INTERNATIONAL JOURNAL OF CLIMATOLOGY
Int. J. Climatol. **36**: 1026–1032 (2016)
Published online 19 July 2015 in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/joc.4399

Royal Meteorological Society

Short Communication

Linkage between the East Asian January temperature extremes and the preceding Arctic Oscillation

Shengping He[a,b]* and Huijun Wang[a,b]
[a] *Climate Change Research Center, Chinese Academy of Sciences, Beijing, China*
[b] *Nansen-Zhu International Research Centre, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China*

ABSTRACT: This study investigated the potential connection of the October–December Arctic Oscillation (AO-OND) with the following January East Asian temperature extremes and the possible mechanisms. It was found that the extreme cold (warm) events are less (more) frequent in January over East Asia following a positive phase of AO-OND, which might be attributed to the intrinsic persistence of AO-OND in stratosphere and the memory of Eurasian snow cover. It is revealed that the barotropic structure of AO-OND could extend to stratosphere. By the following January, wave activities propagate from stratosphere downwards to upper troposphere, then bent equatorwards to mid-latitudes. Consequently, the January circumpolar jet (polar vortex) gets strengthened (colder) because the equatorward-pointing (upwards) wave activities correspond to poleward meridional eddy momentum (eddy heat) flux. In such a way, the anomalous AO-OND persists into the following January. On the other hand, positive phase of AO-OND causes significant surface warming in most parts of Eurasia, leading to Eurasian snow melt. The October–December snow melt has intrinsic climatic memory and further weakens the January Siberian High through reducing surface albedo. Besides, the surface warming in North Europe because of reduced surface albedo caused by snow melt could stimulate a Rossby wave train propagating eastwards across Eurasia, which restrains the blocking events around Ural region. In such a context, the incidence of extreme warm (cold) events over East Asia is more (less) frequent.

KEY WORDS Arctic Oscillation; temperature extremes; snow cover; Rossby wave

Received 18 August 2014; Revised 1 April 2015; Accepted 11 May 2015

1. Introduction

The East Asian continent is much colder in boreal winter than other regions at similar latitudes except for northeastern North America because of a lack of maritime influence from the westerly winds that prevail in Europe. One of the main factors contributing to winter air temperature anomalies in this region is the East Asian winter monsoon (EAWM) (Zhou *et al.*, 2007; Wang *et al.*, 2011; Cheung *et al.*, 2012; Wang and He, 2012; He, 2013; Zhou *et al.*, 2013). Associated with a stronger (weaker) than normal EAWM, the northerly winds along coastal East Asia are more intense (moderate) and bring more (less) cold air to the Asian continent, which leads to colder (milder) winters in East Asia (Boyle and Chen, 1987; Chang *et al.*, 2006; Li and Wang, 2012, 2013a, 2013b).

Numerous previous studies have revealed that the Arctic Oscillation (AO) could significantly impact the EAWM and further causes winter temperature anomalies over East Asia (Gong *et al.*, 2001; Wu and Wang, 2002; Li *et al.*, 2014). A general knowledge of those studies is that, over East Asia, winters are warmer than normal during a positive phase of AO when the EAWM is weaker, while the converse is true during a negative phase of AO. It has been found that the out-of-phase AO–EAWM relationship can be interpreted from the perspective of quasi-stationary planetary waves (Thompson and Wallace, 1998; Chen and Huang, 2002).

Most of the previous work focuses on the impact of winter AO on the simultaneous temperature variability over East Asia. Only a few studies have paid attention to examining the potential influence of the preceding AO on the winter climate over East Asia (Kim and Ahn, 2012; He and Wang, 2013). AO has potential impact on the simultaneous atmospheric circulation over East Asia. The correlation coefficients (CCs) between the AO index during 1951–2013 January and the indices such as the Siberian High, East Asian trough, East Asian jet stream, and 850-hPa meridional wind over East Asia (He and Wang, 2012) are −0.27, 0.45, −0.41, and 0.30, respectively, suggesting a weaker-than-normal monsoon. On the other hand, it seems that the anomalous January AO might persist from October–December. The CCs of AO index in 1951–2013 January with those in the preceding

* Correspondence to: S. He, Climate Change Research Center, Chinese Academy of Sciences, Huayanli 40, Chaoyang district, Beijing 1000029, China. E-mail: hshp@mail.iap.ac.cn

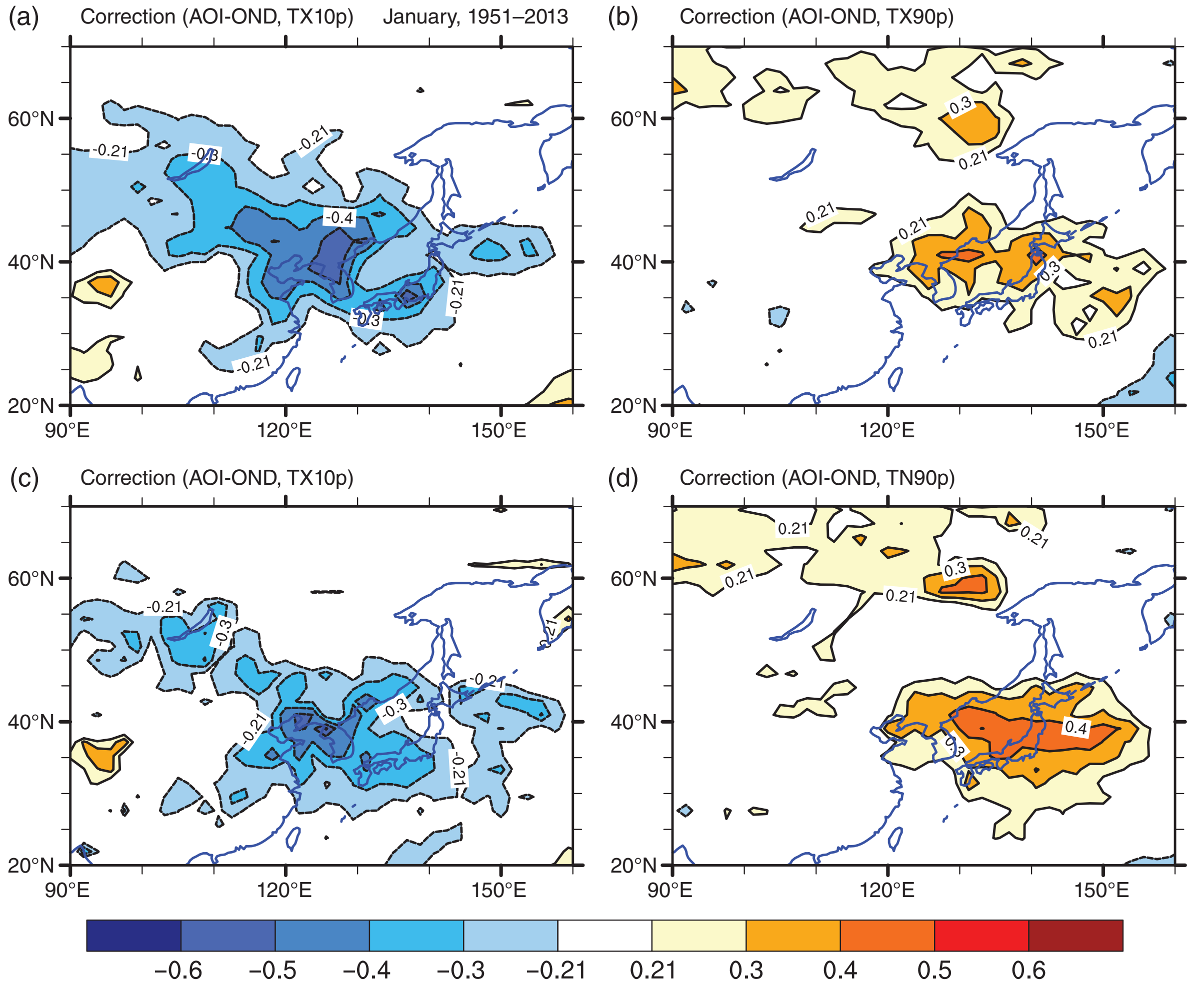

Figure 1. Correlation maps of AOI-OND with the following January (a) cold days (TX10p), (b) warm days (TX90p), (c) cold nights (TN10p), and (d) warm nights (TN90p) during 1951–2013, respectively. Shaded values are significant at 90% level from a two-tailed Student's *t*-test.

October, November, and December are 0.25, 0.24, and 0.45, respectively. Therefore, the East Asian January climate might has potential connection with the preceding October–December AO (AO-OND), as pointed out by He and Wang (2013). However, it is still unclear whether the January temperature extremes over East Asia could be impacted by the preceding AO-OND. The purpose of this study is to uncover this scientific question.

2. Data and method

The datasets used in this study include: (1) daily maximum (T_{max}), minimum (T_{min}) temperature, and monthly mean large-scale atmospheric datasets derived from the National Centers for Environment Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis dataset (Kalnay *et al.*, 1996); (2) snow cover derived from the 20th Century Reanalysis (Compo *et al.*, 2011); (3) monthly mean AO indices obtained from the Climate Prediction Center. When a daily T_{max}/T_{min} is lower (greater) than 10th (90th) percentile value, it is classified as a cold (warm) day/night. In general, January cold/warm extremes over East Asia are about 3–5 days.

The Eliassen–Palm (E-P) flux, which is a measure of the wave activity propagation (Andrews *et al.*, 1987), is employed as a diagnostic tool. In theory, an equatorward (upwards) E-P flux vector corresponds to poleward meridional eddy momentum (heat) flux. The blocking event detection algorithm basically follows the one adopted by Tibaldi and Molteni (1990). All data and indices were detrended before analysis. The period we focus on is 1950/51 to 2012/13.

3. Results

As illustrated in Figure 1, the AO-OND is consistently highly correlated with the following January temperature extremes over East Asia. As to cold days (Figure 1(a)), there are significant negative CCs extending from the central Siberia southeastwards to Northeast Asia including Eastern China, Northeast China, Korea, and Japan. The spatial distribution of CCs of AOI-OND with the cold nights (Figure 1(c)) is much similar to that with the cold days. In contrast, significant positive correlation exists between the AOI-OND and warm days in the following January, mainly registered over Northern Siberia, Northeast China, Korea, and Japan (Figure 1(b)). The situation of warm nights (Figure 1(d)) is analogous to that of warm days; however, the magnitude of the CC is relative larger. It means that East Asia would suffer more (less) extreme warm (cold) events in January when the preceding AO-OND is stronger, and vice versa. In addition, the CCs of the AOI-OND with the number of cold days, warm days, cold nights, and warm nights area-weighted averaged in the domain of (30°–45°N, 110°–150°E) during 1951–2013 January are −0.50, 0.29, −0.50, and 0.44, respectively (above 95% confidence level). Moreover, there are 43, 45 (38, 43), out of 63 years saw the AOI-OND and the frequency of following January cold days and cold nights (warm days and warm nights) over East Asia is to be out-phase (in-phase), which is about 68%, 71% (60%, 68%) of total events. Consequently, the AO-OND might be an important and useful predictor for the occurrence of temperature extremes in the following January over East Asia.

How does the AO-OND influence the following January temperature extremes over East Asia? To address this issue, we presented, in Figure 2(a), the regression of January 850-hPa wind and surface air temperature (SAT) with respect to the AOI-OND. When the AO-OND is stronger, obvious anomalous southerly wind emerges over East Asia in the following January (Figure 2(a): vectors). The southeasterly wind anomaly over East Asia opposes to the January mean northwesterly winds in this region and thus reduces cold air advection, which induces increasing of temperature (Figure 2(a): shaded). The weakening/increasing of January mean northerly wind/temperature is unfavourable (favourable) for the occurrence of cold (warm) events over East Asia (Zhou *et al.*, 2009; Wei *et al.*, 2011; Yan *et al.*, 2011).

The following question is in what process the signal of AO-OND can persist into the following January? The first reason might be the memory of autumn snow cover in Eurasia. As shown in Figure 2(b), October–December SAT is significantly warmer in Eurasia concurrent with positive phase of AO-OND. Such Eurasian warming might be caused by the temperature advection induced by the enhanced westerlies associated with the positive phase of AO, which could bring more warm marine air into the interior of the continents (Thompson and Wallace, 1998). As the surface gets warm, the snow cover over Eurasia decreases dramatically, especially in North Europe and central Siberia (Figure 2(c)). The snow cover anomaly has intrinsic climatic memory and therefore influences the subsequent climate (Li and Wang, 2014). On one hand, the snow melt in central Siberia could lead to a weakened Siberian High in the following January because of reduced surface albedo (Figure 4(a)). On the other hand, the surface warming in North Europe because of reduced surface albedo caused by snow melt might stimulate a Rossby wave train propagating eastwards across Eurasia. To demonstrate this issue, we presented the 500-hPa quasi-geostrophy streamfunction (QGSF: contours) and wave activity flux (WAF: vectors) anomalies in October–December (Figure 3(a)) and the following January (Figure 3(b)) with regard to AOI-OND, which was computed according to Plumb's formulation and describes the propagation of stationary Rossby waves (Plumb, 1985). Indeed, the 500-hPa QGSF anomalies in October–December resemble a Rossby wave train signal, which originates from the North Atlantic Ocean and North Europe. Significant 'negative', 'positive', 'negative', and 'positive' QGSF anomalies are located in the western Atlantic, western Europe, eastern Europe, and Siberia to East Asia, respectively (Figure 3(a): contour). The eastward propagation of the Rossby wave is more visible by inspecting the WAF, which propagates eastwards from the Ural Mountain region to Siberia (Figure 3(a): vectors). By the following January, the Rossby wave train is still obvious (Figure 3(b)). Such an interaction among waves and mean flow might restrain the blocking events around Ural region (Zhou *et al.*, 2009). Figure 3(c) shows the number of January blocking days following positive (red) and negative (blue) phase of AO-OND. The January Ural

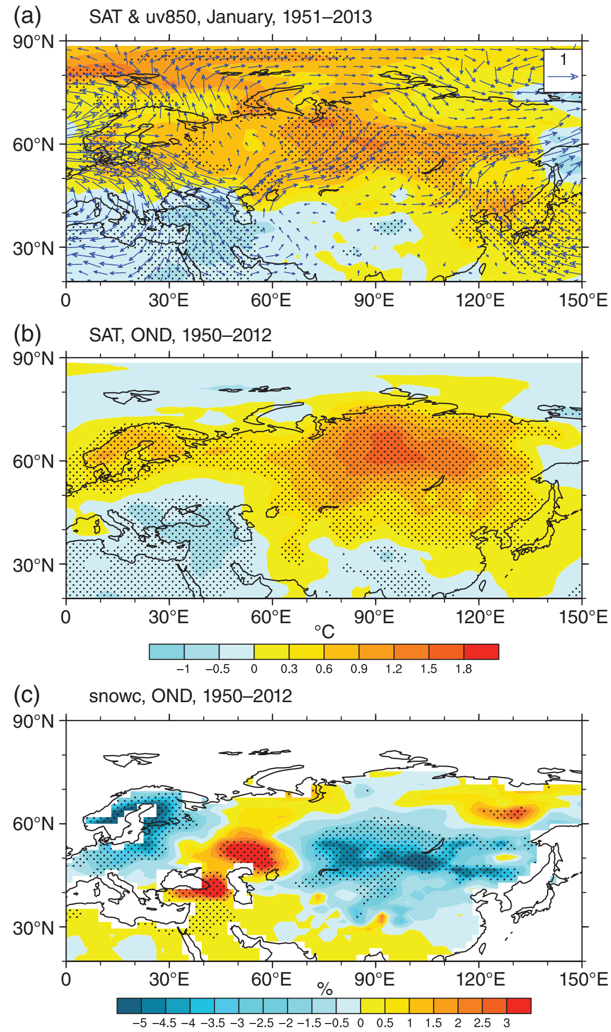

Figure 2. Regression maps of (a) SAT (shaded: °C) and 850-hPa wind (vectors: m s^{-1}) anomalies in 1951–2013 January, (b) SAT and (c) snow cover (shaded: %) anomalies in 1950–2012 October–December with regard to the AOI-OND. Dotted values indicate the SAT/snow cover anomalies significant at the 95% confidence levels based on the Student's *t*-test.

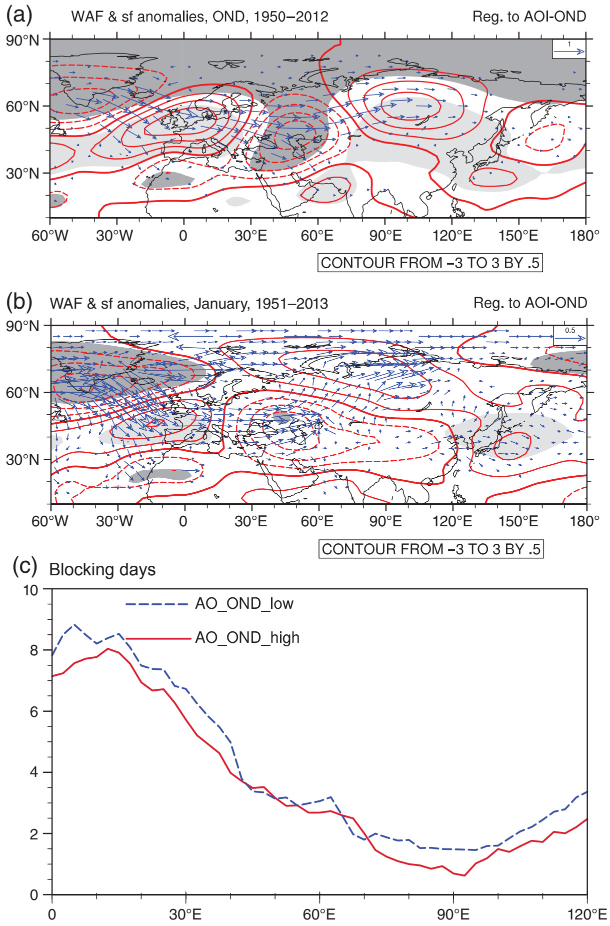

Figure 3. Regression maps of the 500-hPa quasi-geostrophy streamfunction (contours: 10^6 m^2 s^{-1}) and wave activity flux (vectors: scale in m^2 s^{-2}) in (a) October–December, (b) following January with regard to AOI-OND during 1950–2012. Shaded values are significant at the 95% confidence levels based on the Student's *t*-test. (c) January mean blocking frequency (days) following high (red curve) and low (blue curve) AO-OND years (*based on 0.8 standard deviation*).

blocking (50°–90°E) frequency after positive phase of AO-OND is relative less, which is favourable (unfavourable) for the occurrence of warm (cold) extremes over East Asia (Cheung *et al.*, 2012). Meanwhile, the decreasing of Ural blocking events would weaken the Siberian High (Takaya and Nakamura, 2005). This could generate anomalous southeasterly over East Asia which prevents cold air from invading East Asian continent (Figure 2(a): vectors).

On the other hand, the influence of AO-OND in January might come from its own intrinsic memory. The CC between the AO indices in October–December and January during 1950/51 to 2012/13 is about 0.48. Also the January sea level pressure anomalies regressed onto the AOI-OND resemble well the AO pattern (Figure 4(a)). The intrinsic memory of AO-OND might be because of its features in the stratosphere. As displayed in Figure 4(c), the AO-OND shows a barotropic structure, especially in the Arctic and North Pacific where the signal extends to the stratosphere. By the following January, the barotropic AO remains and the anomalous signal in stratosphere propagates downwards to troposphere (Figure 4(d)). To illustrate the interaction between the stratosphere and troposphere, we presented the zonally averaged zonal wind (blue contours), E-P flux cross sections (vectors) and its divergence (red contours) in October–December (Figure 4(e)) and the following January (Figure 4(f)) composited between high and low AOI-OND. In high AO-OND years, significant anomalous E-P flux propagates from stratosphere downwards to upper troposphere, then bents equatorwards to mid-latitudes (Figure 4(e): vectors). According the definition of E-P flux, the anomalous poleward (equatorward) momentum (heat) fluxes in associated with positive phase of AO-OND help to sustain a stronger circumpolar jet (colder polar vortex) (Hartmann *et al.*, 2000). Besides, according to the assumptions of quasi-geostrophic theory and linear perturbations (Andrews *et al.*, 1987, equation 3.5.5a), the zonal-mean zonal flow is decelerated (accelerated) where there is convergence (divergence) of the E-P flux. Consequently, the red contours of positive E-P flux divergence in stratosphere at high latitudes indicate that the eddy flux anomalies drive circumpolar westerly wind anomalies (Figure 4(e): blue contours). Interestingly, the anomalous planetary wave activity propagating downwards from stratosphere in October–December related to the simultaneous AOI-OND could well persist into the following January (Figure 4(f): vector). Such a mechanism is consistent with the results revealed by Cohen *et al.* (2007). Cohen *et al.* (2007) suggested that years with larger autumn snow extent over Eurasia feature strong wintertime (December) upward-propagating planetary wave activity in the lower stratosphere, leading to negative AO in January, which is the converse case of our study. Consequently, following a positive phase of AO-OND, the polar jet stream is stronger than normal in January (Figure 4(b) and (f): blue contours), which implies a positive phase of AO. Meanwhile, the weakening of East Asian jet stream (Figure 4(b)) would weaken the EAWM (Jhun and Lee, 2004; Li and Yang, 2010) and suppress the propagation of the synoptic-scale transient wave which is a main possible cause for winter temperature extremes over China (Chen *et al.*, 2013). In such a context, the incidence of extreme warm/cold events is more/less frequent.

4. Conclusion and discussion

Using the NCEP/NCAR reanalysis data, this study investigated the potential connection between the AO-OND and the following January temperature extremes (warm/cold days, warm/cold nights) over East Asia. It is revealed that the extreme cold (warm) events would be less (more) frequent in January over East Asia when the preceding AO-OND was in its positive phase; and vice versa. The possible reason is that obvious anomalous southerly wind emerges over East Asia in January when the AO-OND is stronger. Therefore, cold air advection from high latitude to East Asia is weakened, which induces increasing of temperature and is unfavourable (favourable) for the

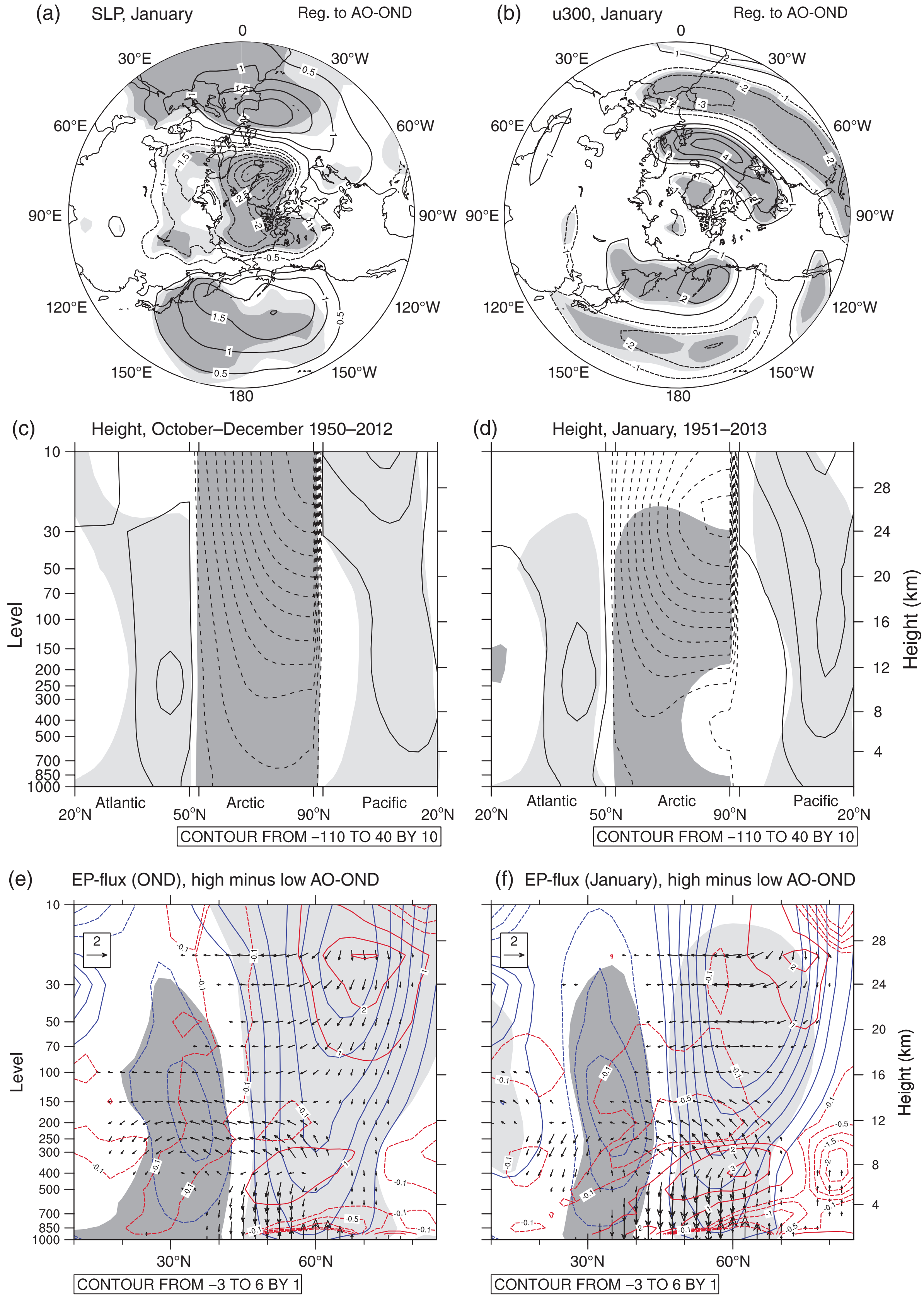

Figure 4. Regression maps of the (a) sea level pressure and (b) 300-hPa zonal wind anomalies in the following January with regard to AOI-OND during 1950–2012. Latitude cross-section for regression maps of (c) October–December and (d) January geopotential height anomalies averaged zonally in the Atlantic (20°–50°N, 90°W–0°), Arctic (60°–90°N, 0°–360°), and Pacific (20°–50°N, 130°E–150°W). Composition of zonally averaged zonal wind (blue contours, interval 1 m s^{-1}), EP flux cross-sections (vector, units: m^2 s^{-2}) and its divergence (red contours, units: m s^{-1} day^{-1}) between high and low AOI-OND (*based on 0.8 standard deviation*): (e) 1950–2012 October–December and (f) 1951–2013 January. Shaded regions (vectors) indicate the anomalies significant at 95% confidence level from a two-tailed Student's *t*-test.

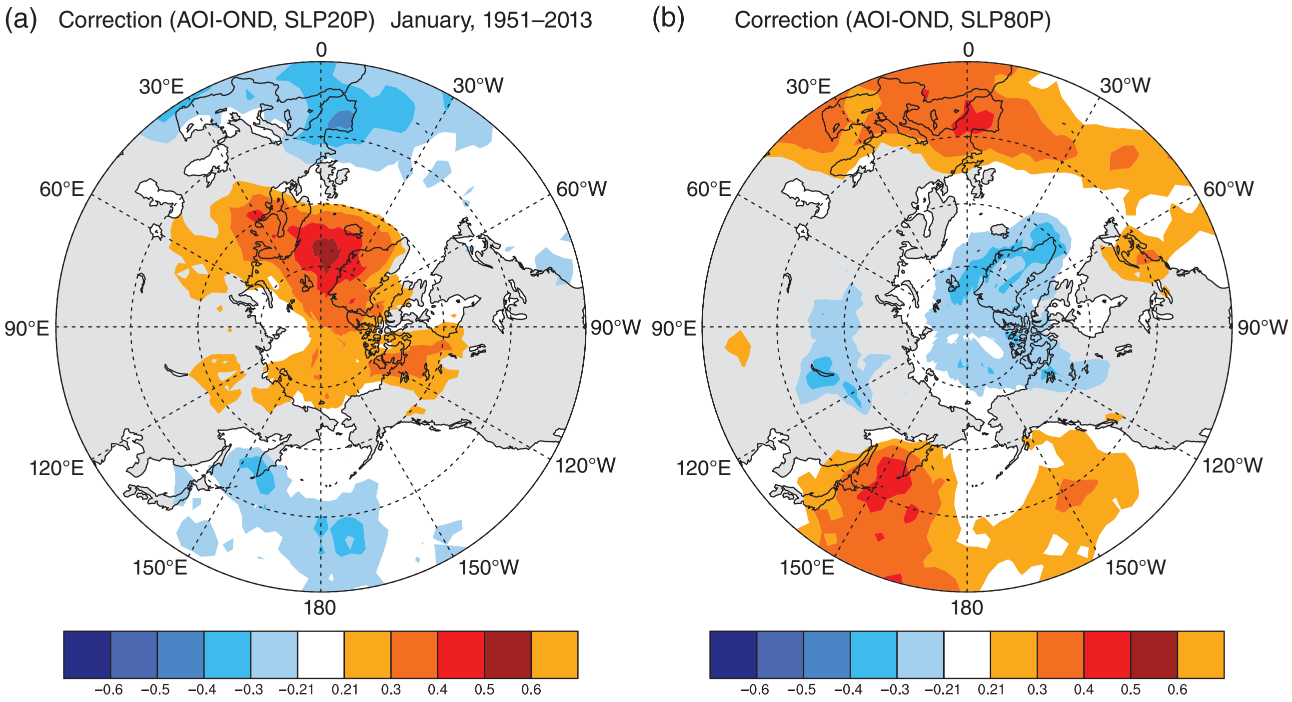

Figure 5. Correlation maps of AOI-OND with the following January (a) extreme low (*lower than 20th percentile value*) and (b) extreme high (*greater than 80th percentile value*) sea level pressure during 1951–2013, respectively. Shaded values are significant at 90% level from a two-tailed Student's *t*-test.

occurrence of cold (warm) events in this region (Zhou *et al.*, 2009; Wei *et al.*, 2011; Yan *et al.*, 2011).

Further analysis revealed that the persistent climatic influence of AO-OND might be because of its own persistence and the memory of Eurasian snow cover. The AO-OND shows a barotropic structure, especially in the Arctic and North Pacific where the signal extends to stratosphere. By the following January, wave activities propagate from stratosphere downwards to upper troposphere, then bent equatorwards to mid-latitudes. Consequently, the January circumpolar jet (polar vortex) gets strengthened (colder). Besides, anomalous E-P flux divergence in stratosphere persists from October–December to the following January, which drives westerly wind anomalies around the polar region. In such a way, the influence of AO-OND persists into the following January. On the other hand, significant surface warming covers most parts of Eurasia in October–December when the simultaneous AO is in positive phase, leading to Eurasian snow melt. As the snow cover anomaly has intrinsic climatic memory (Li and Wang, 2014), the January Siberian High weakens because of reduced surface albedo. Besides, the surface warming in North Europe because of reduced surface albedo caused by snow melt could stimulate a Rossby wave train propagating eastwards across Eurasia, which restrains the blocking events around Ural region. In such a context, the incidence of extreme warm/cold events is more/less frequent.

The above analysis mainly focused on the anomalies of monthly mean atmospheric circulation. The daily atmosphere circulation in January also shows close connection with the preceding AO-OND. As shown in Figure 5, the extreme lower/higher pressure anomalies are much more (less)/less (more) frequent over polar region (North Pacific and North Atlantic) in January following a positive phase of AO-OND. It means that the change in daily atmosphere is consistent with that of temperature extremes over East Asia. Thus, the AO-OND might be an important and useful predictor for the occurrence of temperature extremes in the following January over East Asia.

Acknowledgements

We appreciate NOAA for providing NCEP reanalysis datasets, which is available for free at (http://www.esrl.noaa.gov/psd/). This research was supported by the National Natural Science Foundation of China (Grant Nos. 41210007 and 41130103).

References

Andrews DG, Holton JR, Leovy CB. 1987. *Middle Atmosphere Dynamics*. Academic press: New York, NY.

Boyle JS, Chen T. 1987. *Monsoon Meteorology*. Oxford University Press: Oxford, UK.

Chang C, Wang Z, Hendon H. 2006. *The Asian Monsoon*. Springer Press: Berlin and Heidelberg, Germany.

Chen W, Huang R. 2002. The propagation and transport effect of planetary waves in the Northern Hemisphere winter. *Adv. Atmos. Sci.* **19**: 1113–1126.

Chen HS, Liu L, Zhu YJ. 2013. Possible linkage between winter extreme low temperature events over China and synoptic-scale transient wave activity. *Sci. China Earth Sci.* **56**: 1266–1280.

Cheung HN, Zhou W, Mok HY, Wu MC. 2012. Relationship between Ural-Siberian blocking and the East Asian winter monsoon in relation to the Arctic oscillation and the El Niño-Southern Oscillation. *J. Clim.* **25**: 4242–4257, doi: 10.1175/JCLI-D-11-00225.1.

Cohen J, Barlow M, Kushner PJ, Saito K. 2007. Stratosphere–troposphere coupling and links with Eurasian land surface variability. *J. Clim.* **20**: 5335–5343.

Compo GP, Whitaker JS, Sardeshmukh PD, Matsui N, Allan RJ, Yin X, Gleason BE, Vose RS, Rutledge G, Bessemoulin P, Brönnimann S, Brunet M, Crouthamel RI, Grant AN, Groisman PY, Jones PD, Kruk MC, Kruger AC, Marshall GJ, Maugeri M, Mok HY, Nordli Ø, Ross TF, Trigo RM, Wang XL, Woodruff SD, Worley SJ. 2011. The twentieth century reanalysis project. *Q. J. R. Meteorol. Soc.* **137**: 1–28.

Gong D, Wang S, Zhu J. 2001. East Asian winter monsoon and Arctic oscillation. *Geophys. Res. Lett.* **28**: 2073–2076, doi: 10.1029/2000GL012311.

Hartmann DL, Wallace JM, Limpasuvan V, Thompson DWJ, Holton JR. 2000. Can ozone depletion and global warming interact to produce rapid climate change? *Proc. Natl. Acad. Sci. U.S.A.* **97**(4): 1412–1417.

He SP. 2013. Reduction of the East Asian winter monsoon interannual variability after the mid-1980s and possible cause. *Chin. Sci. Bull.* **8**(12): 1331–1338, doi: 10.1007/s11434-012-5468-5.

He SP, Wang HJ. 2012. An integrated East Asian winter monsoon index and its interannual variability. *Chin. J. Atmos. Sci.* **36**(3): 523–538 (in Chinese).

He SP, Wang HJ. 2013. Impact of the November/December Arctic Oscillation on the following January temperature in East Asia. *J. Geophys. Res.* **118**: 12981–12998, doi: 10.1002/2013JD020525.

Jhun JG, Lee EJ. 2004. A new East Asian winter monsoon index and associated characteristics of the winter monsoon. *J. Clim.* **17**(4): 711–726.

Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen J. 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.* **77**(3): 437–471.

Kim HJ, Ahn JB. 2012. Possible impact of the autumnal North Pacific SST and November AO on the East Asian winter temperature. *J. Geophys. Res.* **117**: D12104, doi: 10.1029/2012JD017527.

Li F, Wang HJ. 2012. Predictability of the East Asian winter monsoon interannual variability as indicated by the DEMETER CGCMS. *Adv. Atmos. Sci.* **29**: 441–454, doi: 10.1007/s00376-011-1115-3.

Li F, Wang HJ. 2013a. Relationship between Bering Sea ice cover and East Asian winter monsoon year-to-year variations. *Adv. Atmos. Sci.* **30**: 48–56, doi: 10.1007/s00376-012-2071-2.

Li F, Wang HJ. 2013b. Autumn sea ice cover, winter Northern Hemisphere annular mode, and winter precipitation in Eurasia. *J. Clim.* **26**: 3968–3981.

Li F, Wang HJ. 2014. Autumn Eurasian snow depth, autumn Arctic sea ice cover and East Asian winter monsoon. *Int. J. Climatol.* **34**(13): 3616–3625, doi: 10.1002/joc.3936 s.

Li Y, Yang S. 2010. A dynamical index for the East Asian winter monsoon. *J. Clim.* **23**(15): 4255–4262.

Li F, Wang HJ, Gao YQ. 2014. On the strengthened relationship between East Asian winter monsoon and Arctic Oscillation: a comparison of 1950–1970 and 1983–2012. *J. Clim.* **27**(13): 5075–5091, doi: 10.1175/JCLI-D-13-00335.1.

Plumb RA. 1985. On the three-dimensional propagation of stationary waves. *J. Atmos. Sci.* **42**: 217–229.

Takaya K, Nakamura H. 2005. Mechanisms of intraseasonal amplification of the cold Siberian high. *J. Atmos. Sci.* **62**: 4423–4440.

Thompson DWJ, Wallace JM. 1998. The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.* **25**(9): 1297–1300, doi: 10.1029/98GL00950.

Tibaldi S, Molteni F. 1990. On the operational predictability of blocking. *Tellus* **42A**: 343–365.

Wang HJ, He SP. 2012. Weakening relationship between East Asian winter monsoon and ENSO after mid-1970s. *Chin. Sci. Bull.* **57**(27): 3535–3540, doi: 10.1007/s11434-012-5285-x.

Wang HJ, Yu ET, Yang S. 2011. An exceptionally heavy snowfall in Northeast china: large-scale circulation anomalies and hindcast of the NCAR WRF model. *Meteorol. Atmos. Phys.* **113**: 11–25, doi: 10.1007/s00703-011-0147-7.

Wei K, Chen W, Zhou W. 2011. Changes in the East Asian cold season since 2000. *Adv. Atmos. Sci.* **28**: 69–79, doi: 10.1007/s00376-010-9232-y.

Wu B, Wang J. 2002. Winter Arctic oscillation, Siberian high and East Asian winter monsoon. *Geophys. Res. Lett.* **29**(19): 1897, doi: 10.1029/2002GL015373.

Yan ZW, Xia JJ, Qian C, Zhou W. 2011. Changes in seasonal cycle and extremes in China during the period 1960–2008. *Adv. Atmos. Sci.* **28**: 269–283.

Zhou W, Li CY, Wang X. 2007. Possible connection between Pacific oceanic interdecadal pathway and East Asian winter monsoon. *Geophys. Res. Lett.* **34**: L01701, doi: 10.1029/2006GL027809.

Zhou W, Chan JCL, Chen W, Ling J, Pinto JG, Shao YP. 2009. Synoptic-scale controls of persistent low temperature and icy weather over Southern China in January 2008. *Mon. Weather Rev.* **137**: 3978–3991.

Zhou Q, Chen W, Zhou W. 2013. Solar cycle modulation of the ENSO impact on the winter climate of East Asia. *J. Geophys. Res. Atmos.* **118**: 5111–5119, doi: 10.1002/jgrd.50453.

