Role of Eurasian Snow Cover in Linking Winter-Spring Eurasian Coldness to the Autumn Arctic Sea Ice Retreat

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Abstract An anomalous “north-south” dipole mode of the snow water equivalent (SWE) persisting from winter to spring is detected over the Eurasian mid-to-high latitudes in this study. Using observational data sets and numerical experiments of the Community Atmospheric Model (5.0), we show that this mode contributes to prolonged winter-springtime coldness in midlatitude Eurasia and is closely linked to the declining November Arctic sea ice concentration. The decline in the sea ice concentration over the Barents-Laptev Seas can induce a teleconnection pattern over the mid-to-high latitudes in the following winter, accompanied by an anomalous ridge over the Ural Mountains and an anomalous trough over Europe and East Asia. Such changes in the large-scale circulation lead to more cold surges and heavy snowfall in the midlatitudes and light snowfall in the high latitudes, forming an anomalous north-south dipole mode of the SWE, which further reduces the temperature through thermodynamic feedback. Due to seasonal memory, this SWE pattern can persist into the following spring and can lead to springtime midlatitude coldness via thermodynamic and dynamic processes. For the thermodynamic process, the anomalous SWE condition can lead to anomalous wet soil, reduced incoming surface solar radiation, and cooling air in the midlatitudes. This phenomenon induces an enhanced Siberian High and a deepened East Asian trough via the snow-Siberian high-feedback mechanism, which favors a cold spring in northern East Asia. Further analysis suggests that an empirical seasonal prediction model based on the SWE can reasonably predict East Asian spring temperature anomalies.

1. Introduction

In recent decades, several extreme events, such as record-breaking snowfall and cold spells, have attacked the Eurasian midlatitudes in winter and spring (e.g., Jones et al., 2012; Takano et al., 2008; Wang & Chen, 2010, 2014; World Meteorological Organization, 2009, 2011, 2018). Some of those events lasted from winter to spring under the background of global warming and thus received considerable attention. Long-lasting and continuous heavy snowfall and cold spells often have disastrous impacts on society, the economy, agriculture, and people’s lives. Therefore, it is important to investigate possible controlling factors for the prolonged winter-spring snow cover and the surface air temperature (SAT) in northern Eurasia, particularly their variability.

The snow cover and SAT variations in Eurasia are affected by many factors, including the Arctic Oscillation (AO)/North Atlantic Oscillation (Zveryaev & Sulev, 2009; Cheung et al., 2012), the El Niño–Southern Oscillation (Graf & Zanchettin, 2012), and the Arctic sea ice variability (e.g., Liu et al., 2012). The SAT variability in Eurasia is also related to Eurasian snow cover (e.g., Chen et al., 2003; Handorf et al., 2015; Luo & Wang, 2018; Wang et al., 2009). For instance, Handorf et al. (2015) indicated that changes in Eurasian snow cover may modulate the midlatitude atmospheric circulation throughout winter and spring. Studies also found that excessive autumn Siberian and continental-scale snowfall could modulate the variability in the winter AO (e.g., Gastein et al., 2017; Gong et al., 2003; Saito & Cohen, 2003; Smith et al., 2011). Moreover, anomalous autumn snow cover was also found to cause strengthened East Asian winter monsoons via positive feedback between the thermal conditions over land surfaces and the upper atmospheric circulation system (Luo & Wang, 2018), which would further modulate the variation in local SAT (Chen...
et al., 2003). However, few studies have explored whether any snow cover anomaly patterns can persistent from winter to spring and whether such persistence may play a key role in modulating the midlatitude atmospheric circulation system as well as the associated SAT variability in spring.

Previous studies have linked the possible role of the dramatic decline in Arctic sea ice, especially over the Barents-Kara Seas, to the anomalous heavy snow cover and coldness over the Eurasian midlatitudes (Honda et al., 2009; Liu et al., 2012). When the sea ice was reduced in the Barents-Kara Seas, both observational and modeling studies have identified several responses in the mid-to-high-latitude atmospheric circulation, such as the occurrence of a negative AO/North Atlantic Oscillation phase (Cohen et al., 2010), the activities of anomalous Rossby wave trains over the Eurasian midlatitudes (Honda et al., 2009), and the modification of the storm track in the North Atlantic sector (Alexander et al., 2004; Magnusdottir et al., 2004). Such circulation changes would in turn affect winter Eurasian snowfall and SAT variability in the midlatitudes (e.g., Alexander et al., 2010; Budikova, 2009; Deser et al., 2010; Francis et al., 2009; Jaiser et al., 2012; Overland & Wang, 2010; Petoukhov & Semenov, 2010). However, for the SAT variability in spring, Chen and Wu (2017) highlighted the dynamic influence of autumn Arctic sea ice changes in the Pacific sector on modulating the trans-seasonal troposphere-stratosphere coupling. The reduction in Arctic sea ice could fuel stronger winter snowstorms and yield extensive snow cover over Eurasia via modulating large-scale circulation anomalies and increasing the atmospheric water vapor content (e.g., Liu et al., 2012). However, it should be noted that the link between Arctic sea ice and the Eurasian climate is still under debate (e.g., Gao et al., 2015; Vihma, 2014). Not all studies have found a significant relationship between Arctic sea ice loss and midlatitude cooling, because the latter may be offset by the advection of warmer air masses (e.g., McCusker et al., 2016; Screen, 2017; Sorokina et al., 2016). Moreover, most studies focused on the Arctic sea ice-winter snow linkage, while the Arctic sea ice influence on the subsequent springtime snow has rarely been investigated, not to mention the long-lasting snowy and colder conditions from winter to spring. The question then arises whether autumn Arctic sea ice can have a thermodynamic influence on the spring SAT variability through the persistence of snow cover from winter to the subsequent spring. In the present study, we will address this question.

Considering the lack of reliable observational and reanalysis snowfall data (Broxton et al., 2016), most studies were conducted by using a compensatory analysis of snow cover data (e.g., Cohen et al., 2012; Liu et al., 2012). In particular, Liu et al. (2012) focused on snow extent in the midlatitudes but failed to identify signals in the snow depth as well as those in the high latitudes. In this study, we use several snow data sets to verify our conclusions. The data sets, methods, and numerical model applied are described in section 2. We depict the winter and spring Eurasian snow water equivalent (SWE) variation and its potential linkage with the reduced Arctic sea ice condition in section 3; possible climatic influences of the SWE variation on persistent cooling from winter to spring over mid-to-high-latitude Eurasia are also explored. In section 5, we discuss possible physical mechanisms behind the Arctic sea ice influence on the winter-spring SWE variation and their impacts on the climate. In section 6, we establish a prediction model for the East Asian spring temperature using the winter Eurasian SWE. A summary and discussion are presented in section 6.

2. Data Sets, Methods, and the Numerical Model

The data sets analyzed in this study include the monthly ERA-Interim data set (Dee et al., 2011). The variables from this reanalysis product include air temperature, wind, and geopotential height at various levels, plus latent heat flux, surface solar radiation, snowfall, and soil moisture. The monthly sea surface temperature (SST) and sea ice concentration (SIC) data are from the Hadley Centre SST data set (Rayner et al., 2003). The monthly Northern Hemisphere SWE (GlobSnow SWE) data are from the Finnish Meteorological Institute (Takala et al., 2011). We also use the weekly Northern Hemisphere Equal-Area Scalable Earth Grid 2.0 snow cover data set from Rutgers University (Robinson et al., 2012).

In the present study, the winter mean refers to the December-January-February (DJF) mean, and the spring mean refers to the March-April (MA) mean because the SWE and its variance are relatively small after April. An empirical orthogonal function (EOF) analysis is performed to identify the major modes of the Eurasian SWE in boreal winter. An observational analysis is performed for the period of 1979–2015,
with linear trends removed prior to the analysis. The statistical significance of the regression and correlation coefficients and the ensemble-mean differences of the model outputs are determined by two-tailed Student's t tests.

The numerical model we used in this study is the National Center for Atmospheric Research Community Atmospheric Model version 5 (CAM5), with a horizontal resolution of approximately 1.9° latitude and 2.5° longitude. It has 30 levels, with the topmost level at 3.6 hPa. More information on this model will be provided in section 5.3.

3. Dipole Mode of the Winter-Spring SWE and Its Association With Midlatitude Temperature and Arctic Sea Ice Retreat

3.1. Persistent Eurasian SWE Mode From Winter to Spring

To depict the major modes of SWE variability, an EOF analysis is applied to the winter SWE anomalies over mid-to-high-latitude Eurasia (20°–140°E, 40°–70°N). The first EOF mode (EOF1) accounts for approximately 13.4% of the total variance and is characterized by a west-east dipole pattern (Figure 1a). The
second EOF mode (EOF2) accounts for 11.8% of the total variance and is characterized by a “north-south” dipole pattern, with less SWE over the region north of 60°N and more SWE to the south (Figure 1b). This feature is different from that in previous studies due to different snow parameters and analysis periods; for example, Dash et al. (2005) used the Eurasian snow depth obtained from the Historical Soviet Daily Snow Depth version II data set to show a same-sign anomaly over Eurasia. As seen in Figures 1c and 1d, the first principal component (PC1) of the SWE displays an evident interdecadal variation, while the second principal component (DJF-PC2 SWE) is predominantly characterized by interannual fluctuation. According to the North significance test (North et al., 1982), the eigenvalues in the first two EOF modes are significantly separated, and they are thus independent of each other. Similarly, the second mode of the DJF ERA-Interim monthly accumulated snowfall amount data shows a north-south dipole pattern as well as an interannual variation in its corresponding PC2, which is highly correlated with the DJF-PC2 SWE ($R = 0.56$; Figure S1). The aforementioned results indicate that the dipole mode is the major mode representing the interannual variation in the winter snow cover. Since the first mode of the SWE can hardly persist into the following spring, only the second mode is discussed in the present study.

Figure 2a shows the spring SWE anomalies correlated with the DJF-PC2 SWE. The winter north-south dipole SWE pattern is followed by an identical SWE pattern in spring, with a positively correlated SWE in midlatitudes south of 60°N and negatively correlated SWE in the north. This similarity reflects a seasonal footprint characteristic of the winter-spring SWE. Note that this pattern resembles the EOF1 mode of spring SWE anomalies, which accounts for 15.3% of the total variance (Figure 2b). The correlation coefficient between the PC1 of the spring SWE (MA-PC1) and the DJF-PC2 SWE is as high as 0.70 (Figure 1b). These results illustrate that the north-south dipole SWE pattern and its variation can persist from winter into the subsequent spring.

To investigate the persistence characteristics of the snowfall and snow cover fraction parameters related to the variation in the SWE, we present their regression anomalies in Figure 3. The temporal regression anomaly of snowfall represents an inclination (referred to as A) from a mathematic sense, which means that for

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**Figure 2.** (a) Correlation coefficients between MA-mean SWE and the DJF-PC2 SWE. (b) Spatial pattern of the first EOF mode of MA-mean SWE. Stippling in (b) denotes significant correlation coefficient between MA-PC1 SWE and MA-mean SWE. Correlation coefficients significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively.

**Figure 3.** Linear regression of (a) DJF-mean and (b) MA-mean ERA-Interim snowfall amount onto the DJF-PC2 SWE during 1979–2015 (units: mm/month). (c and d) The same as in (a) and (b) except for Rutgers snow cover fraction. Values significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively.
the DJF-PC2 SWE equal to 1, there should, in a statistical sense, be A mm of excess snowfall. If the DJF-PC2 SWE equals 1, the dipole mode of the winter SWE is accompanied by approximately 6–8 mm less snowfall over the northern continent and 5–7 mm excess snowfall over the midlatitudes, particularly over Europe and northern/central East Asia (Figure 3a). Such a snowfall pattern would lead to a similar spatial distribution in Rutgers’s snow cover fraction (SCF) in Eurasia, with a broader above-normal SCF (above 5%) in large parts of the midlatitudes, including Southern Europe and northern/central East Asia (Figure 3c). Following the winter SWE variation, the spring snowfall pattern shows persistent negative anomalies (1.6–2.4 mm) in northern Eurasia (Figure 3b), and the spring SCF shows persistent positive anomalies (above 5%) to the south (Figure 3d). This finding indicates that the winter snowfall and the SCF reveal identical footprint characteristics as the SWE.

To examine whether the abovementioned north-south dipole SWE patterns have any connection with the Eurasian SAT, Figure 4 presents regression anomalies in the SAT from winter to spring onto the DJF-PC2 SWE. In winter, continent-scale coldness with approximately 1.2 °C lower temperature prevails over mid-to-high-latitudes stretching eastward from Europe to northern East Asia (Figure 4a). In the following spring, the midlatitude cold situation with 0.5 °C lower temperature
3.2. Relationship With the Preceding Autumn SIC

To investigate the possible link between the winter Eurasian SWE with the Arctic sea ice, we use correlation maps of the September-December Arctic SIC with the DJF-PC2 and the MA-PC1 (Figure 5). The SWE dipole pattern is negatively correlated with the September SIC, which is confined to a quite small region over the Pacific sector (Figures 5a and 5e) but is significantly and negatively correlated with the October-December SIC over the Barents-Laptev Seas (Figures 5b–5d and 5f–5h). The November SIC signals are particularly robust with respect to the DJF-PC2 and MA-PC1 SWE (Figures 5c and 5g); hence, we define an Arctic SIC index (SICI) using the regionally averaged November SIC over the region within 30°–100°E and 67°–85°N. The November SICI is closely correlated with the DJF-PC2 SWE ($R = -0.59$) and with the MA-PC1 SWE ($R = -0.46$), indicating a significant and coincident relationship between the November SICI and winter-spring SWE variations (Figure 1d). However, the October SIC and the DJF mean SIC are not significantly correlated with the DJF-PC2 and the MA-PC1 SWE (figure not shown).

Figure 6 displays the SWE and SAT anomalies in winter and spring, respectively, regressed against the preceding November SICI. The variation in the November SIC is accompanied by a north-south dipole pattern of the winter SWE that is quite similar to the EOF2 of the winter SWE (Figure 6a versus Figure 1d). This winter dipole pattern is followed by a similar spring SWE pattern, which largely resembles the EOF1 of the spring SWE (Figure 6b versus Figure 2b). These relations imply a possible role of the November SIC in affecting the concordant Eurasian winter-spring SWE variability, especially through changes in snowfall (Figure S2). Similarly, the SAT pattern associated with the declining November SIC reveals broad surface cooling with 1 °C lower temperature in mid-to-high-latitudes in winter (Figure 6c) and with 0.6 °C lower temperature in the following spring (Figure 6d). Then, we checked the variation in the spring SAT and its link with the November SIC (not shown). We find that the EOF1 of the spring SAT reflects midlatitude coldness, similar to that in Figure 4b, and has an intimate relationship with the DJF-PC2 SWE ($R = 0.34$) and the

Figure 6. Linear regression of (a) DJF-mean and (b) MA-mean SWE anomalies (units: cm) onto the SICI index. (c and d) The same as (a) and (b) except for SAT (units: °C). Linear regression of MA-mean SAT anomaly (units: °C) onto the MA SWE-PC1 index is shown in (e). Values significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively.
MA-PC1 SWE ($R = 0.50$) but a weak relation with the November SIC ($R = 0.17$). Hence, the spring SWE (Figure 6e) seems to play a leading role in determining the spring temperatures in Europe and northern East Asia. These results imply that the autumn SIC may influence spring coldness in midlatitudes through the prolonged SWE condition.

4. Possible Physical Mechanism

4.1. Winter Responses to SIC Forcing

Figure 7 displays the winter 500-hPa (H500) and 300-hPa geopotential height (H300) fields associated with the variations in the DJF-PC2 SWE and the November SICI. These results show that associated with the anomalously low SIC and north-south dipole SWE, a notable feature occurs; that is, the H500 is 20–40 m higher over a broad region of the Arctic Ocean and northern Eurasia, along with an anomalously higher upper level ridge extending from the Barents-Laptev Seas into the Ural Mountains, which is compensated by 8–32 m lower height in the midlatitude North Atlantic Ocean and Eurasia (Figures 7a and 7b). The H300 fields largely resemble the H500 field in the Arctic and high latitudes (Figures 7c and 7d). Note that these patterns are similar to the negative phase of the winter AO, which has been reported in several studies.
Figure 8. Same as Figure 7 except for DJF-mean (a and b) 300-hPa zonal wind (units: m/s), (c and d) 850-hPa wind (units: m/s), and (e and f) surface shortwave radiation (units: W/m²). Magnitude in (e) and (f) is divided by 2.
with a more meridional meandering in the Eurasian midlatitudes and an anomalous low trough in northern East Asia. Under such a circulation change, the prevailing westerly wind in the upper troposphere blowing across the subpolar region from the North Atlantic to Europe and the Far East is weakened by approximately 2 m/s.

Figure 9. Linear regression of (a) MA-mean sea level pressure (units: hPa), (b) 500-hPa geopotential height (units: m), and (c) 300-hPa zonal wind (units: m/s) fields onto the DJF-PC2 SWE. Values significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively. (d–f) and (g–i) The same as (a)–(c) except for linear regressions onto the SICI and MA-PC1 SWE, respectively. Magnitudes in the middle column are divided by 10.

(Jaiser et al., 2012; Overland & Wang, 2010), with a more meridional meandering in the Eurasian midlatitudes and an anomalous low trough in northern East Asia.

Under such a circulation change, the prevailing westerly wind in the upper troposphere blowing across the subpolar region from the North Atlantic to Europe and the Far East is weakened by approximately 2 m/s.
As noted by Eichelberger and Hartmann (2007) and by Sun et al. (2016), the variation in the westerly jet stream can directly affect the Eurasian winter climate, particularly the increase in blocking events (Davini et al., 2012; Hoskins et al., 1983). In our results, an anomalous upper level blocking ridge is observed over the Barents Sea–Ural Mountains, and such a blocking-like structure favors more frequent incursions of cold spells from the Arctic into the midlatitudes. Moreover, the 850-hPa wind field is characterized by an anomalous anticyclone over the region from the Arctic to the Ural Mountains and by two anomalous cyclones over the North Atlantic section and East Asia (Figures 8c and 8d). The cyclonic anomalies bring water vapor from the North Atlantic to Southern Europe and from the Northwest Pacific to northern East Asia, resulting in higher snow efficiency in these regions (Figures 1b, 3a, and 3c). In contrast, easterly wind anomalies prevail over central and northern Europe, leading to less warm air advection from the Atlantic, and therefore colder SAT and less snowfall in central and northern Europe (Figures 4a and 6c). These results suggest that the wavy pattern over the midlatitudes linked to the diminishing Arctic SIC is probably an important contributor to the winter SWE variability.

In addition to the direct influence of the tropospheric wavy structure that boosts cold waves into the midlatitudes, the prominent positive height anomalies over the Siberian Plateau representative of a stronger-than-normal Siberian-Mongolian High (figure not shown) are also responsible for winter surface cooling. More extended snow cover in the midlatitudes acts to thermodynamically reduce the solar heat flux by approximately 1.6 W/m² (Figures 8e and 8f) and the turbulent heat fluxes, which in turn increase the surface cooling over the Siberian Plateau and favor the accumulation of intense cold air masses (Cohen et al., 2012; Ghatak et al., 2012; Sun, Yang, et al., 2016). This combined effect of blocking-like circulation and heavy SWE anomalies will result in cold winter in the northern continents (Figure 4a).

4.2. Spring Circulation and SAT Responses

A question that remains unclear is how the November Arctic SIC or associated winter Eurasian SWE causes persistent surface cooling in the subsequent spring. To answer this question, we primarily examined the possible role of spring atmospheric circulation. Figure 9 presents the regression fields of tropospheric circulation onto the DJF-PC2 SWE, SICI, and MA-PC1 SWE indices. Apparently, circulation anomalies over Eastern
Figure 11. CAM5 composite differences between sensitivity experiments and control experiments. (a) DJF 1,000-hPa geopotential height (units: m), (b) 850-hPa wind (units: m/s), (c) 500-hPa geopotential height (units: m), (d) 300-hPa zonal wind (units: m/s), (e) MA 500-hPa geopotential height (units: m), and (f) MA 300-hPa zonal wind (units: m). Values significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively.
Europe and East Asia become insignificant and extremely weak in terms of the sea level pressure, H500, and U300 in response to the DJF-PC2 SWE (Figures 9a–9c) and SICI (Figures 9d–9f), except for the 10 m slightly stronger high over Siberia linked to a decreased SIC (Figure 9e). It seems that the preceding winter negative AO-like pattern (Figure 7), subjected to Barents-Kara SIC loss, could not persist into the subsequent spring. The cross section of the potential height averaged over 70°–90°N further indicates that such AO-like responses evolve from November but disappear in February (Figure S3). This result is somewhat different from that from Chen and Wu (2017), which indicated that the loss of Pacific SIC would excite the long-standing negative phase of the AO to cause uniform Eurasian cooling in early spring. In regard to the spring SWE, conversely, the circulation responses to the MA-PC1 SWE are particularly significant than that to the DJF-PC2 SWE (Figures 9a–9c), with a 1.2–1.5-hPa stronger Siberian High, a deepened trough with a 1.5 m lower height over northern East Asia as well as a 1.5 m/s slower upper level zonal flow, prefiguring more frequent cold waves (Figures 9g–9i). This enhanced East Asian trough will in turn increase the snow accumulation over these regions, which is called the snow-monsoon feedback mechanism (Luo & Wang, 2018).

Other important factors affecting the spring SAT might be the spring surface hydrological and thermal conditions. To verify this possibility, we check the regression of spring soil moisture and heat flux fields (i.e., the sum of surface solar radiation and the latent heat flux). The winter (Figure 10a) and spring (Figure 10c) north-south SWE dipole patterns correspond to identical soil patterns in Eurasia in spring, with 6–8 kg/m² wetter soils in the midlatitudes and 6–10 kg/m² drier soils to the north, while the November SIC loss does not correspond well to the midlatitude wet soil (Figure 10b). This is consistent with that Zhang et al. (2017), which unveiled the SWE influence on soil moisture. The extended snow cover in spring tends to increase the surface albedo and thus reduces the incoming surface solar radiation in Europe and northern East Asia. In addition, the SWE in late spring begins to thaw, and the latent heat flux starts to increase; the sum of the solar radiation and the latent heat flux is thereby reduced by 8 W/m² over the two regions (Figures 10d–10f), contributing to the anomalously colder SAT in Europe and northern East Asia (Figure 4b). This result suggests that the spring SWE plays a crucial role in maintaining springtime coldness over Europe and northern East Asia through modulating the surface thermal conditions and the associated circulation anomalies. This finding is consistent with that from Chen et al. (2003), which indicated the positive role of the winter SWE in boosting cold spells; however, their study was conducted for spring. If we remove the influence of the spring dipole SWE, the November SIC and the winter SWE would show no significant correlation with the spring SAT. Therefore, only if the winter SWE dipole pattern persists into spring will the springtime coldness over Europe and northern East Asia be affected by the November SIC and the winter SWE. This phenomenon also implies that the persistence of the SWE dipole pattern in spring may be due to seasonal memory.

### 4.3. Model Experiments

We verify the physical processes through which the declining SIC affects the north-south SWE anomaly dipole mode by conducting simulations of the CAM5 model. For these simulations, the SST and the SIC are specified as boundary conditions based on the Hadley Centre data sets. Two sets of experiments are carried out with different seasonally varying SICs and fixed SSTs. The control experiment is forced by the annually repeating monthly climatological SIC that represents the climatology for the 1979–2015 period. The sensitivity experiment perturbed by the basin-wide Arctic SIC is integrated with the prescribed SICs from November to the following February obtained by adding regression anomalies onto the DJF-PC2 SWE to the climatological SIC in the control experiment. The prescribed SICs from March to April in the sensitivity experiment are the same as those in the control experiment to exclude the influence of spring SIC and to highlight the effect of the spring SWE. Each experiment is integrated from 1 November to 30 April. To obtain reliable atmospheric circulation responses to sea ice loss (Mori et al., 2014; Screen et al.,

![Figure 12. CAM5 composite differences of (a) DJF-mean and (b) MA-mean SWE (units: cm) between sensitivity experiments and control experiments. Values significantly exceeding the 95% and 99% confidence levels are shown as purple and blue dots, respectively.](image-url)
2013), each experiment consists of 50 ensemble members with slightly varying initial atmospheric conditions derived from the control experiment. The response of the model to the prescribed SIC loss is examined by the ensemble-mean difference between the sensitivity and control experiments.

Figure 11 displays the DJF mean large-scale circulation responses to the anomalously low SIC over the Arctic. In the 1,000-hPa field, we see that the diminishing SIC forces an anomalous pattern over the mid-to-high latitudes of Eurasia, signifying a positive anomaly extending southward from the Arctic to Northern Siberia and negative anomalies over Europe and the North Pacific (Figure 11a). The apparent wavy patterns in the mid-to-high latitudes are also reproduced in the simulated 850-hPa wind (Figure 11b) and H500 (Figure 11c) fields. Correspondingly, a weakened westerly wind prevailing over the subpolar region, which favors the development of the wavy pattern due to its dynamical role in Rossby wave activity (e.g., Francis & Vavrus, 2012), is simulated in the U300 field (Figure 11d). Although the simulation shows good similarity to the observation, some differences are identified over the Ural Mountains and northern East Asia. Indeed, more meridional meandering of the jet stream is simulated over the regions south of the Ural Mountains, and the simulated low pressure tends to shift south-eastward when compared to the observation. These discrepancies are possibly related to the exaggerated Arctic amplification in the CAM5 simulations, which do not properly represent the Arctic clouds and ignore the coupled processes with the SST and sea ice models (Kay et al., 2012).

Associated with the abovementioned circulation anomalies, the winter snowfall deficit north of 56°N and snowfall increase to the south is largely simulated. Correspondingly, northern reductions in the SWE are fairly well reproduced, but the southern increases, seen in Figures 1b and 2b, are not significant, particularly for the negative SWE anomaly over central Siberia (Figure 12a). This negative SWE anomaly is probably due to the stronger upper level ridge anomaly over the Ural Mountains. Nevertheless, the model is able to capture the overall distribution of SAT anomalies, with uniform cold anomalies in mid-to-high-latitude Eurasia and warm anomalies over the Arctic (Figure 13a). In spring, negative SWE anomalies persist in high latitudes, while positive SWE anomalies recover slightly over Europe and East Asia (Figure 12b), which reproduces the deceleration of westerly winds, the development of an anomalous high over the Arctic extending to the Siberian region and the enhanced East Asian trough (Figures 11e and 11f). Consequently, a colder SAT is simulated over Europe and East Asia, with a warmer SAT in between these regions (Figure 13b). Although the regional details differ somewhat between the responses of the modeled SWE and SAT and the observations, the model simulations do reproduce most of the observed responses. The consistency between the model simulations and the observations support the hypothetical linkages among reduced November Arctic SIC conditions and the persistent SWE anomaly dipole mode and cold SAT from winter to spring.

5. Seasonal Prediction

The aforementioned relationships illustrate that the DJF north-south dipole SWE condition over Eurasia may provide potential predictability for springtime coldness in northern East Asia. Thus, we establish a statistical model for seasonal forecasting and assessing the potential risk of
such cold events. An empirical statistical prediction model is designed using only DJF-PC2 of SWE as the predictor in a simple linear regression equation with an independent-sample-test method to hindcast the East Asian (110°–130°E, 40°–55°N) spring SAT anomalies during 1979–2015. The prediction skill of the statistical model is evaluated using temporal correlation coefficients between the observation and the hindcast. Figure 14a compares the time series of the regional-averaged SAT anomalies calculated from the ERA-Interim data and the hindcast result of the empirical model. The hindcast SAT index displays variation similar to that from the ERA-Interim data, and their correlation coefficient is 0.54. Figure 14b further shows the spatial distribution of temporal correlations between the ERA-Interim and hindcast SAT anomalies over Eurasian continent during 2006–2015. They are positively correlated in the mid-to-high-latitude regions, with maximum correlation coefficients greater than 0.6 over Northern Europe and northern East Asia, exceeding the 99% significance level. On the other hand, we also obtain significantly negative correlations in northern/central Russia. This result implies that the new empirical model captures the variation in the spring SAT reasonably well in the European and northern East Asian regions but fails to reproduce the SAT variability in northern/central Russia. This finding confirms that the preceding wintertime SWE is a notable factor for the potential predictability of the spring SAT variability in Europe and northern East Asia.

6. Conclusions and Discussion

The DJF mean SWE over the Eurasian mid-to-high latitudes displays a large interannual variability in the EOF2, which accounts for 11.8% of the total variance and predominantly depicts a north-south anomalous dipole mode. Such SWE dipole pattern in winter tends to persist into the following spring. We present observed evidence and numerical experimental results to show that this persistent SWE dipole pattern plays an important role in linking the prolonged midlatitude coldness from winter to spring to the diminishing SIC in the Barents-Laptev Seas in the preceding autumn. This finding indicates that the winter SWE and/or the preceding November SIC are potential predictors for the winter-spring SAT variability in mid-to-high-latitude Eurasia. Detailed physical processes are summarized in Figure 15 and are described below.

The declining November SIC in the abovementioned Arctic regions forces a basically barotropic response in winter with a negative AO-like pattern in the high latitudes and a pronounced wave train pattern in the mid-latitudes. These patterns feature an anomalous upper level blocking ridge over the Barents Sea-Ural Mountains, along with two troughs over Europe and East Asia. This wavy pattern benefits the occurrence of excessive and frequent snowfall over the Eurasian midlatitudes, causing much SWE there and thus forming a north-south SWE dipole pattern. The SWE dipole pattern is accompanied by continental-scale lower temperatures over most parts of mid-to-high-latitude Eurasia. Moreover, the cold surface temperature in winter tends to continue into the subsequent spring, particularly in Europe and northern East Asia.

To investigate plausible physical processes for the influence of the preceding November Arctic SIC on the following winter-spring Eurasian SAT variability, we examined the circulation response and the surface condition response. The winter negative AO-like response, however, fails to persist into spring. In contrast, the SWE dipole pattern has a cross-seasonal persistence from winter to spring. When the north-south SWE dipole pattern emerges in winter, wetter soil and associated lower heat fluxes in the midlatitudes tend to occur in the subsequent spring, along with drier soils and higher heat fluxes in the high latitudes. Such
soil moisture and surface heat flux changes help the winter SWE and low temperatures persist until spring. Moreover, the long-standing SWE dipole pattern in turn induces a strengthened Siberia High and a deepened East Asian trough, favoring cold surges into northern East Asia. Finally, the spring SWE-related atmospheric circulation change contributes to cold northern East Asian springs by modulating the surface thermal conditions rather than the AO responses as reported by Chen, & Wu, et al. (2017).

Furthermore, we established a new empirical model for hindcasting the spring SAT anomalies in northern East Asia based on the winter DJF-PC2 SWE. The higher correlation between the empirical model and the observed spring SAT index implies that the winter SWE is indeed a key factor for the potential prediction of the spring SAT variability.

Our new findings include the influence of the autumn Arctic SIC loss on the prolonged SWE and lower SAT patterns that can be sustained from winter until the following spring. For example, most studies noted the SIC feedback on the winter AO (Jaiser et al., 2012; Liu et al., 2012; Peings & Magnusdottir, 2014; Screen et al., 2013) and the prolonged winter-spring AO (Chen et al., 2017). Luo and Wang (2018) focused on autumn and winter and found that the EOF1 pattern of the autumn snow cover extent, with a uniform anomaly in the Eurasian continent, can persist to winter and is linked to cold Eurasian winters. Chen et al. (2017) focused on the relaying role of the AO in affecting the spring SAT, while we investigate the relaying role of the EURASEW. Our findings provide additional precursor factors for predicting the spring SAT variability in addition to using the winter and spring AO.

Note that the designed experiment in the present study is very ideal and simplified compared to, for example, those of Screen (2017) and Ogawa et al. (2018), who performed coordinated experiments with several atmospheric general circulation models. Furthermore, current climate models have considerable uncertainties in terms of capturing the SIC-winter Eurasian climate connection (Gao et al., 2015; Liu et al., 2012; McCusker et al., 2016; Sorokina et al., 2016; Sun, Perlwitz, et al., 2016; Vihma, 2014). This underestimation of the modeled atmospheric response may be related to the partial representation of aerosols and clouds in CAM5, which are critical for explaining Arctic climate responses (Kay et al., 2012). In addition to Arctic sea ice loss, the natural variability in the midlatitude atmosphere also favors cold surface conditions in winter and spring (Liu et al., 2012; Screen et al., 2013; Sun et al., 2016; Wang et al., 2019). Considering the importance of SIC-atmosphere coupled processes, the current model response of midlatitude SWE anomalies to Arctic sea ice with Atmospheric Model Intercomparison Project-type simulation may be largely underestimated. Thus, to better understand the interaction between declining Arctic sea ice and the midlatitude atmosphere, Coupled Model Intercomparison Project-type simulations are needed in our further research. The nonlinear impacts of autumn Arctic sea ice and winter snow cover on the Eurasian climate will also be investigated.

Acknowledgments
The ERA-Interim data set is publicly available for download at https://apps.ecmwf.int/datasets/data/interim-full-moda/levtype=sfc/. The Hadley SIC data set is available at https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html. The GlobSnow SWE data set is available at http://www.globsnow.info/. The Rutgers snow cover data set is available at https://climate.rutgers.edu/snowcover/. The authors thank the three anonymous reviewers whose constructive comments are helpful for improving the overall quality of the paper. This research is supported by the National Key Research and Development Program of China (during the 13th Five-Year Plan; grant 2016YFA0601501), the National Natural Science Foundation of China (grants 41790472, 41730959, and 41505053), the National Basic Research Program of China (grant 2013CB430203), and the Basic Scientific Research and Operation Foundation of CAMS (grants 2018ZQ06, 2018KJ029, 2018KJ030).

References
Alexander, M. A., Bhatt, U. S., Walsh, J. E., Timlin, M. S., Miller, J. S., & Scott, J. D. (2004). The atmospheric response to realistic Arctic sea ice anomalies in an AGCM during winter. Journal of Climate, 17(5), 890–905. https://doi.org/10.1175/1520-0442(2004)017<0890: TARTRA>2.0.CO;2
Alexander, M. A., Tomas, R., Deser, C., & Lawrence, D. M. (2010). The atmospheric response to projected terrestrial snow changes in the late 21st century. Journal of Climate, 23(23), 6430–6437. https://doi.org/10.1175/2010JCLI3899.1
Broxton, P., Zeng, X., & Dawson, N. (2016). Why do global reanalyses and land data assimilation products underestimate snow water equivalent? Journal of Hydrometeorology, 17, 2743–2761. https://doi.org/10.1175/JHM-D-16-0056.1
Budikova, D. (2009). Role of Arctic sea ice in global atmospheric circulation: A review. Global and Planetary Change, 68(3), 149–163. https://doi.org/10.1016/j.gloplacha.2009.04.001
Cheung, H. N., Zhou, W., Mok, H. Y., & Wu, M. C. (2012). Relationship between Ural–Siberian blocking and the East Asian winter monsoon in relation to the Arctic Oscillation and the El Niño–Southern Oscillation. Journal of Climate, 25(2), 4242–4257.
Chen, H. S., Sun, Z. B., & Zhu, W. J. (2003). The effects of Eurasian snow cover anomaly on winter atmospheric general circulation: Part II. Model simulation. Chinese Journal of Atmospheric Sciences, 27(5), 848–860. (in Chinese)
Chen, S., & Wu, R. (2017). Impacts of early autumn Arctic sea ice concentration on subsequent spring Eurasian surface air temperature variations. Climate Dynamics, 30, 1–20.
Cohen, J., Foster, J., Barlow, M., Saito, K., & Jones, J. (2010). Winter, 2009–2010: A case study of an extreme Arctic Oscillation event. Geophysical Research Letters, 37, L17707. https://doi.org/10.1029/2010GL044256
Cohen, J. L., Furtado, J. C., Barlow, M. A., Alexeev, V. A., & Cherry, J. E. (2012). Arctic warming, increasing snow cover and widespread boreal winter cooling. Environmental Research Letters, 7(1), 14007–14014(8). https://doi.org/10.1088/1748-9326/7/1/014007
Dash, S. K., Singh, G. P., Shekhari, M. S., & Vernekar, A. D. (2005). Response of the Indian summer monsoon circulation and rainfall to seasonal snow depth anomaly over Eurasia. Climate Dynamics, 24(1), 1–10. https://doi.org/10.1007/s00382-004-0448-3
Davini, P., Cagnazzo, C., Gualdi, S., & Navarra, A. (2012). Bidimensional diagnostics, variability, and trends of Northern Hemisphere blocking. Journal of Climate, 25(19), 6496–6509. https://doi.org/10.1175/JCLI-D-12-00332.1
Sun, C. H., Yang, S., Li, W. J., Zhang, R. N., & Wu, R. G. (2016). Interannual variations of the dominant modes of East Asian winter monsoon and possible links to Arctic sea ice. *Climate Dynamics, 47*(1-2), 483–496. https://doi.org/10.1007/s00382-015-2851-3

Sun, L. T., Perlwitz, J., & Hoerling, M. (2016). What caused the recent “warm arctic, cold continents” trend pattern in winter temperatures? *Geophysical Research Letters, 43*, 5345–5352. https://doi.org/10.1002/2016GL069024

Takala, M., Luojus, K., Pulliainen, J., Derksen, C., Lemmetyinen, J., Kärnä, J. P., et al. (2011). Estimating northern hemisphere snow water equivalent for climate research through assimilation of space-borne radiometer data and ground-based measurements. *Remote Sensing of Environment, 115*(12), 3517–3529. https://doi.org/10.1016/j.rse.2011.08.014

Takano, Y., Tachibana, Y., & Iwamoto, K. (2008). Influence of large-scale atmospheric circulation and local sea surface temperature on convective activity over the Sea of Japan in December. *Scientific Online Letters on the Atmosphere, 4*, 113–116.

Vihma, T. (2014). Effects of arctic sea ice decline on weather and climate: A review. *Surveys in Geophysics, 35*(5), 1175–1214. https://doi.org/10.1007/s10712-014-9284-0

Wang, L., & Chen, W. (2010). Downward Arctic Oscillation signal associated with moderate weak stratospheric polar vortex and the cold 2009 December. *Geophysical Research Letters, 37*, L09707. https://doi.org/10.1029/2010GL042659

Wang, L., & Chen, W. (2014). The East Asian winter monsoon: Re-amplification in the mid-2000s. *Chinese Science Bulletin, 59*(4), 430–436. https://doi.org/10.1007/s11434-013-0029-0

Wang, L., Deng, A., & Huang, R. (2019). Wintertime internal climate variability over Eurasia in the CESM large ensemble. *Climate Dynamics, 52*(11), 6735–6748. https://doi.org/10.1007/s00382-018-4542-3

Wang, L., Huang, R., Gu, L., & Chen., & Kang, W. (2009). Interdecadal variations of the East Asian winter monsoon and their association with quasi-stationary planetary wave activity. *Journal of Climate, 22*(18), 4860–4872. https://doi.org/10.1175/2009JCLI2973.1

World Meteorological Organization (2009). WMO statement on the status of the global climate in, 2008 (World Meteorological Organization, Geneva) WMO-No. 1039.

World Meteorological Organization (2011). WMO statement on the status of the global climate in, 2010 (World Meteorological Organization, Geneva) WMO-No. 1074.

World Meteorological Organization (2018). WMO statement on the status of the global climate in, 2017 (World Meteorological Organization, Geneva) WMO-No. 1212.

Zhang, R., Zhang, R., & Zuo, Z. (2017). Impact of Eurasian spring snow decrement on East Asian summer precipitation. *Journal of Climate, 30*(9), 3421–3437. https://doi.org/10.1175/JCLI-D-16-0214.1

Zveryaev, I. I., & Gulev, S. K. (2009). Seasonality in secular changes and interannual variability of European air temperature during the twentieth century. *Journal of Geophysical Research, 114*(D2). https://doi.org/10.1029/2008JD010624