Impact of Sea Ice Reduction in the Barents and Kara Seas on the Variation of the East Asian Trough in Late Winter

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ABSTRACT: Using the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) dataset and the Specified Chemistry Whole Atmosphere Community Climate Model (WACCM-SC), the impacts of sea ice reduction in the Barents–Kara Seas (BKS) on the East Asian trough (EAT) in late winter are investigated. Results from both reanalysis data and simulations show that the BKS sea ice reduction leads to a deepened EAT in late winter, especially in February, while the EAT axis tilt is not sensitive to the BKS sea ice reduction. Further analysis shows that the BKS sea ice reduction influences the EAT through the tropospheric and stratospheric pathways. For the tropospheric pathway, the results from a linearized barotropic model and Rossby wave ray tracing model reveal that long Rossby wave trains stimulated by the BKS sea ice loss propagate downstream to the North Pacific, strengthening the EAT. For the stratospheric pathway, the upward planetary waves enhanced by the BKS sea ice reduction shift the subpolar westerlies near the tropopause southward. With the critical lines displaced equatorward, the poleward transient eddies break at lower latitudes, shifting the eddy momentum deposit throughout the troposphere equatorward. Tropospheric westerlies maintained by eddy momentum deposit are also shifted southward, inducing the cyclonic anomalies over the North Pacific and deepening the EAT in late winter. Nudging experiments show that the tropospheric pathway only contributes to around 29.7% of the deepening of the EAT in February induced by the BKS sea ice loss, while the remaining 70.3% is caused by stratosphere–troposphere coupling.

KEYWORDS: Monsoons; Stratosphere-troposphere coupling; Climate variability

1. Introduction

During the recent decades, several highly populated countries in East Asia, including China, Japan, and South Korea, have suffered from a large number of casualties and great economic loss due to increasing severe winter weather, such as the 2008 cold spell in South China (e.g., Lee et al. 2010; Xie et al. 2013; Ma et al. 2013). Arctic sea ice reduction, particularly over the Barents–Kara Seas (BKS), has been considered as an important potential factor inducing the frequent, long-lasting, and strong cold air outbreak (CAO) events over the East Asian region (e.g., Wu et al. 2011; Tang et al. 2013; Gao et al. 2015; Zhang et al. 2016; Chen and Luo 2017; McKenna et al. 2018). Some studies found that the BKS sea ice loss can influence the East Asian winter climate through modulating the Arctic Oscillation/North Atlantic Oscillation (AO/NAO; e.g., Hopsch et al. 2012; Wu et al. 2015) or the “warm Arctic–cold Eurasia” teleconnection (e.g., Mori et al. 2014, 2019). In recent years, the key role of stratosphere–troposphere coupling in the East Asian cooling in late winter induced by the BKS sea ice loss has been highlighted in several studies (e.g., Kim et al. 2014; Zhang et al. 2016; Zhang et al. 2018a,b). However, some other studies found no significant cooling over East Asia in model simulations forced by the BKS sea ice reduction and argued that there is no robust causality between the BKS sea ice reduction and the East Asian cooling (e.g., Barnes and Screen 2015; McCusker et al. 2016; Blackport et al. 2019). Further, some studies argued that the frequent severe cold events over East Asia in recent years are mainly due to natural variability, rather than the BKS sea ice loss (e.g., Overland et al. 2015; Luo et al. 2019). Therefore, a question arises as to whether there is a robust mechanism linking the BKS sea ice reduction and the East Asian cooling, which is very important for clarifying the causality between them.

The East Asian winter climate is directly influenced by the East Asian winter monsoon (e.g., Kim et al. 2019; Wang et al. 2019). As an important part of the East Asian winter monsoon, the East Asian trough (EAT) is characterized by a 500-hPa low center of geopotential height over East Asia and the northwest Pacific (Zhang et al. 1997; Wang et al. 2009; Song et al. 2016). The formation of the EAT is closely related to the large-scale topography (e.g., the Tibetan Plateau; Holton 2004) and land–sea thermal contrast during the boreal winter (Nakamura et al. 2010). Many researchers have suggested that the variation of the EAT (e.g., in the intensity and the axis tilt) is one of the most important factors influencing the winter weather and climate over East Asia (e.g., Zhang et al. 1997; Cui and Sun 1999; Chen et al. 2005; Wang and Chen 2010; Wang and Feng 2011; Huang et al. 2012; Kim et al. 2019). For example, Zhang et al. (1997) and many other studies have found that the deepening of the EAT could strengthen the CAO events over East Asia (e.g., Cui and Sun 1999; Huang et al. 2012). Some other studies reported connections between the variation of
the EAT intensity and the North Pacific storm track, the precipitation over South China, and so on (e.g., Wang and Chen 2010; Wang and Feng 2011). In addition, Wang et al. (2009) reported that the variation of the EAT axis tilt could modulate the East Asian winter monsoon and affect weather conditions over East Asia.

Various climate factors can impact the EAT, including its intensity and axis tilt. Previous studies pointed out that ENSO is a major factor influencing the variation of the EAT (e.g., Zhang et al. 1996; Tao and Zhang 1998; Sakai and Kawamura 2009). Tao and Zhang (1998) found that El Niño events in winter are more likely accompanied by a weakened EAT. Sakai and Kawamura (2009) reported that the joint influence of ENSO and NAO plays a major role in modulating the EAT. In addition to the impact of ENSO events in the tropics, the EAT can also be influenced by some climate systems at middle and high latitudes (e.g., Wu and Huang 1999; Gong et al. 2001; Wu and Wang 2002a,b; Nakamura et al. 2015). Gong et al. (2001) found that the negative phase of the AO is favorable for a stronger EAT. Moreover, Wu and Wang (2002b) pointed out that the impact of the Siberian high on the EAT is different from that of the AO/NAO. Song et al. (2016) diagnosed the respective contributions of low-frequency Rossby waves and synoptic transient eddies to the intraseasonal variation of the strength of the EAT and investigated its impacts on boreal winter climate, yet the role of external forcings (e.g., ENSO events and Arctic sea ice variation) was not discussed in their study. Chen et al. (2014) found that the variation of autumn Arctic sea ice can influence the EAT and further lead to changes in the East Asian winter monsoon. However, there is no robust dynamical pathway linking the Arctic sea ice loss and the EAT in their work and further work is needed to improve our understanding on the underlying mechanisms. Also, most related studies mainly focused on the atmospheric response of the winter mean state to the BKS sea ice reduction (e.g., Mori et al. 2014; McKenna et al. 2018), while the intraseasonal evolution of the atmospheric response to the BKS sea ice loss still needs to be fully investigated. Particularly, the important role of stratosphere–troposphere coupling in the atmospheric response to the BKS sea ice loss has been gradually recognized in recent years (e.g., Kim et al. 2014; Zhang et al. 2016, 2018a,b). Therefore, it is worthwhile to clarify whether the BKS sea ice reduction in winter can exert an impact on the EAT through stratosphere–troposphere coupling (referred to as the stratospheric pathway for simplicity). Furthermore, quantifying the relative contributions of the tropospheric and stratospheric pathways will improve our understanding of the respective roles of the two pathways in the EAT variation induced by the BKS sea ice loss, as in Nakamura et al. (2016), Wu and Smith (2016), Zhang et al. (2016), and Zhang et al. (2018a).

In this study, how and to what extent the BKS sea ice reduction influences the intensity and axis tilt of the EAT are preliminarily investigated using ERA-Interim reanalysis data. Furthermore, underlying mechanisms and possible pathways by which the BKS sea ice reduction influences the EAT are diagnosed using the WACCM-SC model, a linearized barotropic model, and Rossby wave ray tracing model. This “hierarchy of model” approach improves our understanding of the mechanisms responsible for the impacts of Arctic sea ice loss on the midlatitude winter weather and climate. Moreover, the respective roles of the tropospheric and stratospheric pathways and the interaction between the two pathways are also analyzed by applying a nudging method in the model simulations. The remainder of this paper is organized as follows. Section 2 describes the data and methods. Section 3 presents the results on the relationship between the BKS sea ice reduction and EAT variations in late winter. Sections 4 and 5 discuss the tropospheric and stratospheric pathways for the BKS sea ice reduction to influence the EAT. Section 6 investigates the relative contributions of the tropospheric and stratospheric pathways to the EAT variation induced by the BKS sea ice reduction and the interaction between the two pathways. The conclusions and discussion are given in section 7.

2. Data and methodology

a. Data and numerical model

The European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) dataset from January 1979 to December 2018 is used in this study. The horizontal resolution of this dataset is 1° × 1° and there are 37 vertical levels ranging from 1000 to 1 hPa. In this study, daily mean results are derived from the 6-hourly ERA-Interim reanalysis dataset. Monthly mean sea ice concentration (SIC) data used in this study are from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST1; Rayner et al. 2003). The horizontal resolution of this dataset is 1° × 1° and this dataset covers the time period from 1870 to the present day. The surface heat flux dataset from 1919 to 2017 (including sensible heat flux and latent heat flux) is obtained from the NCEP–NCAR reanalysis and the horizontal resolution of this dataset is 1.9° × 1.9°. The BKS region in this study is outlined as the area of 20°–90°E, 65°–85°N. In this study, early winter refers to November and December and late winter refers to the following January and February.

Two numerical experiments forced by BKS high- and low-SIC boundary conditions (hereafter HICE and LICE runs for simplicity) are performed using the Specified Chemistry Whole Atmosphere Community Climate Model version 4.0 (WACCM-SC). This model is the component of the Community Earth System Model version 1.2.0, developed and maintained by the National Center for Atmospheric Research (NCAR). The horizontal resolution is 1.9° × 2.5° and there are 66 vertical levels ranging from 1000 to around 0.0006 Pa. Considering that chemical interaction is not focused on in this study, the specified chemistry version of WACCM4 with fewer chemical interactions is applied. More details on the model can be found in Smith et al. (2014). The two numerical experiments are forced by the same initial conditions and repeating seasonal cycle of boundary conditions, except for the SIC boundary condition. The BKS high- and low-SIC boundary conditions are composited from HadISST1 with respect to high- and low-SIC years, respectively, during the period 1979–2017. The corresponding SST boundary condition is applied where the SIC boundary condition is changed. The details of the selection criteria are
shown in section 2b. Both the experiments are integrated for 85 years, and the first 10 years are discarded as the model spinup.

To separate the impact of the BKS sea ice reduction on the EAT through the tropospheric and stratospheric pathways, another pair of HICE and LICE runs is performed with a nudging method applied (Labe et al. 2020) and the details are as follows: first, a control run (hereafter CTRL) driven by the climatological forcing (e.g., sea surface temperature, sea ice concentration, and chemical concentration) runs for 85 years to obtain a climatological seasonal cycle of zonal wind \( u \), meridional wind \( v \), and air temperature \( T \). The output of the CTRL run is every 3 h. The simulated climatology acts as the reference profile for the HICE/LICE nudging experiments (hereafter HICE_ndg and LICE_ndg). To constrain the nudge region in this study, namely the northern polar stratosphere (above 100 hPa, north of 66°N), the target area is “masked” in the HICE_ndg and LICE_ndg runs and the magnitude of this mask coefficient denotes the nudging strength, with 1 representing full nudge and 0 no nudge applied. In this study, the mask coefficient above 100 hPa and north of 66°N is set to 1 and gradually decreases to 0 from 100 to 200 hPa vertically and from 66° to 60°N meridionally, following a hyperbolic tangent function. The remaining area is set to 0. The area with varying coefficient acts as a transition zone to minimize the artificial effect of the abruptly changing coefficient. In the HICE_ndg and LICE_ndg runs, the meteorological fields (i.e., \( u, v \), and \( T \)) in the northern polar stratosphere are nudged toward the climatological state every 3 model hours and stratosphere–troposphere coupling is suppressed. The nudging method applied in this study has the following benefits. First, considering the zonal asymmetry of the climatological winter Arctic polar vortex (e.g., Zhang et al. 2016, 2018, 2019), this method can reproduce a more realistic state of the Arctic stratospheric polar vortex; second, nudging the model atmosphere toward the climatological state can reduce the interannual variability of the stratospheric polar vortex and help us focus on the intraseasonal variability of atmospheric response to the BKS sea ice reduction. For more details on the simulation configurations, please refer to Table 1.

### Table 1. Description for model experiment configurations.

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>SIC forcing</th>
<th>SST forcing</th>
<th>Use nudge?</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>Climatological season cycle of HadISST1 1979–2017 data</td>
<td>Climatological season cycle of HadISST1 1979–2017 data</td>
<td>No</td>
</tr>
<tr>
<td>HICE</td>
<td>Composited season cycle of BKS high-SIC years from HadISST1 1979–2017 data</td>
<td>Same as SIC forcing, but only applied where SIC changes, the rest is set to climatology</td>
<td>No</td>
</tr>
<tr>
<td>LICE</td>
<td>Composited season cycle of BKS low-SIC years from HadISST1 1979–2017 data</td>
<td>Same as SIC forcing, but only applied where SIC changes, the rest is set to climatology</td>
<td>No</td>
</tr>
<tr>
<td>HICE_ndg</td>
<td>Composited season cycle of BKS high-SIC years from HadISST1 1979–2017 data</td>
<td>Same as SIC forcing, but only applied where SIC changes, the rest is set to climatology</td>
<td>Yes</td>
</tr>
<tr>
<td>LICE_ndg</td>
<td>Composited season cycle of BKS low-SIC years from HadISST1 1979–2017 data</td>
<td>Same as SIC forcing, but only applied where SIC changes, the rest is set to climatology</td>
<td>Yes</td>
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### b. Methodology

To do the composite analysis associated with the BKS sea ice reduction, first early winter mean BKS SIC time series from 1979 to 2017 are detrended, and then one standard deviation (std) of the detrended time series is chosen as the threshold to select BKS high- and low-SIC years; that is, a year is regarded as a BKS high (low) SIC year when the BKS SIC of this year is above (below) one std of the detrended time series. Figure 1a shows the detrended time series of the early winter BKS SIC. According to this selection criterion, 7 years are selected as the BKS high-SIC years (1988, 1997, 1998, 2002, 2003, 2010, and 2014) and 3 years as the BKS low-SIC years (1984, 2012, and 2016). The composited differences of SIC, sensible heat fluxes, and latent heat fluxes in early winter between these selected BKS low- and high-SIC years are shown in Figs. 1b–d. Note that the Arctic sea ice reduction is mainly concentrated in the BKS region (Fig. 1b). With the massive sea ice loss in the BKS region, more open water areas occur and more sensible and latent heat fluxes are transported into the atmosphere above (Figs. 1c,d). In this study, a two-tailed Student’s \( t \) test is applied to assess the statistical significance of the composite analysis and a Kolmogorov–Smirnov test (Siegel 1957) is applied to assess the statistical significance of the difference in probability distribution functions (PDFs) between two samples.

The EAT is a stationary 500-hPa large-scale atmospheric circulation system over East Asia and the North Pacific (hereafter EANP) in the boreal winter. In this study, the intensity and axis tilt of the EAT are defined following the approach in Wang et al. (2009); that is, an empirical orthogonal function (EOF) analysis is first performed on the 500-hPa geopotential height field in the region of 25°–50°N and 100°E–180°, and then the intensity of the EAT and the tilt of the EAT axis are inferred from the first two leading modes of EOF analysis. The first principal component (PC1) is used as an index for the EAT intensity and the larger the PC1 index is, the stronger the EAT is, and vice versa. The second principal component (PC2) is used to infer the tilt of the EAT axis and the larger the PC2 index is, the more northeast–southwest the EAT axis tilts. In our study, EOF analysis is applied to the...
February 500-hPa geopotential height field of ERA-Interim re-analysis from 1979 to 2017 and the corresponding geopotential height field from the CTRL simulation and the results are shown in Fig. 2. The geopotential potential height fields are normalized before the EOF analysis to avoid the possible influence of remote large variance in the target area. PC1 and PC2 from ERA-Interim reanalysis explain around 46.5% and 22.9%, respectively, of the total variance of the 500-hPa geopotential height field over the EANP region, and 47.8% and 18.7% for PC1 and PC2 of the 500-hPa geopotential height field from the CTRL simulation, close to the results from Wang et al. (2009). It should be pointed out that PC1 and PC2 from ERA-Interim reanalysis and WACCM-SC simulations all pass the criterion of North et al. (1982), implying that all these principal components are well separated from other components. In the criterion of North et al. (1982), two principal components are considered as significantly separated if the sampling error of one principal component is comparable to the distance to the other principal component.
The obtained spatial patterns of PC1 and PC2 in February derived from ERA-Interim reanalysis are regressed upon 1980–2017 daily 500-hPa geopotential height field in February. Before the regression, the daily 500-hPa geopotential height fields are detrended and normalized, with the seasonal cycle also removed to exclude the autocorrelation in daily data. PC1 and PC2 derived from the CTRL simulation are also regressed upon the simulated February daily 500-hPa geopotential height field for the HICE, LICE, HICE_ndg, and LICE_ndg simulations. The simulation data are processed in the same way as ERA-Interim reanalysis data before the EOF analysis.

Atmospheric waves can influence the geopotential height of quasi-stationary flow through horizontal divergence/convergence of eddy vorticity and eddy heat fluxes. According to Lau and Holopainen (1984), the wave-induced geopotential height tendency can be written as

$$\left[ \frac{g}{f} \nabla^2 + f g \frac{\partial}{\partial \phi} \left( \frac{1}{\sigma} \frac{\partial}{\partial \phi} \right) \right] \frac{\partial Z}{\partial t} = D,$$

where \(Z\) is geopotential height, \(g\) is gravitational acceleration, \(f\) is the Coriolis parameter, and \(\sigma\) is the static stability. The \(D\) term on the right-hand side of Eq. (1) is the eddy forcing term and can be decomposed into two parts \(D^{\text{HEAT}}\) and \(D^{\text{VORT}}\), which arise from eddy heat flux and eddy vorticity flux, respectively, as follows:

$$D = D^{\text{HEAT}} + D^{\text{VORT}} = f \frac{\partial}{\partial \phi} \left( \frac{\nabla \cdot \nabla \phi}{\nabla} \right) + (-\nabla \cdot \nabla \zeta),$$

where \(\theta\) is potential temperature, the overbars denote time average, and the primes denote the deviation from the corresponding time-averaged quantity; \(\nabla\) is the hemispheric average of \(-\partial T/\partial \phi\) and \(\zeta\) is the perturbation relative vorticity. In this study, Eq. (2) is applied to diagnose the wave-induced geopotential height tendency over the trough region of the EAT (30°–50°N, 110°–170°E; Song et al. 2016).

To clarify the potential role of Rossby waves in the connection between the geopotential height anomalies over the BKS region and the EANP region, a linearized barotropic model is used in this study (Shaman and Tziperman 2007). This model uses 500-hPa streamfunction climatology in February from the CTRL simulation as the background and an initial relative vorticity forcing is set over the BKS region (i.e., 65°–85°N, 20°–90°E). The linearized barotropic vorticity equation is solved using spherical harmonics and triangular 24 truncation (T24). For more details on the solution process, please refer to Branstator (1983). To analyze the propagation of Rossby waves, a Rossby wave ray tracing model is used to obtain the wave propagation paths. The background flow field is obtained from the climatology of 500-hPa wind field in February from the CTRL simulation. For more details, please refer to Karoly (1983) and Zhang et al. (2015).

It is known that transient eddies induced by atmospheric baroclinicity propagate toward the high latitudes and dissipate at their critical lines (Randel and Held 1991). The westerly eddy momentum fluxes are transported meridionally by these transient eddies and hence influence the structure of the tropospheric zonal wind. Previous studies showed that the meridional transport of eddy momentum fluxes is mostly pronounced in the upper troposphere because typical baroclinic eddies can only reach the tropopause (e.g., Ait-Chaalal and Schneider 2015). To explain the tropospheric westerly anomalies induced by BKS sea ice reduction, the phase speed spectra of eddy momentum flux in February are analyzed. The phase speed spectra analysis is conducted as follows: first, calculate the space–time cospectra of February time series of zonal and meridional wind fields from ERA-Interim reanalysis and the WACCM-SC simulations using the method of Hayashi (1971); then transform the obtained (frequency, wavenumber) spectra into (angular phase speed, wave-number) spectra using the method of Randel and Held (1991); finally, sum the spectra over wavenumber and obtain the phase speed spectra as a function of latitude and angular phase speed. For more details on the phase speed spectrum analysis, please refer to Randel and Held (1991).
3. The connection between BKS sea ice reduction and EAT variation

Figures 3a–c show the geopotential height differences between BKS low- and high-SIC years as well as the climatological geopotential height fields from December to the following February during 1979–2017 derived from ERA-Interim reanalysis data. The climatological 500-hPa geopotential height fields from December to the following February during 1980–2017 are also shown (deviations from zonal mean; contour lines with solid and dashed lines indicating positive and negative values, respectively; gpm). Figures 3d–f show the corresponding geopotential height differences between the LICE and HICE simulations. The simulated 500-hPa climatological geopotential height field from December to the following February in the CTRL run is similar to that derived from ERA-Interim reanalysis data. Furthermore, the simulated geopotential height anomalies induced by the BKS sea ice reduction are overall in good agreement with the results derived from ERA-Interim reanalysis data. In the results of model simulations, there exist weak geopotential height anomalies over the EANP region without statistical significance in December (Fig. 3d) from the WACCM-SC simulations, and these anomalies grow stronger and statistically significant in the following January (Fig. 3e) and peak in February (Fig. 3f). It can be concluded from the results of both ERA-Interim reanalysis data and the WACCM-SC simulations that the BKS sea ice reduction can induce the negative geopotential height anomalies over the EANP region during winter, and these anomalies grow stronger in the following January and February. These negative geopotential height differences over the EANP region indicate the strengthening of the EAT. We will discuss this progressively developing
atmospheric response to the BKS sea ice reduction later in this work. In the following, the variations of the EAT in February induced by BKS sea ice reduction will be discussed in detail.

Figure 4 shows the PDFs of February EAT daily intensity and axis tilt indices in BKS high- and low-SIC scenarios derived from ERA-Interim reanalysis data and the WACCM-SC simulations. The results derived from ERA-Interim reanalysis data indicate that the PDF of the EAT intensity in low-SIC years is skewed toward higher values compared with that in high-SIC years (Fig. 4a). The differences in the EAT intensity between BKS high- and low-SIC years are significant at a 99.5% confidence level under the Kolmogorov–Smirnov test. The results from the WACCM-SC simulations also indicate that the EAT tends to be stronger in the BKS low-SIC scenario than that in the high-SIC scenario and this result is statistically significant at a 95% confidence level under the Kolmogorov–Smirnov test (Fig. 4c). These results suggest that the BKS sea ice reduction can significantly strengthen the intensity of the EAT in February. For the EAT axis tilt, the results from both ERA-Interim reanalysis and WACCM-SC simulations show no obvious skew in the PDFs between the BKS low- and high-SIC scenarios, though the differences between these PDFs are statistically significant at a 99% and 90% confidence level under the Kolmogorov–Smirnov test for the reanalysis data and model simulations, respectively (Figs. 4b,d). Considering the insignificant differences of the EAT axis tilt index between
BKS high- and low-SIC scenarios from both ERA-Interim reanalysis data and WACCM-SC model simulations, the following discussion will focus mainly on dynamical mechanisms by which the BKS sea ice reduction impacts on the EAT intensity in the following February.

4. Tropospheric linkage between the BKS sea ice reduction and the EAT

Figure 5 shows the composited 500-hPa wind and geopotential height differences in February between the BKS low- and high-SIC scenarios derived from ERA-Interim reanalysis data and WACCM-SC simulations. Note that there is an anticyclonic anomaly located over the region of 60°–75°N, 60°W–60°E associated with the BKS sea ice reduction, consistent with the positive center of the statistically significant geopotential height anomalies (Fig. 5a). It is worth noting that this anticyclonic anomaly is much larger than the BKS region and also extends to the Eurasian continent due to the development of the atmospheric response to the sea ice anomalies, as mentioned in some previous studies (e.g., Ambaum and Hoskins 2002; Deser et al. 2007; Ruggieri et al. 2016, 2017). There also exists a cyclonic flow (negative geopotential height) anomaly over the EANP region and an anticyclonic (positive geopotential height) anomaly over the east Pacific. This spatial pattern of wind anomalies shows a wavelike pattern from the BKS region to the east Pacific, implying that wave trains could originate from the BKS region and propagate toward the EANP region and hence strengthen the EAT. The results derived from the WACCM-SC simulations are overall in accord with those from ERA-Interim reanalysis data (Fig. 5b). However, the cyclonic wind anomalies downstream of the BKS region are located a little more eastward and the anticyclonic anomalies are relatively weaker in the results derived from the WACCM-SC simulations compared with those from ERA-Interim reanalysis data.

A linearized barotropic model is applied to diagnose whether the local positive geopotential height anomalies induced by the BKS sea ice reduction could stimulate the above-mentioned wave train and further lead to the negative geopotential height anomalies over the EANP region. To make the magnitude of initial forcing for the linearized barotropic model representative, the difference in relative vorticity over the BKS region in February between the LICE and HICE runs is used (i.e., $-1.38 \times 10^{-6}$ s$^{-1}$). This initial forcing is added to the linearized barotropic model and the climatological streamfunction at 500 hPa in February from the CTRL simulation is used as the background field. The steady-state solution to the barotropic vorticity equation for this initial forcing is shown in Fig. 6. Note that there is a positive vorticity response...
downstream of the initial negative vorticity forcing over the BKS region. This positive vorticity response peaks from the Mongolian Plateau to northeastern China and extends eastward to the North Pacific, with a negative vorticity response also over the east Pacific. The spatial pattern of the vorticity response in Fig. 6 is similar to the spatial pattern of the 500-hPa wind anomalies induced by the BKS sea ice reduction (Fig. 5).

The solution of the linearized barotropic vorticity equation implies that the wave trains stimulated by the BKS sea ice reduction propagate southeastward to the EANP region and induce the negative geopotential height anomalies over this region, as shown in Fig. 3. It should be pointed out that the steady-state solution is not sensitive to small changes in the magnitude and size of the initial perturbation. Song et al. (2016) and some other studies have discussed the key role of the Rossby wave trains in the upstream region of the EANP in the deepening of the EAT (Joung and Hitchman 1982; Park et al. 2014). Here, the contribution of the Rossby wave trains to the negative geopotential height tendency over the trough region of the EAT in February is diagnosed using wave trains to the negative geopotential height tendency over the BKS region. This positive vorticity response peaks from the Mongolian Plateau to northeastern China and extends eastward to the North Pacific, with a negative vorticity response also over the east Pacific. The spatial pattern of the vorticity response in Fig. 6 is similar to the spatial pattern of the 500-hPa wind anomalies induced by the BKS sea ice reduction (Fig. 5). The solution of the linearized barotropic vorticity equation implies that the wave trains stimulated by the BKS sea ice reduction propagate southeastward to the EANP region and induce the negative geopotential height anomalies over this region, as shown in Fig. 3. It should be pointed out that the steady-state solution is not sensitive to small changes in the magnitude and size of the initial perturbation. Song et al. (2016) and some other studies have discussed the key role of the Rossby wave trains in the upstream region of the EANP in the deepening of the EAT (Joung and Hitchman 1982; Park et al. 2014). Here, the contribution of the Rossby wave trains to the negative geopotential height tendency over the trough region of the EAT in February is diagnosed using the method of Lau and Holopainen (1984). Equation (2) (see section 2b) is applied to diagnose the wave-induced geopotential height tendency over the trough region of the EAT. The results show that in the reanalysis data, the difference of eddy forcing term $D$ over the trough region between the low- and high-BKS SIC years is $2.37 \times 10^{-8}$ s$^{-1}$ located at 65°–85°N, 20°–90°E (black box), which is the 500-hPa relative vorticity differences over the BKS region in February between the LICE and HICE runs.

5. Stratospheric linkage between the BKS sea ice reduction and the EAT

The previous section has investigated the tropospheric pathway for the BKS sea ice reduction to influence the EAT. In addition, some other studies reported that stratosphere–troposphere coupling is important for the lead–lag linkage between the BKS sea ice reduction in early winter and the midlatitude atmospheric circulation anomalies in late winter (e.g., Kim et al. 2014; Nakamura et al. 2016; Zhang et al. 2016, 2018a,b). This significant stratospheric pathway for the BKS sea ice loss in early winter to influence the EAT in late winter is investigated.

Figure 8 shows time–height cross sections of the midlatitude geopotential height and vertical component of wave flux differences between BKS low- and high-SIC scenarios derived from both ERA-Interim reanalysis and WACCM-SC simulations. Although some differences exist between Figs. 8a and 8b, both the reanalysis data and model simulations show positive geopotential height anomalies in the lower stratosphere in January due to the enhanced upward propagating planetary waves induced by the BKS sea ice reduction. These positive geopotential height anomalies persist in the stratosphere for around 2 months and then extend downward to the surface in February, consistent with some previous studies (Kim et al. 2014; Zhang et al. 2016, 2018b). Although the time for the positive anomalies to extend to the surface in WACCM-SC simulation is a bit later than that in the reanalysis data, the whole process of “stratosphere–troposphere–stratosphere” linkage can be found in both ERA-Interim reanalysis and WACCM-SC simulations, implying a stratospheric pathway for the BKS sea ice reduction in early winter to influence the EAT in late winter.

With more planetary waves induced by the BKS sea ice reduction propagating into and breaking in the stratosphere (Fig. 8), the subpolar westerlies in the UTLS region are decelerated due to wave–mean flow interactions (Holton 2004), as shown in Figs. 9b and 9f (blue/red lines). Linear theory...
indicates that transient eddies can hardly propagate meridionally beyond their critical lines in which $\bar{u} = c$, where $\bar{u}$ denotes zonal mean zonal wind and $c$ denotes the phase speed of atmospheric waves (Randel and Held 1991; Holton 2004). Figures 9b and 9f show the latitude–phase speed cross sections of the eddy momentum flux differences at 200 hPa between BKS low- and high-SIC scenarios. In the reanalysis data, with the subpolar zonal wind decelerated in the UTLS region, poleward eddy momentum fluxes in the UTLS region at 60°N are decreased, while the fluxes between 15° and 45°N are increased, suggesting that the transient eddies in the UTLS region break at lower latitudes when the BKS sea ice is reduced (Fig. 9b). The corresponding results derived from the WACCM-SC simulations (Fig. 9f) are overall consistent with that from ERA-Interim reanalysis data, except that the divergence and convergence anomalous centers slightly shifted southward. This dipolar structure of eddy momentum flux anomalies further favors a southward shift of the westerly jet stream in the UTLS region (Limpasuvan and Hartmann 2000).

According to the mechanism of interaction between two counterpropagating Rossby waves, the phase speed of transient eddies in the troposphere is decelerated in response to the slower waves in the UTLS region (Chen and Held 2007; Wittman et al. 2007). Figures 9d and 9h show the differences in phase speed spectra of the eddy momentum flux at 500 hPa between BKS low- and high-SIC scenarios. Note that there also exists a dipolar pattern of eddy momentum flux anomalies at 500 hPa, similar to that in the UTLS region (Figs. 9b,f). In addition, it is worth noting that the phase speed of the negative anomalies is larger than that of positive anomalies in both observation and model simulations, also found in the UTLS region (Figs. 9b,f). This indicates the deceleration of transient eddies in troposphere in response to the slower waves in the UTLS region. As a result, tropospheric transient eddies also break at lower latitudes, according to Chen and Held (2007). That is, the deceleration of the phase speed of the tropospheric eddies could shift the eddy momentum flux equatorward.

Figure 10 shows the differences of the zonal wind at 500 hPa in February between the BKS low- and high-SIC scenarios derived from ERA-Interim reanalysis data and the WACCM-SC simulations. It is worth noting that there are negative anomalies at 55°N and positive anomalies at 35°N, indicating the southward

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**Fig. 7.** Time evolution of Rossby ray paths (yellow curves) on days 1, 3, 5, 7, 9, and 11, with wavenumber 1 along the rays, with ray sources located south of BKS region (black box) and geopotential height differences (color-filled contours; gpm) between BKS low- and high-SIC scenarios on 500 hPa in February, derived from WACCM-SC simulations. The values over dotted regions are statistically significant at the 95% confidence level according to a two-tailed Student’s $t$ test.
shift of tropospheric westerlies. This is caused by the equatorward shift of poleward tropospheric eddy momentum fluxes in response to the decelerated transient eddies in the UTLS region as mentioned above. With the tropospheric westerlies shifted equatorward, a cyclonic pattern of the zonal wind anomalies is formed, leading to negative geopotential height anomalies over the EANP region, further enhancing the intensity of the EAT.

6. Relative contributions of the two pathways to EAT strengthening in February induced by the BKS sea ice loss

The above-mentioned results have confirmed that both tropospheric and stratospheric processes induced by the BKS sea ice reduction can influence the EAT intensity. In this study, the nudging experiments (see section 2a) are used to separate the contributions of the tropospheric and stratospheric pathways to the strengthening of the EAT by the BKS sea ice loss. Figure 11 shows the 500-hPa geopotential height anomalies induced by the BKS sea ice reduction through the tropospheric and stratospheric pathways, respectively. Note that there are negative geopotential height anomalies over the EANP region through the tropospheric pathway during the whole winter (Figs. 11a–c), which could be due to the persistent low sea ice condition in the BKS region as is shown in Figs. 1b and 12. Note that the geopotential height anomalies induced by tropospheric processes show no statistical significance at 95% confidence level. The absence of statistical significance could be due to the low signal-to-noise ratio caused by the atmospheric internal variability in the model simulation, mentioned by many other studies (e.g., Screen et al. 2014; Mori et al. 2019; Warner et al. 2020). In addition, the negative geopotential height anomalies over the EANP through the stratospheric pathway first occur in January and grow stronger in February, consistent with many previous studies (e.g., Kim et al. 2014; Zhang et al. 2016, 2018a,b). These negative geopotential height anomalies constructively interfere with the anomalies associated with the tropospheric pathway, leading to a larger amplitude of atmospheric response over the EANP region in February to the BKS sea ice reduction (Figs. 3c,f). This could explain why the strengthening of the EAT induced by the BKS sea ice reduction peaks in February.
The tropospheric and stratospheric pathways described in sections 4 and 5 are dynamically linked with each other. Geopotential height response to the BKS sea ice reduction through the tropospheric pathway is favorable for the enhancement of upward propagating wave fluxes (Fig. 8), which is the key part of the stratospheric pathway, as is described in section 5. So the tropospheric pathway is an important prerequisite for the stratospheric pathway. However, the BKS sea ice reduction

Fig. 9. Composited eddy momentum fluxes (color-filled contours; m² s⁻²) in (a) BKS low-SIC scenario and (b) differences between BKS low- and high-SIC scenarios at 200 hPa in February as a function of latitude and angular phase speed derived from ERA-Interim reanalysis data. (c), (d) As in (a) and (b), but at 500 hPa. (e)–(h) As in (a)–(d), but for the WACCM-SC simulations results. The zonal mean zonal wind divided by cosine value of latitude in BKS low-SIC scenario (blue line) and high-SIC scenario (red line) is also shown. The x axis denotes angular phase speed, and the y axis denotes latitude.

Fig. 10. Composited 500-hPa zonal wind differences between BKS low- and high-SIC scenarios (color-filled contours; m s⁻¹) in February derived from (a) ERA-Interim reanalysis dataset and (b) WACCM-SC simulations. The climatological 500-hPa zonal wind fields in February are also shown (solid contour lines; m s⁻¹). The values over dotted regions are statistically significant at the 95% confidence level according to a two-tailed Student’s t test.
can influence the EAT through the tropospheric pathway regardless of the stratospheric pathway, as is shown in Figs. 11a–c. To quantify the relative contributions of the processes associated with the tropospheric and stratospheric pathways to the strengthening of the EAT in February, 500-hPa February geopotential height fields in the region of 25°–50°N, 100°E–180° from HICE, LICE, HICE_ndg, and LICE_ndg runs are projected onto PC1 derived from the CTRL run. Considering

**Fig. 11.** Geopotential height differences between BKS low- and high-SIC scenarios (color-filled contours; gpm) at 500 hPa through the tropospheric pathway in (a) December, (b) January, and (c) February, derived from the WACCM-SC simulations. The climatological 500-hPa geopotential height field (CTRL) is also shown (deviations from zonal mean; contour lines with solid and dashed lines indicating positive and negative values, respectively; gpm). (d)–(f) As in (a)–(c), but through the stratospheric pathway. The values over dotted regions are statistically significant at the 95% confidence level according to a two-tailed Student’s t test.

**Fig. 12.** Composited SIC (%) anomalies in (a) January and (b) February between BKS low- and high-SIC scenarios. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student’s t test.
the approximately linear feature between the impacts of tropospheric and stratospheric processes (Wu and Smith 2016; Zhang et al. 2018a), the relative contribution of the tropospheric pathway is calculated based on the following equation:

$$R_{tropo} = \frac{TSLICEnudge - TSHICEnudge}{TSLICE - TSHICE} \times 100\%, \quad (3)$$

where TS is the projected time series of PC1 and the overbar denotes time mean. The result shows that the influence of the BKS sea ice reduction through the tropospheric pathway contributes around 29.7% to the strengthening of the EAT induced by the BKS sea ice loss. Hence, 70.3% of the strengthening of the EAT induced by the BKS sea ice loss is mainly caused by stratosphere–troposphere coupling. The quantitative results suggest that the tropospheric pathway only contributes to a small part of the strengthening of the EAT in February induced by the BKS sea ice loss, although it is an important prerequisite for the influence of the BKS sea ice reduction through the stratospheric pathway. A large part of the strengthening of the EAT induced by the BKS sea ice reduction is actually contributed by stratosphere–troposphere coupling, consistent with the results of Zhang et al. (2016), Wu and Smith (2016), and Zhang et al. (2018b).

7. Conclusions and discussion

Using ERA-Interim reanalysis data and WACCM-SC simulations, this study investigates the potential impacts of the BKS sea ice reduction on the EAT in late winter. The results show that the BKS sea ice reduction can induce the negative geopotential height anomalies over the EANP region during winter, especially in February, indicating the deepening of the EAT. In addition, the tilt of the EAT axis shows no significant variation associated with the BKS sea ice reduction both in ERA-Interim reanalysis data and the WACCM-SC simulations.

Song et al. (2016) found that the upstream Rossby wave trains play a key role in the intraseasonal variation of the EAT intensity. Based on their study, we investigated whether the BKS sea ice reduction can stimulate Rossby wave trains and further influence the strength of the EAT. Using a linearized barotropic model, the atmospheric response to the positive geopotential height anomalies over the BKS region is diagnosed and the results show a wavelike pattern with negative relative vorticity over the BKS region and positive over the EANP region. To further investigate the role of atmospheric waves in the linkage between the BKS sea ice reduction and the deepening of the EAT, the eddy forcing term over the EANP region is diagnosed, according to Lau and Holopainen (1984). Both the reanalysis data and model simulations show that the negative geopotential height tendency over the trough region of the EAT is mostly contributed by long Rossby waves, which is supported by the Rossby wave ray tracing model.

In addition, previous studies found that stratosphere–troposphere coupling plays an important role in the potential linkage between the BKS sea ice reduction and the midlatitude winter climate (e.g., Kim et al. 2014; Zhang et al. 2016, 2018a,b). The deepened EAT induced by the BKS sea ice reduction through the tropospheric pathway is favorable for the enhanced upward propagating planetary waves. With more planetary waves propagating into the stratosphere, the circumpolar westerlies in the UTLS region are decelerated through wave–mean flow interaction. Due to the southward shift of the critical lines, the transient eddies in the UTLS region break at lower latitudes, leading to the equatorward shift of eddy momentum deposit. The tropospheric eddies are also decelerated and break at lower latitudes in response to the slower waves in the UTLS region, further shifting the tropospheric eddy momentum deposit southward. So the tropospheric westerlies maintained by these momentum fluxes are shifted southward, leading to the cyclonic anomalies over the EANP region and further enhancing the EAT intensity. Figure 13 depicts the dynamical mechanisms...
linking the BKS sea ice reduction and the deepened EAT through the tropospheric and stratospheric pathways described in this study.

To quantify the relative contributions of the tropospheric and stratospheric pathways in the linkage between the BKS sea ice reduction and the deepening of the EAT, another two nudging experiments with stratosphere–troposphere coupling suppressed are performed. The results show that with stratosphere–troposphere coupling suppressed, the tropospheric pathway only contributes around 29.7% of the strengthening of the EAT in February induced by the BKS sea ice loss, while the remaining part is mainly due to stratosphere–troposphere coupling. This indicates that although the tropospheric pathway is an essential prerequisite for the stratospheric pathway, stratosphere–troposphere coupling plays a more important role in the linkage between the BKS sea ice reduction and the strengthening of the EAT in late winter. Our results further support the important role of stratospheric processes in the linkage between the Arctic sea ice loss and midlatitude weather and climate pointed out in some studies (Nakamura et al. 2016; Zhang et al. 2016, 2018a).

Although many previous studies have discussed how some climate factors (e.g., ENSO events, AO/NAO, and the Siberian high) influence the EAT (e.g., Gong et al. 2001; Wu and Wang 2002b; Sakai and Kawamura 2009), little is known regarding the potential linkage between the BKS sea ice reduction and the wintertime EAT and underlying dynamical mechanisms. The details of the intraseasonal atmospheric response to the BKS sea ice loss also need a thorough investigation and this study is trying to narrow these knowledge gaps. In addition, a significant East Asian cooling in late winter induced by the BKS sea ice loss can be found in our simulations, as shown in Fig. S3b. So it can be supposed that the response of the EAT to the BKS sea ice loss could play an important role in the linkage between the BKS sea ice reduction and the frequent East Asian cold winters. Furthermore, the inconsistency in the simulated East Asian cooling associated with the BKS sea ice loss in this study with that in some previous studies (e.g., McCusker et al. 2016; Ogawa et al. 2018; Blackport et al. 2019; Screen and Blackport 2019) can be due to the state of model atmosphere/ocean/land (e.g., Peings and Magnusdottir 2014; Osborne et al. 2017; Labe et al. 2019; Nakamura et al. 2019; Warner et al. 2020) and the geographic pattern of sea ice (Screen et al. 2018), among other things. Also, atmospheric internal variability is found to cause uncertainties in the Arctic–midlatitude connection in the model simulations (e.g., Screen et al. 2014; Mori et al. 2019; Warner et al. 2020). The above-mentioned factors call for multimodel experiments with large ensembles to minimize the influence of natural variability.

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