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ABSTRACT

In this paper, the authors use NCEP reanalysis and 40-yr ECMWF Re-Analysis (ERA-40) data to document the strengthened relationship between the East Asian winter monsoon (EAWM) and winter Arctic Oscillation (AO) on the interannual time scale with a comparison of 1950–70 and 1983–2012. Their connection was statistically insignificant during 1950–70, whereas it was statistically significant during 1983–2012. The latter significant connection might be attributed to the East Asian jet stream (EAJS) upstream extension: the EAJS signal is relatively confined to the western North Pacific before the 1970s, whereas it extends westward toward East Asia after the 1980s. This upstream extension leads to the rearrangement of eastward-propagating Rossby waves with a much wider horizontal structure, thereby bonding the EAWM and the AO.

Furthermore, the authors present observational evidence and model simulations demonstrating that the reduction of autumn Arctic sea ice cover (ASIC) is responsible for the strengthened EAWM–AO relationship after the 1980s by producing the EAJS upstream extension. After the 1980s, a strong anticyclonic anomaly over the polar ocean and anomalous easterly advection over northern Eurasia are generated by the near-surface heating over the Barents–Kara (B–K) Seas caused by the reduction of ASIC. This further induces cold anomalies over northern Eurasia, altering the meridional temperature gradient between the midlatitude and tropical region and consequently leading to westward penetration of the EAJS.

1. Introduction

The most distinct feature of the East Asian winter monsoon (EAWM) is its cold surge activity. As these cold surges move southward with cold, dry continental air, they cause intense drops in temperature over East Asia (Tao and Chen 1985; Sun et al. 2009, 2010; Wang et al. 2011; Li and Wang 2012, 2014; Wang and He 2013).

Early studies indicated that the frequency of the cold surges is related to the variability of the cold Siberian high (SH), the warm Aleutian low (AL), the East Asian trough (EAT) in the middle troposphere, and the East Asian jet stream (EAJS) in the upper troposphere.

Factors that may influence the EAWM include El Niño–Southern Oscillation (ENSO; Webster and Yang 1992; Zhang et al. 1996; Lau and Nath 2000; Wang and He 2012; He and Wang 2013a; Wang et al. 2013), the Arctic Oscillation (AO; Gong et al. 2001; Wu and Wang 2002; He and Wang 2013b), the Antarctic Oscillation (Fan and Wang 2004, 2006), conditions on the Tibetan Plateau and Eurasian snow cover (Walland and Simmonds 1996;
The Arctic climate is rapidly changing, as documented by the Arctic warming and Arctic sea ice decline. Early studies have suggested that the climate changes in Arctic can potentially lead to climate changes in the extratropics in the Northern Hemisphere (Wu et al. 1999; Budikova 2009; Bader et al. 2011; Li and Wang 2013a, b). For example, Liu et al. (2012) detected the substantial impacts of autumn (September–November) Arctic sea ice cover (ASIC) on the EAWM, winter temperature, and snowstorm activity. Li and Wang (2013c) documented the impacts of ASIC on the AO and winter precipitation in Eurasia. More recently, Li et al. (2014) revealed a strengthening of the AO–ENSO relationship in January after the mid-1990s. They suggested that after the mid-1990s the reduction of September Arctic sea ice cover (SIC) accounts for a large-scale wave train across the Pacific in January. The AO is strongly coupled to the circulation in the Pacific as a result. Motivated by Li et al. (2014), the objective of this paper is to explore a strengthening of the EAWM–AO relationship and test the role of ASIC on this strengthened relationship.

The rest of the paper is organized as follows: Section 2 describes the datasets used. Section 3 illustrates the strengthened EAWM–AO relationship and associated atmospheric variability. The role of ASIC is investigated in section 4. Finally, we conclude the paper with a summary of the findings in section 5.

2. Data

The datasets employed in this research include the National Centers for Environmental Prediction (NCEP) atmospheric reanalysis (1948–2013) with 2.5° × 2.5° resolution (Kalnay et al. 1996), using variables including air temperature at 2 m and 1000 hPa (T2m and T1000); wind vector at 850 hPa (UV850); geopotential height at 1000, 500, and 300 hPa (Z1000, Z500, and Z300); zonal wind component at 200 hPa (U200); the 40-yr European Center for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) (1957–2002) with 2.5° × 2.5° resolution (Uppala et al. 2005); and the Met Office Hadley Centre Sea Ice and Sea Surface Temperature dataset, version 1 (HadISST1) (1870–2013) with 1.0° × 1.0° resolution (Rayner et al. 2003). The AO index is defined as the first principal component of 1000-hPa height anomalies poleward of 20°N (available from http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/ao.shtml). The common time period is set to 1950–2012. The winter of 1950 refers to the 1950/51 winter and so on. The months of December–February are used in calculating the winter mean for all variables (e.g., atmosphere and sea ice). In this paper, all data and indexes were detrended before analysis.

3. The strengthened EAWM–AO relationship from observation

a. Time evolutions

Various indexes have been used to describe the EAWM. In this paper, we use three categories of the EAWM index [EAWMI (1–3)], which describe the EAWM in the lower, middle, and upper troposphere, respectively. The indexes include wind speed at 850 hPa [averaged within (25°–50°N, 115°–145°E); Wang and Jiang 2004], East Asian trough [Z500 averaged within (25°–45°N, 110°–145°E); Wang and He 2012], and upper-tropospheric zonal wind shear [U200 averaged within (30°–35°N, 90°–160°E) minus half-U300 averaged within (50°–60°N, 70°–170°E) and half-U300 averaged within (5°S–10°N, 90°–160°E); Li and Yang 2010]. All three indexes are normalized. Additionally, we define an ensemble EAWMI (4) by evenly averaging of these three categories of the normalized EAWMI.

The time evolutions of EAWMI (1–4), together with the negative AO index (−AOI) are presented in Fig. 1 (left). Although there are some differences in amplitude, for each EAWMI, a strong EAWM to −AO in-phase correlation after the mid-1970s can be identified. The 25-yr sliding correlation coefficients between EAWMI (1–4) and −AOI (Fig. 1, right) indicate that the EAWM to −AO in-phase correlations are statistically insignificant before the mid-1970s but become statistically significant afterward. Additionally, we also performed a 21-yr sliding correlation (figure not shown). The results confirmed that the strengthened EAWM–AO relationship after the mid-1970s is robust. Based on the identified change in the mid-1970s in the EAWM–AO relationship and in order to avoid the effect of...
choosing the sliding window width, we selected two subperiods, 1950–70 (P1) and 1983–2012 (P2), by removing 12 consecutive years (equal to the half-width of the sliding window) around the year 1976, to explore the possible mechanism.

Table 1 shows the relationships (correlation coefficients) between EAWMI (1–4) and EAWMI (1–3), as well as between EAWMI (1–4) and AOI. Clearly evident is that EAWMI (4) has significant relationships with EAWMI (1–3) for the whole period and two subperiods (1950–70 and 1983–2012). It also confirms that the relationships between EAWM (1–4) and AOI all tend to be unstable with time. More significant relationships appear in the later period.

b. Lower-tropospheric structure of EAWM and AO

To illustrate the common characteristics of the EAWM-related lower-tropospheric anomalies, we present in Fig. 2 the spatial distributions of linear regression coefficients of Z1000 and T1000 with UV850 derived from the NCEP reanalysis and ERA-40 datasets upon the EAWMI (4) during the two periods. The results reveal two important features. First, for each period, high EAWMI (4) is associated with a robust SH, a deep AL, predominant lower-level northwesterlies, and significant cold anomalies centered over the Far East. Second, for the later period only, the typical behavior of the monsoon circulation occurs with a striking dipolar pattern over the Europe–Atlantic sector (Fig. 2b) and a zonally elongated cold signature from Europe to the Far East (Fig. 2d). This result suggests a stronger teleconnection with the AO: that is, the strengthened EAWM–AO relationship during recent decades. A parallel analysis has been performed using Z1000, T1000, and UV850 derived from the ERA-40 dataset during 1957–70 and 1983–2001. The results based on ERA-40 (Figs. 2e–h) are similar to
those based on the NCEP reanalysis (Figs. 2a–d) with spatial correlation coefficients ranging from 0.71 to 0.99.

Furthermore, we performed empirical orthogonal function (EOF) analyses on Z1000 north of 20°N derived from the NCEP reanalysis dataset during the two periods (Figs. 3a,b). Two leading EOF modes of sea level pressure (SLP EOF1) show quite different patterns. For P1, the SLP EOF1 mode clearly shows a Europe–Atlantic dipole (anomalies over the Pacific are less associated with this mode). For P2, the SLP EOF1 mode shows a polar–midlatitude seesaw with a high degree of annularity. That is, there are more occurrences of action centers in the Pacific.

The spatial distributions of linear regression coefficients of T1000 and UV850 upon the –AOI during the two periods (Figs. 3c,d) indicate the following: For P1, the T1000 and UV850 regression map is characterized with anticyclonic anomalies centered over Greenland and with significant cold anomalies from the Europe to west Siberia. For P2, the T1000 and UV850 regression map is accompanied by an adjustment of the large-scale circulation system over the Asia–Pacific sector with cyclonic anomalies over the Pacific (a deep AL), thereby favoring a southeastward intrusion of the cold air from Siberia. The EAWM is intensified as a result. The AO shows the significant correlation with winter surface air temperature in the Eurasian continent during the two periods (Figs. 3c,d), whereas the EAWMI (4) shows the similar correlation only in the later period (Figs. 3d, 2h).

It should also be noted that the results from NCEP (Figs. 3a–d) are similar with those from ERA-40 (Figs. 2e–h), with spatial correlation coefficients from 0.93 to 0.98.

c. Upper-tropospheric structure of EAWM and AO

EAWM is an important component of EAW. Yang et al. (2002) reveals the association of the Asian–Pacific–American climate with the EAWSI. They found that a strong EAWSI is associated with an intensification of the weather and climate systems in Asia and over the Pacific because of the related changes in stationary wave patterns. Watanabe (2004) analyzed anomalous atmospheric circulation and temperature fields associated with the North Atlantic Oscillation (NAO) on the interannual and intraseasonal time scales. They suggested that the NAO signal extends toward East Asia and the North Pacific in February and the meridional wind anomaly shows a wave train along the EAWSI, collocated with an anomalous vorticity source near the jet entrance. The time evolutions of the annual EAWSI index [EAWSI; U200 averaged within (30°–35°N, 130°–160°E); Yang et al. 2002] and its 11-yr high-pass filtered values (HP-EAWSI) (Fig. 4a) show that the amplitude of HP-EAWSI is reduced on the decadal time scale over the past 60 yr with the standard deviations of 1.03 for 1950–70 and 0.74 for 1983–2012. The 25-yr sliding correlation coefficients between EAWSI and –AOI (Fig. 4b) indicate that the relationship between EAWSI and –AOI tends to be unstable as well. The EAWSI to –AO in-phase correlations become statistically significant after the 1980s. As will be discussed below, the EAWSI plays as a bridge linking EAWM and AO, particularly after the 1980s.

To explore spatial feature of the EAWSI associated with EAWSI, we present in Fig. 5 the spatial distributions of linear regression coefficients of U200 and T1000–UV850 upon the EAWSI during the two periods. For P1, the EAWSI signal concentrated on the western North Pacific (Fig. 5a), which was similar with the results from Yang et al. (2002). However, for P2, the EAWSI signal extended westward toward East Asia (Fig. 5b). The near-surface air temperature anomaly closely matched such an extension, called the EAWSI upstream extension throughout the paper. The main difference between P2 and P1 is zonally elongated cold signature from Europe to the Far East in P2 (Fig. 5d).

We next investigate the source and propagation of wave activity that is associated with change in EAWSI. Figure 6 illustrates the spatial distributions of linear regression coefficients of Z300–stationary wave activity flux (SWAF; formulated by Plumb 1985) and Z700 upon the EAWSI during the two periods. For P1, the strengthened EAWSI was associated with eastward-propagating stationary wave and with distinct strong
FIG. 2. Linear regression coefficients of (a),(b) Z1000 (units: gpm; contour) and (c),(d) T1000 (units: °C; contour) and UV850 (units: m s\(^{-1}\); arrow) derived from the NCEP reanalysis dataset upon the EAWMI (4) during (a),(c) 1950–70 and (b),(d) 1983–2012. Solid (dashed) contour lines indicate positive (negative) values. Areas with positive (negative) values that are statistically significant at the 95% confidence level, as estimated using Student’s \(t\) test, are shaded dark (light). (e)–(h) As in (a)–(d), but for the ERA-40 dataset.
FIG. 3. The leading EOF modes of Z1000 north of 20°N derived from the NCEP reanalysis dataset during (a) 1950–70 and (b) 1983–2012. Linear regression coefficients of T1000 (units: °C, contour) and U/V850 (units: m s⁻¹, arrow) derived from the NCEP reanalysis dataset upon the AOI during (c) 1950–70 and (d) 1983–2012. Solid (dashed) contour lines indicate positive (negative) values. Areas with positive (negative) values being statistically significant at the 95% confidence level, as estimated using Student’s t test, are shaded dark (light). (e)–(h) As in (a)–(d), but for the ERA-40 dataset.
source over the eastern Pacific (Fig. 6a). For P2, the eastward propagation of SWAF is weaker over the eastern Pacific, whereas it is stronger over the Atlantic with a much wider horizontal structure (Fig. 6b). The corresponding Z700 anomaly in Fig. 6d has a dipolar pattern over the Europe–Atlantic sector, rather similar to those attributed to AO variability.

To confirm the results from linear regression computations, we performed composite analysis with strong-minus-weak EAJSI years (based on the criterion that EAJSI exceeds 1.0 standard deviation or is less than −1.0 standard deviation). It can be seen from Fig. 7 that the composite features are consistent with those of Figs. 5 and 6. Thus, the strengthened EAJS–AO relationship after the 1980s is robust.

4. Possible reason for strengthened EAWM–AO relationship from observation

Satellite observations show substantial reduction in ASIC (Stroeve et al. 2007) and overall thinning in conjunction with a loss of older, thicker sea ice (Maslanik et al. 2007; Kwok et al. 2009). Such changes have been accompanied by the prominent extratropical atmospheric variability, such as the EAWM and AO (Liu et al. 2012; Li and Wang 2013c; Li et al. 2014). Recent model studies have suggested that lower-tropospheric heating over the Barents–Kara (B–K) Seas in the eastern Arctic caused by the sea ice reduction results in strong anticyclonic anomaly over the polar ocean and anomalous easterly advection over northern Eurasia (Petoukhov and Semenov 2010). To understand the mechanism behind the strengthening of the EAWM–AO relationship, we will explore the potential impacts of the reduction in ASIC. We performed linear regression coefficients of ASIC upon the AOI (Fig. 8a) and the EAJSI (Fig. 8b) to identify the potential regions of interest. Both AOI and EAJSI are negatively correlated with change in ASIC over the B–K Seas. The temporal variations of the ASIC over the B–K Seas (67°–85°N, 30°–135°E), defined as the ASIC index (ASICI), are presented in Fig. 8c. The amplitude of HP-ASICI shows much stronger interannual variations after the 1980s. The standard deviations during the two periods (1950–70 and 1983–2012) are 0.68 and 1.20, respectively.

Figure 9 illustrates the spatial distributions of linear regression coefficients of wintertime Arctic SIC (WSIC) and T1000 north of 20°N upon the ASICI during the two periods to depict the evolution of the ASIC signal. For P1, decreasing ASIC corresponds to a significantly below-normal WSIC from the Greenland Sea to the Barents Sea (Fig. 9a). For P2, the significantly below-normal WSIC exists only over the B–K Seas in the eastern Arctic (Fig. 9b), a regime change that has been attributed to the reduction and thinning of ASIC in recent decades (Li and Wang 2013c). It is known that low SIC can decrease surface albedo and strengthen the amount of ocean heat released to the atmosphere. This implies large impacts on air–sea exchanges (Koenigk et al. 2009). It is thus reasonable to expect that warm anomalies occur over a region, where sea ice had actually retreated. For P1, strong linked response of the T1000 to ASIC decrease are warm anomalies from the Greenland Sea to the Barents Sea and over northern Eurasia but cold anomalies in the Canadian Archipelago (Fig. 9c). For P2, the pattern of T1000 response to ASIC decrease is drastically different. This pattern has warm anomalies mainly over Greenland and the B–K Seas but cold anomalies over northern Eurasia (Fig. 9d). Such considerable continental cooling was also identified by Outten and Esau (2012). They suggested that this cooling is related to extreme warming around the Kara Sea caused by the sea ice decrease through changes in the meridional temperature gradient and large-scale atmospheric flow.

The potential atmospheric circulation responses to change in ASIC are illustrated in Fig. 10. For P1, the pattern of Z700 response to ASIC decrease has a strong cyclonic anomaly over the western Arctic, collocated
with a robust SH (Fig. 10a). The corresponding Z300 anomaly pattern (Fig. 10c) looks rather similar to the Z700 response. The eastward propagation of SWAF is weaker with distinct strong source over the western Arctic. In addition, anomalous westerly advection develops around the polar ocean (Fig. 10e) that transports warming Atlantic air masses and sharpens warm anomalies over northern Eurasia (Fig. 9e).

However, for P2, the pattern of Z700 response to ASIC decrease resembled the negative phase of the AO, which has a Europe–Atlantic dipole (Fig. 10b). The corresponding Z300 anomaly pattern (Fig. 10d) has much in common with the Z700 anomaly pattern. Like those attributed to AO variability, the physical reason for these patterns might be a modulation in the strength of the polar vortex aloft. The eastward propagation of SWAF has a wider horizontal structure with a distinct strong source over the Atlantic. This suggests that the patterns of remote teleconnections result from excitation of the large-scale quasi-stationary Rossby waves, consistent with findings reported by Honda et al. (2009). The U200 anomaly pattern (Fig. 10f) actually shows the EAJS upstream extension. In addition, the considerable continental cooling is consistent with westward
penetration of the EAJS toward East Asia, which is likely caused by the intensified meridional temperature gradient between the midlatitude and tropical regions (Fig. 9f).

In summary, we have shown, using the NCEP reanalysis and ERA-40, the strengthened EAWM–AO relationship after the 1980s and the consistent changes in EAWM-related circulations in the upper and lower troposphere. We hypothesize that the reduction of ASIC is responsible for the strengthened EAWM–AO relationship after the 1980s by inducing a strong anticyclonic anomaly over the polar ocean and anomalous easterly advection over northern Eurasia. This further causes a wider horizontal structure of the eastward propagation of SWAF with distinct strong source over the Atlantic and cold anomalies over northern Eurasia, altering the meridional temperature gradient between the midlatitude and tropical region and consequently leading to the EAJS upstream extension.

5. Results from numerical simulation

To test our hypothesis, we conducted simulations using the National Center for Atmospheric Research
Fig. 7. Strong-minus-weak EAJSI composites of (a),(b) U200 (units: $\text{m s}^{-1}$; contour); (c),(d) T1000 (units: $\text{C}^\circ$; contour) and UV850 (units: $\text{m s}^{-1}$; arrow); (e), (f) Z300 (units: gpm; contour) and SWAF (units: $\text{m}^2 \text{s}^{-2}$; arrow); and (g),(h) Z700 (units: gpm; contour) derived from the NCEP reanalysis dataset during (left) 1950–70 and (right) 1983–2012. Solid (dashed) contour lines indicate positive (negative) values. Areas with positive (negative) values that are statistically significant at 90% confidence level, as estimated using Student’s $t$ test, are shaded dark (light).
Community Atmospheric Model, version 3.1 (Collins et al. 2006). In these simulations, sea surface temperatures (SSTs) and sea ice concentrations were specified as boundary conditions based on a merged product of the Hadley Centre Sea Ice and Sea Surface Temperature dataset and the National Oceanic and Atmospheric Administration weekly optimum interpolation SST analysis (Hurrell et al. 2008). The experimental design was similar to that of Liu et al. (2012).

The simulation configuration had a horizontal resolution of approximately 2.8° and 26 vertical levels extending up to 3.5 hPa. The impact of diminishing Arctic sea ice during the freeze up on atmospheric circulation was assessed by comparing two experiments with different seasonally varying sea ice distributions, while all other external variables remained fixed. The control experiment was run with seasonally varying Arctic sea ice based on the climatology of the Hadley Centre sea ice concentrations for 1979–2010. The perturbed experiment was integrated with seasonally varying ice loss in the area (65°–85°N, 30°W–90°E). Sea ice losses were calculated as SIC in 1983–2012 minus those in 1950–70. Global SSTs for both experiments were set to their climatological monthly values based on the merged SST dataset for the same period as the sea ice climatology in the control experiment. For the perturbed experiment, in areas where the sea ice was removed, the SST was set to the freezing point of seawater, −1.8°C. To help gauge confidence in the model’s response to sea ice losses, each experiment consists of 20 ensemble members with slightly different initial conditions.

The model response to ASIC loss was examined by calculating the mean difference in the Z300, SLP and

![Figure 8](https://example.com/figure8.png)

**Figure 8.** Linear regression coefficients of autumn (September–November) ASIC (units: 10^5 km^2; shaded) upon (a) the −AOI and (b) the EAJSI from 1983 to 2012. Dotted values are statistically significant at the 95% confidence level, as estimated using Student’s t test. (c) Time series of the ASIC in the area (67°–85°N, 30°–135°E), outlined in (a),(b), from 1950 to 2012.
FIG. 9. Linear regression coefficients of WISC (December–February; units: 10^5 km^2; shaded) upon the −ASICI during (a) 1950–70 and (b) 1983–2012. Dotted values are statistically significant at the 95% confidence level, as estimated using Student’s t test. Linear regression coefficients of (c),(d) T1000 (units: °C; contour) and UV850 (units: m s^-1; arrow) and (e),(f) T300 (units: °C; contour) and UV200 (units: m s^-1; arrow) north of 20°N upon the −ASICI during 1950–70 and 1983–2012. Solid (dashed) contour lines indicate positive (negative) values. Areas with positive (negative) values that are statistically significant at 95% confidence level, as estimated using Student’s t test, are shaded dark (light).
Fig. 10. Linear regression coefficients of (a),(b) Z700 (units: gpm; contour); (c),(d) Z300 (units: gpm; contour) and SWAF (units: m$^2$s$^{-2}$; arrow); and (e),(f) U200 (units: m$s^{-1}$; contour) derived from the NCEP reanalysis dataset upon ASIC during (left) 1950–70 and (right) 1983–2012. Solid (dashed) contour lines indicate positive (negative) values. Areas with positive (negative) values that are statistically significant at the 95% confidence level, as estimated using Student’s $t$ test, are shaded dark (light).
SAT between the perturbed and control experiments. The modeled Z300 and SLP anomalies consist of an AO dipolar pattern, an intensified EAJS, and a deep AL. As shown in Fig. 10c, diminishing ASIC in the region of interest generated a zonally elongated cold signature from Europe to the Far East. The encouraging consistency between model simulations and observations supports our hypothesis that there is a dynamical consistency between ASIC and EAJS, and the latter acts as the bridge for the strengthened relationship between EAWM and AO.

6. **Discussion and conclusions**

The weakening of the EAWM has been documented by both the reanalysis data (Wang and Chen 2010) and the ensemble hindcasts (Li and Wang 2012). Early studies indicated that the interannual variability of EAWM decreased after the 1980s (Wang and He 2012; He et al. 2013). He (2013) found that the decrease is due to global warming, which reduces the land–sea contrast variability on both the interdecadal and interannual time scales. The impact of AO on the EAWM was
studied by Gong et al. (2001) and Wu and Wang (2002). In this paper, we illustrated (Fig. 1 and Table 1) that the EAWM–AO relationship is not stable and was strengthened during recent decades. Observational evidence and model simulations were further presented in this paper to document the possible mechanism responsible for the strengthening of the EAWM–AO relationship, with a comparison of 1950–70 and 1983–2012. It is revealed that the rapid reduction of ASIC might be an important factor. The reduction in ASIC can cause the warming in the Arctic and cooling in the Eurasian continent (Arctic warming and Eurasian cooling; Figs. 9d, 11). Such Arctic warming and Eurasian cooling responses can also be found in the phase 5 of the Coupled Model Intercomparison Project (CMIP5) models that are related to the Arctic sea ice reduction in the Barents and Kara Sea sectors (Yang and Christensen 2012). The warmed Arctic can change the air pressure between the Arctic and the midlatitude and therefore change the EAJS (Fig. 11c). This speculation can be supported by the earlier studies that say the warming in the Arctic can cause the slowdown of the westerly at midlatitudes (Francis and Vavrus 2012; Guo et al. 2013). Furthermore, the Arctic sea ice loss can lead to increased winter precipitation and negative surface air temperature gradient between the midlatitude and high latitudes of East Asia (Wu et al. 2013).

We now summarized our main findings as follows:

1) The in-phase correlations between EAWMI (1–4) and AOI are statistically insignificant during 1950–70 but statistically significant during 1983–2012.

2) The spatial distribution of the EAWM-associated atmospheric variability shows distinct difference. For P1, stronger EAWM is associated with a robust SH, a deep AL, predominant lower-level northwesterlies, and significant cold anomalies centered over the Far East. Comparatively, for P2, stronger EAWM is associated with a zonal symmetry Arctic–Atlantic SLP anomaly dipole resembling AO signature and with a cold signature zonally elongated from Europe to the Far East.

3) The spatial structure of AO was also found to undergo important interdecadal variations. For P1, the SLP EOF1 mode clearly shows an AO dipolar pattern with two centers located in the Atlantic and the Arctic, respectively. For P2, the SLP EOF1 mode is featured with a polar–midlatitude seesaw, accompanied by more occurrences of a center of action in the Pacific.

4) Numerical simulations and observations showed that the reduction in ASIC is responsible for the strengthened EAWM–AO relationship after the 1980s. The mechanism can be described as follows: After the 1980s, the near-surface heating over the B–K Seas in the eastern Arctic caused by the reduced ASIC is associated with a strong anticyclonic anomaly over the polar ocean and anomalous easterly advection over the northern continents. This causes a wider horizontal structure of the eastward propagation of SWAF with a distinct strong source over the Atlantic. Furthermore, the considerable Eurasian cooling is consistent with the westward penetration of the EAJS toward East Asia by the intensified meridional temperature gradient between the midlatitude and tropical regions (Fig. 9f).

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