Pacific Water Transport in the Western Arctic Ocean Simulated by an Eddy-Resolving Coupled Sea Ice–Ocean Model

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ABSTRACT

The process of the Pacific water transport in the Chukchi Sea and the southern Canada Basin is investigated by using an eddy-resolving coupled sea ice–ocean model. The simulation result demonstrates that the Pacific water flows into the basin by mesoscale baroclinic eddies, which are generated and developed as a result of the instability of a narrow and intense jet through the Barrow Canyon. Each eddy has a baroclinic anticyclonic structure, and its horizontal and vertical scales grow up by being merged with other ones during August and September, they separate into anticyclones whose diameters are about 50 km in October, and then they gradually shrink in early winter. The Pacific water transport across the Beaufort shelf break reaches maximum (about 0.3 Sv, where 1 Sv = 10^6 m^3 s^-1) during late summer and early autumn when the eddy activities are enhanced. The sensitivity experiments indicate that the shelf-to-basin transport differs depending on the sea ice condition in the Chukchi Sea during summer. The difference is found to be associated with the jet strength, which is closely related to the location of the sea ice margin. When the sea ice margin is located in the Canada Basin, the jet is stronger, and mesoscale eddy activities and corresponding inflow of the Pacific water into the basin are enhanced. When sea ice remains in the shelf even in late summer, sea ice ocean stress plays a great role in braking the jet and the consequent suppression of the shelf-to-basin transport. The freshwater and heat transports into the basin associated with the Pacific water inflow depend on not only the volume flux but also on surface buoyancy flux in the shelf, which varies according to sea ice condition. The freshwater transport referenced to 34.8 psu is 259 km^3 yr^-1 in the medium sea ice extent case. Although the Pacific water becomes freshened as a result of its mixing with sea ice meltwater in the large extent case, the freshwater transport is still less than in the other cases. The heat transport is promoted by preferable absorption of solar heat in addition to energetic eddy-induced transport in the small extent case. The heat amount provided into the basin is equivalent to the reduction of sea ice thickness by about 1 m yr^-1 north of the Chukchi and Beaufort shelf breaks.

1. Introduction

Mechanism of freshwater accumulation and release in the Canada Basin is one of the principal concerns in recent Arctic Ocean modeling (Proshutinsky et al. 2005). About 10 times as much freshwater as the total annual river discharge to the entire Arctic Ocean is stored in the Beaufort gyre. Its decadal variation is supposed to affect global thermohaline circulation by freshwater transport to the deep-convection area in the northern North Atlantic (Komuro and Hasumi 2007). To obtain detailed understanding of its effect, realistic simulation by using a sophisticated numerical model is indispensable. However, most previous models commonly produce a significant high-salinity bias in the central Canada Basin (Steele et al. 2001a; Steiner et al. 2004). Although a few models underestimate the salinity to the contrary, such models adopt restoring of salinity in the surface mixed layer or induce unrealistic warm bias in the basin interior.

The Pacific water, which flows through the Bering Strait, is one of the predominant freshwater sources in the Canada Basin. The freshwater flux of the Bering Strait throughflow referenced to the average salinity in the Arctic Ocean (34.8 psu) reaches about 2500 km^3 yr^-1, which is comparable with the total river discharge into

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the entire Arctic Ocean (Serreze et al. 2006). It is supposed that the Pacific water becomes even fresher by being mixed with sea ice meltwater in the Chukchi Sea during summer and provides plenty of freshwater for the Canada Basin. Because both the eastward Alaskan coastal current and the westward Beaufort gyre flow along the Beaufort shelf break, some mesoscale processes are necessary for the transport of coastal waters into the central basin. Although the contribution of mesoscale eddy activities over the shelf break to the shelf-to-basin transport was previously discussed (Pickart 2004), the formation mechanism and the behavior of the eddies have not fully been clarified.

In the central Labrador Sea, where deep convection develops during winter, it is indicated that warm eddies generated by the instability of buoyant coastal currents contribute to the spring restratification in the convection area, although other mechanisms are also proposed (Katsman et al. 2004). Such a restratification process may occur in the Pacific side of the Arctic Ocean. It is possible that warm and low salinity eddies spawned from the Alaskan coastal current transport the Pacific water across the shelf break and then maintain the minimum salinity in the central Canada Basin. However, the internal Rossby radius of deformation is so small (about 10 km) in the polar region that such mesoscale eddies are hardly resolved by basin-scale models. It is plausible to think that the insufficient eddy-induced transport of the Pacific water across the Beaufort shelf break is an essential factor for the salinity bias arising in most previous models.

Pickart (2004) constructs hydrographic composites along the Beaufort shelf break in each season by using the mooring and CTD data from 1950 to 1987. The composites indicate that surface-intensified jet over the shelf break transports warm freshwater eastward during late summer and early autumn. In the offshore side of the shelf break, silicate-rich warm eddies appear to originate from the Pacific Ocean. He focuses on the close similarity of hydrographic structure between the Beaufort shelf break and the Middle Atlantic Bight shelf break, where the surface-intensified jet enhances mesoscale-eddy generation by mixed baroclinic and barotropic instability. These fields imply that buoyant eddies formed by the instability of the surface-intensified jet play an important role in the Pacific water transport to the basin, whereas quantitative estimation of the shelf-to-basin transport and arguments about its interannual variability have not been performed yet. Although Spall (2007) and Spall et al. (2008) discuss water mass transformation and shelfbreak eddies in the western Arctic Ocean by using a high-resolution numerical model, these studies focus on the overall circulation in the Chukchi Sea, dense water flow, and corresponding cold-core eddies originated from a bottom-intensified shelfbreak jet. Shimada et al. (2006) focus on the dynamic coupling between sea ice and ocean circulation and propose that the reduced internal ice stress causes intense westward flow in the southern Beaufort Sea. Thus, it is expected that volume, heat, and freshwater fluxes associated with the Pacific water transport into the Canada Basin would also differ depending on the sea ice extent, which seasonally and interannually varies.

In this study, mechanisms controlling the Pacific water transport from the Chukchi shelf to the Canada Basin and their relationship with sea ice condition are investigated by using an eddy-resolving coupled sea ice–ocean model with realistic experimental design. Especially, we focus on surface-intensified buoyant eddies near the Beaufort shelf break. The paper is organized as follows. Model configuration and experimental design are outlined in section 2. The basic process of the Pacific water transport is described in section 3. The dependence of the transport on sea ice condition is focused on in section 4. Discussions and summary are presented in sections 5 and 6, respectively.

2. Experimental design
   a. Model description

The coupled sea ice–ocean model used in this study is the Center for Climate System Research Ocean Component Model, version 3.4 (Hasumi 2006). The sea ice part has both thermodynamic and dynamic components. The zero-layer formulation of Semtner (1976) is adopted for the thermodynamic part. In the dynamic part, equations for momentum, mass, and concentration are taken from Mellor and Kantha (1989). The internal sea ice stress is calculated based on the elastic–viscous–plastic rheology (Hunke and Dukowicz 1997). The ocean part is a free surface ocean general circulation model formulated on the spherical coordinate system. The model incorporates the uniformly third-order polynomial interpolation algorithm (Leonard et al. 1994) for tracer advection. The turbulence closure scheme of Noh and Kim (1999) is applied for calculating vertical viscosity and diffusion coefficients. The background vertical diffusion coefficient varies with depth from $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ at the top level to $3.0 \times 10^{-4}$ m$^2$ s$^{-1}$ at the bottom level. To represent mesoscale eddy activities well, the horizontal eddy viscosity coefficient is diagnosed by using the Smagorinsky-type biharmonic formulation whose constant coefficient is set to 3 (Griffies and Hallberg 2000), and an enstrophy preservation scheme is adopted (Ishizaki and Motoi 1999). The background horizontal viscosity and diffusion coefficients are $5.0 \times 10^2$ and 1.0 m$^2$ s$^{-1}$, respectively.
The model domain contains the Chukchi Sea and the southern area of the Canada Basin (Fig. 1). The bathymetry is constructed from the International Bathymetric Chart of the Arctic Ocean (Jakobsson et al. 2000). The model’s spherical coordinate system is rotated so that the singular points are on the equator. The horizontal resolution is about 2.5 km, and there are 25 vertical levels. The vertical grid width varies from 2 m at the top level to 1000 m at the bottom level (Table 1).

b. Boundary conditions

The atmospheric forcing components are constructed from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis monthly climatology from 1948 to 2006 (Kalnay et al. 1996). The wind stress is calculated from the sea-level pressure by following the formulation adopted in the Arctic Ocean Model Intercomparison Project (AOMIP; available online at http://fish.cims.nyu.edu/project_aomip/). The river water discharge is prescribed as surface freshwater flux based on Perry et al. (1996). In the marginal region of the model domain, a sponge boundary condition is applied for the ocean part and an open boundary condition is applied for the sea ice part. The horizontal viscosity and diffusion coefficients are increased, and the temperature and salinity at all the depths are restored to the monthly mean of the Polar Science Center Hydrographic Climatology (PHC) data (Steele et al. 2001b) in the sponge region.

The Pacific water inflow with seasonal cycle is provided at the Bering Strait based on the hydrographic observation of Woodgate et al. (2005). The annual mean inflow and salinity of the Pacific water at the strait are set to 0.8 Sv (1 Sv = 10^6 m^3 s^-1) and 31 psu, respectively. The inflow reaches maximum in June and minimum in December, and its seasonal amplitude is 0.4 Sv. The salinity reaches maximum in March and minimum in September, and its seasonal amplitude is 1 psu. The temperature is kept at freezing point from January to June and reaches maximum (5°C) in September. The zonal gradient of temperature and salinity is also taken into account, so that warmer and fresher water passes along the Alaskan side. The total freshwater transport through the Bering Strait is 2755 km^3 yr^-1, which is almost equal to the estimation of Woodgate and Aagaard (2005).

c. Experiments and integration methods

To visualize the Pacific water pathway, a virtual tracer is provided at the Bering Strait so that the Pacific water concentration is always kept at 100% at the strait in all cases performed in this study. Advection and diffusion of the tracer are calculated by the same formulation as ocean temperature and salinity. As shown in section 3a, the Pacific water transport across the Beaufort shelf break reaches maximum during late summer and early autumn and becomes minimum in midwinter. Hence, the model is integrated for one year from March to February.
March and February are regarded as the beginning and ending of the energetic transport, respectively. The initial temperature and salinity fields are derived from the PHC March data. To assess the dependence of the Pacific water transport on sea ice condition, several initial distributions of sea ice thickness are constructed by multiplying arbitrary factors to the annual mean sea ice concentration derived from the National Ice Center (NIC) Arctic sea ice chart (Fetterer and Fowler 2006). The experiments where 100, 200, and 400 cm are chosen as the factor are named as the small (SICE), medium (MICE) and large (LICE) ice cases, respectively. Because the sea ice and ocean hydrographic fields in the MICE case are the closest to observational climatology among the experiments (shown in section 3a), this case is chosen as a control case.

3. Process of the Pacific water transport into the Canada Basin

a. Basic states

The simulation result in the MICE case is compared with observational data. The sea ice margin retreats offshore from the Alaskan northern coast in September (Fig. 2). In winter, sea ice covers the Chukchi Sea and thinner ice area appears as a result of the intensified westward wind stress along the western coast of Alaska, where coastal polynya is often formed (Martin et al. 2004). The seasonal cycle of sea ice extent almost corresponds to the NIC climatology (Fig. 2e). The sea ice thickness in the Canada Basin reaches maximum (about 200 cm) in May and minimum (about 70 cm) in September. The sea ice velocity field shows weak cyclonic...
circulation in summer and the strong anticyclonic Beaufort gyre in winter, as shown in the Polar Pathfinder sea ice motion vectors (Fowler 2003). Driven by energetic eddies in the ocean surface layer, random sea ice flow appears locally along the Beaufort shelf break during late summer and early winter. Mechanisms for the eddy generation are described in section 3b.

The northward Bering Strait throughflow bifurcates roughly into three branches following major features of bottom topography over the broad, shallow Chukchi Sea. One branch travels northwest toward the Herald Canyon. The other two branches trace northeast through the Central Channel and along the Alaskan coastline, respectively. The latter two currents eventually join and then create a strong, narrow jet in the Barrow Canyon. This overall circulation pattern in the Chukchi Sea is consistent with previous observations (Weingartner et al. 2005) and realistic regional model experiments (Winsor and Chapman 2004; Spall 2007). The northward velocity in the Barrow Canyon is comparable with the absolute geostrophic velocity estimated from the ADCP and CTD data during summer (Pickart et al. 2005; Fig. 3a). This intense flow is referred to as the Barrow Canyon jet hereafter.

The temperature and salinity averaged in the top 100 m show a warm, fresh bias compared with the PHC data along the Alaskan coast during summer (not shown). This inconsistency may be because the Pacific water signal is not fully captured in the coarse-resolution PHC fields. Recent hydrographic observations depict a steep density front in the Barrow Canyon and along the Beaufort shelf break (Pickart et al. 2005; Pickart 2004). In these observations, chemical properties suggest the warm part is originated from the Pacific Ocean, and the temperature minimum near freezing point under the warm part is recognized as a signal of the Chukchi shelf water, which became denser by surface cooling and sea ice brine rejection during the previous winter. Besides, another warm mass is observed near the bottom of the canyon, where the Atlantic-origin water occasionally intrudes from the Canada Basin (Münchow and Carmack 1997). The model result broadly captures these structures, such as the Pacific-origin warm mass over the isopycnal front and the underlying cold mass, although the simulated front is somewhat steeper than the observed profile (Pickart et al. 2005) and no warm signal is located in the deepest layer (Fig. 3b).

The distribution of virtual tracer associated with the Pacific water provided at the Bering Strait demonstrates that the Pacific water primarily passes through the Barrow Canyon during summer, whereas it is transported toward the western side of the shelf and the Northwind Ridge by robust easterly wind during winter (Fig. 4). The Pacific water content is calculated by a following formulation:

$$\int_{-H}^{0} C_{PW} dz,$$

where $C_{PW}$ is the concentration of the Pacific water tracer and $H$ is the water depth in each grid. The transport weighted by the Pacific water concentration through the canyon reaches 0.59 Sv in August. The Pacific water flow from the Chukchi shelf into the Canada Basin is concentrated mainly in the vicinity of the Barrow Canyon, where mesoscale eddies appear. Figure 5 displays the transition of kinetic energy in the upper layer from August to November. The kinetic energy is represented by the monthly mean of $\rho (u^2 + v^2)/2$, where

\begin{align*}
\int_{-H}^{0} C_{PW} dz,
\end{align*}
where \( \rho_0 \) is the water density \((1.0 \times 10^3 \, \text{kg m}^{-3})\), and \( u \) and \( v \) are the zonal and meridional ocean velocities in each time step, respectively. The maximum kinetic energy, which reaches \(40 \, \text{J m}^{-3}\), is located along the shelf break north of the Barrow Canyon during August and September. The energetic region is distributed into several maximum energy zones in October and then disappears in November. The daily mean horizontal velocity field on 15 October shows a few anticyclonic eddies over the shelf break and is likely reflected to the spatial pattern of the kinetic energy (Fig. 6a). These eddies have baroclinic structures, and their vertical scale is about 200–300 m in October (Fig. 7). These results suggest that each eddy grows up by being merged with other ones from August to

![Figure 5](image_url)

**FIG. 5.** Monthly mean kinetic energy (J m\(^{-3}\)) averaged in the top 100 m in the MICE case in (a) August, (b) September, (c) October, and (d) November. Contour interval is 12 J m\(^{-3}\).
September, separates into anticyclones whose typical diameter are about 50 km in October, and then gradually shrinks in early winter. Temporal and spatial variations of the shelf-to-basin transport of the Pacific water almost coincide with the eddy fields. The net transport increases during late summer and early autumn when the eddy activities are enhanced, and the maximum transport reaches 0.3 Sv (Fig. 9a). The Ekman-induced component of the transport analytically estimated from surface stress fields is significantly smaller than the total transport. Therefore, the Pacific water inflow into the basin is mainly realized by the mesoscale eddies along the Beaufort shelf break.

The overall circulation, the hydrographic structure, and the Pacific water pathway described above are broadly common among the SICE, MICE, and LICE cases. The detailed comparison among the experiments is provided in section 4.

b. Mechanism for mesoscale eddy generation

It is suggested that the halocline layer between 50 and 300 m in the Canada Basin is full of a number of mesoscale anticyclonic eddies, and several kinds of sources and generation mechanisms of the eddies have been proposed (Manley and Hunkins 1985). Chao and Shaw (2003) hypothesized that subsurface anticyclones along the Beaufort shelf break were originated from the winter-transformed Pacific water and suggested that an interaction between dense-water plume and offshore baroclinic current was necessary for the generation and seaward migration of the cold-core eddies by idealized numerical experiments. The detailed physical and chemical properties of a cold-core eddy observed on the Chukchi Sea continental slope were surveyed by Mathis et al. (2007) and Kadko et al. (2008). Both studies found that the bottom-intensified shelf-edge current formed such eddies, which play a significant role in the carbon, oxygen, and nutrients transport into the upper halocline of the Canada

![Fig. 6](image-url) (a) Daily mean northward velocity averaged in the top 100 m on 15 Oct in the MICE case (color shades, cm s\(^{-1}\)). Energy conversion rates from (b) APE \( T_{bc} \) and (c) MKE \( T_{bt} \) to EKE in September in the MICE case (KJ m\(^{-2}\) month\(^{-1}\)). Region corresponds to white rectangular in Fig. 1. Vectors show ocean velocity averaged in the top 100 m, and their unit vector is 50 cm s\(^{-1}\). Refer to Fig. 7 for meaning of white line in (a).

![Fig. 7](image-url) (a) Daily mean northward velocity on 15 Oct in the MICE case (cm s\(^{-1}\)). Horizontal axis corresponds to the white line in Fig. 6a (right side is west). Vertical axis denotes depth (m).
Basin. Spall et al. (2008) combines an idealized numerical model and high-resolution mooring data in the southern Beaufort Sea; their analysis of energetics also indicates that anticyclonic cold-core eddies are generated by the baroclinic instability of the shelfbreak jet. Timmermans et al. (2008) reported that other kinds of eddies were captured by ice-tethered profilers in the central basin. These eddies are located at the shallower depth than previously observed in the southern basin, and their hydrographic properties are similar to those in the Eurasian side of surface density front between the Beaufort gyre and the Transpolar Drift. Correspondingly, it is likely that the southward migration of buoyant eddies toward the central Canada Basin also exists.

In this subsection, the mechanism for the generation of the mesoscale buoyant eddies that transport the Pacific water to the Canada Basin during late summer and early autumn is investigated using the simulation result in the MICE case. The mean kinetic energy (MKE) and the eddy kinetic energy (EKE) are evaluated as

\[ r_0 \left( \frac{u^2}{2} + \frac{v^2}{2} \right) \quad \text{and} \quad r_0 \left( \frac{u^2}{2} + \frac{v^2}{2} \right), \]

respectively, where bars indicate monthly means and primes denote deviations from the monthly means. The EKE increases especially along the Barrow Canyon and the Beaufort shelf break during late summer. To specify the EKE source, the energy conversion rates are calculated by following the formulations of Eden and Böning (2002). The conversion rate from the available potential energy (APE) to the EKE induced by baroclinic instability is

\[ T_{bc} = -\rho_0 \left\{ \frac{u'}{u} \frac{\partial \Pi}{\partial x} + \frac{v'}{v} \left( \frac{\partial v}{\partial x} + \frac{\partial \Pi}{\partial y} \right) + \frac{\partial v'}{\partial y} \right\} dz, \]

where \( \rho \) is the in situ density, \( \sigma_\theta \) is the horizontally averaged potential density, and \( g \) is the gravitational acceleration. These conversion rates suggest that the baroclinic instability is dominant in the vicinity of the Barrow Canyon, and \( T_{bc} \) is also significantly large there (Figs. 6b,c). The minimum potential vorticity is located more than 30-m depth of the onshore side of the Barrow Canyon, and such a profile satisfies the necessary condition for baroclinic instability (Fig. 8a). The growth time scale of the instability \( \sqrt{\frac{\Pi}{Ri}} \), where \( Ri \) is the Richardson number and \( f \) is the Coriolis parameter (Eady 1949), is locally just several hours. These circumstances fully account for the development of baroclinic instability. On the other hand, Lozier et al. (2002) discuss the Middle Atlantic Bight shelfbreak jet using linear instability analysis and indicate that the jet, whose horizontal velocity shear is of the same order of magnitude as the local Coriolis parameter, becomes unstable within several days. The ratio of the simulated horizontal velocity shear to the local Coriolis parameter, which is defined as the Rossby number, exceeds 0.6 in the Barrow Canyon during late summer (Fig. 8b). This result also supports that the jet strength is enough for eddy formation by barotropic instability. These disturbances produced by baroclinic and barotropic instability contribute to the development of the mesoscale eddies during late summer and early autumn, rather than generate new eddies, and each eddy has a lifetime of several months (Fig. 5). Thus, we can conclude that the instability of the Barrow Canyon jet generates and develops the mesoscale eddies that play a great role in the Pacific water transport from the Chukchi shelf to the Canada Basin, whereas the detailed mechanisms controlling the eddy properties have not been fully clarified yet.
4. Dependence on sea ice conditions

Because sea ice becomes a source or sink of momentum and buoyancy for the ocean surface layer, it is expected that the strength of the Barrow Canyon jet and the corresponding eddy-induced transport of the Pacific water depend on the sea ice condition. To investigate the dependence, the simulation results integrated from the different initial states of sea ice thickness are compared. The sea ice margin retreats far from the Alaskan northern coast during summer in the SICE case, whereas it remains in the Chukchi Sea throughout the year in the LICE case (Fig. 2). The difference of sea ice margin between these two cases is comparable with the range of interannual variations observed by the Special Sensor Microwave Imager (SSMI; Cavalieri et al. 2006). The Pacific water transport across the Beaufort shelf break and the Barrow Canyon jet are both more prominent in the SICE case than in the LICE case during late summer and early autumn (Figs. 9a,b and 10a,b). Note that the transport is mainly realized by the mesoscale eddies in both the cases. The maximum jet velocity exceeds 100 cm s\(^{-1}\) in the SICE case, whereas it is below 70 cm s\(^{-1}\) in the LICE case. The energy conversion rates by baroclinic and barotropic instability, the Richardson number, and the horizontal velocity shear in the Barrow Canyon also support the more active generation and development of mesoscale eddies in the SICE case than in the LICE case. These results suggest that the Barrow Canyon jet and the eddy-induced transport of the Pacific water are promoted when summer sea ice extent shrinks significantly.

The hydrographic structure and the Pacific water profile in the Barrow Canyon also differ between these three experiments. The Pacific water gathers toward the Alaskan side of the steep density front in the SICE case, whereas it extends westward over the flattened front in the LICE case (Figs. 10c,d). The profiles of temperature in the SICE case and salinity in both cases almost correspond to that of density, whereas the temperature in the LICE case is uniform at freezing point (Figs. 10c–h). To investigate the relationships of these differences between the experiments with the sea ice condition, the contribution of surface momentum and buoyancy fluxes to the ocean fields is evaluated in following subsections.

a. Surface momentum flux

The relationship between the Barrow Canyon jet and the surface momentum flux is investigated. In the LICE case, the maximum surface stress exceeds 0.2 Pa in the Barrow Canyon, which is an order of magnitude larger than in the SICE case (Fig. 11). The difference is explained by the location of the sea ice margin. Because the sea ice margin remains in the shelf in the LICE case, the sea–ice ocean stress has the potential to control the current. In the Barrow Canyon, the velocity difference between sea ice and ocean reaches more than 50 cm s\(^{-1}\). Hence, the jet is significantly braked by the sea ice ocean stress, which is proportional to the square of the velocity difference. On the other hand, the jet strength is maintained in the SICE case where the sea ice margin is located in the Canada Basin. The time series of jet velocity, surface stress, and sea ice concentration in the Barrow Canyon also indicate their close relationship (Figs. 9b–d). In all the cases, the abrupt decline of surface stress and the consequent acceleration of jet are induced just after the exposure of open water in the canyon. If sea ice freely drifts, it is anticipated that the velocity difference between sea ice and ocean and the corresponding surface stress are suppressed as a result of strong dynamic coupling. The prolonged large surface stress for several months indicates that internal ice stress balances with the sea ice ocean stress in the vicinity of the canyon. The sea ice dragged by the jet inevitably converges north of the canyon, so the internal stress becomes mostly dominant there. The Pacific water and density profiles in the Barrow Canyon are also significantly altered by the surface stress. The Pacific water is transported mainly along the northward current following the major features of bottom topography under ice-free condition, whereas it diffuses westward accompanied by deceleration of the jet under ice-covered condition. This is the reason why the steep density front is kept in the SICE case and becomes gentle in the LICE case.

It is expected that the sea ice ocean drag coefficient differs between undeformed ice and ridged ice. In our simulation, the coefficient is set to 5.0 \times 10^{-3} following McPhee (1980). This value is obtained from the free-drift force balance at manned stations, so it may be overestimated for deformed ice. To investigate the sensitivity of the jet strength to the drag coefficient, the coefficient is set to 5.0 \times 10^{-2} as an additional experiment. The result shows the maximum jet velocity in each case decreases by about 20 cm s\(^{-1}\), respectively, in ice-covered early summer. The observational estimate of the drag coefficient around the Barrow Canyon would be helpful for the model validation.

b. Surface buoyancy flux

The surface buoyancy flux in the Chukchi Sea is also a matter of concern for the jet strength because its baroclinic component is controlled by density gradient between the warm, fresh Pacific water and the cold, saline Chukchi shelf water. The buoyancy flux is composed of surface heat and freshwater flux. Both components are closely related to the location of sea ice margin. A major part of downward shortwave radiation is absorbed in the
FIG. 9. Monthly averaged (a) transport of virtual Pacific water tracer across the dashed line in Fig. 4a in the SICE, MICE, and LICE cases (Sv). Its Ekman-induced component in the MICE case is also shown. (b) Spatially maximum velocity in the Barrow Canyon (cm s$^{-1}$); (c) surface stress (Pa); (d) sea ice concentration (unitless); (e) Pacific water salinity (psu); and (g) Pacific water temperature (°C) averaged in the Barrow Canyon. The salinity and temperature are averaged over only the grids where Pacific water exists, so no value is plotted during early spring. Transport of (f) freshwater (km$^3$ month$^{-1}$) and (h) heat (10$^{19}$ J month$^{-1}$) associated with Pacific water tracer across the dashed line in Fig. 4a. Reference salinity and temperature are 34.8 psu and –1.8°C, respectively.
FIG. 10. (a),(b) Northward velocity (cm s$^{-1}$), (c),(d) the Pacific water concentration (unitless), (e),(f) potential temperature (°C), and (g),(h) salinity (psu) in the Barrow Canyon in August in the (left) SICE and (right) LICE cases. Black dashed contours show potential density, and the interval is 0.4 kg m$^{-3}$.
ice-free ocean surface layer, whereas it is reflected and ocean heat content could be reduced by melting sea ice in the ice-covered area. The surface freshwater flux is composed of sea ice meltwater, precipitation ($P$), evaporation ($E$), and river runoff ($R$). The sea ice melting during summer, which can be estimated from the sea ice thickness at the beginning of the ice-melt season, is a few meters in the Chukchi Sea. In this region, the precipitation and evaporation derived from the NCEP–NCAR reanalysis data are about 20 and 15 cm yr$^{-1}$, respectively. The Colville River, which has the largest discharge in the model domain, provides only 18 km$^3$ yr$^{-1}$ (Perry et al. 1996). This discharge is equivalent to surface freshwater flux of about 7 cm yr$^{-1}$ in the entire Chukchi Sea. Thus, the sum of $P$, $E$, and $R$ is negligible compared with the sea ice melting during summer. Therefore, the surface freshwater flux is provided mainly in the ice-covered area.

In the SICE case, more absorption of solar heat by the Pacific water following rapid sea ice retreat in the shelf sharpens the horizontal temperature gradient in the Barrow Canyon (Fig. 10c). On the other hand, in the LICE case, the Pacific water loses its heat by melting sea ice and the temperature in the canyon becomes uniform at the freezing point regardless of the Pacific water profile (Fig. 10f). The salinity change of the Pacific water brought by surface freshwater flux also depends on the location of sea ice margin. In the SICE case, the Pacific water, which passes through the Barrow Canyon during late summer, obtains little freshwater because it is transported after the sea ice retreats (Fig. 10g). On the other hand, the Pacific water is continuously mixed with the sea ice meltwater and becomes gradually fresher in the LICE case (Fig. 10h). The Pacific water salinity averaged in the Barrow Canyon is 31.4 (30.6) psu in the SICE (LICE) case in August (Fig. 9e). Here, an amount of sea ice melting that accounts for the salinity difference between these cases is roughly estimated by using the following budget analysis:

$$HS_1 + \delta h S_{ice} = (H + \delta h)S_2,$$

where $S_1$ and $S_2$ are seawater salinity before and after, respectively, mixed with sea ice meltwater whose amount is $\delta h$ (in m), sea ice salinity $S_{ice}$ is 5 psu in this model, and the Pacific water content in the Barrow Canyon $H$ is about 30 m in both the cases. When it is assumed that the salinity difference between the SICE and LICE cases could be substituted for $S_1 - S_2$,

$$\delta h = \frac{S_1 - S_2}{S_2 - S_{ice}} H = \frac{31.4 - 30.6}{30.6 - 5} 30 = 1.19 \text{ m}.$$
This simple estimation is comparable with sea ice melting in the upstream region of the Barrow Canyon during summer in the LICE case and supports the conclusion that the freshening of the Pacific water is caused by mixing with sea ice meltwater in the shelf.

To specify the individual contribution of sea ice ocean stress and surface buoyancy flux to the jet strength, sensitivity experiments where the sea ice ocean stress term is excluded from the surface momentum flux throughout the integration period are performed. The other conditions in these experiments—small (SNIS), medium (MNIS), and large (LNIS) ice extent cases with no sea ice–stress condition—are the same as the SICE, MICE, and LICE cases, respectively. The results show that the jet strength and the Pacific water and density profiles in the Barrow Canyon are almost the same regardless of the sea ice conditions (Figs. 12a–d). The contribution of the horizontal temperature gradient to the density gradient counterbalances that of the salinity gradient (Figs. 12e–h). Thus, it is found that the dependence of the jet strength on sea ice conditions is insignificant under no sea ice–ocean stress condition. The simulated maximum jet velocity in these cases reaches about 130 cm s$^{-1}$, which is equal to that in the SICE case (Fig. 13). These results support the conclusion that the Barrow Canyon jet is essentially braked by the sea ice ocean stress and that it is suddenly accelerated just after the disappearance of sea ice from the canyon, whereas the generation of mesoscale eddies and the corresponding eddy-induced transport are caused in both ice-free and ice-covered conditions.

5. Discussion

a. Freshwater transport across the Beaufort shelf break

It is conceivable that the insufficient transport of the Pacific water across the Beaufort shelf break is one of the important factors for the salinity bias arising in most of the Arctic Ocean models. To quantify the contribution, the freshwater transport from the Chukchi shelf to the Canada Basin is assessed. A lateral freshwater flux associated with the Pacific water transport is represented by the following formulation:

\[
\int_{-H}^{0} \frac{S_{ref} - S}{S_{ref}} u C_{pw} \, dz,
\]

where $S$ and $u$ are ocean salinity and cross-shelf velocity, respectively, and $S_{ref}$ is the reference salinity. The deepest depth $H$ of the vertical integration is set to a water depth in each grid. The lateral freshwater transport referenced to 34.8 psu across the black dashed line in Fig. 4 is 259 km$^3$ yr$^{-1}$ in the MICE case. The transport is promoted according to both enhanced eddy activities and the freshening of the Pacific water during late summer (Figs. 9e,f). The eddy-induced freshwater transport may cancel out the high-salinity bias in the Canada Basin arising in previous coarse-resolution models, while the model domain needs to be extended to the whole Arctic Ocean to evaluate the actual salinity change. It is expected that the lateral freshwater flux into the basin differs depending on not only the volume flux but also on the summer sea ice melting in the Chukchi shelf. The sensitivity experiments reveal that the freshwater flux in the SICE case (292 km$^3$ yr$^{-1}$) is larger than that in the LICE case (226 km$^3$ yr$^{-1}$) despite the lower Pacific water salinity in the Barrow Canyon in the latter case throughout the integration period (Fig. 9e). These fluxes are almost the same even when $H$ is set to 100 m. This result suggests that the volume transport across the shelf break is still dominant for the lateral freshwater flux compared to the summertime mixing of the Pacific water and sea ice meltwater in the shelf in these experiments.

b. Interannual variation of freshwater transport

The number of years when open water is exposed in the entire Barrow Canyon is counted for each month using the SSM/I sea ice concentration data from 1979 to 2005. Figure 14 shows that the ice-free period interannually varies between July and November. The sea ice regularly covers the canyon from December to June, and even occasionally in September. Therefore, it is conceivable that the Pacific water transport across the Beaufort shelf break also varies interannually. Previous hydrographic observations indicate the decadal variation of the freshwater content in the Canada Basin, which has not been reproduced in the AOMIP models yet (Proshutinsky et al. 2005). A freshening trend from the 1980s to the 1990s is hardly interpreted by the wind stress variation related to the Arctic Oscillation (Thompson and Wallace 1998), because the intensified polar vortex is expected to release the freshwater accumulated in the central Canada Basin. The finding obtained in this study proposes that the shelf-to-basin transport of the Pacific water is promoted following the northward shift of the summer sea ice margin observed during this period. If the change of Pacific water salinity through the period is small, the enhanced transport would contribute to the freshening in the central basin. However, if the summer sea ice melting in the shelf is significantly reduced, the freshwater transport to the basin may decrease. Therefore, to evaluate the actual interannual variation of the freshwater transport, realistic decadal simulations and detailed analyses of the freshwater budget in the Chukchi Sea are necessary.
FIG. 12. Same as Fig. 10, but in the NIS cases.
c. Possible contribution of ocean heat transport to sea ice retreat

In this section, the influence of ocean heat transport—accompanied by the Pacific water inflow into the Canada Basin—on thermodynamic response of sea ice is examined. In the SICE case, high ocean heat content appears not only in the central Chukchi shelf but also on the offshore side of the Beaufort shelf break where the Pacific water abundantly spreads in winter season (Fig. 15a). Minimum of the thermodynamic sea ice growth over the warm pool indicates that ocean heat provided by the eddy-induced shelf-to-basin transport of the Pacific water delays sea ice production in the basin interior (Fig. 15c). We estimate the lateral ocean heat flux associated with the Pacific water transport as follows:

\[ \int_{-H}^{0} \rho_0 C_{po} (T - T_{ref}) u C_{pw} \, dz, \]

where \( \rho_0 \) is the water density, \( C_{po} \) is the specific heat of seawater (3.99 \times 10^3 \text{ J K}^{-1} \text{ kg}^{-1}), T and \( u \) are ocean temperature and horizontal velocity. The reference temperature \( T_{ref} \) is \(-1.8^\circ\text{C}\). Here \( H \) is set to a water depth that is the same as the freshwater flux in section 5a. The shelf-to-basin heat transport is clearly large during late summer in the SICE and MICE cases (Fig. 9h). When we enclose the Pacific water area in the southern Canada Basin by a red line in Fig. 15c, which is composed of the shelf-basin boundary (i.e., the dashed line in Fig. 4) and a 20-m contour level of the Pacific water content in February, the annual lateral net heat influx into this area is \( 2.42 \times 10^{19} \text{ J} \) in the SICE case (Table 2). The provided heat amount is equivalent to a reduction of sea ice thickness by about 1 m yr\(^{-1}\) in the area. Even when the deepest depth \( H \) is set to 100 m, these values are almost same. On the other hand, in the LICE case, spatial patterns of the ocean heat content and the sea ice growth are independent of the Pacific water distribution (Fig. 15b). The ocean temperature in the upper layer of the basin is uniformly at the freezing point throughout winter. The lateral heat influx is one order of magnitude smaller than in the SICE case, and the equivalent change in sea ice thickness is negligible compared with the simulated winter growth (Table 2). Even when the contour level of the Pacific water content used for the definition of the Pacific water area is changed to 10 or 30 m, the similar difference is obtained. These results suggest that sea ice growth in the southern Canada Basin during winter is restricted by the energetic shelf-to-basin transport of the Pacific water; this water becomes warmer by the absorption of shortwave radiation during summer in the Chukchi Sea when the extent of sea ice is less condition, while the Pacific water inflow into the basin has a minor contribution to affect sea ice growth offshore side of the shelf break in the large sea ice extent condition. Shimada et al. (2006) propose that ocean heat transport accompanied by the anticyclonic Beaufort gyre is promoted by the intensified westward surface stress under fragile ice. The finding in this study provides an additional process that preferable absorption of solar heat in the shelf and the enhanced eddy-induced inflow of the Pacific water into the basin also play an important role in the increase of the ocean heat transport toward the Northwind Ridge when the sea ice margin retreats far from the Alaskan northern coast. Thus, there may be several positive feedback processes between sea ice extent and ocean heat transport. The detailed analyses...
of such mechanisms should be conducted in the near future to discuss the recent sea ice variations.

6. Summary

The process of the Pacific water transport in the Chukchi Sea and in the southern Canada Basin is investigated using an eddy-resolving coupled sea ice–ocean model. Most of the Pacific water that passes through the Bering Strait during summer is transported by the northward currents following major features of bottom topography over the Chukchi Sea and then flows into the Canada Basin by mesoscale baroclinic eddies over the Beaufort shelf break. These buoyant eddies are generated and develop as the result of the instability of the Barrow Canyon jet and have a lifetime of several months. The sensitivity experiments reveal that the Pacific water transport across the shelf break differs depending on the sea ice condition. When the sea ice margin retreats toward the basin interior, the Barrow Canyon jet is intensified and the corresponding enhanced eddy activities promote the shelf-to-basin transport. When sea ice remains in the shelf even in late summer, the enlarged sea ice ocean stress plays a great role in braking the jet and consequent restraint of the eddy activities. The dependence of the jet strength on surface buoyancy flux, such as absorption of shortwave radiation and sea ice melting, is negligible compared with that on the surface momentum flux. Dependence of the freshwater and heat transports associated with the Pacific water inflow into the basin on surface buoyancy flux in the Chukchi shelf

| Lateral heat influx (J yr\(^{-1}\) and W) into the Pacific water area (km\(^2\)) and the equivalent reduction of sea ice thickness in the area (cm yr\(^{-1}\)). Definition of these variables is described in section 5c. |
|-------------------------------------------------|--|---|------|
| Lateral heat influx (J yr\(^{-1}\) | Pacific water area (km\(^2\)) | Equivalent thickness reduction (cm yr\(^{-1}\)) |
| SICE 2.42 × 10\(^{19}\) (7.67 × 10\(^{11}\) W) | 7.84 × 10\(^4\) | 101 |
| MICE 0.80 × 10\(^{19}\) (2.54 × 10\(^{11}\) W) | 6.07 × 10\(^4\) | 43 |
| LICE 0.13 × 10\(^{19}\) (0.41 × 10\(^{11}\) W) | 4.92 × 10\(^4\) | 9 |
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