Severe Cold Winter in North America Linked to Bering Sea Ice Loss

MIZUKI IIDA, SHUSAKU SUGIMOTO, AND TOSHIO SUGA

Department of Geophysics, Graduate School of Science, Tohoku University, Sendai, Japan

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ABSTRACT

North America experienced an intense cold wave with record low temperatures during the winter of 2017/18, at the time reaching the smallest rank of sea ice area (SIA) in the Bering Sea over the past four decades. Using observations, ocean reanalysis, and atmospheric reanalysis data for 39 winters (1979/80–2017/18), both the Bering SIA loss and cold winters in North America are linked robustly via sea level pressure variations over Alaska detected as a dominant mode, the Alaska Oscillation (ALO). The ALO differs from previously identified atmospheric teleconnection and climate patterns. In the positive ALO, the equatorward cold airflow through the Bering Strait increases, resulting in surface air cooling over the Bering Sea and an increase in Bering SIA, as well as surface warming (about 4 K for the winter mean) for North America in response to a decrease of equatorward cold airflow, and vice versa for negative phase. The northerly winds with the cold air over the Bering Sea result in substantial heat release from ocean to atmosphere over open water just south of the region covered by sea ice. Heating over the southern part of Bering Sea acts as a positive feedback for the positive ALO and its related large-scale atmospheric circulation in a linear baroclinic model experiment. Bering SIA shows no decreasing trend, but has remained small since 2015. CMIP6 climate models of the SSP5–8.5 scenario project a decrease of Bering SIA in the future climate. To explain severe cold winters in North America under global warming, it is necessary to get an understanding of climate systems with little or no sea ice.

1. Introduction

Surface warming in the Arctic is enhanced (Holland and Bitz 2003; Serreze and Francis 2006; Serreze et al. 2009; Screen and Simmonds 2010; Cohen et al. 2014; Barnes and Polvani 2015) with a rapid decline of Arctic sea ice (Comiso et al. 2008; Polyakov et al. 2012; Stroeve et al. 2012; Wang and Overland 2012; Perovich et al. 2017; Dai et al. 2019). This process is referred to as the Arctic amplification of global warming (Screen and Simmonds 2010; Serreze and Barry 2011). Contrary to the general prospect of global warming (IPCC 2013), extreme cold winters have frequently occurred in recent years in midlatitude regions of the Northern Hemisphere (Cohen et al. 2014; Horton et al. 2015; Johnson et al. 2018). These are induced by large-scale atmosphere circulation patterns, similar to Arctic Oscillation (AO)

and blocking events, which are caused by Arctic sea ice loss (Screen and Simmonds 2013, 2014; Cohen et al. 2014; Vihma 2014; Overland et al. 2015), such as central Eurasian cooling attributable to sea ice loss in the Barents–Kara Seas (Honda et al. 2009; Petoukhov and Semenov 2010; Inoue et al. 2012; Nakanowatari et al. 2014; Sato et al. 2014; Mori et al. 2014, 2019; Chen et al. 2016) and eastern Asian cooling from an ice-free Chukchi Sea (Tachibana et al. 2019).

North America also has experienced extreme cold winters over the past 10 years (Wallace et al. 2014). These extreme conditions are related to an increased waviness of westerly jet streams. Some studies pointed out that the wavier jet stream is secondary in effect to the Arctic amplification of global warming (Francis and Vavrus 2012, 2015; Liu et al. 2012; Tang et al. 2013), whereas other studies suggest that changes to the jet stream result from natural variability (Barnes 2013; Barnes et al. 2014; Hassanzadeh et al. 2014). Recent atmospheric general circulation model experiments demonstrated that the wavier jet stream that triggered the severe winter in 2013/14 in North America was caused by the retreat of sea ice in the Bering Sea (Fig. 1)

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Corresponding author: Shusaku Sugimoto, shusaku.sugimoto.d7@tohoku.ac.jp

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FIG. 1. Winter climatology of sea ice concentration from COBE-SST (shading) and sea level pressure from JRA-55 (contours); see details in text. The orange rectangle represents the Bering Sea region (BSR; 55°–65°N, 165°E–155°W).

(Lee et al. 2015). The sea ice area (SIA) in the Bering Sea is correlated with the winter surface air temperature (SAT) in North America (Nakanowatari et al. 2015). In recent years, the effects of Bering SIA on winter conditions in North America have gained attention. Four major factors have been suggested as potential causes of winter Bering SIA variations. The first is local winds (Niebauer 1980, 1998; Walsh and Sater 1981; Cavalieri and Parkinson 1987; Fang and Wallace 1998; Honda et al. 1999; Deser et al. 2000; Sasaki and Minobe 2005; Yamamoto et al. 2006; Ukita et al. 2007), as the SIA tends to be large when northerly wind anomalies occur. The second is storms (Overland and Pease 1982), as frequent storms in the Bering Sea are related to large SIA. The third is ice influx from the Arctic Ocean (Zhang et al. 2010). The fourth is high sea surface temperature (SST) in the Bering Sea resulting from heat advection through the Alaskan Coastal Current forced remotely by atmospheric circulation changes from subtropical regions during the previous summer (Nakanowatari et al. 2015), as warm SSTs do not favor SIA increase. However, the dominant control on variations in Bering SIA has yet to be determined.

The Aleutian low (AL) is located in the central North Pacific during winter (Fig. 1). Some studies denoted that the North Pacific Oscillation (NPO)–west Pacific (WP) pattern, which represents a meridional movement of the AL (Sugimoto and Hanawa 2009), results in a seesaw fluctuation of SIA between the Bering and Okhotsk Seas in the northwestern North Pacific (Fang and Wallace 1998; Honda et al. 1999). However, other studies have reported that the SIA in these two regions often fluctuates in phase (Liu et al. 2007; Linkin and Nigam 2008). It has been suggested that the NPO–WP only influences Bering SIA (Matthewman and Magnusdottir 2011). Relationships among the Bering Sea, Okhotsk Sea, and

North America and air–sea–ice coupled system connecting them are not fully understood.

In this study, we focus on the Bering Sea, which is located between the Okhotsk Sea and North America. We investigate the temporal behavior of Bering SIA and its cause using observational, ocean reanalysis, and atmospheric reanalysis data. We investigate large-scale atmospheric circulation patterns as causes of Bering SIA variations, and explore whether Bering SIA variations can act as a positive feedback on the atmosphere or not conducting a numerical experiment. Here, we present a new perspective on the relationships among Bering SIA, Okhotsk SIA, and winter conditions in North America.

2. Data and methods

a. Gridded data

We use five monthly-mean sea ice concentration dataset: 1) the Centennial In Situ Observation-Based Estimates of the Variability of SST and Marine Meteorological Variables (COBE-SST) product (Ishii et al.

FIG. 2. Winter climatology of CAM amount (shading) and CAM flux (vectors) below $\theta_T = 280$ K.

FIG. 3. (a) Winter SIA time series, for the BSR from five datasets: COBE-SST, ERAI, CFSR, ORAS5, and OISST. (b) The Morlet wavelet transform coefficient for the normalized winter Bering SIA time series from COBE-SST in (a). Shading indicates the amplitude of the real part of the wavelet coefficient. The black line represents the local wavelet spectra, which are defined as the square of the absolute wavelet transform coefficient, and exceed the 10% significance level. The significance level of the wavelet amplitude is evaluated using a Monte Carlo simulation based on a red noise [AR(1)] model for the observed lag -1 correlation coefficient using a 10 000-point surrogate time series. The light shading represents the cone of influence where edge effects become important.

2005), which provides boundary conditions for the Japanese Meteorological Agency (JMA) Japanese 55-year Reanalysis (JRA-55) product (Kobayashi et al. 2015) on a 1.25° longitude $\times$ 1.25° latitude grid; 2) boundary conditions from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim global atmospheric reanalysis product [ERA-Interim (ERAI); Dee et al. 2011] on a $0.75^\circ \times 0.75^\circ$ grid; 3) the National

FIG. 4. The first EOF mode for winter sea ice concentration over the BSR. (top) Time coefficient (black line) with the percentage of explained variance in parentheses, and the winter Bering SIA time series (red dashed line). (bottom) Distribution of regression coefficients between the time coefficient and winter sea ice concentration field with black lines denoting regions exceeding the 5% significance level. The orange dashed rectangle in the bottom panel represents the BSR.

Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) product (Saha et al. 2010, 2014) on a $0.5^\circ \times 0.5^\circ$ grid; 4) the National Oceanic and Atmospheric Administration (NOAA) optimum interpolation SST (OISST) product (Reynolds et al. 2007) on a $0.25^\circ \times 0.25^\circ$ grid; and 5) the ECMWF Ocean Reanalysis System 5 (ORAS5) product (Zuo et al. 2018) on a $0.25^\circ \times 0.25^\circ$ grid. Three monthly-mean SST datasets are also used: 1) the COBE-SST product, 2) the OISST product, and 3) the ORAS5 product.

For monthly atmosphere variables of SAT, sea level pressure (SLP), surface pressure, surface winds, surface latent heat flux, surface sensible heat flux, and geopotential heights at 250, 500, and 850 hPa (Z250, Z500, and Z850), we use three global atmospheric reanalysis products: 1) the JRA-55 product, 2) the ERAI product, and 3) the CFSR product. In the following sections, we show only results using JRA-55, as ERAI and CFSR showed almost identical results. We use monthly net diabatic heating rate in vertical from the JRA-55. These are the sum of large-scale condensation heating rate, convective heating rate, vertical diffusion heating rate, solar radiative heating rate, and longwave radiative heating rate. The vertical diffusion heating rate and the sum of the large-scale condensation and convective heating rates are referred to as the sensible heating and latent heating, respectively.

FIG. 5. (a) Winter sea ice concentration in (top) large SIA periods (1995, 1998, 2000, 2002, 2006, 2009, 2012, 2013) and (middle) small SIA periods (1985, 2001, 2015, 2017, 2018), and (bottom) the difference between the two periods. The orange rectangle denotes the BSR. Black lines in the bottom row represent regions exceeding the 95% confidence limit. (b),(c) As in (a), but for surface pressure (white contour with an interval of 4 hPa in regions <992 hPa) and Z500, respectively. Arrows in the bottom row of (b) represent the difference in surface winds; arrows smaller than 2 m s^{-1} are not shown for clarity.

Our analysis period is the 39 winters (December–February) during 1980–2018, and climatological value is an average over the 39 winters. Note that, in the time series figures, winter values are plotted as the calendar year for midwinter conditions (January–February); for example, winter values for December 1999–February 2000 are plotted as the year 2000.

b. Teleconnection pattern index and climate index

As an indicator of large-scale organized atmospheric perturbations, we use the following indices: the AO index (a leading mode of the empirical orthogonal function (EOF) analysis of SLP; Thompson and Wallace 1998); the North Pacific index (NPI; the SLP averaged within the area of 30° – 65°N , 160°E – 140°W ; Trenberth

and Hurrell 1994), which indicates changes in AL intensity (Sugimoto and Hanawa 2009); the Niño-3.4 index (SST anomalies averaged for 5°S – 5°N , 170° – 120°W ; Trenberth 1997), which is used as an indicator of El Niño events; and eight atmospheric teleconnection pattern indices (Barnston and Livezey 1987; Bell and Halpert 1995): the North Atlantic Oscillation (NAO), east Atlantic (EA), WP, Pacific–North American (PNA), east Atlantic–west Russian (EAWR), Scandinavia (SCAND), tropical–Northern Hemisphere (TNH), and polar–Eurasia (PE) patterns.

c. Cold-air outbreaks

Cold-air outbreaks are sudden intrusions of polar cold air masses into the midlatitudes and bring severe winters.

TABLE 1. Correlation coefficients of eight teleconnection pattern indices, three climate indices, and the Bering SIA time series (B-SIA; red line in Fig. 3a) to the Bering SIA time series and the leading three modes of SLP (SLP1, SLP2, SLP3). Values in boldface type exceed the 5% significance level.

	Teleconnection pattern index								Climate index			
	TNH	PNA	WP	EAWR	NAO	EA	PE	SCAND	NPI	AO	Niño-3.4	B-SIA
B-SIA	−0.35	−0.03	0.06	0.11	−0.08	−0.09	0.30	−0.07	−0.02	0.10	0.03	
SLP1	0.64	−0.51	0.46	−0.16	0.35	0.04	−0.23	−0.10	0.71	0.35	−0.60	−0.01
SLP2	−0.24	0.77	0.41	0.25	−0.20	0.28	0.18	0.01	−0.58	−0.20	0.43	−0.08
SLP3	−0.38	0.08	0.32	0.30	0.13	0.04	−0.13	−0.16	−0.04	0.13	0.20	0.67

FIG. 6. (a) First, (b) second, and (c) third REOF modes extracted for winter SLP over a domain covering the North Pacific and Alaska. (top) Time coefficient. The percentage of explained variance is represented in parentheses. (bottom) Distribution of regression coefficients between the time coefficient and the winter SLP field with black lines denoting regions exceeding the 5% significance level. The brown dashed rectangle in the bottom panels represents the domain.

We use the recently derived cold airmass (CAM) index (Iwasaki et al. 2014), which is regarded as a good indicator of cold-air outbreaks (Yamaguchi et al. 2019). The CAM amount (DP) is the vertical (pressure) depth of the atmosphere below a designated potential temperature θ_T , and is calculated as follows:

$$DP \equiv p_s - p(\theta_T), \quad (1)$$

where p_s is the ground surface pressure, and $p(\theta_T)$ is the pressure on isentropes $\theta = \theta_T$. We set $\theta_T = 280$ K, following previous studies (e.g., Iwasaki et al. 2014; Abdillahi et al. 2017; Kanno et al. 2017; Yamaguchi et al. 2019). The CAM flux is given by

$$\mathbf{F} \equiv \int_{p(\theta_T)}^{p_s} \mathbf{v} dp, \quad (2)$$

where $\mathbf{v}$ is the horizontal wind vector. We calculated the CAM amount and CAM flux using the 6-hourly JRA-55 product and the 6-hourly ERAI product. We checked that the results are insensitive to the product and θ_T , revealing that almost identical results were obtained using other values such as 285 K.

Cold-air outbreaks have two major equatorward streams (Iwasaki et al. 2014): the East Asian stream and the North American stream (Fig. 2). Figure 2 shows another equatorward CAM stream through the Bering Strait. The recently defined CAM index allows us to

FIG. 7. Winter (a) CAM amount and (b) SAT in (top) the positive ALO phase (1983, 1992, 1998, 1999, 2006, 2008, 2009, 2012, 2013) and (middle row) the negative ALO phase (1985, 1993, 2001, 2014, 2015, 2018), and (bottom) the difference between the two phases. Black lines in the bottom row represent regions exceeding the 95% confidence limit. Arrows in (a) represent the CAM flux. Arrows smaller than 400 hPa m s⁻¹ in the top and middle rows and 200 hPa m s⁻¹ in the bottom row are not shown for clarity.

provide a new perspective on Bering SIA and its relationship with the winter climate.

d. Degrees of freedom for correlation analysis

The degrees of freedom are estimated by dividing the data length by the characteristic time scale (a zero-cross scale) calculated according to a lagged autocorrelation function. We evaluate statistical significance using a two-sided Student's *t* test.

3. Temporal behavior of SIA in the Bering Sea

We investigate the temporal behavior of SIA in the Bering Sea region (BSR; 55°–65°N, 165°E–155°W; orange rectangle in Fig. 1). The winter Bering SIA time series shows an increase and a decrease of up to $300 \times 10^3 \text{ km}^2$ in the five SIA datasets (Fig. 3a), with a long-term period of 10–15 years (Fig. 3b). The SIA increases in the early 1990s and around 2010, and decreases in the mid-1980s and early 2000s. In addition to this low-frequency variation, a high-frequency variation with a

period of about 3 years is identified after 2000 (Fig. 3b). Bering SIA does not show a significant decreasing trend over the past four decades, although overall sea ice cover in the Arctic sea is decreasing (e.g., Comiso et al. 2008). Interestingly, Bering SIA has been small since 2015 and three of the five smallest-SIA years have occurred since 2015, with 2018 having record-breaking small SIA. In the following analyses, we show only results using COBE-SST, as all datasets yielded similar results.

We examine the usefulness of the regional-mean Bering SIA time series. Figure 4 displays the results of an EOF analysis using a covariance matrix for winter sea ice concentrations. The first mode, which explains most of the total variance (64%), is dominated by variance in the eastern part of the Bering Sea. Note that significant variability in the Okhotsk Sea is not detected in the long-term analysis for the past 39 winters, although previous work pointed out relationships between sea ice variability in the Bering and Okhotsk Seas (Fang and Wallace 1998; Liu et al. 2007; Linkin and Nigam 2008).

FIG. 8. (a) Winter sea ice concentration in (top) the positive and (middle) the negative ALO phase, and (bottom) the difference between the two phases. Black lines represent regions exceeding the 95% confidence limit. (b),(c) As in (a), but for upward sensible heat flux and upward latent heat flux, respectively.

The time coefficient is well correlated with the regional-mean Bering SIA time series, with a correlation coefficient R of 0.99. It is reasonable to accept that the regional-mean values well represent SIA variations in the Bering Sea.

We explore the spatial variability of sea ice concentrations in the Bering Sea by performing a composite analysis. We define two categories based on the Bering SIA time series (red line in Fig. 3a): during large (small) SIA periods, the SIA anomaly is greater (less) than plus (minus) one standard deviation. We checked that nearly identical results were obtained in the cases in which data since 2015 (the small Bering SIA) were included or excluded. During small SIA periods, there is little sea ice extent in the Bering Sea (middle row in Fig. 5a). A comparison between the two periods shows a well-defined monopole structure in the eastern part of the Bering Sea (bottom row in Fig. 5a), consistent with the EOF results (Fig. 4).

4. Large-scale atmospheric circulation linking Bering SIA loss and severe winter in North America

To detect the cause of Bering SIA variations, we explore atmospheric fields by performing a composite

analysis. A comparison of SLP between the two periods (bottom row in Fig. 5b) shows an east–west seesaw pattern whose boundary approximately corresponds to the Bering Strait (around 170°W), accompanied by northerly wind anomalies through the strait. This means that local northerly winds exert a large influence on Bering SIA variations, as pointed out by previous works (e.g., Niebauer 1980; Walsh and Sater 1981; Honda et al. 1999; Ukita et al. 2007). The amplitude and size of the negative SLP anomaly east of the Bering Strait, over Alaska, are much larger than those of the positive anomaly west of the Bering Strait (Fig. 5b). Thus, the surface pressure anomaly around Alaska is primarily responsible for Bering SIA. We next investigate the vertical structure of the pressure pattern over Alaska. Figure 5c represents result for the middle troposphere, revealing negative height anomalies (i.e., cyclonic circulation anomalies) over Alaska during large SIA periods. The similar patterns are detected for the upper and lower troposphere (not shown). This indicates that the pressure pattern over Alaska has a primary equivalent-barotropic structure in the vertical direction.

To understand whether the pressure pattern over Alaska inducing Bering SIA variations is related to previously identified climate and atmospheric teleconnection patterns, such as NPO–WP, as pointed out by

Difference in Geopotential Height

(a) Z250

(b) Z500

(c) Z850

FIG. 9. Differences in winter (a) Z250, (b) Z500, and (c) Z850 between the positive and negative ALO phase. Black lines represent regions exceeding the 95% confidence limit. Arrows represent the wave activity flux formulated by Takaya and Nakamura (1997, 2001). Arrows smaller than $1 \text{ m}^2 \text{ s}^{-2}$ are not shown for clarity.

previous works (Walsh and Johnson 1979; Fang and Wallace 1998; Liu et al. 2007; Linkin and Nigam 2008; Matthewman and Magnusdottir 2011), we investigate potential relationships of Bering SIA with climate/teleconnection pattern indices. Results clearly show that Bering SIA is not correlated with any previously identified pattern (Table 1).

Next, we explore whether atmospheric variations over Alaska that lead to Bering SIA variations can be represented by a dominant mode by performing a rotated EOF (REOF) analysis that was used to identify the teleconnection patterns (Barnston and Livezey 1987; Bell and Halpert 1995). The REOF analysis is performed using a varimax orthogonal rotation, which is

widely accepted as the most accurate analytic algebraic orthogonal rotation (Kaiser 1958, 1959; Richman 1986). This analysis was applied to the first 20 EOF modes (which account for more than 95% of the original variance) extracted using a covariance matrix of winter SLP over a domain covering the North Pacific and Alaska (brown dashed rectangle in Fig. 6). The first mode (SLP1; Fig. 6a) is dominated by variance in the north-eastern North Pacific, and the time coefficient is highly correlated with the TNH index, NPI, and Niño-3.4 index (Table 1); El Niño influences the AL intensity (Zhang et al. 1996) and can form atmospheric circulation patterns that are similar to the TNH (Barnston and Livezey 1985; Mo and Livezey 1986). In the second mode (SLP2; Fig. 6b), larger variances are found in the central North Pacific, and the time coefficient is correlated with the PNA index and NPI (Table 1); both the PNA and NPI well represent changes in AL intensity (Sugimoto and Hanawa 2009). The first two leading modes are not correlated with Bering SIA (Table 1), revealing their independence of one another. The third mode (SLP3; Fig. 6c) exhibits a large variance around Alaska, which is spatially similar to the surface pressure pattern related to Bering SIA variations (Fig. 5b). The time coefficient is highly correlated with the Bering SIA time series (Table 1). Furthermore, it is found that SLP3 is not significantly correlated with any previously identified atmospheric index. The explained variance of SLP3 (13.7%) is comparable to those of major teleconnection patterns (14.4% for SLP1 and 13.7% for SLP2). This strongly suggests the importance of SLP variations over Alaska. In the following subsections, we denote SLP3 as the Alaska Oscillation (ALO) for convenience, investigate surface atmosphere and ocean responses attributable to the ALO, and explore large-scale atmospheric circulation patterns associated with the ALO.

a. Surface atmosphere and ocean responses in the Bering Sea to ALO

The equatorward CAM flow exists through the Bering Strait (Fig. 2) in the western part of the ALO. We investigate the effects of the ALO on the surface atmosphere and ocean in the Bering Sea in the viewpoint of CAM. We define two additional categories (positive and negative ALO phases) using the time coefficient of the ALO; in the positive (negative) ALO phase, the time coefficient is greater (less) than plus (minus) one standard deviation. Large-scale pattern of CAM and CAM flow are very similar to those in positive and negative ALO phases (Fig. 7a), but there is a significant difference locally around the Bering Strait between the two phases; an equatorward CAM flow through the Bering Strait strengthens, resulting in an increase in CAM

FIG. 10. (a) Differences in vertical profile of winter net diabatic heating rate (black), sensible heating rate (red), latent heating rate (blue), and the sum of solar and longwave radiative heating rates (pink) between the positive and negative ALO phases over the southern region of the Bering Sea (58.75° – 61.25° N, 175° E– 170° W). (b) Z500 response for 30–40 days in LBM experiment using idealized heating over the Bering Sea. Gray contours represent positive heating anomalies at a model sigma level of 0.98, with a contour interval of 2 K day^{-1} ; the zero contour is omitted. The $\times$ symbol represents a central position of the idealized heating. (c) Vertical profile of idealized heating on the sigma coordinate at the central position of the heating set in LBM experiment [$\times$ symbol in (b)].

amount over the Bering Sea in positive ALO phase. The SAT, in the positive ALO phase, decreases over the Bering Sea (Fig. 7b), consistent with the CAM (Fig. 7a) and sea ice pattern (Fig. 8a). Figure 8a provides confirmation

that Bering Sea ice is not related to Okhotsk Sea ice, and this is reasonable because of the absence of ALO-related atmospheric variations around the Okhotsk Sea (Fig. 6c).

FIG. 11. Severe cold event in winter of 2017/18. (a) Daily SAT time series in Minnesota in the United States [45°N , 92.5°W ; purple circle in (b)]. (b) SAT anomalies and (c) Z500 anomalies, averaged during the cold event [shaded period in (a)]. The daily climatological data were obtained by applying a 31-day running-mean filter to calendar-day means throughout the analysis period.

It is expected that northerly winds with cold air would influence the heat exchange between the ocean and atmosphere over open water in the southern part of the Bering Sea. We explore the upward sensible and latent heat fluxes for positive and negative ALO phases. A comparison between the two phases reveals an increase of heat release from the ocean to the atmosphere of 100 W m^{-2} in a narrow zonal band (55° – 60°N) of open water just south of the region covered by sea ice during positive ALO phase (upper row in Figs. 8b,c). The sensible heat flux is about 3 times larger than the latent heat flux because of the cold northerly winds. Negative heat flux anomalies are seen in a region north of the positive anomalies. These indicate the role of sea ice as

thermal insulation; that is, no upward heat release exists over the northern part of the Bering Sea in positive ALO phase because of the sea ice coverage. The amplitude and size of the northern region are considerably smaller than those of the southern region (about one-third for sensible heat flux).

b. Severe cold winters in North America due to negative ALO

We examine vertical structures of ALO. A composite map shows negative height anomalies throughout the troposphere over Alaska (Fig. 9), revealing the primary equivalent-barotropic structure in the vertical direction, consistent with the composite results for Bering SIA (Figs. 5b,c). Interestingly, there are positive height anomalies over North America, associated with a stationary Rossby wave propagation from Alaska (arrows in Fig. 9). We label the ALO-related atmospheric circulation as the Alaska–North America (ANA) pattern for convenience. The ALO–ANA pattern is most noticeable at high altitudes (Fig. 9).

Changes in atmospheric circulation associated with the ALO–ANA pattern affect the winter climate of North America, as well as the Bering Sea. In negative ALO phase (Fig. 7), equatorward CAM flow strengthens, resulting in the CAM amount increase and a large SAT decrease over North America, with a maximum cooling of 4 K even on the 3-month mean. The occurrence of the negative ALO–ANA pattern brings severe cold winters to North America.

c. Comment on possible positive feedback of Bering Sea ice on atmospheric circulation

In positive ALO phase, northerly winds with cold air over the Bering Sea are responsible for a large release of heat from the ocean to the atmosphere over open water of about 100 W m^{-2} compared with negative ALO phase (Fig. 8). At midlatitudes, this difference affects overlying atmospheric fields, including the vertical temperature structure, storm-track activity, and large-scale atmospheric circulations (Ma et al. 2015; Sugimoto et al. 2017). We investigate the vertical structure of diabatic heating over the southern part of the Bering Sea to detect effects of air–sea heat exchange on the overlying atmosphere. In positive ALO phase, the diabatic heating is remarkable below 900 hPa and reaches a maximum value of 10 K day^{-1} at near surface (Fig. 10a). Sensible heating makes the largest contribution to net heating (Fig. 10a), which reflects the large heat release in the form of sensible heat flux.

How does the heating from the ocean affect the atmosphere? To obtain an implication for this point, we conduct an experiment using a linear baroclinic model

FIG. 12. Differences in winter (a) sea ice concentration and (b) SST between positive and negative ALO phase. Black lines represent regions exceeding the 95% confidence limit. (c),(d) As in (a) and (b), but for the preceding autumn (September–November).

(LBM). The LBM consists of primitive equations exactly linearized about a basic state [see [Watanabe and Kimoto \(2000, 2001\)](#) for mathematical expressions] and has a spatial resolution of T42 with vertical 20 sigma levels with vorticity, divergence, temperature, and surface pressure as model variables. The winter climatology derived from JRA-55 is adopted as the basic state. We set idealized heating over the Bering Sea, with a Gaussian horizontal distribution and a sinusoidal vertical pattern having a maximum at a model sigma level of 0.98 (10 K day^{-1} ; see [Figs. 10b,c](#)). The simulated atmospheric response to the heating over the Bering Sea roughly captures the ANA pattern ([Fig. 10b](#)), with a positive height anomaly over North America and a negative height anomaly around Alaska. The former bears a good resemblance to JRA-55 ([Fig. 9b](#)) in terms of amplitude, location, and spatial size. The amplitude and spatial size of the latter are similar to those of JRA-55, but its center of action is slightly farther northward. This experiment strongly suggests that vigorous heat release from the ocean can act as a positive feedback to the ALO–ANA pattern.

The LBM experiment shows another positive height anomaly, unrelated to the ALO–ANA pattern, around the Okhotsk Sea ([Fig. 10b](#)). Our experiment only introduced heating over the Bering Sea, but cooling also exists over the northern Bering Sea in positive ALO phase ([Fig. 8](#)). Including this cooling effect could improve reproducibility in numerical experiments.

5. Summary and concluding remarks

Our analyses of five datasets show that winter Bering SIA variations with a period of 10–15 years are not related to variations in sea ice concentration variations in the Okhotsk Sea. Variations of SLP are present over Alaska [Alaska Oscillation (ALO)], are characterized by an equivalent-barotropic structure throughout the troposphere, differ from previously identified atmospheric patterns, and are primarily responsible for Bering SIA variations and severe winters in North America. In the positive ALO phase, equatorward CAM flow through the Bering Strait strengthens, resulting in surface air cooling over the Bering Sea and an

FIG. 13. Differences in monthly Z500 between the positive and negative ALO phases from (a) October of the previous year to (f) March. Black lines represent regions exceeding the 95% confidence limit.

increase in Bering SIA and, conversely, surface air warming of about 4 K even on winter mean values in North America in response to the decrease in equatorward CAM flow. The opposite takes place during the negative ALO phase. The northerly winds with cold air result in a large release of heat from the ocean to the atmosphere over open water just south of the region covered by sea ice. The LBM experiment suggested that this heating over the southern Bering Sea can act as a positive feedback for the positive ALO–ANA pattern. Our results provide a new perspective on the role of sea ice in the climate system, including its thermal insulation effects. Understandings of the cause-and-effect relationships among CAM, SAT, sea ice, and surface heat flux are necessary to establish an air–sea ice interaction system.

A prolonged and intense cold wave from late December 2017 to early January 2018 hit North America in which record low temperatures gripped much of the central–eastern United States and central–eastern Canada (WMO

2018) (Figs. 11a,b). This severe cold winter was caused by a prominent atmospheric circulation anomaly associated with a positive height anomaly over Alaska and a negative anomaly over North America (Fig. 11c), consistent with the ALO–ANA pattern identified in our study. During this winter, the Bering SIA was extremely small (Fig. 3a), and this is also evidence of a link between severe cold winters in North America and Bering Sea ice loss. Wavier westerly jets associated with blocking are often observed around Alaska (e.g., Bao and Wallace 2015). To better understand the mechanisms for the severe winters in North America, the atmospheric response on weekly and biweekly time scales to Bering SIA should be further explored.

We attempt to obtain an implication for generation, evolution, and decay of the ALO–ANA pattern. Past works pointed out that the winter Bering Sea ice extent would be influenced by ice flux from the Arctic Ocean (e.g., Zhang et al. 2010) and by a warm ocean flow from the southern region in the preceding autumn (e.g.,

TABLE 2. List of the 12 CMIP6 climate models used in this study.

Model name	Institute	Country	Ocean grid (lon $\times$ lat)
CanESM5	Canadian Centre for Climate Modelling and Analysis	Canada	360 $\times$ 291
CESM2	National Center for Atmospheric Research	United States	320 $\times$ 384
CESM2-WACCM			320 $\times$ 384
EC-Earth3	EC-Earth Consortium	Sweden	362 $\times$ 292
EC-Earth3-Veg			362 $\times$ 292
GFDL-CM4	National Oceanic and Atmospheric Administration (Geophysical Fluid Dynamics Laboratory)	United States	1441 $\times$ 1081
GFDL-ESM4			721 $\times$ 577
IPSL-CM6A-LR	Institute Pierre Simon Laplace	France	362 $\times$ 332
MIROC6	Japan Agency for Marine-Earth Science and Technology/Atmosphere and Ocean Research Institute (The University of Tokyo)/National Institute for Environmental Studies/RIKEN Center for Computational Science	Japan	360 $\times$ 256
MPI-ESM1-2-HR	Max Planck Institute for Meteorology	Germany	802 $\times$ 404
MRI-ESM2-0	Meteorological Research Institute	Japan	360 $\times$ 363
NESM3	Nanjing University of Information Science and Technology	China	320 $\times$ 384

Nakanowatari et al. 2015). We investigate ocean conditions in winter and preceding autumn. During winter in the positive ALO phase, the SST has negative anomalies in the Bering Sea (Fig. 12b), a suitable condition for a sea ice generation and its evolution (Fig. 12a).

However, in the preceding season, there are no significant signals for both SST in the Bering Sea and sea ice concentration in Bering Sea/Arctic Sea (Figs. 12c,d). Next, we describe the seasonal evolution of atmospheric fields (Fig. 13). Negative height anomalies over Alaska, related

FIG. 14. Time series of winter SIA for the BSR from the 12 CMIP6 climate models of SSP5-8.5 scenario listed in Table 2. All outputs are interpolated to a common 1° longitude $\times$ 1° latitude grid.

to ALO, become significant in November, and both the amplitude and spatial size gradually increase during the winter. Positive height anomalies over North America emerge 1–2 months later, and then the large-scale atmospheric circulation of ANA pattern takes a mature phase in January–February. Understanding how this pressure pattern forms over Alaska in the late autumn will aid efforts to improve forecasting accuracy for the North America climate in mid-/late winter. In addition, in January and February (Figs. 13d,e), positive geopotential height anomalies are detectable in a region south of negative anomalies over Alaska, and the positive SST anomalies are distributed in eastern tropical region, which is similar to the El Niño event, despite no significant signals (Fig. 12b). These might imply influences of remote forcing associated with tropical heating on the evolution and persistence of the ALO–ANA pattern.

Sea ice in the Arctic Sea has significantly decreased under global warming (Simmonds 2015). Bering SIA has shown no decreasing trend over the past four decades, but it may decrease in the future. In fact, Bering SIA has remained very small over the last five years with a record-breaking small value in 2018. We examine the Bering SIA under a future climate scenario based on the highest-emission scenario (SSP5–8.5, rapid economic growth driven by fossil fuels) of the 12 climate models that participated in phase 6 of the Coupled Model Intercomparison Project (CMIP6; Table 2; Eyring et al. 2016), which will inform the International Panel on Climate Change (IPCC) Sixth Assessment Report. Although the Bering SIA differs among the CMIP6 runs, all the climate models consistently project a decrease in Bering SIA (Fig. 14) and predict the complete loss of Bering SIA as early as the late 2030s and by 2075 in more than 50% of the predictions. Our study under the present climate shows that severe cold winters in North America are linked to Bering SIA loss via the negative ALO–ANA pattern. In the near future, with little or no sea ice in the Bering Sea, this negative ALO–ANA pattern may occur more frequently, or a completely different climate mode may appear. To predict severe cold winters in North America under global warming, it will be necessary to model regional and general climate systems with little or no sea ice.

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