A negative phase shift of the winter AO/NAO due to the recent Arctic sea ice reduction in late autumn

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Abstract

This paper examines the possible linkage between the recent reduction in Arctic sea ice and the wintertime Arctic Oscillation (AO)/North Atlantic Oscillation (NAO). Analysis of the ERA interim reanalysis and merged Hadley/OI-SST data indicate that a reduced (increased) sea ice area in November leads to more negative (positive) phases of the AO and NAO in early and late winter, respectively. We simulate the atmospheric response to observed sea ice and sea surface temperature (SST) anomalies using an atmospheric general circulation model (AGCM for Earth Simulator, AFES version 4.1). The recent Arctic sea ice reduction gives rise to cold winters to mid-latitude continental regions, namely, Europe, eastern Siberia, and North America, linked to an anomalous circulation pattern from the negative phase of AO/NAO. The reduction in sea ice approximately doubles the frequency of large negative AO events in the seasonal mean time scale. In association with the seasonal mean negative AO, high frequent cold air out break from the Arctic to the mid-latitudes also increases during winter. In comparison, SST anomalies in the tropics and middle–high latitudes mask this continental cooling signal. Model-based analysis reveals that the stationary Rossby wave response to sea ice reduction in the Barents Sea induces an anomalous meridional circulation. We found that this anomalous meridional circulation cools the mid-latitudes and warms the Arctic, adding extra Arctic heating equivalent to about 25% of the direct surface heat release from the sea ice reduction.
1. Introduction

The recent decrease in Arctic sea ice extent (SIE) is a marked signature of global warming in the troposphere, which is likely a combined result of anthropogenic radiative forcing by increasing greenhouse gases (GHGs) and the natural variability and feedbacks within the Earth’s ice–ocean–atmosphere coupled system on a decadal timescale [Comiso et al., 2008; Schweiger et al., 2008; Sereeze et al., 2007, 2009; Kay et al., 2011]. Studies have examined the atmospheric response to changes in Arctic SIE to gain a better understanding of the underlying feedback mechanisms and interactions of the Arctic air–sea–ice coupled system. Anomalously cold winters in the mid-latitudes over Eurasia have been connected to an atmospheric response to Arctic summer sea ice reduction [Honda et al., 2009; Francis et al., 2009; Petoukhov and Semenov, 2010; Inoue et al., 2012; Liu et al., 2012; Tang et al., 2013]. The dynamical processes associated with this remote response have been identified in both observation-based analysis and atmospheric general circulation model (AGCM) experiments. Francis et al. [2009] pointed to changes in Northern Hemisphere (NH) atmospheric circulation such as the weakening of the polar jet stream being associated with Arctic sea ice reduction. Honda et al. [2009] showed that the stationary Rossby wave response to the increase in turbulent heat flux associated with summer Arctic ice reduction intensifies the wintertime Siberian high that brings cold air outbreaks.

Previous studies have noted that atmospheric variations associated with the Arctic Oscillation (AO)/Northern Hemisphere Annular Mode (NAM), which is a dominant mode of the Northern Hemisphere atmospheric variability [Thompson and Wallace, 1998, 2000], have impacts on Arctic sea
ice [e.g., Rigor et al., 2000, 2002; Ogi et al., 2008, 2010; Ogi and Yamazaki, 2010; Stroeve et al., 2011]. Other large-scale variations, such as the dipole pattern in the Arctic, also have impacts on Arctic sea ice [Wu et al., 2006; Maslanik et al., 2007; Stroeve et al., 2007; Holland et al., 2008; L’Heureux et al., 2008; Wang and Overland, 2009]. On the other hand, some studies presented that the negative phase of the AO/NAM appears as a response to the Arctic sea ice reduction through AGCM experiments with prescribed sea ice or SST conditions [Magnusdottir et al., 2004; Deser et al., 2004; Alexander et al., 2004]. Recently, the relationship between summer Arctic ice reductions and the negative phase of the AO/NAM and North Atlantic Oscillation (NAO) signals in winter has been reported [Liu et al., 2012; Jaiser et al., 2012]. Observation-based studies have reported that the negative trends in the AO/NAM and NAO are related to the decreasing trend in Arctic sea ice from summer to autumn [Overland and Wang, 2010; Hopsch et al., 2012; Rinke et al., 2013] and the increasing trend in Eurasian snow cover [Cohen et al., 2012]. Results from numerical simulations using atmosphere–ocean coupled models suggest that Arctic sea ice variability and the modulation of the AO/NAM are linked [Sokolova et al., 2007; Cohen et al., 2012]. Recognizing a potentially critical role of this link in the NH climate system, the primary goal of this study is to determine underlying processes connecting Arctic sea ice variability, its atmospheric response, and modulation of the AO/NAM and the NAO. Our understanding of the atmospheric response to Arctic ice variability is often hindered by internal atmospheric variability [Deser and Phillips, 2009; Screen et al., 2013]. Recognizing this, we will evaluate changes in the probability distribution of AO/NAM index related to sea ice reduction while taking internal atmospheric variability into account. We used an AGCM to
address these issues because it allowed us to conduct sensitivity experiments by changing boundary
conditions representing sea ice variations. In this paper, we first examine the observed atmospheric
variability in the NH associated with Arctic sea ice variability. Next, we investigated the
climatological impacts of recent Arctic ice reduction using AGCM experiments, and finally we
examine modulations of the simulated AO/NAM through an analysis of the model results.

2. Data and indices

2.1. Sea surface temperature (SST) and sea ice concentration (SIC)

We used the Merged Hadley-National Oceanic and Atmospheric Administration (NOAA)/Optimum Interpolation (OI) Sea Surface Temperature (SST) and Sea ice Concentration (SIC) [Hurrell et al., 2008] datasets for the period 1979–2011. We obtained time series of the Arctic sea ice area for each calendar month, which is defined as SIC (%) × grid-area (m²) integrated northward of 65°N. In our analysis, we removed the linear trend for the period 1979–2011 prior to evaluating the interannual relationship between Arctic sea ice and atmospheric variables.

The same data were also used as boundary conditions for sensitivity experiments using an AGCM. To examine the influences of the recent Arctic ice reduction, we used 5-year monthly varying climatologies for the early period (1979–1983) and the late period (2005–2009). As an example of changes in SST and sea ice thickness (SIT) (converted from SIC, see detail in section 2.4), Figure 1 shows anomalies of SST (Figure 1a) and SIT (Figure 1b) for the late and early periods in September. Figure 1d and e (1g and h) show SST and SIT anomalies in November (January). While during boreal
summer SST warms globally and SIT largely decreases in the Pacific sector of the Arctic Ocean, during the boreal winter, SST shows La Niña type anomalies in the Pacific region and warm anomalies across the whole Atlantic region. SIT shows large ice reductions in the East Siberia Sea in September; the Barents-Kara Sea and the Bering Strait in November; and the Barents Sea, the Nordic Sea, and the Sea of Okhotsk in January. Figure 1c, 1f, and 1i show regression coefficients of detrended SIT anomaly against the normalized sea ice area index in September, November, and January, respectively. Spatial patterns of late minus early anomalies are quite similar to the regressions. Magnitudes of late minus early anomalies are about four times larger than the regression. This indicates the recent Arctic sea ice reduction is nearly equivalent to 4 sigma of the interannual natural variability of the total Arctic sea ice area. We thus supposed that similar physical process dominates the atmospheric responses to the Arctic sea ice changes on the interannual and decadal time scale.

2.2. ERA interim

We used monthly mean data from the European Centre for Medium-Range Weather Forecast (ECMWF) ERA-interim reanalysis for the period 1979–2012 [Dee et al., 2011]. Data were averaged for 3-month periods (e.g., December, January, and February or DJF) to reduce noise caused by the high frequent internal variability of atmospheric fields. We used linear regression methods to evaluate the impacts of SIC variations over interannual timescales. Atmospheric variables detrended for the period 1979–2012 were regressed on the time series of the Arctic sea ice area.
2.3. Annular mode indices

We used AO and NAO indices provided by the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) (http://www.esrl.noaa.gov/psd/data/climateindices). The indices were averaged for three months prior to use. Furthermore, we calculated empirical orthogonal function (EOF) of 3-month mean of geopotential height at 500 hPa (Z500) northward of 30°N. A spatial pattern of the primary mode was quite similar to Z500 anomaly regressed on the AO index in the individual 3-month mean. EOF1 score of this Z500-EOF was for an annular mode index (Z500-EOF1 score) in the middle troposphere.

2.4. Model and simulations

We used the AGCM for Earth Simulator (AFES) [Ohfuchi et al., 2004, 2007; Enomoto et al., 2008; Kuwano-Yoshida et al., 2010] version 4.1 with a triangular truncation at horizontal wavenumber 79 (T79; approximately 1.5° horizontal resolution), and with 56 vertical levels and an elevation of about 60 km at the model top. The AFES version 4.1 used here is a major update of the version used by Honda et al. [2009], as the horizontal resolution and model top are both higher (increasing from T42 to T79 and 30 to 60 km, respectively). Note that because the model does not treat SIC directly, namely, a model grid is either ice-covered or ice-free. To include the partial ice cover effects on turbulent heat flux, SIC data is converted into SIT. We assumed that the maximum ice thickness in the Arctic is 50 cm, so that SIC from 0 to 100 % was linearly converted into SIT from 0 to 50 cm. We then manipulated all grids where converted SIT was less than 5 cm to have no ice (i.e., 0 cm). With this
method, turbulent heat flux over sea ice is reasonably well simulated, though 50 cm maximum ice thickness is small compared with reality. Perpetual model runs were performed with boundary conditions of mean seasonal cycles of the global SST and SIT. Other external forcing were all fixed as 380 ppm for CO$_2$, 1.8 ppm for CH$_4$, and monthly climatological mean of O$_3$ for 1979-2011 period obtained from the Japanese 25-year Reanalysis (JRA-25)/Japan Meteorological Agency (JMA) Climate Data Assimilation System (JCDAS) reanalysis data [Onogi et al., 2007] for O$_3$. Default values of aerosol and incident solar radiation were used. The control run (CNTL) was performed with a 5-year monthly-varying climatology of SST and SIT for the early period (1979–1983). The Global run was performed with the SST and SIT climatology for the late period (2005–2009). We also ran other cases that were similar to CNTL. The Northern case used SIT and SST for the late period in the region northward of 50°N. The N.ICE case used changes in SIT only for the region northward of 50°N (i.e., SST is same as CNTL). These four cases are summarized in Table 1. All runs used the same initial conditions (the monthly mean of January 1979 from JAR25/JCDAS). We carried out 60-year integrations for the CNTL and N.ICE runs, and 20-year integrations for the other runs after a 1-year spin-up. All variables of daily mean output were averaged for three months, and we mainly focused on the boreal winter (DJF).

3. Results

3.1. Observed evidence of the ice–atmosphere linkage

First, we examined observational evidence of the relationship between wintertime
atmospheric and Arctic sea ice variations. As sea ice has a much longer memory compared with the atmosphere, we calculated the lag-correlations between the DJF mean AO index and the Arctic sea ice area in the leading months. After detrending both the AO index and sea ice area, the correlations were 0.14, 0.28, 0.48, and 0.26 for the sea ice area in September, October, November, and December, respectively. The November sea ice area has the maximum correlation with the DJF mean AO index. The time series of November sea ice area and the DJF mean AO and NAO indices are shown in Figure 2. Both of the DJF averaged AO and NAO indices are significantly related to the November sea ice time series, not only over an interannual timescale, but also at decadal timescales (e.g., negative trends after 1989).

Figure 3 (top row) shows anomalies of geopotential height at 500 hPa (Z500) from October-November-December (OND) mean to January-February-March (JFM) mean regressed on the time series of the November Arctic sea ice area. Note that the sign is reversed so that height anomalies are associated with a decrease in Arctic sea ice area. Most of the regressed anomalies show positive anomalies in the Arctic Ocean region, but negative anomalies at mid-latitudes. This structure of the geopotential height anomaly resembles the negative phase of the AO/NAM pattern. Consistent with this result, regressions of sea level pressure (SLP) anomalies shown in Figure 3 (2nd row) have persistent positive anomalies in the Arctic Ocean region and negative anomalies at mid-latitudes. In comparison, Figure 4 shows sign-reversed correlation coefficients of the November sea ice area and the AO index for August-September-October (ASO) to March-April-May (MAM) (black and solid line). The correlations are negatively high from −0.42 in ASO, to −0.53 in DJF, and the rapidly
decreases to a minimum in February-March-April (FMA). While the correlations of Z500-EOF1 score (black and dashed line) are similar to those of AO index, the correlations of NAO index (gray line) are negatively high in the wintertime. This strong relationship between the annular mode from late summer to mid-winter and November sea ice variability may suggest that summer annular mode affects on the summer-to-autumn Arctic sea-ice and the Arctic sea-ice in the autumn affects on the winter annular mode, because the annular mode signal does not persist beyond the season. While the negative anomalies of Z500 in central to eastern Siberia are significant in early winter (OND and November-December-January (NDJ)), those in Europe and North Atlantic are significant in late winter (NDJ–JFM). Such a shift in the spatial pattern from Siberia to the North Atlantic is also found in the SLP anomalies, (Figure 3, 2nd row). During OND, positive anomalies in the Arctic Ocean extend toward Eurasia around 90°E. Afterwards, from NDJ to JFM, negative anomalies in the North Atlantic gradually develop as extensions of the positive anomalies around 90°E gradually vanish. Anomalies in the temperature at 2 m height (T2m) also show this shift (Figure 3, 3rd row). That is, during OND and NDJ, cold anomalies are found in eastern Siberia associated with the high pressure anomalies in northern Siberia (i.e., intensification of the Siberian high), and during DJF and JFM, cold anomalies are found in Eastern Europe.

These results indicate that sea ice variability in late autumn is strongly related to the AO/NAM-like variations from late summer to mid-winter. Furthermore, interannual variations of wintertime atmospheric fields show a gradual shift from negative (positive) AO-like signals in early winter to negative (positive) NAO-like signals in mid-winter associated with the decrease (increase) in
November sea ice extent. The above results imply the possible influence of Arctic ice reduction in late autumn on the recent negative trend in winter annular modes. In the following sections, we examine the changes in NH climate associated with recent sea ice reduction, and their impacts on annular modes such as the AO/NAM and NAO through an analysis of the results AGCM sensitivity experiments.

3.2. Climatological impacts of recent ice reductions

3.2.1. Model results and comparisons with observations

In this section, we analyze the differences between the CNTL run and the other three runs to evaluate the climatological impacts of recent Arctic sea ice reduction. We also compare model results with regressed fields upon the November sea ice area from the ERA interim reanalysis data.

We examined Z500 and T2m because they are typical variables that represent atmospheric variability in the troposphere and near the surface, respectively. Figure 5a–c compares the Z500 anomalies for Global, Northern, and N.ICE with CNTL. Positive anomalies are commonly found over the Barents Sea where sea ice reductions are large. On the other hand, anomalies in the mid-latitudes are different among the three experiments. This is due to differences in SST anomalies. In particular, in Global, the negative phase of the Pacific North America (PNA)-like pattern is obvious, corresponding to La Niña-like SST anomalies. In Northern, this PNA-like pattern vanishes. In N.ICE, although only changes in sea ice extent in the Arctic are responsible for the anomalous atmospheric fields, significant anomalies are found in the mid-latitudes (i.e., negative anomalies in Europe and
eastern Siberia are found in addition to positive anomalies over Alaska and the Arctic). A wave-train
like structure along the Eurasian Arctic coast (i.e., Europe–Barents–Siberia) resembles the Eurasian
(EU) pattern [Wallace and Gutzler, 1981; Ohhashi and Yamazaki, 1999]. Figure 5d–f shows the T2m
anomalies that correspond to anomalies in Z500 for each experiment. In particular, cold anomalies in
eastern Siberia found only in N.ICE correspond to the negative anomalies of Z500 over the same
region.

We compared the pattern similarity of each model result with the regressed fields of ERA
interim data upon the November sea ice area (Table 2). Spatial correlations of Z500 in the NH
(northward of 30°N) against the ERA interim anomalies are the lowest in Global and the highest in
N.ICE. Spatial correlations of T2m are similar. Large correlations in N.ICE with regressed fields
suggest that the model accurately simulates the atmospheric response to ice reduction. Note that the
correlation is slightly larger with regression fields in NDJ than in DJF. We use the regression fields in
NDJ for comparison with the model results in the remainder of this section.

Figure 6a and b shows the NH temperature response in N.ICE relative to CNTL and its zonal
mean, respectively. Cold anomalies are found in Eastern Siberia, Europe, and North America, and
warm anomalies exist in the Arctic Ocean and the Sea of Okhotsk. The zonal mean temperature
response shows a large warm anomaly in the polar region. In the mid-latitudes, while the zonal mean
temperature response shows small anomalies, anomalies averaged only over land are clearly cold. The
area-weighted temperature response over the NH mid-latitudes (30–60°N) is nearly zero. However,
averaging only over land it becomes −0.14 K, with a minimum of −0.42 K at 53.0°N. In comparison,
Figure 6c presents regressed fields of NDJ mean temperatures at 850 hPa from the ERA interim data associated with the normalized time series of November sea ice area. Warm anomalies in the Arctic and cold anomalies over mid-latitude land are also found. Continental cold anomalies are located in North America, Europe, and eastern Siberia, resembling the N.ICE case. The area-weighted temperature response averaged in the NH mid-latitudes (30–60°N) is $-0.09 \, \text{K}$, with a minimum of $-0.17 \, \text{K}$ at 45.0°N; averaged only over land this becomes $-0.18 \, \text{K}$, with a minimum of $-0.30 \, \text{K}$ at 49.5°N. The results indicate that the recent ice reduction makes some contribution that cools NH continental regions. Therefore, we next consider the results from the N.ICE case in more detail to explain the mechanism of this continental cooling as a response to ice reductions.

3.2.2. Stationary Rossby response and associated modulation of meridional circulations

Having recognized the statistical relationships between Arctic sea ice reduction and the atmospheric response, we next consider possible mechanisms responsible for the occurrence of cold anomalies over the NH land regions. Figure 7 shows DJF averaged 3D temperature advection anomalies (N.ICE minus CNTL) caused by climatological temperature advected by anomalous winds (shaded). Contours and arrows display the corresponding temperature climatology and the anomalous horizontal wind vectors, respectively (vertical wind anomalies are not shown, but are taken into account for temperature advection). Cold air advected from the climatological cold-core by the anomalous wind leads to the near-surface temperature anomalies in eastern Siberia, North America, and Europe shown in Figure 6a. The anomalous wind at 850 hPa corresponds to a geostrophic
relationship with the geopotential height anomaly in the upper levels (e.g., 500 hPa in Figure 5c). The upper-level geopotential anomalies (which are strongly associated with the potential vorticity anomalies) induce low-level wind anomalies [Lau and Holopainen, 1984; Lau and Nath, 1991] and are consistent with the results of Honda et al. [2009].

We next examine how Arctic ice reduction affects changes in the atmospheric circulation in the middle to upper troposphere. To achieve this, and following Honda et al. [2009], we used the wave activity flux (WAF) developed by Takaya and Nakamura [2001], which indicates the 3D group velocity of a quasi-geostrophic wave packet and thus propagation of the stationary Rossby wave activity. Figure 8a shows anomalous fields of the N.ICE experiment (N.ICE minus CNTL) including the turbulent heat flux (i.e., the sum of sensible heat and latent heat fluxes from the surface to the atmosphere), the vertical component of the WAF at 700 hPa, and the horizontal vectors of the WAF at 300 hPa. The southward propagation of the Rossby wave activity that originates in the Arctic region (thick arrows) can be seen in eastern Siberia and eastern North America. In particular, the disturbance in eastern Siberia (around Lake Baikal, 110°E and 50°N) originates from an upward propagation of the Rossby wave activity associated with turbulent heat flux anomalies over the Barents Sea. Figure 8b shows geopotential height anomalies at 300 hPa (Z300) from the N.ICE experiment (N.ICE minus CNTL) and corresponding WAF. The wave activity propagates from the Barents Sea region to eastern Siberia where cyclonic anomalies are found. Figure 8c shows vertical–meridional cross sections of geopotential height anomalies, and zonal and vertical components of the WAF along the brown line drawn in Figure 8b. The Rossby wave activity propagates upwards and eastwards from the wave
source near the surface in the Arctic region (northward of 65°N). This is similar to the result of Honda et al. [2009] who found cyclonic anomalies in eastern Siberia formed by a stationary Rossby wave response to ice reduction in the Barents Sea. The Rossby wave propagation from the Arctic to the mid-latitudes is found not only in eastern Siberia but also in northern North America. These two pathways are located west of the climatological troughs (Figure 9a) and are collocated with the southeastward extension of the jet stream. This is consistent with the notion that the Rossby wave activity prefers to propagate along a wave-guide where the meridional gradient of the absolute vorticity is large (red hatching in Figure 9b). Therefore, this result implies that Arctic ice reduction induces a modulation of the climatological planetary wave. The modulation of the planetary wave is accompanied by changes in the meridional circulation that are related to heat exchange between the Arctic and mid-latitudes.

We next examine changes in the meridional circulation based on the transformed Eulerian mean (TEM) diagnosis, and we estimate the feedbacks of these circulation changes on the Arctic region. Figure 10a shows DJF zonal mean wind anomalies of N.ICE against CNTL. A dipole pattern from the weakened subpolar jet (60–70°N) and intensified subtropical jet (30–40°N) are evident, although only the former is significant. Figure 10b shows anomalies in the Eliassen–Palm (EP) flux [Andrews and McIntyre, 1976] and its divergence. Upward wave propagation is enhanced in the lower to middle troposphere, and the upper tropospheric zonal flow at high latitudes (around 65°N at 300 hPa) is decelerated by the resulting wave drag. Figure 10c shows the anomalous residual mean circulation associated with the calculated EP flux divergence. To balance the zonal flow deceleration
due to the anomalous wave drag, a northward residual circulation is induced in the upper troposphere (50–80°N between 400 and 300 hPa) accompanied by upward motion in the mid-latitudes (40–50°N) and downward motion at high latitudes and in the Arctic (around 60°N and 85°N). Figure 11a shows the longitudinal distribution of DJF averaged anomalies of eddy momentum flux (u´v´) at 300 hPa (N.ICE minus CNTL). Note that here the eddy is defined as the anomaly from the zonal mean of a daily variable, according to the TEM formulation. Negative anomalies of the eddy momentum flux corresponding to a northward EP flux anomaly are found in central Siberia and north of the Okhotsk Sea. Figure 11b shows the longitudinal distribution of the anomalous eddy heat flux (v´T´) at 500 hPa. Positive anomalies corresponding to an upward EP flux anomaly are found in central Siberia, and negative anomalies corresponding to a downward EP flux anomaly are found in the Okhotsk Sea. Figure 11c and d shows the eddy momentum and heat fluxes due to the stationary eddy (i.e., anomalies from the zonal mean of DJF averaged variables), respectively. The flux anomalies due to the stationary eddy are dominant in central Siberia (around 90°E and 60°N), implying that contributions of flux anomalies due to the transient eddy (i.e., total minus stationary) are less dominant. This result suggests that, in central Siberia, the stationary wave response to ice reduction is likely to be the driver of the anomalous meridional circulation. The adiabatic processes associated with the anomalous meridional circulation as shown in Figure 10c cools the mid-latitudes and warms the Arctic region throughout the troposphere, suggesting that changes in the mean flow associated with the modulation of the planetary wave have a positive feedback on the impacts of Arctic ice reduction. We estimated the heating rate induced by the residual mean vertical motion (w*) in the troposphere (850 to 300 hPa; see Appendix).
The heating rate averaged over the Arctic region (northward of 75°N) was 1.59 W m\(^{-2}\), and averaged over the mid-latitudes (40–60°N) it was −0.24 W m\(^{-2}\). This heating rate induced by the secondary circulation amounts to about 25% of the direct heat release from the ice reduction itself, which is 6.23 W m\(^{-2}\) (obtained from the turbulent heat flux anomaly averaged over the Arctic).

3.2.3. Modulation of the AO/NAM

As shown in Section 3.2.2, the downward and upward motion induced by ice reduction warms the Arctic and cools mid-latitudes. Such an anomalous meridional circulation caused by stationary wave drag is accompanied by a weakening of the polar vortex, which shows variability that is strongly related to AO/NAM [Limpasuvan and Hartmann, 2000]. We estimated the contribution of Arctic ice reduction on the AO/NAM signals. We applied EOF analysis to the 3-month mean Z500 data from the 60-year results of the CNTL and N.ICE runs. We also applied EOF analysis to the Z500 data from N.ICE combined with that from CNTL (i.e., a total sample size of 120 years). Using this combined EOF analysis, we evaluated the modulation of AO/NAM as a response to recent sea ice reduction.

Figure 12 a–c shows the EOF1 loading patterns of the DJF mean Z500 from CNTL, N.ICE, and CNTL+N.ICE, respectively. There are notable differences in the EOF1 patterns from CNTL and N.ICE: while the Arctic low-pressure anomalies in CNTL have a single-cell structure, in N.ICE they have a three-cell structure in which individual low-pressure cells pair with a high-pressure anomaly in the mid-latitudes. The combined EOF1 pattern from CNTL+N.ICE shows a mix of the CNTL and N.ICE patterns. The score of the combined EOF1 can be used to simplify the quantification of
modulation of the primary mode. Figure 12d shows the time series of the EOF1 scores of the combined
EOF. The score shifts to be more negative from the CNTL period to the N.ICE period. The difference
between these two periods is 0.494 $\sigma$, which exceeds the 99% significance level. This indicates the
modulation of the primary mode in the atmosphere due to changes in the boundary conditions (i.e., the
AO/NAM shift to the negative phase is associated with Arctic ice reduction). This may be one of
reasons for the change in the EOF1 pattern from single-cell in CNTL to three-cell in N.ICE. Due to the
meandering of the high latitude jet stream, cold air outbreaks from the Arctic to the mid-latitudes occur
more frequently during the negative phase of AO/NAM than during the positive phase. Thus, dipole
patterns associated with the local meridional heat exchange will dominate in EOF1 of N.ICE.

For a more quantitative estimation of the modulation of the primary mode, we examined the
probability density function (PDF) of the EOF1 score using the nonparametric density estimation
technique [Kimoto and Ghil, 1993]. Figure 13 shows histograms and associated PDFs of combined
EOF1 scores for the CNTL and N.ICE periods. The probability density of the CNTL period is larger
than that of the N.ICE period for the positive score range. The probability of positive scores larger than
1.0 $\sigma$ was 20.4% and 10.8% for the CNTL and N.ICE periods, respectively. On the other hand, the
probability for scores less than $-1.0 \sigma$ was 9.3% and 18.2% for the CNTL and N.ICE periods,
respectively. Furthermore, while the PDF of CNTL skews to the right (skewness = 1.023), that of
N.ICE does not (skewness = 0.001). This finding supports the more frequent appearances of a large
negative phase of AO/NAM during the N.ICE period. The results indicate that the large positive
(negative) phase of AO/NAM occurs half (twice) as frequently in association with a negative shift of
the AO/NAM due to Arctic ice reduction.

Next we examined the PDF change in daily score. Figure 14 shows PDF of daily scores, which is obtained by daily Z500 anomalies projected onto individual EOF1 patterns of CNTL and N.ICE periods (i.e., Figure 12a and 12b), respectively. Z500 anomalies against daily climatological seasonal cycle of individual runs were used and daily scores were normalized by 1.0 standard deviation (σ) of individual EOF1 scores of 3-month mean fields. Although standard deviations of two periods are similar, skewness is negatively larger and kurtosis is positively larger in N.ICE than in CNTL, supporting that large negative AO/NAM occurs more frequently in association with the Arctic sea ice reduction in addition to the negative shift of the mean. Figure 15 shows composite of the daily SLP and T2m anomalies for individual daily EOF1 scores less than -2.0 σ in DJF. SLP anomalies corresponding to negative AO and associated continental cooling are found for both CNTL (Figure 15a and 15d) and N.ICE (15b and 15e) runs. Difference of SLP anomalies of N.ICE minus CNTL shows large negative anomaly in the North Atlantic and positive in the Atlantic side of the Arctic Ocean, resembling negative NAO pattern (Figure 15c). Associated T2m difference shows negative anomalies in the eastern North America and North Europe, although in East Asia only small negative anomaly is found northwest of Korean Peninsula. Because daily anomalies used for the composite analysis were departures from daily climatology of individual runs, the results indicate more frequent cold air outbreak in N.ICE than in CNTL, and thus supports the explanation of that the local meridional heat exchange dominates in EOF1 of N.ICE.

Finally, we will consider the seasonal evolution of the impact of ice reduction on NH climatic
Figure 16a–d compare the Z500 anomalies of the N.ICE run with the CNTL run from the OND mean to the JFM mean. Only positive anomalies are found surrounding the Arctic Ocean in OND. However, negative anomalies appear in eastern Siberia in NDJ, and an anomalous pattern forms (which is similar to an annular mode) in DJF, with negative anomalies in Europe and eastern North America. The simulated seasonal evolution partly resembles the observed one, in which Siberian negative anomalies first appear in OND, and negative anomalies in Europe and eastern North America appear after that (1st row in Figure 3). Furthermore, a deepening of the annular mode is found in the simulation as indicated by the seasonal evolution of the EOF1 scores shown in Figure 16e. The EOF1 negative scores gradually increase in magnitude from late autumn (ASO) to late winter (JFM) and rapidly return towards zero in spring (MAM). This resembles the observed evolution of the annular modes (Figure 4), although we note that observational results are based on the AO/NAM index (EOF of 1000 hPa pressure), while the simulated results are based on the EOF of Z500.

4. Discussion

4.1. Winter annular modes and November sea ice

In section 3.1, we showed that the wintertime AO/NAM- and NAO-like signals are strongly related to Arctic sea ice variability in November. Many studies have reported a significant relationship between Arctic summer sea ice loss, in particular September sea ice, and the negative trend of AO/NAM based on both observations [Overland and Wang, 2010; Hopsch et al., 2012; Liu et al., 2012; Jaiser et al., 2012; Rinke et al., 2013] and numerical simulations [Sokolova et al., 2007; Liu et
Based on our results, the autumn Arctic sea ice variability is more strongly related to the winter AO/NAM. This is reasonable because Arctic sea ice variability is large in the Barents Sea in autumn and the East Siberia Sea in summer, and thus the associated surface turbulent heat flux anomaly over the Arctic Ocean is much larger in autumn than that in summer. As we discuss below, this supports a possible mechanism for the modulation of the AO/NAM. Furthermore, the November sea ice variation is also related to summer AO/NAM. As shown in Ogi et al. [2008, 2010], this is due to wind-driven sea ice anomaly in the summer Arctic Ocean and its memory in the autumn. Because in general the atmospheric variation cannot persist beyond the season, the results might imply longer persistency of the AO/NAM through the Arctic sea ice variation.

### 4.2. Climatological impacts of recent sea ice and SST anomalies

In section 3.2, we showed the climatological impact of recent ice reductions on the NH climate by comparing results from three model experiments. The winter atmospheric response to NH ice reduction (*N.ICE* case) shows geopotential height anomalies similar to the EU pattern. Surface temperature anomalies associated with the EU pattern show continental cooling structures similar to the regression fields from the ERA interim reanalysis dataset (Figure 6). Spatial correlations between the model responses and the ERA interim anomalies are highest in *N.ICE* and lowest in *Global* (Table 2), indicating the importance of influences from SST anomalies. Changes in tropical SSTs cause a PNA-like pattern (*Global* case). Changes in SSTs in the mid- and high latitudes generally warm the NH (e.g., see the comparison between T2m anomalies in the *Northern* and *N.ICE* cases). These
impacts of SSTs significantly mask the impacts of ice reductions (Figure 4; Table 2).

4.3. A possible mechanism for the modulation of the AO/NAM

The negative phase shift in the primary AO/NAM-like mode is found in both the model experiments and the reanalysis data. The modulation of the AO/NAM results from the direct (i.e., generation of Rossby waves) and indirect (i.e., changes in the meridional circulation) responses to sea ice reduction. The stationary Rossby wave response to sea ice reduction in the Barents Sea first appears in eastern Siberia in the early winter (Figures 3 and 16), which is similar to the results of previous studies [Francis et al., 2009; Honda et al., 2009; Liu et al., 2012; Tang et al., 2013]. Later, an annular-mode-like pattern, anticyclonic anomalies in the polar region and cyclonic anomalies in the mid-latitudes, emerges in the mid- and late-winter. Our TEM analysis shows that stationary eddies corresponding to the Rossby wave response in eastern Siberia induce an anomalous meridional circulation that cools the mid-latitudes and warms the polar region (Figure 10). Finally, anticyclonic anomalies over the polar region induce or intensify the cyclonic anomalies in the mid-latitudes, especially in Europe, eastern Siberia, and eastern North America through the preferred wave-guide of the climatological jet stream (Figure 9).

Our results in the model experiments are consistent at some points with the previous studies that showed the negative AO-like response to the Arctic sea ice reduction [Magnusdottir et al., 2004; Deser et al., 2004; Alexander et al., 2004]. Our findings of that the indirect response and its physical
process through the stationary Rossby response to the ice reduction could explain the AO-projected component of the response to the sea ice variation shown by Magnusdottir et al. [2004] and Deser et al. [2004]. Honda et al. [1996, 1999] and Alexander et al. [2004] presented that the sea ice variation in the Sea of Okhotsk generates wave train-like response approaching North America. Even though, our model simulation does not show apparent wave train-like response to the Okhotsk sea ice reduction, cyclonic anomaly and associated cold anomaly are found in North America. This suggests that the Rossby wave response to the Arctic anticyclonic anomaly due to the additional warming in the Arctic has impacts on the North America as well as the Okhotsk sea ice reduction does. This is consistent with that the Arctic sea ice reduction excites a modulation of AO/NAM as the indirect influences.

Petoukov and Semenov [2010] showed the strong non-linearity of the linkage between sea ice reductions in the Barents-Kara seas and associated AO/NAM signals. For example, an intermediate amount of sea ice reduction observed in recent years induces a negative AO. However, either smaller or larger reductions in sea ice can induce a positive AO through non-linear responses to surface heat sources. This is critical to our results as it is difficult to infer future projections of NH climate change from a continuous decline in Arctic sea ice. It would be difficult to infer which phase of the AO would be more dominant in the far future after complete sea ice loss.

4.4. Seasonal evolution

Our results from both observation and model indicate a transition of the atmospheric response
to sea ice reduction that is more annular pattern associated with significant anomalies in the Atlantic region in the late winter (Figure 3 and 16). The upward surface heat flux associated with the sea-ice reduction in Barents Sea becomes large as season advances from late summer to mid-winter. After that, the heat flux forces the atmosphere during winter continuously. Then, the climatological wave guide for the Rossby wave propagation is located close to Barents Sea during winter (Figure 9). Due to this continuous forcing, the negative AO-like anomaly evolves from autumn to mid-winter (Figure 16). However, the modeled results show quite different with the late winter NAO-like pattern found in the observation results. Yamamoto et al., [2006] presented an appearance of the negative NAO in the late winter corresponding to the sea ice seesaw pattern between reductions in the Barents/Okhotsk and increases the Labrador/Bering Sea. AGCM experiments performed with prescribed sea ice anomaly that decrease in the Barents Sea paired with increase in the Labrador Sea also showed the negative NAO-like response more apparent than our model did [Magnusdottir et al., 2004; Deser et al., 2004; Alexander et al., 2004]. This might explain the difference of NAO signal in late winter between the observation and simulation, because the recent sea ice anomaly (i.e., Late minus Early) used in our experiments is smaller in the Labrador Sea (Figure 1h). This strongly suggests that local responses to the sea ice variation in the Labrador Sea as well as the Barents Sea plays crucial role to generate NAO-like response in the Atlantic.

Previous studies have suggested the stratospheric influence as a possible mechanism for the teleconnection between the variations in tropical SST and the NAO [Scaife et al. 2005; Cagnazzo and Manzini, 2009]. Takaya and Nakamura [2008] pointed to the importance of the fluctuations of the
planetary wave in November to the annular mode in mid-winter through modulation of the stratospheric polar vortex. It is thus not unrealistic to consider a possible link between sea ice reduction and negative phase shift of the NAO in late winter via the stratospheric path. However given the complexity, the topic of a climatic link between the Arctic, the tropics, and the stratosphere is beyond the focus of this study, and the roles of the stratosphere in the seasonal evolution of the climate impacts of sea ice reduction should be studied in the future.

5. Summary

In this paper, we used AGCM experiments to investigate the impacts of sea ice reduction on NH climate, and have demonstrated that in the early winter ice reductions in the Barents Sea cause tropospheric cyclonic anomalies and associated surface cooling anomalies in eastern Siberia. Moreover, the planetary wave modified in central to eastern Siberia induces an anomalous meridional circulation that shifts the AO/NAM-like pattern to the negative phase. Quantitative estimations derived from our modeling showed that atmospheric feedbacks from ice reduction induce a heating rate that is equivalent to 25.5% of the direct heating associated with ice reduction. Associated with the negative AO/NAM-like circulation, near-surface temperature anomalies tend to indicate warm-Arctic, cold-mid-latitudes (WACOM) conditions. This implies the broad influence of the variability of the Arctic ice extent on climate. PDF analysis showed that the frequency of the large negative (positive) AO/NAM has roughly doubled (halved) because of recent ice reductions. In daily basis, extremely negative AO tends to take place more frequently and the associated temperature anomalies in East
Asia, Europe, and North America become more severe. Our results imply some contribution from Arctic sea ice reduction to the severe cold weather outbreaks experienced in East Asia, North America, and Europe during the past few years. Our analysis of the ERA interim reanalysis dataset supports the negative phase shift of the AO/NAM due to recent ice reductions.

The recent atmospheric circulation changes in the real world reflect not only the impact of ice reduction, but also the impacts of the decadal and interannual variations in SST, aerosols, ozone, solar heating, etc. In particular, tropical SST variations have a large effect. For example, our model experiments showed that a PNA-like pattern is dominant if changes in tropical SSTs are taken into account. To estimate the climatic impacts of ice reductions more accurately, a quantitative estimation of the uncertainty associated with the impacts of other climate variability, besides ice reduction, is required. Therefore, in our future work we plan to perform ensemble-based GCM experiments in which historical external conditions are used for the recent past period (e.g., from 1979 to the present).

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To access our AFES simulation data, contact to the corresponding author. We thank to AFES development team at JAMSTEC and the Nakamura laboratory team at the University of Tokyo for their help setting up and tuning the AFES simulations. We also thank K. Nishii, K. Takaya, and A. Kuwano-Yoshida for their helpful discussions and comments. We would like to thank three anonymous reviewers for their valuable remarks to improve the quality of the paper. This study was supported by the Green Network of Excellence Program (GRENE Program) Arctic Climate Change Research Project.

Appendix: Estimation of the heating rate due to residual mean vertical motion

We estimated the heating rate associated with the residual mean vertical motion (w*) obtained from our TEM analysis. First, we calculated potential temperature advection due to the vertical motion from:

\[
\frac{\partial \theta}{\partial t} = -(w^*) \frac{\partial \bar{\theta}}{\partial z^*} \quad (1)
\]

where \( \theta \) is potential temperature, \( w^* \) is the residual mean vertical velocity, and \( z^* = -H \log(p/p_0) \). \( H \) is the scale height (assumed to be 7 km), \( p \) and \( p_0 \) are the pressures at a given level and the surface pressure (1000 hPa), respectively, (') indicates a 3-month average anomaly (N.ICE minus CNTL) and \( \bar{\cdot} \) indicates the climatology of the respective 3-month averages, which are defined as 120-year averages of CNTL and N.ICE. Note that we ignored second order terms of the potential temperature advection because fluctuations in the vertical gradient of the potential temperature are smaller than its...
climatological average.

We obtained the heating rate per unit mass from:

\[ Q^* = Cp \left( \frac{p}{p_0} \right) ^\kappa \left[ \frac{\partial \theta}{\partial t} \right] \]  
(2)

where \( Cp \) is the specific heat at constant pressure and \( \kappa \) is the Poisson constant. Finally, to compare with the turbulent heat flux due to ice reductions, we obtained the heating rate per unit area from vertical integration of \( Q^* \) as:

\[ J = \frac{1}{g} \int_{850 \text{hPa}}^{300 \text{hPa}} Q^* dp \]  
(3)

where \( g \) is the gravitational acceleration. \( Q^* \) is integrated between 850 and 300 hPa, and \( J \) indicates the dynamical heating rate in the free troposphere.

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Table 1. Description of the perpetual model simulations. ‘Early’ (‘Late’) indicates that monthly averaged SST or SIT for a 5-year period covering 1979–1983 (2005–2009) was used for the boundary conditions. Note that SST between 30°N and 50°N was gradually connected in the Northern case by e-folding.

<table>
<thead>
<tr>
<th>Integration period (yr)</th>
<th>SST and ICE south of 30°N</th>
<th>SST north of 50°N</th>
<th>ICE in the NH</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNTL</td>
<td>60</td>
<td>Early</td>
<td>Early</td>
</tr>
<tr>
<td>Global</td>
<td>20</td>
<td>Late</td>
<td>Late</td>
</tr>
<tr>
<td>Northern</td>
<td>20</td>
<td>Early</td>
<td>Late</td>
</tr>
<tr>
<td>N.ICE</td>
<td>60</td>
<td>Early</td>
<td>Early</td>
</tr>
</tbody>
</table>

Table 2. Spatial correlation coefficients between model fields (see Figure 5a–f) and ERA interim anomalous fields in NDJ and DJF regressed on the November sea ice area (1st and 3rd columns of Figure 3). The correlations were calculated using anomaly fields northward of 30°N.

<table>
<thead>
<tr>
<th></th>
<th>Geopotential height at 500 hPa</th>
<th>Temperature at 2 m height</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ERA in NDJ</td>
<td>ERA in DJF</td>
</tr>
<tr>
<td>Global</td>
<td>0.00</td>
<td>0.04</td>
</tr>
<tr>
<td>Northern</td>
<td>0.46</td>
<td>0.25</td>
</tr>
<tr>
<td>N.ICE</td>
<td>0.68</td>
<td>0.50</td>
</tr>
</tbody>
</table>
Figure 1. (a) SST [K] (b) SIT (cm) anomalies of Late (2005–2009) minus Early (1979–1983) periods in September. (c) Sign-reversed regression coefficients of detrended SIT anomaly against normalized sea ice area index in September. Note that the signs of the regression coefficients have been reversed so that they correspond to the decrease in Arctic sea ice. (d-f) for November. (g-i) for January.
Figure 2. Time series of DJF averaged AO and NAO indices (obtained from NOAA CPC at http://www.esrl.noaa.gov/psd/data/climateindices), together with the normalized time series of Arctic sea ice area in the previous November.
Figure 3. Top row: Three-month mean geopotential height anomalies at 500 hPa in OND, NDJ, DJF, and JFM from left to right. These anomalies are lag regression coefficients of the Arctic sea ice area in November. Note that the sign of the coefficients is reversed so that red (blue) corresponds to positive (negative) anomalies when Arctic sea ice decreases. The contour interval is 5 m, and the zero line is omitted. Light, moderate, and heavy shading indicate statistical significance of over 90%, 95%, and 99%, respectively. Middle row: Regression coefficients of sea level pressure with a contour interval of 0.5 hPa. Bottom row: Regression coefficients of 2-m temperature with a contour interval of 0.5 K.
Figure 4. Changing correlation between the November sea ice area and 3-month mean AO/NAO indices and Z500-EOF1 score from ASO to MAM. Note that the signs of the correlation coefficients have been reversed so that they correspond to the decrease in Arctic sea ice.
Figure 5. (a) Anomalies of geopotential height at 500 hPa in DJF for the Global run relative to the CNTL run. Contours indicate the differences in meters. Light, moderate, and heavy shading indicate statistical significance (t-test) at the 90%, 95%, and 99% confidence levels, respectively. (b) and (c) As for (a), but for the Northern and N.ICE runs, respectively. (d–f) As for (a–c), but for 2-m temperature. Contour interval is 1.0 (0.5) K for positive (negative) anomalies.
Figure 6. (a) DJF mean temperature anomalies at 850 hPa of N.ICE against CNTL. (b) Zonal mean temperature anomalies. Black and light blue lines indicate the global mean and land-only mean, respectively. (c) and (d) As for (a) and (b), but for the regressed field of ERA interim in NDJ upon November sea ice area. Values corresponding to $-1.0 \sigma$ of sea ice area are shown.
Figure 7. DJF averaged climatology of temperature (K, contours), vectors of the horizontal wind anomaly (arrows), and 3D temperature advection (K d$^{-1}$, shading) at 850 hPa. The climatology and anomalies are calculated as ($CNTL$ plus $N.ICE$)/2 and $N.ICE$ minus $CNTL$, respectively. Temperature advection is obtained from the products of 3D wind anomalies and the climatological potential temperature gradient. The arrow length corresponding to 1.0 m s$^{-1}$ is indicated in the top-right corner of the panel.
Figure 8. (a) DJF averaged anomalies of \textit{N.ICE} compared with the \textit{CNTL} run. Shading, contours, and arrows indicate turbulent heat flux (W m\(^{-2}\); i.e., sensible heat plus latent heat, upward positive), the vertical component of wave activity flux (10\(^{-2}\) m\(^2\) s\(^{-2}\)) at 700 hPa, and the horizontal wave activity flux vector (m\(^2\) s\(^{-2}\)) at 300 hPa, respectively. The arrow length corresponding to 1.0 m\(^2\) s\(^{-2}\) is indicated in the top-right corner of each panel. Note that the horizontal wave activity fluxes in the southward (northward) direction are drawn as thick (thin) arrows. (b) Geopotential height anomalies at 300 hPa in DJF for \textit{N.ICE} compared with the \textit{CNTL} run. Contours and shading are the same as Figure 4a. Associated wave activity fluxes are shown by green arrows. The corresponding scale and units are displayed at the top-right of the panel. (c) Vertical cross-sections of geopotential height anomalies and the zonal and vertical components of wave activity flux along the brown line in (b).
Figure 9. DJF averaged climatologies of (a) geopotential height (m), and (b) absolute vorticity ($10^6$ s$^{-1}$) at 300 hPa. The climatology is defined as the 120-year average of the CNTL and N.ICE runs. Places where the meridional gradient of the absolute vorticity exceeds $10^{-8}$ s$^{-1}$ km$^{-1}$ are hatched in red in (b).
Figure 10. DJF average zonal mean anomalies of N.ICE compared with the CNTL run. (a) Zonal wind anomalies (m s$^{-1}$). (b) EP flux (m$^2$ s$^{-2}$, green arrows) and its divergence (m s$^{-1}$ day$^{-1}$, contours). (c) Residual mean circulation (m s$^{-1}$, black arrows) and stream function ($10^{10}$ m$^2$ s$^{-1}$, contours). For all panels, light, moderate, and heavy shading indicate statistical significance at the 90%, 95%, and 99% confidence levels, respectively. For (b) and (c), the vertical component of the vectors is multiplied by a factor of 400, and arrow lengths corresponding to 5.0 m$^2$ s$^{-2}$ and 0.05 m s$^{-1}$ are displayed in the top-right corner of panels (b) and (c), respectively.
Figure 11. (a) As for Figure 4a, but for the eddy momentum flux \((u'v')\) obtained from daily mean data. An eddy is defined as an anomaly from the zonal mean field according to the TEM formulation. The units are \(\text{m}^2\ \text{s}^{-2}\). (b) Eddy heat flux \((v'T')\, \text{K m s}^{-1}\). (c) and (d) As for (a) and (b) but for fluxes obtained from DJF mean data. Only the fluxes due to the stationary eddies are shown.
Figure 12. The EOF1 loading pattern of geopotential height at 500 hPa in DJF for: (a) CNTL, (b) N.ICE, and (c) N.ICE combined with CNTL. The contribution of EOF1 is indicated at the bottom-right of each panel. (d) Standardized EOF1 scores for the combined EOF. The left and right halves of (d) correspond to the 60-year periods of CNTL and N.ICE, respectively.
Figure 13. Histogram of the EOF1 score from the combined EOF (0.2 σ bins); red and blue bars indicate the CNTL and N.ICE periods, respectively. The horizontal axis shows scores for the center of each bin. The vertical axis of the left hand side indicates number of count for each bin. Lines indicate the probability density function (PDF) estimated from the EOF1 score for the CNTL (red) and N.ICE (blue) periods, respectively. The vertical axis of the right hand side indicates probability density. The mean score and the integration of the PDF that is more (less) than 1.0 σ (−1.0 σ) are shown in the panel in the colors corresponding to CNTL and N.ICE (red and blue, respectively).
Figure 14. PDF of the daily EOF1 scores obtained by daily Z500 anomalies projected onto the individual EOF1 patterns. Red and blue lines indicate the CNTL and N.ICE periods, respectively, and black line indicates their difference (N.ICE minus CNTL). The horizontal axis shows scores normalized by the standard deviation of 3-month mean EOF. The vertical axis indicates probability density. Standard deviation, skewness, and kurtosis are shown in the panel in the colors corresponding to CNTL and N.ICE (red and blue, respectively), and their differences of N.ICE minus CNTL (Δ) are shown in black.
Figure 15. (a) DJF composite of daily SLP anomalies for daily EOF1 score of \textit{CNTL} less than -2.0 standard deviation of 3-month mean EOF. (b) for daily EOF1 score of \textit{N.ICE}. (c) Difference of composite anomalies of (b) minus (a). (d-f) As in (a-c) but for T2m anomalies.
Figure 16. (a–d) Z500 anomalies of \textit{N.ICE} compared with \textit{CNTL} in OND, NDJ, DJF, and JFM, respectively. Contours and shading intervals are as in Figure 4a. (e) Time series of the mean difference of the EOF1 scores (i.e., \textit{N.ICE} minus \textit{CNTL}).