 80 m near the Kunashiri Strait. The deepening of the mixed layer is thought to be attributed to sea surface cooling and a combination effect of the northwesterly wind of the monsoon and the density front accompanied with ESC (the downfront wind effect; Thomas 2005). The increases of \( T \) and \( S \) in the mixed layer from 0\(^\circ\) to 0.4\(^\circ\)C and from 32.1 to 32.2, respectively, reflect entraining of relatively warm and salty water located below. With respect to stratification, it is also characteristic that isothermals and isohalines are nearly vertical in the bottom layer at the Kunashiri Strait, the lower part of COW (60–120 m at 146.7\(^\circ\)E; Figs. 14b,c). This is interpreted as a result of the formation mechanism of the shelfbreak front. That is, low-density water is transported downslope within the BBL on the shelf, resulting in destruction of stratification.

Figure 14d shows the potential vorticity \( q \) given by

\[
q = -\frac{f + \zeta}{\rho_0} \frac{\partial p}{\partial z},
\]

where \( \zeta \) is the relative vorticity. Near 144\(^\circ\)E, warm and salty water exists in the lower layer on the shelf, leading to strong stratification. As a result, the low potential vorticity water of \( q < 1.0 \) exists only above the 20-m depth. To the east of 144\(^\circ\)E, \( q \) decreases gradually in the subsurface layer with a depth of 20–100 m, corresponding to the deepening of the mixed layer. Furthermore, \( q \) decreases rapidly at the bottom of the Kunashiri Strait because of the destruction of stratification. Therefore, it is concluded that modification from ESCW with large \( q \) to COW with small \( q \) consists of the two processes, at least: deepening of the mixed layer in the Okhotsk Sea and bottom mixing at the Kunashiri Strait.

6. Conclusions and discussion

Results of two numerical experiments using a western North Pacific GCM with horizontal resolutions of approximately 2 and 6 km are used to investigate dynamics and seasonal variation of the Coastal Oyashio (CO), which is the coastal current appearing in the Doto area from winter to spring. In the 2-km resolution model, realistic properties and distributions of water masses are reproduced: that is, the Coastal Oyashio Water (COW) with temperature less than 2\(^\circ\)C, salinity less than 32.6, and potential density less than 26.05 kg m\(^{-3}\) on the shelf and warmer, more saline, and denser waters in the offshore region. As a result, a sharp surface-to-bottom density front with a width of 10 km and a potential density difference of 0.2 kg m\(^{-3}\) is formed near the shelf break between COW and the offshore waters. The CO is reproduced as a baroclinic jet along the front with a maximum velocity of approximately 40 cm s\(^{-1}\), which is distinguished from the Oyashio offshore. Examining thoroughly the velocity and density structures, the front location (water depth), and downslope transport in the BBL, we have found that the density front and the CO can be understood in terms of the simplified dynamical model proposed by Chapman and Lentz (1994) called the shelfbreak front model. We have also found that the transport rate and the density anomaly of COW satisfy the condition for formation of the shelfbreak front (Avicola and Huq 2002).

Comparison between the two models with horizontal resolutions of approximately 2 and 6 km implies that the
horizontal resolution should be higher than the internal deformation radius of the shelfbreak front to represent the sharp frontal structure and the accompanying CO. In the 6-km model, whose grid interval is insufficient to resolve the deformation radius, the CO cannot be reproduced as a shelfbreak frontal jet, leading to unrealistic offshore outflow of COW across the shelf break. The result that the increase in the model resolution improves representation of water-mass distributions in the Doto area suggests that it would be also effective for modeling biological production there.

The origin of COW and the mechanism of the seasonal variation of the CO are revealed based on the 2-km model results. COW is formed through vertical mixing processes of the East Sakhalin Current Water in the southern Okhotsk Sea. Subsequently, COW flows into the Pacific Ocean through the Nemuro and Kunashiri Straits from January to April with an inflow rate of 0.5 Sv. Corresponding to the seasonal variation of the COW inflow, the CO appears in January and disappears in May. Thus, the present results show that the COW inflow and CO are related to circulation in the Okhotsk Sea, which suggests that the seasonal variation of the CO is eventually controlled by the monsoon over the Okhotsk Sea.

To clarify the present achievements, the results about the CO dynamics are compared with previous studies. Tsuji et al. (2008) postulated that the CO is formed by southward propagation of the arrested topographic
wave (ATW) excited by the monsoon in the Okhotsk Sea. However, Chapman and Lentz (1994) pointed out that the along-shelf current formed by the ATW would leak from the shelf in the downstream region because of downslope transport in the BBL. The shelfbreak front model solved this problem by incorporating buoyancy advection to provide a mechanism confining the along-shelf current on the shelf. From this viewpoint, the result that the CO is regarded a shelfbreak frontal jet complements Tsujino et al. (2008), who insisted that wind signals in the Okhotsk Sea are propagated along the Hokkaido coast. Nakamura et al. (2003) reported that the CO is a current accompanied with horizontal density gradient near the coast, based on a GCM simulation with a horizontal resolution of approximately 12 km. This is consistent with the present results, but a sharp frontal structure was not found and the CO was not reproduced as a shelfbreak frontal jet in their model. Using the model with a resolution 6 times as fine as their model, we have clarified the dynamics of the CO. Kono et al. (2004) stated that the CO is not an independent coastal current and that COW is transported by the Oyashio over the slope, based on a hydrographic observation in April 1989. This conclusion is inconsistent with our results, possibly because the CO was difficult to be identified at that time because of approaching of the Oyashio and/or because their observational resolution (only one station on the shelf) was insufficient to capture the horizontal density gradient of the shelfbreak front.

It is worth while to compare the CO with the coastal current in the Middle Atlantic Bight (MAB), because both of them are coastal currents driven by freshwater from high latitudes. The coastal current in the MAB is a baroclinic jet with a width of 15–20 km and a maximum velocity of 20–30 cm s^{-1} accompanied with a density front located at a water depth of 150 m (Linder and Gawarkiewicz 1998). These characteristics are very similar to those of the CO (Figs. 5, 6). In addition, the formation mechanism can be commonly understood in terms of the shelfbreak front model (e.g., Pickart 2000). However, in contrast to the CO region, where COW is supplied only from January to April, the freshwater in the MAB is carried throughout the year (Khatiwala et al. 1999). As a result, there are distinct differences between the seasonal variations.

To focus on basic dynamics of the CO, its short time variability is hardly examined. However, undulating perturbations with considerable amplitudes are often seen along the shelfbreak front (Fig. 3). Similar perturbations have been observed in the MAB and examined with an emphasis on their contribution to water exchange between the shelf and offshore (Garvine et al. 1988). Investigation of instabilities of the shelfbreak front would contribute to understanding of water exchange processes in the Doto area.

In addition, note that interannual variability is not examined, either. Actually, Oguma et al. (2008) reported that composition ratio of the Oyashio water in COW was different between 2005 and 2006. In 2005, the isotopic character of COW was the same as that in the Nemuro Strait, which is consistent with the present tracer experiment (Fig. 12). In contrast, COW in 2006 included a lot of the offshore Oyashio water. Interannual variabilities in the wind field and location of the Oyashio might modify the COW inflow through straits and/or water exchange processes across the shelfbreak front.

Because they are based on the model simulation, the present results are necessary to be verified by fine observations. In this respect, it is encouraging that a similar structure of the CO was found in a fine velocity observation carried out by Kuroda et al. (2006). We believe that fine observations in the future will verify the present results and refine them further.

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APPENDIX

Settings of the Tracer Experiment

A passive tracer experiment was performed to reveal the origin of COW on the southeastern shelf of Hokkaido. Here, we summarize the procedure. First, the initial value of the tracer concentration C was set zero everywhere. We set C = 1.0 for the water that passed through the Nemuro and Kunashiri Straits satisfying the COW definition (T < 2°C and S < 32.6) and calculated its time evolution based on the advection–diffusion equation of the model. The experiment was started on 1 January, when COW had not yet appeared, and finished on 14 March. The value of C in the result indicates the composition ratio of COW originating in the two straits in the water.
The model results obtained in advance were used for the physical fields (a so-called offline approach). To be more precise, every 6 h we recorded the instantaneous fields of velocity and the vertical diffusivity coefficients estimated by the Mellor–Yamada scheme and interpolated them in time. It was verified that almost identical results are obtained using daily averages for the physical fields.

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