Arctic and East Asia Winter Climate Variations Associated with the Eastern Atlantic Pattern

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ABSTRACT

The wintertime Arctic temperature (T ; surface–400 hPa) decreased from 1979 to 1997 and increased rapidly from 1998 to 2012, in contrast to the global mean surface air temperature. Here aspects of circulation variability that are associated with these temperature changes are examined using the NCEP–NCAR reanalysis and ERA-Interim products. It is found that the Nordic–Siberia seesaw of meridional winds near 70°N is associated with two-thirds of the variance of the Arctic winter mean T , possibly contributing to the cooling and warming trends. It is suggested here that the seesaw accounts for much of the difference in Arctic amplification between observations and climate models. Growth of sea ice in winter is hindered by southerly winds over the Nordic region (0°–60°E). Through modulation of the wind seesaw, the eastern Atlantic (EA) pattern is found to be significantly associated with Arctic and East Asia winter climate variations. In one phase of the EA pattern, a midlatitude North Atlantic ridge anomaly is associated with a poleward shift of the mean storm track, a weakened eddy-driven jet over Eurasia, and above-normal sea level pressure (SLP) over Siberia, most significantly in the region to the northwest of Lake Baikal. The EA pattern is associated with two-thirds of the variance of winter-average SLP over Siberia.

1. Introduction

The global mean surface air temperature (SAT) increased between 1979 and 1997, followed by a hiatus (Foster and Rahmstorf 2011; Kosaka and Xie 2013). By contrast, during the months of December–February (DJF), the tropospheric temperature (T) and humidity (q) in the Arctic decreased in the first period and increased rapidly between 1998 and 2012 (see Fig. S1 in the supplemental material; Ding et al. 2014; Jun et al. 2016). Alternatively, the Arctic winter may have shifted from a

cool period during 1979–2003 to a warm period during 2004–12 (Feng and Wu 2015). Large inter-winter variations of tropospheric T and q have also been observed in the polar region (Fig. S1). It has been suggested that global sea surface temperature (SST) patterns are responsible for the warming in the Arctic troposphere and the decrease of sea ice cover is mainly responsible for the warming near the surface (Screen et al. 2012; Perlwitz et al. 2015). The Arctic climate is more sensitive to North Atlantic than Pacific SST variations (Deser 2000). A recent study suggests a possible role of the Pacific Ocean decadal oscillation in regulating wintertime climate in the Arctic (Screen and Francis 2016). However, the dynamic processes linking the global SST and the Arctic climate remain unclear. Locally as warm Atlantic water flows through the Barents Sea, it loses heat to the Arctic atmosphere. Warm Arctic periods are associated with high northward ocean heat transport and reduced Arctic sea ice cover (Smedsrud et al. 2013; Onarheim et al. 2015; Delworth et al. 2016).

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It has also been suggested that the decrease of Arctic sea ice in the last few decades may have caused more frequent cold winters in Eurasia in recent years; however, other studies have suggested that their relationships remain uncertain (Honda et al. 2009; Screen et al. 2013; Tang et al. 2013; Wu et al. 2013; Cohen et al. 2014; Kim et al. 2014; Mori et al. 2014; Kug et al. 2015; Li et al. 2015; Nakamura et al. 2015; Overland et al. 2015; Semenov and Latif 2015; Wu et al. 2015; Sun et al. 2016). Cold-air outbreaks in the midlatitudes have been found to be far more frequent in the negative phase of the North Atlantic Oscillation (NAO) than in the positive phase (Thompson and Wallace 2001). On time scales shorter than a month, atmospheric waves emanating from the North Atlantic propagate eastward crossing Eurasia at various speeds (Joung and Hitchman 1982; Blackmon et al. 1984). On longer time scales, meteorology in Eurasia responds to perturbations over the Atlantic in well-defined geographical patterns (Wallace and Gutzler 1981), which correspond to four weather regimes: the positive and negative phases of NAO, the Atlantic ridge, and the Scandinavian blocking (Cassou 2008).

In particular, the eastern Atlantic (EA) pattern of geopotential height at 500 hPa (z_{500}) is characterized (in its negative phase) by a ridge near 50°N between Newfoundland and Ireland, followed by a trough centered over the Black Sea and a second ridge over Siberia (Wallace and Gutzler 1981). The amplified pressure ridge (EA−) is accompanied by a storm track with a southwest–northeast orientation, with much of the eddy activity being confined upstream and poleward of the ridge, while a broad Atlantic trough (EA+) corresponds to a more zonally oriented storm track (Lau 1988). The eastward extension of the storm track axis to western and central Siberia would result in a low pressure anomaly in the region. These previous results imply that the influence of Arctic sea ice cover on Eurasia must be investigated in connection with the North Atlantic weather regimes and the teleconnection patterns, and that the relation between the Atlantic weather regimes and cold air outbreaks in East Asia is time scale dependent.

Numerous studies have shown the influence of the Siberian high (SiHi) on the winter monsoon in East Asia (e.g., Ding 1990; Cohen and Entekhabi 1999). Winter temperature over East Asia may be influenced by remote forcing and internal variability through large-scale teleconnections that impact the SiHi intensity (e.g., Yang et al. 2002; Cheung et al. 2012; Lim and Kim 2013; Kim et al. 2014; Smoliak and Wallace 2015). Interannual variability and long-term trends of the SiHi intensity have been reported (Panagiotopoulos et al. 2005; Jeong

et al. 2011; Zhang et al. 2012), while their causes remain unclear. The variation of SiHi in winter may be related to snow cover in the area in autumn (Cohen et al. 2001, 2012) and to sea ice cover in the Barents Sea and the Kara Sea (Inoue et al. 2012; Mori et al. 2014). It remains to be determined to what extent the SiHi intensity varies in association with snow cover, Arctic sea ice, and atmospheric teleconnections.

Here, we analyze ERA-Interim (Dee et al. 2011) and NCEP–NCAR reanalysis (Kalnay et al. 1996) products to investigate Arctic winter climate variability and associated changes in East Asia. The objectives of this study are to assess how much natural–internal variability has contributed to climate changes in these regions from 1979 to 2012 and to understand what role teleconnections play in the Siberian SLP variations on the seasonal (DJF) time scale. The data and analysis methods are described in section 2. Results of correlation analysis for time series are presented in section 3. Teleconnections based on empirical orthogonal function (EOF) analysis of SLP are presented in section 4. Conclusions are summarized in section 5.

2. Data and analyses

This study uses monthly mean atmospheric temperature (T), specific humidity (q), zonal wind (u), meridional wind (v), and sea level pressure (SLP) for the period of 1979–2012 from the NCEP–NCAR reanalysis (Kalnay et al. 1996) and the ERA-Interim (hereafter ERAI; Dee et al. 2011). Values for T and q are averaged, weighted by grid size and air density, from 400 to 700 hPa (i.e., middle troposphere) and from 700 to 1000 hPa (lower troposphere), respectively, over the polar cap (north of 70°N) for the months of December–February. High-frequency data are also acquired from the NCEP–NCAR reanalysis. Time fraction (F_{hv}) of high vorticity ($\zeta > 10^{-5} \text{ s}^{-1}$) is derived from 6-hourly local vorticity data, and synoptic variance of v is calculated using a bandpass filter (2.5–6 days), both of which are used to assess transient eddy activities (Lau 1988; Sorteberg and Walsh 2008). Values of F_{hv} and v are averaged for the same vertical layers as for T and time periods for twelve 30° polar wind sectors (Fig. S2 in the supplemental material; index 1–12 for longitude from 0° to 360°E) at 66° and 72°N.

Two sets of sea ice concentration data are used in this study: 1) assimilated data from the Hadley Centre Sea Ice and SST dataset (HadISST; Rayner et al. 2003) and 2) merged satellite data from the NOAA National Snow and Ice Data Center and NASA Goddard Space Flight Center (Meier et al. 2013; Peng et al. 2013). Monthly sea ice concentrations are averaged for the polar cap

(>70°N). The North Atlantic Oscillation index is available online at <http://www.esrl.noaa.gov/psd/data/climateindices/list/>. Correlation coefficients are calculated based on the mean meteorological parameters and sea ice concentrations, and will be shown only for meteorological parameters of the lower troposphere.

Based on previous analysis of the teleconnection patterns (Wallace and Gutzler 1981; Deser 2000), it is hypothesized that atmospheric perturbations over the North Atlantic are driving much of the climate changes in the Arctic and Eurasia in winter. To test this hypothesis, EOF analyses are done for winter mean SLP for the area of 30°–88°N, 60°W–150°E (from the North Atlantic to East Asia). The temporal variations of SLP according to the EOF patterns, or principal components (PCs), are calculated for the period 1979–2012. The NCAR Command Language (<http://www.ncl.ucar.edu>) is used for the EOF analysis of sea level pressure.

3. Arctic climate variations associated with the Nordic–Siberian seesaw of meridional winds

The advection of the climatological-mean temperature field by the anomalous wind field plays an important role in forcing the temperature anomalies associated with major teleconnections (Smoliak and Wallace 2015). The correlation coefficients (r) are calculated between Arctic T (or q) and mean v for twelve 30° polar wind sectors and are shown in Figs. 1a and 1b. Significant ($p < 0.01$) positive correlations are found with v in Nordic regions (0°–60°E), and significant negative correlations are found with v in northern Siberia (90°–150°E). The correlations are not significant in other sectors. Similar results are obtained using ERAI and NCEP–NCAR reanalysis products. Winds vary from year to year in all sectors (standard deviations of the mean v at 72°N ranging from 1.4–2.6 and 0.8–2.1 m s^{-1} in the middle and lower troposphere, respectively). However, the impact on the mean state is largest due to changes in the Nordic and Siberian sectors where the warmest and coldest air masses originate, respectively (see Fig. S3 in the supplemental material). Figure S4 in the supplemental material shows similar patterns of correlation obtained using T , q , and v fields from a 1000-yr segment of a preindustrial control simulation using a Geophysical Fluid Dynamics Laboratory climate model (GFDL CM2.1; Delworth et al. 2006), suggesting that the Nordic–Siberian dipole influence is intrinsic to the Arctic climate system (from interannual to decadal and century time scales).

We also calculated r between T (or q) and F_{hv} for the same 12 sectors (Fig. 1c). The largest negative correlations are found for sectors between 60° and 120°E, which

FIG. 1. Correlation coefficients between pairs of interannual variations: v – q (blue) and v – T (red) based (a) on ERA-Interim data and (b) based on NCEP–NCAR data, and (c) F_{hv} – q (blue) and F_{hv} – T (red) based on NCEP–NCAR data. Values of T and q represent the DJF mean north of 70°N, and values of F_{hv} and v are at about 70°N, all for 700–1000 hPa and between 1979 and 2012. A correlation of 0.4 is significant at $p = 0.01$ (one sided).

is shifted westward by 30° relative to that for v (Fig. 1b). The value of F_{hv} represents the accumulated time under the influence of cyclonic circulation or storms in a season (Sorteberg and Walsh 2008). A small F_{hv} in sector 3 (60°–90°E), for instance, indicates that the region experiences below-normal cyclonic days or above-normal blocking days. As a result, F_{hv} (and ζ) of sector 3 is negatively correlated with v in the Nordic sectors to the west and positively with v in the Siberian sectors to the east (Table S1 in the supplemental material). Transient events in sector 3 would set up a Nordic–Siberian v dipole, and similarly transient events in sector 7 (Alaska) would set up a v dipole in neighboring sectors.

Figure 2 shows the time series of anomalies (normalized by the standard deviation) of Arctic T and q in the lower troposphere between 1979 and 2012 in comparison with that of v for 0°–60°E and for 90°–150°E, respectively. These anomalies show significant correlations with occurrences of matching year-to-year variations as well as parallel trends in the periods of

FIG. 2. Time series of wintertime average q and T (700–1000 hPa, 70°–90°N) compared to v winds (700–1000 hPa, near 70°N) over the Nordic (0°–60°E) and Siberian (90°–150°E) sectors. Anomaly data are normalized by their respective standard deviations. Linear lines of least squares fit are drawn to guide the eye (blue dotted line for T and q , green dotted line for v). Shown are (a) Arctic q (black) and Nordic v (red), (b) Arctic q (black) and Siberian v (red), (c) Arctic T (black) and Nordic v (red), and (d) Arctic T (black) and Siberian v (red). The y axis is reversed for the v anomaly in (b) and (d).

1979–97 and 1998–2012. The trends of T and q are steeper than that of v in the period of 1998–2012 (see below). The Nordic and Siberian circulation changes together explain two-thirds of the interannual variance of Arctic T (and q) in winters from December 1979 to February 2012 in a two-variable linear regression model (Fig. S5 in the supplemental material).

The residual of Arctic T below 700 hPa in winter that is not explained by the two-variable linear model is compared in Fig. S6 of the supplemental material with the annual mean lower tropospheric temperature between 70°S and 82.5°N based on satellite measurements (available at www.remss.com/msu/msu_data_description.html#zonal_anomalies), which were compared to surface air temperature measurements (Foster and Rahmstorf 2011). The residual T is possibly contributed by feedback effects associated with temperature, water vapor, clouds, and surface albedo (e.g., Pithan and Mauritsen 2014), and with changes in sea ice concentration and air–sea heat exchange (Burt et al. 2016). A linear trend ($0.024 \pm 0.008 \text{ K yr}^{-1}$) is estimated for the Arctic T residual based on the ERAI data, which is half of the total trend ($0.049 \pm 0.013 \text{ K yr}^{-1}$), and is a factor of 1.8 greater than that for the global trend ($0.013 \pm 0.003 \text{ K yr}^{-1}$). A similar analysis based on the NCEP–NCAR data gives a linear trend of $0.035 \pm 0.009 \text{ K yr}^{-1}$

in the residual T and an Arctic amplification (AA; defined as the ratio of Arctic trend to global trend) factor of 2.7. The total observed AA in winters from 1979 to 2012 is 3.8 and 4.3 based on the ERAI and NCEP–NCAR data, respectively. In other words, the Arctic warming would be slower than observed if the Nordic–Siberian seesaw did not have a long-term change. In comparison, the mean AA factor based on SAT is predicted to be about 1.6 ± 0.3 (annual), 1.7 (January–March), and 1.8 (October–December) by CMIP5 models for near-term (2020–44 minus 1980–2004) projections (Barnes and Polvani 2015). Similar results are obtained for long-term projections (2076–99 minus 1980–2004). The AA factor is predicted to be 2.6 (annual) and 3.6 (winter) by the same models for 4 times present-day CO_2 (Pithan and Mauritsen 2014). The AA factor is estimated to be 3–4 based on paleoclimate proxies (Miller et al. 2010).

Northward transport of atmospheric moist static energy crossing the Arctic Circle has decreased in winters from 1979 to 2012 (trend = $-0.26 \pm 0.10 \text{ W m}^{-2} \text{ yr}^{-1}$ when divided by the area within the Arctic Circle; Fan et al. 2015), while the top-of-atmosphere outgoing longwave radiation in the Arctic has increased during the same period (trend = $0.23 \pm 0.05 \text{ W m}^{-2} \text{ yr}^{-1}$). This implies an increase of ocean–atmosphere transfer of

FIG. 3. Streamlines based on monthly mean winds interpolated to a potential temperature (PT = 280 K) surface shown in pressure coordinates (hPa, color shading). (a) Composite for DJF 1980/81, 1983/84, 2004/05, 2005/06, and 2011/12; (b) composite for DJF 1981/82, 1986/87, 1996/97, 1997/98, and 2006/07.

energy in the Arctic (total trend = $0.49 \text{ W m}^{-2} \text{ yr}^{-1}$), which is associated with the long-term decline of sea ice (Comiso et al. 2008) and would also contribute to the observed AA. Surplus heat gained in summer through the ice–albedo feedback could contribute 25% of the energy for Arctic winter warming (Bintanja and van der Linden 2013). A reduction of sea ice also increases air temperature, water vapor, and cloudiness, leading to an increase of surface downward longwave radiation and Arctic amplification in winter (Kapsch et al. 2014; Burt et al. 2016).

The “greenhouse effect” of water vapor and clouds may amplify the effect of winds on Arctic winter climate. Figure 3 shows the streamlines based on monthly mean winds interpolated to a potential temperature (PT = 280 K) surface, represented in pressure coordinates, when the air currents are assumed to be isentropic. In warm Arctic winters (January ± 1 month, 1981, 1984, 2005, 2006, and 2012; Fig. 3a), southerly winds advect warm and moist air masses from the North Atlantic over the Nordic region, which then rise over the cold Arctic air and produce clouds. Along the path, downward longwave flux is enhanced by the warm tropospheric air and by clouds trapping planetary radiation (Stramler et al. 2011). Loss of heat from the Arctic Ocean is also suppressed by the warm and moist air flowing above the surface. In cold Arctic winters (1982, 1987, 1997, 1998, and 2007; Fig. 3b), by contrast, southerly winds from Siberia bring cold and dry continental air, causing clear air and a cooling of the surface. Indeed, the downward longwave radiation averaged over the Arctic (70° – 90° N) is positively correlated with v over the Nordic region and

negatively with v over northern Siberia in winter (Fig. S7 in the supplemental material), and is negatively correlated with F_{hv} over west and central Siberia (Fan et al. 2015).

Figure 4 shows the correlation coefficients between Arctic sea ice concentration and v by wind sectors. Significant negative correlations are found between DJF mean v over the Nordic regions and February sea ice, indicating that warm air advection suppresses ice formation in the Arctic. This occurs in concert with ice motion in the Barents Sea driven by surface wind stress (Wu et al. 2006; Overland and Wang 2010). Positive correlations are found between DJF v over central Siberia (sector 4) and sea ice in December as well as February. For monthly data, the largest positive and negative correlations are found between February sea ice and January v (results not shown). This suggests that sea ice growth and motion are possibly responding to circulation changes. The positive correlations result more from the downward trends than from year-to-year variations in v and sea ice (Fig. S8 in the supplemental material).

4. Arctic and East Asia climate variations associated with the eastern Atlantic pattern

In this section, we discuss the association of atmospheric teleconnections with Arctic and East Asia wintertime climate variations. The meridional winds are regulated by three local SLP patterns: the Arctic dipole (Wu et al. 2006; Overland and Wang 2010), the Barents Oscillation (Skeie 2000; Chen et al. 2013), and the SLP monopole over Siberia adjacent to the Kara Sea

FIG. 4. Correlation coefficients between sea ice anomalies (1979–2012, 70°–90°N) and the v wind (700–1000 hPa, near 70°N) by sectors: (a) HadISST ice and ERA-Interim v data, and (b) NSIDC ice and ERA-Interim v data. December (February) ice anomalies and are in black (red) and v values are for DJF. A correlation of 0.4 is significant at $p = 0.01$ (one sided).

(Smoliak and Wallace 2015). These patterns were identified over different analysis domains. Here we present an EOF analysis of the wintertime SLP for the Atlantic–Eurasian region (30°–88°N, 60°W–150°E) to show the interrelated seasonal climate anomalies over the North Atlantic, the Arctic, and East Asia.

The first three EOF patterns are shown in Fig. 5 and their PC time series in Fig. 6. The first EOF (EOF1 in Fig. 5) corresponds to the NAO. The PC1 (Fig. 6) is highly correlated ($r = 0.82$; Table 1) with the NAO index (Hurrell et al. 2003). The second EOF (EOF2) has a trough over the midlatitude North Atlantic and a trough over west-central Siberia (WCS), which is reminiscent of the eastern Atlantic pattern (Wallace and Gutzler 1981). Henceforth we use the EA pattern and the EOF2 pattern interchangeably. The third EOF (EOF3) corresponds to the Scandinavian blocking pattern. The Arctic winter mean T and q show small correlations with the NAO index and with the PC1 and PC3 time series, but significant correlations with the PC2 (Table 1).

The PC2 (Fig. 6) is also significantly correlated with the mean SLP in winter calculated for the region 50°–70°N,

FIG. 5. The spatial patterns of sea level pressure (SLP; 30°–88°N, 60°W–150°E; DJF) anomaly associated with (top)–(bottom) the first three empirical orthogonal functions (EOFs). The fraction of variance explained by each EOF is indicated at top of each panel. Arbitrary scales are used.

50°–110°E (i.e., WCS; $r = -0.81$; Table 1), similar to the result shown in Zhang et al. (2012), and with the mean SLP for the region 40°–55°N, 40°–20°W over the North Atlantic ($r = -0.66$). However, the correlation of SLP is low between the two regions ($r = 0.12$) because the North Atlantic variability is mostly due to the NAO. This begs the question of why a minor mode of variability over the North Atlantic projects a major impact over Siberia.

Figure 7 shows the composite differences calculated between five Atlantic ridge years and five trough years selected based on the highest amplitudes in the PC2 (Fig. 6). The root-mean-square (rms) variance of v at 300 hPa filtered with a bandpass of 2.5–6 days (Fig. 7a) is a measure of synoptic variability (Lau 1988). From ridge years to trough years (from EOF2– phase to EOF2+ phase), synoptic variability increased between 30° and 45°N over the North Atlantic and between 40° and 55°N over Asia, whereas it decreased to the north of these regions (Fig. 7a). In winters of EOF2+ (EA+), a broad Atlantic trough persists over the midlatitudes, storms tend to move more zonally and the eddy-driven jet is extended eastward (Athanasiadis et al. 2010; Wettstein and Wallace 2010; Woollings et al. 2010). The mean storm track and the eddy-driven jet over Eurasia are located between 45° and 50°N (not shown). In winters of EOF2– (EA–), a broad Atlantic ridge persists and the storm track is shifted poleward of its climatological-mean position, passing north of the British Isles toward

FIG. 6. (top)–(bottom) The principal components (PCs) corresponding to the EOFs shown in Fig. 5. Arbitrary scales are used.

the coast of Norway (Athanasiadis et al. 2010). The eddy-driven jet is also shifted poleward as shown by u wind at 300 hPa (Fig. 7b). These shifts in jet stream and storm track from EOF2+ to EOF2– (trough–ridge in the Atlantic sector) favor increases in SLP over Siberia. As a result, two-thirds of the variance of winter-mean SLP in the region northwest of Lake Baikal (WCS) is associated with the PC2 variability or with phase changes of the EA pattern.

The effect of horizontal advection associated with the EOF1 and EOF2 patterns may be shown by correlations of the PC1 and PC2 with air temperature at 2-m height (T_{2m} ; Fig. 8). Significant positive correlations (>0.4) between the PC1 and T_{2m} are shown from Europe to eastern Siberia and Japan (Fig. 8a). When the PC1 is positive a stronger zonal wind brings warm Atlantic air to the interior of the continent and moves more quickly over Siberia, resulting in higher T_{2m} in its path (Wallace et al. 2012). Between the PC2 and T_{2m} , significant

TABLE 1. Correlation coefficients between time series of winter mean variables. SLP is sea level pressure; NAO is North Atlantic Oscillation; PC1–PC3 is principal components 1–3; q is specific humidity; T is temperature; WCS is west-central Siberia (50° – 70° N, 50° – 110° E, i.e., northwest of Lake Baikal); Arctic is north of 70° N, below 700 hPa; and northeast Asia is 40° – 48° N, 100° – 130° E, below 1000 m. Boldface font indicates a correlation coefficient significant at $p < 0.01$.

	NAO	WCS SLP	Atlantic–Eurasian SLP		
			PC1	PC2	PC3
NAO			0.82	–0.24	–0.25
WCS SLP	–0.23		–0.61	–0.81	–0.14
Arctic q	0.16	0.67	–0.04	–0.72	–0.11
Arctic T	–0.05	0.71	–0.31	–0.59	–0.31
Northeast Asia T	0.16	–0.84	0.48	0.75	–0.04

negative correlations (<-0.4) are shown in a swath from the Greenland Sea to the Laptev Sea and positive correlations (>0.4) in the midlatitude Asia from the Caspian Sea to Japan (Fig. 8b).

We also calculated the winter mean air T from the surface to 1000-m altitude for a region from Mongolia to northeastern China (40° – 48° N, 100° – 130° E) and its correlations with various time series. The correlations are significant with the PC1 (0.48) and PC2 (0.75) but insignificant with the PC3 and NAO (Table 1). The PC1 and PC2 together explain 74% of the interannual variance in the winter T over northeast Asia (Fig. S9 in the supplemental material). Although cold winters in East Asia are associated with the Atlantic ridge, extremely cold days may also result from downstream development of Atlantic troughs on the synoptic time scale (results not shown; Joung and Hitchman 1982; Blackmon et al. 1984). Our analysis also confirms the strong influence of SiHi on Arctic and northeast Asian temperatures as indicated by their significant correlations with winter-average SLP in WCS (Table 1). It appears that the EOF2– (EA–) pattern is associated with warm winters in the Arctic and cold winters in East Asia because of fewer cyclones and/or more anticyclones over Siberia, which increase poleward transport over northern Europe and southward transport over East Asia, respectively (Zhang et al. 2012).

What excites the EA pattern from synoptic to seasonal time scales is a topic of great interest in light of its association with climate variability in the Arctic and Eurasia. The EA pattern was found to be correlated to SST in the equatorial Pacific, and to be more responsive to La Niña conditions (Cassou and Terray 2001a). Enhanced convective heating over the Maritime Continent (i.e., Indonesia and adjacent regions) may lead to enhanced excitation of poleward-propagating Rossby waves, which possibly contribute to Arctic warming

FIG. 7. Composite differences (PC2+ minus PC2- or Atlantic trough minus ridge; five winters selected for each phase as in Fig. 3): (a) band-passed (2.5–6 days) v variance at 300 hPa (rms values are shown with a contour interval of 0.5 m s^{-1}), and (b) zonal wind at 300 hPa (contour interval of 2 m s^{-1}). Solid red (dashed blue) contours represent positive (negative) values and the dotted green is the 0 contour.

(Yoo et al. 2011, 2012) and Eurasian cooling (Vecchi and Bond 2004). Convective heating over the Maritime Continent may trigger the Pacific–North American (PNA) pattern (Mori and Watanabe 2008; Adames and Wallace 2014; Bao and Hartmann 2014). An eastward extension of the PNA was suggested as a possible mechanism for exciting and maintaining the EA pattern (Cassou and Terray 2001b). The EA pattern was also found to influence European land carbon sink (Bastos et al. 2016). Its environmental and ecological impacts in the Arctic and central Asia–East Asia remain to be investigated.

5. Conclusions

In summary, the correlation analyses presented in this paper shows a natural mode of Arctic winter variability resulting from the Nordic–Siberian seesaw of meridional winds, which is sensitive to the frequency and persistence of North Atlantic weather regimes and to the frequency and persistence of cyclonic circulations from

FIG. 8. Correlation coefficients (a) between PC1 and winter-average air temperature at 2 m (T_{2m}) and (b) between PC2 and T_{2m} . The color shading for correlation is shown below the panels.

the Ural Mountains to central Siberia (60° – 120° E). The wind seesaw is associated with two-thirds of the interannual variance of winter-mean Arctic temperature between 1979 and 2012, and possibly contributed a substantial fraction of the observed Arctic amplification in this period.

On a seasonal time scale, the EA pattern appears to be the teleconnection most significantly associated with winter climate variations in the Arctic and East Asia, while the NAO is not. About two-thirds of the winter-mean SLP variance over western and central Siberia is associated with the EA pattern. This may suggest an important mechanism for the interannual variations of the Siberian SLP in winter due to internal atmospheric dynamics and/or external forcing. We are conducting further analysis of the physical mechanisms underlying the EA teleconnection, its variability, and associated environmental impact. However, it should be remembered that correlations do not assure causality and that the NAO may influence Arctic climate with a time lag. Future studies should explore the interactions between the Atlantic ridge and sea ice growth from the northern Atlantic to the Kara Sea, and their combined effect on climate in the Arctic and East Asia.

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