1. Introduction

For the climate system, one of the important features of sea ice is the heat insulation effect between atmosphere and ocean. The heat insulation effect is greatly reduced in the case of thin ice. Thus, in the sea ice zone, the heat flux between atmosphere and ocean depends strongly on both ice concentration and thickness. For example, in a coastal polynya, which is a typical thin-ice area formed by divergent ice drift due to prevailing winds or oceanic currents (Morales Maqueda et al. 2004), huge amounts of heat flux occur to the atmosphere; the coastal polynya can be regarded as a hot spot for the atmosphere.

The huge heat loss for the ocean in coastal polynyas leads to high ice production there. Large amounts of brine rejection associated with the high ice production form cold and saline dense water, which is the origin of the intermediate and deep water. The thermohaline circulation (overturning) by sinking of this dense water is one of the important components of the global climate system. The sinking of the dense water in the coastal polynya regions also leads to biogeochemical cycles, including the exchange of CO₂ between the atmosphere and the deeper ocean (Miller and DiTullio 2007; Hoppema and Anderson 2007).

Sea ice formed in coastal polynyas is advected by wind and ocean currents. Melting of the advected ice causes

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cooling of the upper ocean because of the latent heat absorption and supplies freshwater to the ocean. This means a freshwater and negative heat transport by ice. The sea ice formation, its transport, and its melting cause the redistribution of heat and salt. In the marginal ice zone, it was shown that the intensification of stratification in the upper ocean associated with ice melting leads to phytoplankton growth (Alexander and Niebauer 1981; Smith and Nelson 1985). Despite the importance of the redistribution of heat and salt in the climate and biogeochemical systems, these sea ice processes have not been well understood.

In the present study, we tried to create a heat and salt flux dataset in which sea ice processes are included in the Sea of Okhotsk, where the redistribution of heat and salt by the sea ice processes is considered to be particularly important, as described in the next section. This dataset is based on a daily heat budget analysis using ice concentration from the Advanced Microwave Scanning Radiometer for Earth Observing System (EOS) (AMSR-E) on the Aqua satellite. The spatial resolution of the AMSR-E (12.5 km) is about 2 times finer than that of the Special Sensor Microwave Imager (SSM/I), which has been observing sea ice for more than 20 yr. In the Sea of Okhotsk, we can use thin-ice thickness from AMSR-E (Nihashi et al. 2009) as well as the AMSR-E ice concentration for the heat flux calculation. The salt flux of this study consists of salt supply due to brine rejection associated with freezing and freshwater supply associated with melting. The effect of ice advection is taken into account using ice drift derived from AMSR-E. This heat and salt flux dataset will be useful for the validation and boundary conditions of modeling studies. The heat and salt flux dataset created in this study is archived at the website of the Institute of Low Temperature Science, Hokkaido University (http://wwwod.lowtem.hokudai.ac.jp/polar-seaflux).

The air–ocean heat flux in the sea ice zone is provided as a component in some global meteorological datasets. However, the treatment of sea ice is not appropriate in their heat flux calculation. For example, sea ice was treated as weekly ice cover, without considering its concentration, in the National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) reanalysis dataset (Kalnay et al. 1996) and in the 40-yr European Centre for the Medium-Range Weather Forecasts Re-Analysis (ERA-40) dataset (Uppala et al. 2005). Ice concentration was taken into account in the Ocean Model Intercomparison Project (OMIP) dataset (Rößke 2006), which was created as the forcing data for ocean general circulation models (OGCMs). In this dataset, ice concentrations were derived from the Scanning Multichannel Microwave Radiometer (SMMR) on the Nimbus-7 satellite and from the SSM/I on satellites of the Defense Meteorological Satellite Program (DMSP). In the sea ice zone, the OMIP heat flux is larger than that calculated by neglecting the ice concentration. Daily ice concentration has also been used in an air–ice/ocean heat flux dataset created by Large and Yeager (2009) and in the ERA-Interim dataset after January 2002 (Poli 2010). Inoue et al. (2011) showed that ERA-Interim reproduces the temperature profile (turbulent heat fluxes) better than the Japan Meteorological Agency Climate Data Assimilation System (JCDAS) and NCEP–NCAR in the marginal ice zone because of the explicit treatment of ice concentration.

Although there are some heat flux datasets that consider ice concentration, there is no heat flux dataset in which the ice thickness is explicitly taken into account. Further, because the typical width of the coastal polynya from the coast to the offshore is <100 km, the heat flux dataset with the spatial resolution of >100 km cannot resolve the coastal polynyas. Ohshima et al. (2003) estimated the heat flux in the Sea of Okhotsk with a spatial resolution of ~25 km, which is sufficient to resolve a large coastal polynya. In their study, as well as ice concentration, ice thickness was taken into consideration using an SSM/I ice-type algorithm that discriminates among new ice, young ice, and first-year ice (Kimura and Wakatsuchi 1999). In the coastal polynya region, Ohshima et al.’s heat flux dataset showed a large heat flux to the atmosphere, which had not been shown in the previous datasets.

2. Sea ice and its role in the Sea of Okhotsk

The Sea of Okhotsk (Fig. 1) is the southernmost sea in the Northern Hemisphere with a sizeable seasonal ice cover. The initial freezing occurs in the northern part of the sea in November. The ice cover becomes maximum (~1.0 × 10^6 km^2 on average) from the end of February to the beginning of March and melts away by June.

In the northwest shelf (NWS), northern shelf, Gizhiga Bay, northeastern Sakhalin, and Terpeniya Bay regions (Fig. 1), coastal polynyas are formed by divergent ice drift associated with prevailing cold offshoreward wind (Martin et al. 1998). In the North Pacific region, the densest water is formed in the NWS polynya (Shcherbina et al. 2003), and the sinking of this dense water creates vertical circulation (overturning) of the North Pacific down to intermediate depths (approximately 200–800 m deep). Nakano et al. (2007) suggested that, during the past 50 yr, warming of the intermediate water and weakening of the overturning have occurred in the northwestern North Pacific, originating from the Sea of Okhotsk. The production of ice and dense shelf water in the coastal...
polynyas was estimated from a heat budget analysis using a thin-ice thickness algorithm from AMSR-E (Fig. 1; Nihashi et al. 2009).

Based on ice thickness and drift speed measured by moorings in the coastal region off northern Sakhalin, Fukamachi et al. (2009) showed that the southward transport of freshwater due to ice advection by the prevailing northerly wind and by the southward East Sakhalin Current is comparable to or larger than the discharge of the Amur River. For the atmosphere, the ice extent anomaly in the Sea of Okhotsk causes anomalous heat flux to the atmosphere, and it influences the large-scale atmospheric circulation through the propagation of stationary Rossby waves (Honda et al. 1999). Although the Sea of Okhotsk is a marginal sea, the redistribution of heat and salt associated with the sea ice processes has a strong influence on the atmosphere and ocean on a hemispheric scale.

3. Methods and data

A daily heat and salt flux dataset for 9 yr between 2002 and 2010 was created on the AMSR-E polar stereographic grid at a spatial resolution of $\sim$12.5 km (e.g., Fig. 2). In this study, the sea ice zone is set to be the area where the ice concentration is $\geq$30%.

### a. Heat flux calculation

The heat flux calculation is based on that of Ohshima et al. (2003, 2006). The downward heat flux (from the atmosphere to the water and ice surfaces) is set to be a positive value. The net heat flux in a grid cell ($Q$), under the ice concentration ($A$), becomes the sum of the net heat flux at the water surface ($Q_w$) and that at the ice surface ($Q_i$), as follows:

$$Q = (1 - A)Q_w + A Q_i,$$

where $Q_w$ and $Q_i$ are the sum of incoming solar radiation (SW), net longwave radiation (LW), sensible heat flux (SE), latent heat flux (LA), and conductive heat flux in ice ($FC_i$), as follows:

$$Q_w = (1 - \alpha_w)SW_w + LW_w + SE_w + LA_w$$

and

$$Q_i = (1 - \alpha_i)SW_i + LW_i + SE_i + LA_i + FC_i,$$

where $\alpha$ is surface albedo and the subscripts w and i represent the water surface and ice surface, respectively. The heat fluxes in Eqs. (2a) and (2b) are calculated from the bulk and empirical formulas that are suitable for the Sea of Okhotsk, following Ohshima et al. (2003, 2006). In Eq. (2b), $FC_i$ is calculated from the ice bottom temperature ($T_b = -1.8^\circ$C), which is assumed to be at the freezing point at 32.6 psu, and from the ice surface temperature ($T_s$), as follows:

$$FC_i = k_b(T_b - T_s),$$

where $k_b$ is the bulk conductivity, given by

$$k_b = \frac{k_i k_s}{k_i h_i + k_s h_s},$$

where $k_i = 2.03 \text{ W m}^{-1} \text{ K}^{-1}$ and $k_s = 0.31 \text{ W m}^{-1} \text{ K}^{-1}$ are the thermal conductivity of ice and snow, respectively.
and \( h_i \) and \( h_t \) are ice and snow thickness, respectively. In Eq. (3), \( T_s \) is obtained as follows. First, we calculate \( T_s \), which satisfies \( Q_i = 0 \). If the obtained \( T_s \) is \( \leq 0^\circ \text{C} \), then that value is considered to be the surface temperature. If the obtained \( T_s \) is \( >0^\circ \text{C} \), then it is assumed that surface melting occurs. Then, we set \( T_s = 0^\circ \text{C} \) and recalculate \( Q_i \), which is considered to be used for the surface melting. At the ice surface, \( FC_i \) resultant becomes the net heat flux between the atmosphere and the ice surface, except for the case of surface melting.

We used daily mean ice concentration from AMSR-E with \( \sim 12.5 \)-km resolution, based on the enhanced National Aeronautics and Space Administration (NASA) Team (NT2) algorithm (Markus and Cavalieri 2000). The NT2 ice concentration tends to be underestimated in the coastal polynya area (Fig. 2a). Nihashi et al. (2009) showed that ice concentration in the polynya area is \( \sim 100\% \) from the ice surface temperature using clear-sky Advanced Very High Resolution Radiometer (AVHRR) infrared data, which has much finer spatial resolution \( (\sim 1.1 \text{ km}) \) than AMSR-E. In this study, we correct the underestimation of AMSR-E NT2 ice concentration in the coastal polynya area by applying the following method. In each coastal polynya region of the NWS, northern shelf, Gizhiga Bay, northeastern Sakhalin, and Terpenia Bay (Fig. 1), an analysis area that is about 1.5 times larger than the maximum polynya width is set (not shown). If the entire analysis area is judged as the sea ice grid cells and the heat flux between the atmosphere and the ocean surface \( (Q_w) \) is \( \leq -100 \text{ W m}^{-2} \), then the ice concentrations of the coastal polynya grid cells are set to 100\% (Fig. 2b).

In the winter sea ice zone, the heat flux to the atmosphere depends strongly on the ice thickness \( (h_i) \): the
heat flux becomes drastically large in the case of thin ice. To include the effect of $h_i$ in the heat flux calculation, a detection algorithm for thin-ice thickness of $\leq 0.2$ m (Nihashi et al. 2009) using the polarization ratio of AMSR-E brightness temperature at the 36.5-GHz channel ($\text{PR}_{36}$) was used. Because this thin-ice thickness algorithm was developed for coastal polynyas, the ice concentration was assumed to be 100%. The algorithm sometimes judges the lower ice concentration region of the marginal ice zone as thin-ice area because the $\text{PR}_{36}$ value is influenced by the open water fraction (Figs. 2b and 2c). In this study, this misjudgment was corrected using the NT2 ice concentration and the brightness temperature at the open ocean grid cell nearby the ice edge (Fig. 2d; for details of this procedure, see the appendix).

In the case when $h_i$ was estimated to be $>0.2$ m from AMSR-E, $h_i$ was set to 0.8 m. The value of $h_i$ is based on ice draft measured by an ice profiling sonar (IPS; Fukamachi et al. 2006; 2009) and Russian historical data (Petrov 1998). In the Sea of Okhotsk, sea ice is mostly covered with snow, except for new (thin) ice (Toyota et al. 2007). In this study, we assumed an $h_i$ value of 0.16 m, which is 20% of $h_i$, for the case of the thick ice ($h_i = 0.8$ m). This ratio of one-fifth is based on in situ observations of Toyota et al. (2000). No snow ($h_i = 0$) was assumed for the case of thin ice ($h_i \leq 0.2$ m). The surface albedo of water ($\alpha_w$) was assumed to be 0.06. Albedo of ice ($\alpha_i$) was assumed to be 0.27, 0.36, and 0.7 for the thicknesses $h_i$ of $\leq 0.1$, 0.1–0.2, and $> 0.2$ m, respectively, based on the observations of Maykut (1986), Allison et al. (1993), and Toyota et al. (1999, 2002).

In this study, the solar radiation is assumed to be absorbed at the ice and ocean surfaces. Because the extinction coefficient of snow is more than 60 m$^{-1}$ (Mellor 1977), $\sim 95\%$ of the solar radiation is absorbed at the surface snow layer of $\sim 5$ cm. Thus, the assumption is approximately valid in the case of thick ice with snow cover. However, in the case of new ice without snow cover, much more solar radiation is transmitted into ice. In this study, we do not explicitly treat the transmittance because of the complexity: the transmittance depends on many factors, such as ice type (Grenfell and Maykut 1977; Perovich and Grenfell 1981). This simplification will not affect the results of the present study because the effect of the solar radiation is quite small at the season and place where the new ice area (coastal polynya) appears. From our dataset, net solar radiation at the new ice surface in the northwest coastal polynya area in February is $\sim 40$ W m$^{-2}$, which is only 10%–15% of the net heat flux. For the heat flux calculation at the water surface, we assume that the upper ocean is well mixed and that the solar radiation is absorbed in the surface mixed layer.

As the atmospheric and oceanic data for the heat flux calculation, we use the daily mean ERA-Interim data with a spatial resolution of $1.5^\circ \times 1.5^\circ$. Data of air temperature at 2 m, dewpoint temperature at 2 m, wind at 10 m, surface sea level pressure (SLP), cloud cover, and sea surface temperature (SST) are used. In Ohshima et al. (2006), the wind speed at 10 m from ERA-40 was corrected by multiplying by 1.25 based on a comparison with in situ data. Because the wind speeds at 10 m from ERA-Interim are similar to those from ERA-40 with no bias in the Sea of Okhotsk (not shown), we also correct the ERA-Interim wind speed by multiplying by 1.25 for the heat flux calculation.

For the heat flux calculations, the ERA-Interim data are interpolated onto the AMSR-E polar stereographic grid at a spatial resolution of $\sim 12.5$ km with a Gaussian weighting function, where the e-folding scale is roughly set to the gridcell size of $\sim 170$ km. The meteorological variables tend to have steep gradients near the boundary between land and ocean. Thus, the weighting function for land points was reduced to one-fifth for the interpolation of the air temperature at 2 m, dewpoint temperature at 2 m, and wind at 10 m. The SST at grid cells with sea ice was set to the freezing point ($-1.8^\circ$C) for the freezing season (November–March) and set to $-1.0^\circ$C for the melting season (April–May) based on in situ observations (Ohshima et al. 1998; Nihashi et al. 2005).

b. Salt flux calculation

Next, we consider the estimation of salt flux associated with sea ice production. In this study, the amount of ice production is estimated from the heat loss to the atmosphere and ocean surfaces, and then it is converted to the salt flux. We assume that all of the net heat loss to the atmosphere at sea ice grid cells is used for freezing when $Q < 0$ W m$^{-2}$. The oceanic heat flux is assumed to be negligible. In coastal polynya regions, the water temperature is expected to be close to the freezing point over the entire water column, because the polynyas exist on the shallow shelf (water depth of $\leq 200$ m; Fig. 1). In addition, data from bottom moorings in the NWS region reveal that winter water temperature at the bottom layer is close to the freezing point (Shcherbina et al. 2003). The daily rate of ice production ($V_f$ in m day$^{-1}$) in a grid cell is given by

$$V_f = -\frac{Q}{\rho_s L_f} T_d,$$

where $\rho_s (=920$ kg m$^{-3}$) and $L_f (=0.334$ MJ kg$^{-1}$) are the density of sea ice and the latent heat of fusion for ice, respectively, and $T_d$ is 86 400 s. We adopt an $L_f$ value of 0.334 MJ kg$^{-1}$ that has been used in many previous studies. Since a smaller $L_f$ value of 0.234 MJ kg$^{-1}$ was
used in Nihashi et al. (2009), the ice production estimated in this study becomes smaller by ~30% than that estimated in Nihashi et al. (2009). Ice production modified from Nihashi et al. (2009) using an $L_f$ value of 0.334 MJ kg$^{-1}$ is shown in Fig. 1.

The daily salt flux ($S$; kg m$^{-2}$ day$^{-1}$) associated with freezing is obtained from $V_f$ as follows:

$$S = \rho_{fl} V_f (s_w - s_{fl}) \times 10^{-3},$$

where water salinity ($s_w$) is assumed to be a constant value of 32.6 psu (Shcherbina et al. 2003). Ice salinity during freezing is assumed to be $s_{fl} = 0.31s_w$, following Cavalieri and Martin (1994). Thus, $s_{fl}$ is set to ~10 psu. In a unit area, ice production of 1 m corresponds to a salt supply of ~20.7 kg.

c. Freshwater flux calculation

Sea ice melts at its top, bottom, and lateral faces. Because there is no information about ice thickness, the temporal change in ice volume associated with ice melting cannot be estimated directly; consequently, it is more difficult to estimate ice melting than ice production.

From the heat flux calculation, net heat input from the atmosphere averaged over the Okhotsk sea ice zone from April to May, which is the ice melt season, are shown to be ~60 and ~11 W m$^{-2}$ at the ocean and ice surfaces, respectively. This suggests that the amount of the surface melting is relatively small. Nihashi et al. (2011) showed that heat input into the upper ocean through open water fraction is important for sea ice retreat. Within the gridcell scale of ~12.5 km, ice floes with various thicknesses are distributed. The present study does not treat the melting of individual ice floes but considers this process in a bulk fashion. The ice melting is estimated from the daily change in ice concentration by assuming that the average thickness of the individual melting ice is spatially and temporally uniform. The change in ice concentration is also caused indirectly by bottom melting, because melting from the bottom face contributes to make very thin ice or brash ice, which ultimately melts away.

Because sea ice is advected by wind and ocean currents, the daily change in ice concentration must be estimated not on a grid cell (Eulerian manner) but along the drift of ice (Lagrangian manner). The effect of ice drift is especially important at the ice edge, where the advected ice is melted in the relatively warm open ocean. Details of the method used to estimate ice melting are provided below.

An ad hoc assumption of constant ice thickness $\bar{h}_i$ of 0.34 m is made for all melting ice. This $\bar{h}_i$ value is given so that the total ice melting is balanced by the total ice production for the annual cycle in the entire Sea of Okhotsk. Then, the daily rate of ice melt per unit area in a grid cell ($V_m$; m day$^{-1}$) is calculated from the daily change in ice concentration,

$$V_m = \bar{h}_i \frac{dA}{dt}_{melt}.$$  \hspace{1cm} (7)

To estimate $(dA/dt)_{melt}$ in a Lagrangian manner, the effect of ice motion is incorporated as follows. Based on the sea ice data on a given day, we first calculate the ice concentration distribution on the next day, assuming that the concentration is determined solely by the advection of ice without any ice melting and deformation. Specifically, the position of a rectangular grid cell on the next day with a certain ice concentration is calculated through forward tracking using the ice drift velocity on that day. The calculated position of the rectangular grid cell ranges over ~4 grid cells. Then the ice concentration of the original grid cell is divided into the ~4 grid cells according to the area-weighted mean. Such calculation of the ice concentration is accomplished to all the sea ice grid cells. By summing up these concentrations, the ice concentration distribution that would be determined solely by the ice advection is calculated. For a grid cell where the calculated ice concentration exceeds 100%, the concentration is set to 100%. This ice concentration is compared with the ice concentration on the next day observed from the satellite. If the observed ice concentration is smaller than the calculated concentration, then the difference between them is regarded as the result of ice melting, that is, $(dA/dt)_{melt}$.

For the ice concentration, we used the daily mean value from the AMSR-E NT2 algorithm (Fig. 2b). The daily ice drift speed was estimated from the AMSR-E brightness temperature at the 89-GHz channel using the maximum cross-correlation approach [similar to the method described by Kimura and Wakatsuchi (2004)], with the spatial resolution of ~37.5 km. The ice drift speed was temporally averaged using a 3-day running mean (Fig. 2e) because it was shown that the 3-day running mean drift speeds best agree with buoy observations on an ice floe (Kwok et al. 1998; Martin and Augstein 2000). The ice drift speed was interpolated onto a polar stereographic grid at a spatial resolution of ~12.5 km with a Gaussian weighting function.

The negative salt (freshwater) flux ($S$; kg m$^{-2}$ day$^{-1}$) associated with ice melting is obtained from $V_m$ as follows:

$$S = \rho_{im} V_m (s_{im} - s_w) \times 10^{-3},$$ \hspace{1cm} (8)

where ice salinity during the melt season ($s_{im}$) is set to 6 psu, and $\rho_{im}$ is the density of the melting ice and is set to 900 kg m$^{-3}$, corresponding to an ice salinity of 6 psu. During freezing, ice salinity was assumed to be ~10 psu.
The difference in the ice salinity of ∼4 psu is caused by brine drainage for a couple of months from freezing to melting. For simplicity, we neglect the brine drainage because the effect of the gradual salt supply of ∼4 psu is considered to be much smaller than the effect of the rapid salt supply of ∼22.6 psu by freezing (water salinity of 32.6 psu minus new ice salinity of 10 psu). In a unit area, ice melting of 1 m corresponds to a negative salt supply of approximately 23.9 kg. The freshwater flux \( F \) is calculated from \( S \) using the following equation:

\[
F = -\frac{S}{\rho_w} \times 10^{-3},
\]

where \( \rho_w = 1026.25 \text{ kg m}^{-3} \) is seawater density. The freshwater supply of 1 m corresponds to a negative salt supply of approximately –33.45 kg. In the following, the scale of the corresponding freshwater input is also shown in the salt input maps.

### 4. Results

Annual mean net heat flux \( (Q) \) averaged from the 2002/03 to 2009/10 seasons is shown in Fig. 3a. The annual flux is calculated from October to the following September so as not to divide the winter period.
annual mean net heat flux averaged over the entire Sea of Okhotsk is $-22 \text{ W m}^{-2}$. Figure 3a shows a distinct contrast in the western Sea of Okhotsk: large negative values (heat flux to the atmosphere) are shown in the northern coastal polynya regions, whereas the positive values are shown in the south. In the central and eastern Sea of Okhotsk, large negative values are shown. As described in Ohshima et al. (2003), the contrast in net heat flux between the northern and southern parts of the western Sea of Okhotsk suggests the following heat and ice processes: the ice formed in the polynya is advected to the south by the prevailing northerly wind and by the southward East Sakhalin Current; subsequently, melting of the advected ice results in cooling of the upper ocean because of the absorption of latent heat. Thus, the contrast in flux is considered to demonstrate negative heat transport from north to south by sea ice.

A negative value of annual mean net heat flux (i.e., heat flux to the atmosphere) is also caused by the advection of warmer water. The large negative heat flux areas around the Kashevarov Bank (KB; Fig. 1) and the central and eastern Sea of Okhotsk are considered as such cases. The KB region is known as a sensitive he polynya, with relatively low ice concentration (Fig. 2b). Polyakov and Martin (2000) examined the formation and maintenance processes of the KB polynya using an ice-ocean coupled model, and then they indicated that sea ice melting is caused by the oceanic heat flux from below, associated with the residual upward current and tidal mixing. In the central and eastern Sea of Okhotsk, relatively warm water that originates from the East Kamchatka Current comes from the North Pacific through the northern Kuril Straits (Ohshima et al. 2010). This relatively warm water is shown to be one of the factors that control the maximum ice extent (Nakanowatari et al. 2010).

Annual salt input averaged from the 2002/03 to 2009/10 seasons is shown in Fig. 3b. Salt input to the ocean is confined to the northern coastal polynya regions, and freshwater (negative salt) input is shown in most of other sea ice zones. The freshwater input is large, particularly in the southern Sea of Okhotsk. These imply that ice formed in the northern polynyas is advected to the south and subsequently melted there, which corresponds to negative heat transport from north to south, as shown by the annual mean net heat flux (Fig. 3a). The freshwater input is relatively large in the KB region and marginal ice zone of the central and eastern Sea of Okhotsk even though the annual mean net flux (Fig. 3a) shows a heat flux to the atmosphere. These features of the heat flux and freshwater input are considered to be explained by ice melting caused by the advection of warmer water: upwelled water for the KB case, and horizontally advected water for the case of the central and eastern Sea of Okhotsk.

Nonzero values of the annual mean net heat flux (Fig. 3a) are considered to reflect sea ice processes (freezing, advection, and melting) or the oceanic advection. To extract the component of sea ice processes, we calculated the heat flux that corresponds to the annual salt input by ice production and melting (Fig. 3b) from Eq. (5) using $V_f$ and $V_m$ values, and show it in Fig. 3d. The difference between the annual mean heat flux (Fig. 3a) and this sea ice component (Fig. 3d) is considered to represent the heat flux component by the oceanic advection, and is shown in Fig. 3e. This figure clearly demonstrates the effect of the upwelled warmer water in the KB region and the effect of the warmer water advection in the central and eastern Sea of Okhotsk.

Next, we show the annual salt input caused by ice production and melting separately. The freshwater flux associated with ice melting was estimated from the decrease in ice concentration [Eq. (7)]. At the subgrid cell scale (10–20 km), ice melting can be roughly classified into surface melting by direct heat input to the ice from the atmosphere ($\alpha$) and lateral and bottom melting by heat from the ocean ($\beta$). In this study, $\alpha$ was neglected based on the heat flux calculation. The ice melting $\beta$ can be classified in terms of heat source as follows: local heat input into the upper ocean from the atmosphere through open water fraction in the grid cell ($\beta_1$), and horizontal advection of heat ($\beta_2$) and vertical advection of heat ($\beta_3$). Here, we roughly separate the freshwater flux based on whether the melting is caused by the local heat input from the atmosphere ($\beta_1$) or not (advection; $\beta_2$ or $\beta_3$). The classification of $\beta_1$ is defined by $Q > 0 \text{ W m}^{-2}$ and $A > 30\%$, and the ice melting is assumed to be caused by the local heat input. The classifications of $\beta_2$ and $\beta_3$ correspond to the following cases: ice melting is caused by the advection of ice floes from the sea ice zone to the relatively warm open ocean, or ice melting is caused by horizontal or vertical advection of oceanic heat even if $Q < 0 \text{ W m}^{-2}$.

As an example, we show the annual salt and freshwater input for the 2002/03 season. Salt input is shown in most of the sea ice zone with a maximum in the coastal polynya regions (Fig. 4a). Using the scheme described above, freshwater input is classified into that caused by local heat input from the atmosphere ($\beta_1$; Fig. 4b) and that caused by advection ($\beta_2$ and $\beta_3$; Fig. 4c). The freshwater input by local heat input is distributed somewhat evenly over the sea ice zone (Fig. 4b). In contrast, large freshwater input by advection is shown locally, specifically in the KB region and near the ice edge in the central and eastern Sea of Okhotsk (Fig. 4c). In the KB region, the area of the large freshwater input (Fig. 4c) corresponds well with the area within the 200-m isobar (Fig. 1). This finding supports that the ice melting is caused by warmer water upwelled from the deeper ocean due to strong tidal
currents associated with the characterized bottom topography (Ono et al. 2006). The data for other years also show similar results (not shown).

The seasonal evolution of the salt and freshwater input associated with ice production and melting for the 2002/03 winter season are shown in Fig. 5. During the ice advance season (from December to February), the salt input (ice production) is shown in most of the sea ice zone (Fig. 5a). The maximum salt input is shown in the NWS polynya region. From a comparison of the location of the ice edge on the first and last day of each month, the salt input is predominantly shown in the area where the ice advances. From January to March, prominent freshwater input (ice melting) by advection is shown in the KB region and the ice edge areas (Fig. 5c). It is likely that the location of ice edge is significantly determined by this ice melting. From an ice–ocean coupled model in the Sea of Okhotsk, it was shown that a large amount of ice melting occurs around the ice edge during winter (Watanabe et al. 2004). In March, ice melting by local heat input from the atmosphere starts from the south (Fig. 5b). In April, when the ice retreat is largest, the freshwater input by the local heat input is predominant in most of the sea ice zone (~70% of the total ice melt; Fig. 5b). It is noted that relatively prominent freshwater input is shown in the NWS coastal polynya region (Fig. 5d), where the maximum salt input is shown in winter (Fig. 5a).

The coastal polynya becomes open water by April; consequently, solar radiation is efficiently absorbed in the polynya (open water) region, since the surface albedo of water is much lower than that of ice. Because ice floe is advected by wind, the thick ice that remains in offshore areas can be brought to the heated coastal polynya region due to the influence of atmospheric disturbances with a time scale of several days. Thus, the amount of ice melting is considered to be large because ice melting is effectively caused by heat stored in the upper ocean. This indicates that the polynya works as a “meltwater factory” in spring, in contrast to its role as an “ice factory” in winter, as suggested by Ohshima et al. (1998) and Morales Maqueda et al. (2004).

5. An application of the dataset: Relationship between ice melting and spring blooming

In the marginal ice zone of the Arctic and Antarctic Oceans, phytoplankton blooming is associated with ice melting (Alexander and Niebauer 1981; Smith and Nelson 1985). Although the relationship between ice melting and biological activity has not been understood well, the following factors may be important: 1) intensification of the upper-ocean stratification associated with the freshwater supply, 2) release of algae from the ice to the upper ocean (seeding effect), 3) increase in solar illumination, and 4) input of trace nutrients (e.g., iron) into the upper ocean (Smith and Nelson 1985; Sullivan et al. 1993).

Here, we compare the distribution of ice melting with the satellite chlorophyll a concentration, as an application of the present dataset. In the Sea of Okhotsk, Nakatsuka et al. (2004) suggested that the primary production
FIG. 5. Monthly salt and freshwater input associated with (a) ice production and (b), (c) melting from December 2002 to April 2003. The freshwater input associated with ice melting is divided into (b) that caused by local heat input from the atmosphere and (c) that caused by advection. (d) Sum of (a)–(c). Dotted and solid lines indicate ice extent of the month’s first and last day, respectively.
(diatom bloom) is regulated by both ice melting and the Amur River discharge from data collected by sediment traps off Sakhalin. From hydrological and satellite observations, phytoplankton blooming is shown to start precisely after ice retreat, suggesting the effect of ice melting (Zakharkov et al. 2007; Zenkin et al. 2009).

Figure 6a shows maps of monthly freshwater input associated with ice melting in April from 2003 to 2005. In the northwestern Sea of Okhotsk, the amount of ice melting tends to be locally large at the NWS coastal polynya region. Interannual variability of the spatial distribution of the prominent ice melt area is large. Figure 6b shows maps of monthly-mean chlorophyll $a$ concentration in May from Aqua MODIS (Feldman and McClain 2010).

If the correspondence between the ice melting and chlorophyll $a$ concentration is true, then the following scenario is possibly proposed for the phytoplankton blooming in the NWS coastal polynya region. Freshwater supply to the upper ocean associated with ice melting causes the intensification of the upper-ocean stratification. Subsequently, the depth of the surface mixed layer becomes shallower than the euphotic zone, and this leads to effective photosynthesis. The upper-ocean stratification in the coastal polynya region is expected to be well maintained because the effect of waves and swells from the open ocean is damped down by thick ice floes that remain offshore of the coastal polynya region. The release of nutrients and algae (seeding effect) from the ice and the increase in solar illumination associated with the disappearance of ice are considered to be other causes for the correspondence between the ice melting and chlorophyll $a$ concentration (Smith and Nelson 1985; Sullivan et al. 1993).

In the Northeast Water Polynya region of the Arctic, phytoplankton growth is suggested to be accelerated by the intensification of upper-ocean stratification associated
with ice melting (Gradinger and Banmann 1991). Chlorophyll a concentrations from satellites also show that the phytoplankton blooming occurs in the coastal polynya regions, in both the Arctic and Antarctic Oceans (Tremblay and Smith 2007; Smith and Comiso 2008). In each coastal polynya, the physical processes related to phytoplankton blooming are considered to differ because of differences in the large-scale physical processes between the atmosphere and ocean, as summarized in Tremblay and Smith (2007). The freshwater flux dataset associated with the sea ice melting would be potentially useful for such biogeochemical studies in other ice-covered oceans.

6. Discussion and conclusions

The heat flux was calculated based on the method proposed by Ohshima et al. (2003, 2006). For better estimation of the heat flux, the ice concentration from SSM/I was changed to that from AMSR-E with a validated thin-ice algorithm proposed by Nihashi et al. (2009). The spatial resolution of AMSR-E (~12.5 km) is about 2 times finer than that of SSM/I (~25 km). Thus, the coastal polynya regions can be resolved much better than that in Ohshima et al. (2003), and the higher ice production close to the coast is presented. However, the accumulation of the AMSR-E data is still insufficient to discuss the long-term variation because the data are available only from June 2002. The daily SSM/I data are available from July 1987 and have been accumulated until the present day. The heat and salt flux dataset for more than 20 yr might be created using the SSM/I data, although the spatial resolution and the reliability would be rather worse.

From monthly SMMR and SSM/I ice concentration for ~30 yr of 1979–2008 (Cavaliere et al. 1996), the maximum ice area in the Sea of Okhotsk was small during 2004–08; the maximum ice area averaged from 2004 to 2008 is about ~70% of that averaged for the ~30 yr. The minimum was recorded in 2006. Thus, the averaged fluxes shown in the present study (Fig. 3) might be biased to those of less sea ice year.

The dataset of the present study is largely based on ERA-Interim. Because sea ice data with high spatial resolution were taken into consideration in our dataset, the difference of the net heat fluxes between our dataset and ERA-Interim is relatively large in the active freezing season (November–February); the net heat flux from our dataset averaged over the entire Sea of Okhotsk from the 2002/03 to 2009/10 seasons is approximately ~190 W m⁻², while that from ERA-Interim is approximately ~160 W m⁻². The difference in other season is <10 W m⁻². Maps of the net heat fluxes during the active freezing season are shown in Fig. 7. The heat flux of the present study shows larger heat loss to the atmosphere in the coastal polynya regions and the marginal ice zone, indicating that the spatial resolution of the ERA-Interim data (1.5° × 1.5°) is not enough to resolve a coastal polynya whose typical width is <100 km and sea ice distribution in the marginal ice zone (Figs. 2b and 2d).
For the calculation of the salt flux, the ice production was quantitatively estimated from the heat loss to the atmosphere. In contrast, ice melting cannot be estimated from the heat flux; the melting was indirectly estimated from the decrease in ice concentration by assuming constant ice thickness such that the total ice melting is balanced by the total ice production. This assumption is the largest ambiguity of the present study, and thus there is some uncertainty in the spatial distribution of freshwater flux presented in this study. In section 4, it was shown that ice melting in April is mostly caused by local heat input into the upper ocean from the atmosphere. Thus, if the method for estimating ice melt of this study is reasonable, then the amount and distribution of ice melting estimated from the decrease in ice concentration should be similar to those of ice melting that would be done by the local heat input. The amount of ice melting estimated from the decrease in ice concentration (monthly sum of $V_m$) and the amount of ice melting corresponding to the local heat input [calculated from the Eq. (5)] are shown for the case of April 2003 in Figs. 8a and 8b, respectively. The amounts of ice melting averaged over the sea ice zone are $\sim$0.26 and $\sim$0.31 m, respectively. The nearly coincidence of these values and the similar spatial distributions from the independent data might partly support the validity of the estimation of ice melting. The data of other years also show similar results to those of Figs. 8a and 8b (not shown). In the future, the estimation of the amount of ice melting will be improved by introducing more realistic distributions of ice thickness and snow depth from the laser and radar altimeter systems on the Ice, Cloud, and Land Elevation Satellite (ICESat)–second-generation ICESat (ICESat-2) and the second Cryosphere Satellite (CryoSat-2), respectively.

In this study, the daily ice drift speed estimated from AMSR-E brightness temperature was used to explicitly treat the effect of ice advection. In the active ice melt region, ice motion retrievals from passive microwave data are unreliable because of the decorrelation of the passive microwave data resulting from rapid sea ice melting (Kwok et al. 1998). Thus, we also estimate the amount of ice melting using the ice drift speed calculated using a different method. Because ice drift is forced predominantly by the geostrophic wind determined from the surface sea level pressure (SLP) pattern (Kwok et al. 1998), wind-forced ice drift is calculated from the geostrophic wind based on SLP from ERA-Interim, where the ice drift is assumed to be 1.5% of the wind speed and directed 18° to the right (Thorndike and Colony 1982; Kimura and Wakatsuchi 1999). The annual salt input calculated using wind-forced ice drift for the melting is shown in Fig. 3c. In this case, the assumed constant ice thickness $[\overline{h_i}$ in Eq. (7)] is 0.33 m, similar to 0.34 m of the salt flux calculation using the ice drift from AMSR-E. The spatial distribution of ice melting and the $\overline{h_i}$ value estimated using the wind-forced ice drift are similar to those estimated using ice drift from AMSR-E. The wind
factor (a ratio of the ice speed to wind speed) of 1.5% is not always applicable and has a range of 1%–2% in the Sea of Okhotsk (Kimura and Wakatsuchi 2000). We also calculated the amount of ice melting using wind factors of 1% and 2%, and then $h_i$ becomes 0.34 and 0.32 m, respectively. The spatial distributions of the ice melting do not change much (not shown).

The annual mean net heat flux (Fig. 3a) and the annual salt input (Fig. 3b) suggested the transport of freshwater and negative heat from north to south by ice advection associated with the prevailing northerly wind and the southward East Sakhalin Current. From the dataset of the present study, ice production in the NWS and northern shelf coastal polynya regions (Fig. 1) averaged from the 2002/03 to 2009/10 seasons is estimated to be $\sim 4.7 \times 10^{11}$ m$^3$. In the area surrounded by thick dotted lines south of 49°30′N (Fig. 1), a total ice melting of $\sim 2.9 \times 10^{11}$ m$^3$ is estimated. In contrast, from the direct observations of ice thickness and drift speed in the coastal region off northern Sakhalin, the amount of ice transported from north to south is estimated to be $3.1 \times 10^{11}–7.3 \times 10^{11}$ m$^3$ (Fukamachi et al. 2009). These estimations show that the amounts of ice production in the northern coastal polynya regions, the southward transport of ice, and the ice melting in the southern Sea of Okhotsk are roughly comparable, which is consistent with the scenario that freshwater and negative heat are transported from north to south by ice advection.

In the southern Sea of Okhotsk, where sea ice melts, annual freshwater input in a unit area associated with ice melting is estimated to be $\sim 1$ m (Fig. 3b). Climatological annual precipitation minus evaporation $(P - E)$ in a unit area in the southern Sea of Okhotsk is shown to be $\sim 0.5$ m from the precipitation data of the Global Precipitation Climatology Project (GPCP; Adler et al. 2003) and the evaporation data of the objectively analyzed air–sea fluxes (OAFlux; Yu and Weller 2007). The climatological annual amount of discharge from the Amur River was shown to be $3.3 \times 10^{11}$ m$^3$ (Ogi et al. 2001). Annual freshwater input in a unit area by the Amur River discharge in the western half of the Sea of Okhotsk (the area is $\sim 7.5 \times 10^{11}$ m$^2$), where the river discharge likely affect, is calculated to be $\sim 0.4$ m. These suggest that the annual freshwater input associated with ice melting is comparable to or larger than that associated with the $P - E$ and the Amur River discharge.

The heat and salt flux dataset of the present study is useful for validating ice–ocean (and ice–ocean–atmosphere) coupled models, particularly because the redistribution of heat and salt–freshwater contents by sea ice is quite important for thermohaline circulation in the ocean (e.g., Hasumi and Suginohara 1995; Marsland and Wolff 2001) and heat flux anomalies by sea ice significantly affect the atmospheric general circulation (e.g., Honda et al. 1999). The flux dataset can also be used as surface boundary conditions for ice-covered seas in ocean general circulation models (OGCMs). Once the heat and salt flux data associated with ice production and melting are given, OGCMs would be able to represent the thermohaline circulation caused by ice formation, even without a sea ice model. Furthermore, the present dataset is found to be useful for the investigation of the interactions between physical and biochemical processes through the sea ice, such as spring blooming after the ice disappearance. In this study, we have created the heat and salt flux dataset using ice concentration, thickness, and drift from AMSR-E with a relatively high spatial resolution in the Sea of Okhotsk, where we have an AMSR-E thin-ice thickness algorithm and a few validation data. In the Antarctic Ocean, creation of the flux dataset was tried using ice concentration and thickness from SSM/I with a relatively low spatial resolution (Tamura et al. 2011). In principle, it would be possible to develop a flux dataset with high spatial resolution for the Arctic and Antarctic Oceans, with the expectation of using ice thickness from ICESat–ICESat-2 and CryoSat-2 data in the future.

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APPENDIX

AMSR-E Thin-Ice Thickness Algorithm Combined with the Ice Concentration Data

Nihashi et al. (2009) proposed an equation that can estimate thin ice thickness $h_i$ of $\leq 0.2$ m from a comparison between the polarization ratio of AMSR-E brightness temperature at a 36.5-GHz channel PR$_{36}$ and ice...
thickness estimated from clear-sky AVHRR infrared images, as follows:

\[ h_i = -3.78PR_{36} + 0.50. \]  

(A1)

An example of an AMSR-E thin-ice thickness map is shown in Fig. 2c. Thin-ice regions corresponding to coastal polynyas are shown in the NWS region, north shelf region, Gizhiga Bay, coastal regions of northeastern Sakhalin, and Terpenia Bay. From a comparison with a map of AMSR-E ice concentration (Fig. 2b), thin-ice signals are shown also in the marginal ice zone with relatively low ice concentration. The AMSR-E thin-ice thickness algorithm proposed by Nihashi et al. (2009) focuses on coastal polynyas. Because winter coastal polynyas are almost covered with thin ice uniformly, the ice concentration is assumed to be 100% in Eq. (A1). In contrast, the ice condition in the marginal ice zone is not always similar to that in coastal polynyas. The PR36 value of open water is close to that of thin ice. Thus, the algorithm sometimes judges the area of lower ice concentration as the thin-ice area because of the effect of the PR36 value on the open water fraction. Here, we try to improve the AMSR-E thin-ice algorithm to use it even in the marginal ice zone with lower ice concentration.

The AMSR-E brightness temperature at a 36.5-GHz channel for a given grid cell (Tb) is assumed to be the sum of the brightness temperatures of the ice fraction (Tb_i) and that of the open water fraction (Tb_w) under the ice concentration (A), as follows:

\[ Tb = A(Tb_i) + (1 - A)Tb_w. \]  

(A2)

The value of Tb_i for the grid cell can be obtained from Eq. (A2) using the brightness temperature of the open ocean near the ice edge as Tb_w. In this study, we set reference points along the direction of ice advance at intervals of 1° latitude and longitude (Fig. 1). The presence of ice is examined at each point from the land side, and the brightness temperature of the first point judged as ocean is used as Tb_w. In (A2) Tb is obtained for both horizontal and vertical polarizations from Eq. (A2) using the AMSR-E NT2 ice concentration. From the PR36 value calculated using Tb_i values, we obtain the thin-ice thickness for which the effect of the open water fraction is excluded (Fig. 2d). In the coastal polynya region, the distribution of the corrected thin-ice thickness is similar to that of the uncorrected thickness (Fig. 2c). In the marginal ice zone, most of the thin ice shown in the uncorrected thickness map (Fig. 2c) has disappeared in the corrected thickness map (Fig. 2d). This suggests that the sea ice is misjudged as thin ice because of the effect of the PR36 value of the open water fraction. In the
previous studies, the effect of the open water fraction was excluded using the ice concentration, as in the present study [e.g., the algorithm for snow depth on ice proposed by Markus and Cavalieri (1998)].

The method presented here assumes that the AMSR-E NT2 ice concentration is true. Markus and Dokken (2002) compared the ice concentration from the SSM/I NT2 algorithm with that from synthetic aperture radar (SAR) with high spatial resolution in the late summer Arctic Ocean, and they showed a negative bias of <20% in the NT2 ice concentration in the marginal ice zone. Cavalieri et al. (2006) showed a negative bias of ~5% in the AMSR-E NT2 ice concentration in an area of new ice from comparisons with Landsat-7 images in the late winter Arctic Ocean. The results of these studies suggest that the accuracy of the NT2 ice concentration is less reliable in the spring marginal ice zone and in the winter new ice region, which was the focus of the present study. Even so, we consider that the use of the AMSR-E ice concentration is better than assuming 100% ice concentration.

Maps of the corrected AMSR-E thin-ice thickness in the southern Sea of Okhotsk are shown in Fig. A1. The maps show that part of the coastal region of Hokkaido was covered with thin ice from 9 to 11 February 2009. For this period, in situ observations of sea ice were made onboard the icebreaker Soya. The spatial distribution of ice thickness observed by hourly visual observations from the ship’s bridge, based on the Antarctic Sea Ice Processes and Climate (ASPeCt) protocol (Worby and Allison 1999), is shown in Fig. A2. A photograph of coastal Hokkaido taken from a helicopter deployed by the icebreaker Soya shows that the region was widely covered by thin ice (Fig. A3; the ice was nilas, with a thickness of approximately 2–10 cm). These in situ observations reveal that part of coastal Hokkaido was covered by thin ice during this period, confirming that the corrected AMSR-E thin-ice thickness algorithm performs well in the marginal ice zone.

REFERENCES


