Enhanced Linkage between Eurasian Winter and Spring Dominant Modes of Atmospheric Interannual Variability since the Early 1990s

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ABSTRACT

The present study reveals a marked enhancement in the relationship between Eurasian winter and spring atmospheric interannual variability since the early 1990s. Specifically, the dominant mode of winter Eurasian 500-hPa geopotential height anomalies, with same-sign anomalies over southern Europe and East Asia and opposite-sign anomalies over north-central Eurasia, is largely maintained to the following spring after the early 1990s, but not before the early 1990s. The maintenance of the dominant atmospheric circulation anomaly pattern after the early 1990s is associated with a triple sea surface temperature (SST) anomaly pattern in the North Atlantic that is sustained from winter to the subsequent spring. This triple SST anomaly pattern triggers an atmospheric wave train over the North Atlantic through Eurasia during winter through spring. Atmospheric model experiments verify the role of the triple SST anomaly in maintaining the Eurasian atmospheric circulation anomalies. By contrast, before the early 1990s, marked SST anomalies related to the winter dominant mode only occur in the tropical North Atlantic during winter and they disappear during the following spring. The triple SST anomaly pattern after the early 1990s forms in response to a meridional atmospheric dipole over the North Atlantic induced by a La Niña–like cooling over tropical Pacific, and its maintenance into the following spring may be via a positive air–sea interaction process over the North Atlantic. Results of this analysis suggest a potential source for the seasonal prediction of the Eurasian spring climate.

1. Introduction

Atmospheric circulation variations play an important role in weather and climate anomalies over Eurasia (e.g., Ogi et al. 2003, 2005; Sun et al. 2008; Chen et al. 2016; Ye and Wu 2017). For example, Chen et al. (2016) reported that interannual variations in Eurasian surface air temperature (SAT) are primarily controlled by overlying atmospheric circulation anomalies via wind-induced temperature advection and cloud-induced surface shortwave radiation changes. They showed that the first empirical orthogonal function mode (EOF1) of interannual variations of spring Eurasian SAT has a close connection with the Arctic Oscillation (AO), the dominant atmospheric mode over extratropical Northern Hemisphere (Thompson and Wallace 1998). The second EOF mode (EOF2) of Eurasian spring SAT variations is associated with a wave train over the North Atlantic through Eurasia that is induced by the North Atlantic sea surface temperature (SST) changes (Chen et al. 2016). It was reported that the extreme heat waves that occurred over most parts of Eurasia in summer of 2003 were largely due to the atmospheric circulation anomaly pattern related to the positive AO (Beniston 2004; Stott et al. 2004; Ogi et al. 2005). The North Atlantic Oscillation (NAO)-related atmospheric circulation changes may exert substantial influences on climate and weather over Eurasia (e.g., Hurrell and van Loon 1997; Sun et al.
Studies indicated that atmospheric circulation change plays a crucial role in the snow variability over Eurasia (Wu and Chen 2016; Ye and Wu 2017; Ye and Lau 2017). There may be connections between the atmospheric circulation variability during different seasons (e.g., Ogi et al. 2003, 2004; Otomi et al. 2013; Choi et al. 2016; Sun 2017). Ogi et al. (2004) showed that the boreal winter AO has a significant positive correlation with the following summer AO. In particular, a positive (negative) summer AO phase tends to occur when preceding winter AO is in its positive (negative) phase. The mechanism responsible for the connection between winter AO and following summer AO is unclear (Ogi et al. 2004). As atmosphere variability does not have a long memory beyond one season, surface conditions (e.g., SST, snow cover, and Arctic sea ice) may play an important role in maintaining the atmospheric circulation anomalies through seasons. For example, Ogi et al. (2003) reported that SST and sea ice anomalies over the North Atlantic and snow cover anomalies over Eurasia play important roles in connecting the winter NAO-related atmospheric circulation and the subsequent summer atmospheric circulation over subtropical Northern Hemisphere. Through a case study, Otomi et al. (2013) indicated that the North Atlantic triple SST anomaly pattern plays an important role in switching the AO polarity from a large negative value in the winter of 2009/10 to a large positive value in the summer of 2010. Note that the large negative AO in the winter of 2009/10 led to the extreme cold weather over most parts of Eurasia (e.g., Wang and Chen 2010; Otomi et al. 2013). In contrast, the large positive AO in the summer of 2010 resulted in the record-breaking hot summer over the Northern Hemisphere, especially Russia and Japan (Barriopedro et al. 2011; Matsueda 2011; Otomi et al. 2013).

Sun (2017) showed that the dominant mode of Eurasian atmospheric circulation in boreal winter has a close connection with that in the subsequent spring during 1979–2014. However, the mechanisms responsible for this connection are unclear and remain to be unraveled. One goal of this study is to identify the factors for the pronounced linkage between winter and spring Eurasian dominant atmospheric circulation anomalies. Our analysis shows that the North Atlantic SST anomalies may play a key role in maintaining the Eurasian dominant mode of atmospheric interannual variation during winter and spring. Previous studies have demonstrated that the linkages between the SST anomalies in the North Atlantic and atmospheric circulation anomalies over the North Atlantic and Eurasia are unstable (e.g., Walter and Graf 2002; Wu et al. 2011; Chen et al. 2015a; Chen and Wu 2017). Chen et al. (2015a) indicated that the connection between winter tropical northern Atlantic SST and NAO has experienced pronounced interdecadal changes in the past. This indicates that the North Atlantic SST anomalies may not be always able to maintain their role in atmospheric circulation variability during winter through the following spring. Hence, another goal of this study is to unravel whether the relationship between the dominant modes of Eurasian atmospheric interannual variations during the winter and subsequent spring has experienced interdecadal changes in the past.

The rest of this study is structured as follows. Section 2 describes the datasets and methods employed in this paper. Section 3 provides evidences for the decadal changes in the connection between winter and spring Eurasian dominant atmospheric modes. Section 4 compares the temporal evolution of SST anomalies in the North Atlantic before and after the interdecadal change, and the role of the North Atlantic triple SST anomaly pattern in maintaining the atmospheric anomalies over Eurasia. Section 5 investigates the formation and maintenance of the North Atlantic triple SST anomaly pattern. Section 6 summarizes the primary results of this study.

### 2. Data and methods

This study employs monthly geopotential height, horizontal winds, air temperature, and surface heat flux as well as daily geopotential height provided by the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996). The NCEP–NCAR reanalysis dataset is available from 1948 to the present. Surface heat fluxes, including surface latent and sensible heat fluxes, surface shortwave and longwave radiation, have a spectral T62 Gaussian-grid horizontal resolution. The atmospheric variables on pressure levels have a $2.5^\circ \times 2.5^\circ$ horizontal resolution. The daily geopotential height data are used to calculate synoptic-scale eddy (also called storm track). This study also uses monthly mean atmospheric data provided by the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim) from 1979 to the present (Dee et al. 2011) with a horizontal resolution of $2.5^\circ \times 2.5^\circ$.

The monthly SST data used in this study are derived from the National Oceanic and Atmospheric Administration (NOAA) Extended Reconstructed SST, version 4 (ERSST.v4), dataset (Smith et al. 2008). The ERSST.v4 dataset covers the period from 1854 to the present and has a $2^\circ \times 2^\circ$ horizontal resolution. The present analysis also uses monthly surface air temperature...
provided by the University of Delaware (Matsuura and Willmott 2009), which has a horizontal resolution of 0.5° × 0.5° and is available from 1900 to 2014.

This study employs the wave activity flux put forward by Takaya and Nakamura (1997, 2001) to describe stationary Rossby wave propagation. This wave activity flux is generally parallel to local group velocity associated with a stationary wave train in the Wentzel–Kramers–Brillouin approximation (Takaya and Nakamura 1997, 2001). Furthermore, the wave activity flux is independent of the wave phase. The formula to calculate the wave activity flux is as follows:

\[
W = \frac{p}{2|U|} \left\{ U(u'^2 - \psi'v') + V(-u'v' + \psi'u') + U(-u'v' + \psi'u') + V(u'^2 - \psi'u') + \frac{f_o^2}{N^2} [U(u'T' - \psi'T_y') + V(-u'T' - \psi'T_y)] \right\}
\]

where \(\psi', v' = (u', v')\), and \(U = (U, V)\) represent perturbed geostrophic streamfunction, perturbed geostrophic winds, and mean winds, respectively. Here \(H_o, T', P, R_o, f_o, N\) are the scale height, perturbed air temperature, pressure standardized by 1000 hPa, gas constant related to dry air, the Coriolis parameter at 45°N, and the Brunt–Väisälä frequency, respectively. The subscripts \(x\) and \(y\) indicate derivative in the zonal and meridional directions, respectively. Climatological mean flow is calculated based on the period 1958–2014.

The Fisher’s r–z transformation was used to estimate the significance of the difference between two correlation coefficients. First, the two involved correlation coefficients \(r_1\) and \(r_2\) are subjected to a Fisher transform (Fisher 1921) as follows:

\[
z_1 = 0.5 \ln \left\{ \frac{1 + r_1}{1 - r_1} \right\} \quad \text{and} \quad z_2 = 0.5 \ln \left\{ \frac{1 + r_2}{1 - r_2} \right\}.
\]

Then, the standard parametric test was employed to examine the null hypothesis of the equality of the \(z_1\) and \(z_2\). The test statistic \(u = (z_1 - z_2)/[1/(N_1 - 3) + 1/(N_2 - 3)]^{1/2}\) satisfies the normal distribution. Both \(N_1\) and \(N_2\) are the sample sizes corresponding to \(r_1\) and \(r_2\). Finally, the statistical significance was estimated based on the two-tailed Student’s \(t\) test with the confidence level of 95% being used in the present study.

This study focuses on the variations on the interannual time scale. All variables are subjected to a 9-yr high-pass Lanczos filter to obtain their interannual components (Duchon 1979). Significances of the correlation coefficient and composite differences are estimated according to the two-tailed Student’s \(t\) test.

3. Interdecadal change in the linkage between winter and spring dominant atmospheric modes

An EOF method was employed to identify the dominant mode of 500-hPa geopotential height anomalies over the mid-to-high latitudes of Eurasia (40°–70°N, 0°–140°E) during 1959–2014. The anomalies were weighted by the cosine of the latitude to account for the decrease of area toward the North Pole in the EOF analysis. Figures 1a and 1c display the distribution of 500-hPa geopotential height anomalies corresponding to the leading mode (EOF1) during boreal winter |December (−1)–February(0) [D(−1)JF(0)], where (−1) denotes the prior year and (0) denotes the target year| and spring [MAM(0)], respectively, which is obtained by regression upon the corresponding principal component time series (PCI) (Figs. 1b,d). The EOF1 explains about 30.3% (26.7%) of the total variance of 500-hPa geopotential height in winter (spring), and is well separated from the other EOF modes according to the measure of North et al. (1982).

The spatial distribution of the geopotential height anomalies over Eurasia displays a close resemblance between winter and spring. Negative anomalies are observed over Europe and East Asia through the mid-latitude North Pacific and positive anomalies are present over north-central Eurasia through the Eurasian Arctic region (Figs. 1a and 1c, respectively). The pattern correlation coefficient is about 0.81 over the domain of 40°–70°N, 0°–140°E. In comparison, the center of negative anomalies over East Asia and the center of positive anomalies over central Eurasia are located more northwestward in spring compared to those in winter (Figs. 1a,c). This result is generally consistent with that of Sun (2017), whose analysis was based on the NCEP–NCAR reanalysis data during 1979–2014. Sun (2017) showed a significant connection between the winter and spring PCI time series. However, we found that the correlation coefficient between winter and spring PCs is only 0.2 during 1958–2014 (Figs. 1b and 1d, respectively). This implies that there exist interdecadal changes in the connection between the winter and spring dominant atmospheric modes over the mid-to-high latitudes of Eurasia.

The above interdecadal change in the linkage between winter and spring dominant modes is confirmed by a running correlation analysis. Figure 2 displays running correlation coefficients between winter and spring PCs with a 21-yr window. Years labeled in Fig. 2 denote the central year of the 21-yr window. Apparently, the correlation coefficients are small and insignificant before the early 1990s, but become large positive and significant after the early 1990s (Fig. 2). In the following, two 21-yr
periods before and after the early 1990s were selected according to the sliding correlation in Fig. 2 to investigate the interdecadal change. To avoid overlapping and the possible effect of the sliding window width, two periods, 1968–88 and 1994–2014, were selected. It should be mentioned that results obtained in this study are not sensitive to the reasonable change (forward or backward by several years) of the selected two periods. The correlation coefficients are −0.09 and 0.64 during 1968–88 and 1994–2014, respectively. According to the Fisher’s $r$–$z$ transformation, the difference of the correlation coefficient between the above two periods is significant at the 99% confidence level.

We have further compared the winter and spring PC1 time series. During 1968–88, there are 11 same sign and 10 opposite sign years. During 1994–2014, there are 17 same sign and 4 opposite sign years. These statistics are consistent with the difference of correlation between the two periods. These results indicate that the dominant circulation anomaly pattern over the mid-to-high latitudes of Eurasia in boreal winter is largely maintained into the following spring after the early 1990s, but not for the period before the early 1990s. Our analysis shows that this interdecadal change in the connection between Eurasian winter and spring dominant modes of the atmospheric variability around the early 1990s can also be captured by the ERA-Interim data (figures not shown).

In the following, the temporal evolution of anomalies before and after the early 1990s is compared to investigate the interdecadal change. Two 21-yr periods...
(1968–88 and 1994–2014) were selected according to the sliding correlation. Years with positive and negative winter PC1 time series within the above two 21-yr periods are selected according to the 0.5 standard deviation, which are presented in Table 1. There are six (nine) positive years and eight (six) negative years, respectively, during 1968–88 (1994–2014). Composite anomalies in positive and negative PC1 years are constructed during the above two periods separately. Figures 3 and 4 display the differences of composite 500-hPa geopotential height anomalies and Figs. 5 and 6 shows the differences of composite 850-hPa wind anomalies from October–December(−1) [OND(−1)] to MAM(0) between the positive and negative PC1 winter years during 1994–2014 and 1968–88, respectively.

During 1994–2014, a dipole anomaly pattern appears over the North Pacific sector, with positive anomalies over the midlatitude North Pacific and negative anomalies over the Aleutian Islands–mainland Alaska, which tend to be maintained from OND(−1) to February–April(0) [FMA(0)] (Figs. 3a–e). Over the North Atlantic sector, a triple anomaly pattern is visible from OND(−1) to D(−1)JF(0), with positive anomalies over the midlatitudes and negative anomalies over the subtropics and the high latitudes (Figs. 3a–c). With the weakening of the negative anomalies over the high latitudes, the triple pattern changes to a dipole pattern between the subtropics and the mid-to-high latitudes in January–March(0) [JFM(0)] (Fig. 3d). This meridional dipole pattern is maintained to subsequent spring (Figs. 3e,f). A wave pattern appears over Eurasia starting from D(−1)JF(0) and persists to MAM(0), although there are some changes in the location and amplitude of anomalies (Figs. 3e–f). This wave pattern resembles the dominant mode of Eurasian atmospheric variability in winter and spring (Figs. 1a and 1c, respectively).

During 1968–88, an east–west wave pattern is observed over the midlatitude North Pacific through North America to the North Atlantic from OND(−1) to D(−1)JF(0) (Figs. 4a–c). In OND(−1), there is a weak wave train over Eurasia, with negative anomalies over western Europe and East Asia–North Pacific and positive anomalies over north-central Eurasia (Fig. 4a). This wave train is enhanced and maintained to JFM(0), and then it is weakened notably in FMA(0) (Figs. 4b–e). In particular, negative anomalies over western Europe become extremely weak and insignificant in FMA(0) (Fig. 4e). In MAM(0), anomalies over the high-latitude Eurasia are featured by a north–south contrast (Fig. 4f), which is different from those in D(−1)JF(0) (Fig. 4e). This indicates that the atmospheric circulation anomalies related to the dominant mode in winter cannot be maintained to subsequent spring during 1968–88. It should be mentioned that spatial distributions of the 500-hPa geopotential height anomalies in D(−1)JF(0) over the mid-to-high latitudes of the Eurasia are generally similar during 1968–88 and 1994–2014. This implies that there is no obvious interdecadal change around the early 1990s regarding the leading modes of the Eurasian winter atmospheric interannual variability.

Composite 850-hPa wind anomalies show prominent differences between the two periods as well. During 1994–2014, cyclonic anomalies are observed over the subtropics and high latitudes of the North Atlantic and an anticyclonic anomaly is present over the midlatitudes from OND(−1) to D(−1)JF(0), with significant easterly wind anomalies around 35°N and westerly wind anomalies around 60°N (Figs. 5a–c). With the northward movement of the wind anomalies, a meridional dipole pattern prevails over the North Atlantic in JFM(0) (Fig. 5d). This meridional dipole pattern is maintained to subsequent spring (Figs. 5d–f). During 1968–88, an east–west wave train is observed over the North Atlantic from OND(−1) to D(−1)JF(0), with cyclonic anomalies over coastal North America and western Europe and an anticyclonic anomaly over the midlatitudes North Atlantic (Figs. 6a–c). This east–west pattern is in contrast with the meridional dipole pattern during 1994–2014 (Figs. 5a–c). In JFM(0) and FMA(0), wind anomalies over the North Atlantic are relatively weak (Figs. 6d, e). In MAM(0), a dipole pattern forms over the North Atlantic, with a cyclonic anomaly around 40°N and anticyclonic anomaly around 70°N (Fig. 6f).

The above results suggest that the temporal evolution of the atmospheric circulation anomalies over the North Atlantic and Eurasia displays substantial differences between the two periods. During 1994–2014, the North

**Table 1.** List of positive and negative years of the PC1 time series corresponding to EOF1 of winter 500-hPa geopotential height anomalies during 1968–88 and 1994–2014. Positive and negative years are selected according to a 0.5 standard deviation.

|------|---------|-----------|
Atlantic is dominated by a south–north pattern, whereas, during 1968–88, an east–west wave pattern prevails until winter and switches to a south–north pattern in spring over the North Atlantic. Over Eurasia, the zonal wave pattern is maintained from winter to the subsequent spring during 1994–2014, but the wave pattern weakens after winter during 1968–88.

4. Factors for the maintenance of the Eurasian atmospheric circulation anomaly pattern

This section analyzes why the atmospheric wave pattern over Eurasia during winter is maintained to the subsequent spring after the early 1990s, but not before the early 1990s. It is known that atmosphere variability does not have a long memory beyond one season. This indicates that surface conditions (e.g., SST anomalies) may play an important role in sustaining the atmospheric circulation anomaly pattern through seasons. Studies have found that SST forcing is crucial for cross-season linkages and may provide important sources for climate predictability because of the heat capacity of the ocean (e.g., Frankignoul and Hasselmann 1977; Watanabe and Kimoto 2000a; Deser et al. 2003; Wu et al. 2014). We contrast the SST anomalies before and after the early 1990s to find out whether the maintenance of the dominant atmospheric circulation anomaly pattern is related to the role of the SST forcing. Figures 7 and 8 display differences of composite SST anomalies between positive and negative years of D(−1)JF(0) PC1 during 1994–2014. Stippling denotes differences significantly different from zero at the 95% confidence level.
During 1994–2014, a La Niña–like SST anomaly pattern is present in the equatorial central-eastern Pacific from OND(−1) to D(−1)JF(0) (Figs. 7a–c). The SST anomalies weaken with the center shifting to the equatorial central Pacific in the following seasons (Figs. 7d–f). A tripolelike SST anomaly pattern is observed in the extratropical North Atlantic from OND(−1) to D(−1)JF(0) (Figs. 7a–c). It should be mentioned that the triple SST anomaly pattern related to negative SST anomalies in the subtropics and positive SST anomalies in the midlatitudes that are relatively weak in OND(−1) (Fig. 7a), and then increase and become significant in D(−1)JF(0) (Fig. 7c). This indicates that the triple SST anomaly pattern in the North Atlantic Ocean was not fully formed until D(−1)JF(0) (Fig. 7c). The triple SST anomaly pattern is maintained from the winter to the subsequent spring with slightly northward movement, especially, the positive SST anomalies in the midlatitudes (Figs. 7d–f).

During 1968–88, SST anomalies in the equatorial central-eastern Pacific are weak negative from OND(−1) to JFM(0) (Figs. 8a–d), indicating a weak connection of winter dominant mode with preceding ENSO variability. The magnitude of these SST anomalies increases somewhat in the following seasons (Figs. 8e,f). La Niña–like SST anomalies are observed in the tropical central-eastern Pacific in MAM(0) (Fig. 8f). A triple SST anomaly pattern in the North Pacific persists from OND(−1) to MAM(0), with negative anomalies around 40°N and positive anomalies in the western subtropics and along the west coast of North America. There are positive SST anomalies in the tropical North Atlantic and negative SST anomalies along the east coast of North America from OND(−1) to JFM(0) (Figs. 8a–d).
The SST anomalies become weak and insignificant in the whole North Atlantic region from FMA(0) to MAM(0) (Figs. 8e,f).

The above analysis shows a substantial difference in the evolution of SST anomalies in the Pacific and North Atlantic between the two periods. During 1994–2014, a La Niña–like SST anomaly pattern prevails in the tropical Pacific during the preceding fall and winter, and a notable triple SST anomaly pattern persists in the North Atlantic from winter to the subsequent spring. In contrast, during 1968–88, significant SST anomalies only occur in the tropical North Atlantic from the preceding fall to winter. The SST anomalies are very weak in spring in the whole North Atlantic. The SST anomalies in the equatorial central-eastern Pacific remain small, although with some increase in their magnitude from winter to spring.

The differences of SST anomalies in the equatorial central-eastern Pacific, the tropics, and midlatitude of the North Atlantic are further illustrated in Fig. 9, which compares the temporal evolution of the SST anomalies in the Niño-3.4 region (5°S–5°N, 120°–170°W), in the northern tropical Atlantic region (NTA; 5°–25°N, 15°–60°W), and in the midlatitude North Atlantic region (35°–47°N, 25°–60°W) during the two periods. During 1994–2014, the Niño-3.4 SST anomalies are largely negative in winter and weaken thereafter (Fig. 9a). In contrast, during 1968–88, the Niño-3.4 SST anomalies...
are slightly negative in winter and increase somewhat thereafter. The NTA SST anomalies are significant positive in fall and winter, but become small in spring during 1968–88 (Fig. 9b). The NTA SST anomalies are weak during 1994–2014. Furthermore, significant and positive SST anomalies are found in the midlatitude North Atlantic from winter [i.e., D(-1)JF(0)] to the subsequent spring [i.e., MAM(0)] during 1994–2014 (Fig. 9c), corresponding well to the formation of the triple SST anomaly pattern in D(-1)JF(0) and its persistence to the subsequent MAM(0) as identified in Fig. 7. In comparison, the midlatitude North Atlantic SST anomalies are generally weak and insignificant during 1968–88 (Fig. 9c).

The contrast in the temporal evolution of the North Atlantic SST anomalies between the two periods corresponds well to that of the atmospheric circulation anomaly pattern over the North Atlantic and Eurasia. Previous studies have shown that the triple SST anomaly pattern in the North Atlantic may stimulate an eastward-propagating wave train from the North Atlantic to East Asia (e.g., Wu et al. 2009; Wu et al. 2010; Chen et al. 2016). This implies that the maintenance of the atmospheric dipole pattern over the North Atlantic and the atmospheric wave train over the mid-to-high-latitude Eurasia from winter to the subsequent spring during 1994–2014 is likely related to the persistence of the triple SST anomaly pattern in the North Atlantic. In comparison, the SST anomalies in the North Atlantic are generally weak in spring during 1968–88. As such, the atmospheric wave train over Eurasia in winter cannot persist to the following spring. In the following, we
present evidence to demonstrate the important role of the North Atlantic triple SST anomaly pattern in maintaining the atmospheric wave train over Eurasia during 1994–2014. Figure 10 displays the difference of anomalies of composite 200-hPa wave activity flux and streamfunction from D(−1)JF(0) to MAM(0) between positive and negative PC1 years during 1994–2014. An obvious wave train is observed to extend from the North Atlantic first eastward to western Europe, then northeastward to central Eurasian continent, and finally southeastward to East Asia (Fig. 10a). A source of the wave train is located over the subtropical North Atlantic (Fig. 10a), which may be related to anomalous convergence related to the upper anomalous cyclone (not shown). Previous studies have demonstrated that divergence (or convergence) anomalies in the upper troposphere act as sources of Rossby wave activity (Sardeshmukh and Hoskins 1988; Watanabe 2004). Note that the maintenance of anomalous circulation over the North Atlantic may be related to the North Atlantic triple SST anomaly pattern, which will be discussed later. The anomalous wave train over the North Atlantic through Eurasia persists from winter to the following spring, although the wave activity shifts slightly northward over the North Atlantic and weakens over the Eurasian continent in spring (Figs. 10b–d). Hence, the atmospheric wave train originating from the North Atlantic may play an important role in maintaining the atmospheric circulation anomaly pattern over Eurasia.

We perform numerical experiments with an atmospheric general circulation model (AGCM) to validate the role of the North Atlantic triple SST anomaly in maintaining the atmospheric wave train over Eurasia. Two numerical experiments are conducted: one control experiment (EXP_CLIM) and one sensitivity experiment (EXP_NATSST). The atmospheric model used is ECHAM5 (Roeckner et al. 2003), with horizontal
resolution of spectral T63 and 19 vertical levels (T63L19). In EXP_CLIM, the model is forced by global climatological SST with seasonal cycle. In EXP_NATSST, the model is driven by the global climatological SST plus the SST anomalies in the North Atlantic (10°–60°N, 15°–60°W) during spring (Fig. 7f). More details for the model experiments are provided in Table 2. These two experiments are integrated for 30 years and the last 20 years are used in the analysis. The differences between the two experiments represent the response of the model to the imposed SST anomalies. The difference of spring 500-hPa geopotential height between EXP_NATSST and EXP_CLIM is displayed in Fig. 11.

Negative geopotential height anomalies are seen over Europe and East Asia, and positive geopotential height anomalies extend from north-central Eurasia northward to the Arctic region (Fig. 11). Hence, the distribution in 500-hPa geopotential height response over the mid-to-high-latitude Eurasian continent simulated by the AGCM is generally consistent with that in the observations (Fig. 3). This verifies the role of the North Atlantic triple SST anomaly in maintaining the anomalous atmospheric circulation over the Eurasia from winter to the subsequent spring.

It should be mentioned that the geopotential height response over the North Atlantic region shows some differences between the AGCM simulation (Fig. 11) and the observation (Figs. 3 and 10). For example, positive geopotential height anomalies over the northern North Atlantic in the AGCM experiment (Fig. 11) shift southeastward compared to those in the observation (Figs. 3 and 10). Negative geopotential height anomalies over the subtropical central North Atlantic shift slightly southward in the AGCM than those in the observation (Figs. 3, 10, and 11). The reasons for the above differences remain unclear to us. One plausible reason may be because the AGCM experiments cannot capture the coupling processes between atmospheric circulation changes and the underlying SST anomalies over the North Atlantic where there are complicated nonlinear

Fig. 8. As in Fig. 7, but for differences of composite SST anomalies during 1968–88.
interaction processes between the jets and eddies (e.g., Kushnir et al. 2002). Another plausible reason may be that the impacts of other factors (such as the Pacific SST anomalies, Arctic sea ice, or the background atmospheric circulation, etc.) may contribute to the observed atmospheric circulation anomalies over the North Atlantic. Nevertheless, the atmospheric circulation anomalies over the mid-to-high latitudes of Eurasia in response to the North Atlantic SST forcing bear a resemblance to those in the observations (Figs. 10 and 11).

5. Factors for the formation and maintenance of the North Atlantic triple SST anomaly pattern

The previous section showed the role of the North Atlantic triple SST anomaly pattern in maintaining the atmospheric circulation anomaly pattern over Eurasia from winter to spring during 1994–2014. A question is what contributes to the formation and maintenance of the triple SST anomaly pattern in the North Atlantic. According to previous studies, this may be related to the air–sea interaction process in the North Atlantic region. Previous studies have demonstrated that the triple SST anomaly pattern in the North Atlantic has a positive interaction with the overlying meridional dipole atmospheric pattern (Frankignoul et al. 1998; Watanabe and Kimoto 2000b; Marshall et al. 2001; Czaja and Frankignoul 2002; Czaja et al. 2002; Cassou et al. 2004; Pan 2005). On one hand, the meridional dipole atmospheric circulation anomaly may cause the formation of the triple SST anomaly pattern via changes in surface heat fluxes, as demonstrated by previous studies (e.g., Marshall et al. 2001; Czaja et al. 2002; Cassou et al. 2004; Wu and Liu 2005). In particular, it is generally indicated that the atmosphere–ocean coupling over the North Atlantic is most significant when the atmosphere circulation changes leading the triple SST anomaly by several months (e.g., Deser and Timlin 1997; Wu and Liu 2005; Yang and Wu 2015). On the other hand, the triple SST anomaly pattern, in turn, may induce a meridional dipole atmospheric anomaly via eddy feedback processes (Czaja and Frankignoul 2002; Cassou et al. 2004; Pan 2005).

During 1994–2014, obvious meridional dipole atmospheric circulation anomalies are seen over the North Atlantic Ocean (Fig. 5). Accordingly, triple SST anomaly patterns are apparent from the preceding winter to the subsequent spring (Fig. 6). This corresponding relation is generally consistent with previous findings (e.g., Czaja et al. 2002; Cassou et al. 2004; Pan 2005). In comparison, during 1968–88, the North Atlantic is dominated by an east–west wave train from OND(−1) to D(−1)JF(0) (Figs. 6a–c). Furthermore, during JFM(0) and FMA(0), atmospheric circulation anomalies are generally weak and insignificant over the North Atlantic Ocean (Figs. 6d,e). These atmospheric circulation anomalies during 1968–88 are not favorable for formation of the triple SST anomaly pattern and its maintenance (Fig. 8). This indicates that differences in the SST anomalies over the North Atlantic Ocean are closely related to the overlying atmospheric circulation changes. As the North Atlantic triple SST anomaly pattern during 1994–2014 is important for maintaining the atmospheric circulation anomaly pattern over Eurasia
from winter to spring, in the following, the formation
and maintenance of the triple SST anomaly pattern were
further examined via analyzing surface heat flux
changes. Figure 12 displays surface net heat flux anom-
aliies over the North Atlantic from OND(−1) to
MAM(0) during 1994–2014. We have also calculated
anomalies of surface latent and sensible heat flux as well
as surface shortwave and longwave radiation (figures
not shown). It is found that changes in surface net heat
flux over the North Atlantic are dominated by surface
latent heat flux changes (not shown). Note that values of
surface heat fluxes are taken to be positive (negative)
when their directions are downward (upward), which
contribute to ocean warming (cooling).

Surface net heat flux anomalies in OND(−1), ND
(−1)J(0), and D(−1)JF(0) are characterized by a triple
pattern over the North Atlantic, with positive anomalies
over the midlatitudes and negative anomalies over the
subtropics and high latitudes (Figs. 12a–c). This indicates a
contribution of surface net heat flux changes to the for-
mation of the triple SST anomaly pattern before and
during D(−1)JF(0) for the period 1994–2014 (Figs. 7a–c
and 9c). In particular, negative surface net heat flux
anomalies over the subtropical eastern North Atlantic
around 20°N from OND(−1) to D(−1)JF(0) (Figs. 12a–c)
contribute to the formation of the negative SST anomalies
in D(−1)JF(0) (Fig. 7c). In addition, large positive surface
net heat flux anomalies over the midlatitudes of the North
Atlantic between 35° and 55°N in D(−1)JF(0) (Fig. 12c)
also contribute to the generation of the positive SST
anomalies in D(−1)JF(0) (Fig. 7c). Besides surface net
heat fluxes, the oceanic heat transport processes may also
play a role in forming the North Atlantic triple SST
anomaly pattern in D(−1)JF(0). For example, the easterly
wind anomalies around 30°–40°N may induce anomalous
northward Ekman currents, contributing to the SST

![FIG. 10. Differences of composite anomalies of 500-hPa wave activity flux (vectors; m² s⁻²) and streamfunction (shading; 10⁵ m² s⁻²) in (a) D(−1)JF(0), (b) JFM(0), (c) FMA(0), and (d) MAM(0) between positive and negative years of D(−1)JF(0) PC1 during 1994–2014.]

![TABLE 2. Details of the AGCM numerical experiments.]

<table>
<thead>
<tr>
<th>Expt name</th>
<th>SST boundary condition</th>
<th>Integrated period</th>
</tr>
</thead>
<tbody>
<tr>
<td>EXP_CLIM</td>
<td>Climatological global SST during 1979–2008 with seasonal cycle.</td>
<td>30 yr</td>
</tr>
<tr>
<td>EXP_NATSST</td>
<td>As with EXP_CLIM, but SST anomalies in the North Atlantic (10°–60°N, 15°–60°W) during spring are added to the climatology.</td>
<td>30 yr</td>
</tr>
</tbody>
</table>
The above analysis shows that the formation of the North Atlantic triple SST anomaly pattern in winter \([D(-1)JF(0)]\) is associated with changes in surface net heat fluxes. After the triple SST anomaly pattern is formed in winter, it may persist into the following spring because of the large heat capacity of the ocean and through its coupling with the NAO-like meridional atmospheric circulation anomalies. As demonstrated by previous studies (Peng et al. 2003; Cassou et al. 2004; Pan 2005), the triple SST anomaly pattern may contribute to the maintenance of the NAO-like meridional dipole atmospheric circulation anomalies over the North Atlantic via synoptic-scale eddy feedback processes (also called storm track). In the following, we further address the feedback of the triple SST anomaly pattern to the atmospheric circulation changes via analyzing the synoptic-scale eddy activity. Figure 14 displays a latitude–time cross section along 30°–60°W of differences of anomalies of the composite 200-hPa storm track and zonal winds between positive and negative years of winter PC1 during 1994–2014. The storm-track activity is defined as the root-mean-square of the 2–8-day bandpass-filtered geopotential height, following previous studies (e.g., Lee et al. 2012; Chen et al. 2015b).

Pronounced negative anomalies of the storm track are observed around 40°N from OND\((-1)\) and D\((-1)JF(0)\), and shift slightly northward around JFM\((0)\) (Fig. 14a). This northward shift is consistent with the northward movement of the anomalous cyclone over the subtropics and anomalous anticyclone over the midlatitudes in JFM\((0)\) (Figs. 3d and 5d). Weakening of the storm track may be related to the reduction in the meridional SST gradient (Fig. 7). Studies have reported that the reduced meridional temperature gradient may suppress the baroclinic eddy growth (e.g., Lindzen and Farrell 1980; Hoskins and Valdes 1990; Nakamura and Shimo 2004), and hence, weaken the storm-track activity. In addition, as demonstrated by previous studies (Lau 1988; Cai et al. 2007; Chen et al. 2014), weakening of the storm track is accompanied by positive geopotential height tendency to its north and negative geopotential height tendency to its south, as well as weakening of the westerly jet stream (Fig. 14b). Through this process, the meridional dipole atmospheric anomaly pattern with an anomalous cyclone to the south and an anomalous anticyclone to the north persists to the following MAM\((0)\) (Figs. 3 and 5). This confirms the feedback of the triple SST anomaly pattern on the meridional dipole atmospheric anomaly pattern over the North Atlantic.

The above analysis indicates that there exist air–sea interactions over the North Atlantic during 1994–2014. On one hand, atmospheric wind anomalies contribute to the triple SST anomalies through surface heat fluxes. On
the other hand, the SST anomaly distribution modifies the meridional temperature gradient and the storm-track activity that in turn modifies the wind pattern. Thus, the maintenance of the triple SST anomaly from winter to spring is associated with the air–sea interaction process in the North Atlantic region. A question that still needs to be addressed is what contributes to the formation of the North Atlantic triple SST anomaly pattern in winter during 1994–2014.

According to the temporal evolution of SST anomalies in Fig. 7, there is an obvious precursory signal in the equatorial central-eastern Pacific SST before the winter triple SST anomaly pattern during 1994–2014. This is further illustrated in Fig. 15, which compares the correlation of the Niño-3.4 SST from OND(−1) to MAM(0) with an index of the North Atlantic triple SST anomalies in winter during 1968–88 and 1994–2014. The North Atlantic triple SST index (NATI) is defined as follows:

\[
\text{NATI} = \frac{\text{SST}_B - (\text{SST}_A + \text{SST}_C)}{2},
\]

where SST$_B$, SST$_A$, and SST$_C$ represent normalized winter SST anomalies averaged over 35°–47°N, 25°–60°W; 55°–65°N, 25°–60°W; and 20°–30°N, 10°–35°W, respectively. Selections of these regions are based on the spatial distribution of SST anomalies in Fig. 7c. A positive NATI corresponds to positive SST anomalies in the mid-latitudes, and negative SST anomalies in the subtropics and high latitudes of the North Atlantic (Fig. 7c).
During 1994–2014, the Niño-3.4 SST is correlated significantly with the winter NATI during fall through spring (Fig. 15a). The time lead suggests that there may be an influence of preceding ENSO on the triple SST anomalies. The persistence of the significant correlation may be due to the persistence of the ENSO SST anomalies in the tropical central-eastern Pacific. In contrast, the correlation is weak from the preceding fall to the subsequent spring during 1968–88. This indicates that the triple SST anomaly pattern is not related to ENSO before the early 1990s.

Previous studies have indicated that preceding ENSO-related SST changes in the tropical central-eastern Pacific may induce a triple SST anomaly pattern in the North Atlantic one to two seasons later through anomalous atmospheric teleconnections (e.g., Curtis and Hastenrath 1995; Klein et al. 1999; Saravanan and Chang 2000; Alexander et al. 2002; Czaja et al. 2002; Huang et al. 2004; Wu and Zhang 2010; Chen et al. 2015a). To demonstrate the role of preceding ENSO in the formation of the triple SST anomalies in the North Atlantic during 1994–2014, we display SST and 850-hPa winds anomalies in the preceding fall and winter obtained by regression on the normalized fall [OND (−1)] Niño-3.4 SST in Fig. 16. To facilitate comparison, the regression coefficients in Fig. 16 are multiplied by −1 so that the anomalies correspond to La Niña events.

![Fig. 13. Latitude–time cross section (20°–80°W averaged) of differences of composite (a) SST (°C) and (b) surface net heat flux (W m⁻²) between positive and negative years of D(−1)JF(0) PC1 during 1994–2014. Stippling denotes anomalies significantly different from zero at the 95% confidence level.](image)

![Fig. 14. Latitude–time cross section (30°–60°W averaged) of differences of composite 200-hPa (a) storm track (m) and (b) zonal wind anomalies (m s⁻¹) between positive and negative years of D(−1)JF(0) PC1 during 1994–2014. Stippling denotes anomalies significantly different from zero at the 95% confidence level.](image)

![Fig. 15. Correlation coefficients of D(−1)JF(0) NATI with the Niño-3.4 index from OND(−1) to MAM(0) during 1968–88 and 1994–2014. The horizontal solid line indicates correlation coefficient significant at the 95% confidence level. Definition of the NATI is provided in the text.](image)
La Niña–like SST anomalies are clearly present in the tropical Pacific from fall to winter (Figs. 16a,b). The North Atlantic is dominated by a horseshoe-like SST anomaly in fall (Fig. 16a), with significant positive SST anomalies in the midlatitudes and negative anomalies along the west coast of North Africa and western Europe. A triple SST anomaly pattern develops in the North Atlantic in winter (Fig. 16b), similar to that shown in Fig. 7. An atmospheric teleconnection is found from the North Pacific through North America to the North Atlantic in fall and winter, although the anomalies in fall are relatively weak (Figs. 16c,d). In particular, over the North Atlantic, an anticyclonic anomaly is observed over the midlatitudes, and cyclonic anomalies are found over the subtropics and high latitudes, which are similar to the distributions of wind anomalies in Figs. 5a–c. As shown in the previous section, such wind anomaly distribution favors the formation of the triple SST anomalies via surface heat flux changes. This indicates that the preceding ENSO plays an important role in the formation of the triple SST anomaly pattern over the North Atlantic via inducing atmospheric teleconnection. This is generally consistent with previous studies (Alexander et al. 2002; Czaja et al. 2002; Huang et al. 2004; Chen et al. 2015a).

6. Summary and discussion

The present study investigates the connection between the dominant mode of interannual variations of Eurasian atmospheric circulation between boreal winter and the subsequent spring. The distribution of atmospheric anomalies over the mid-to-high latitudes of Eurasia associated with the leading mode of 500-hPa geopotential height anomalies displays a close resemblance between winter and spring. The leading mode features same-sign height anomalies over western Europe and East Asia and opposite-sign height anomalies over north-central Eurasia in both winter and spring. However, the correlation coefficient between the winter and spring PCI time series shows a notable interdecadal change around the early 1990s. This indicates an interdecadal change in the linkage between the winter and spring dominant atmospheric circulation anomaly pattern. This change implies that the dominant atmospheric circulation anomaly pattern tends to be...
maintained from winter to the following spring after the early 1990s, but not so before the early 1990s.

A comparative analysis shows the role of the North Atlantic SST anomalies in the above interdecadal change. During 1994–2014, a triple SST anomaly pattern in relation to the winter dominant atmospheric mode, with same-sign anomalies in the subtropics and the high latitudes and opposite-sign anomalies in the mid-latitudes, starts to appear in fall and is sustained to the following spring. This triple SST anomaly pattern triggers an eastward-propagating atmospheric wave train over the North Atlantic through mid-to-high-latitude Eurasia. Because of the persistence of the triple SST anomalies, the atmospheric circulation anomaly pattern is maintained over Eurasia from winter to the following spring, which accounts for the connection between the winter and spring dominant atmospheric mode during 1994–2014. By contrast, during 1968–88, the winter dominant atmospheric mode-related SST anomalies are weak in the extratropical North Atlantic during fall through spring. The SST anomalies in the tropical North Atlantic are significant in winter, but become weak in spring. Correspondingly, the atmospheric circulation anomaly pattern over mid-to-high-latitude Eurasia in winter cannot be maintained to the following spring during 1968–88.

Further analysis suggests that the formation of the North Atlantic triple SST anomaly pattern during 1994–2014 is related to surface net heat flux change induced by a meridional dipole atmospheric anomaly pattern over the North Atlantic. This atmospheric anomaly pattern is likely related to a La Niña–like SST cooling in the tropical Pacific. The maintenance of the triple SST anomaly pattern from winter to spring is associated with its interaction with overlying atmospheric circulation changes. After the triple SST anomalies are induced, they have feedback on the atmospheric circulation through modulating the meridional temperature gradient and the storm-track activity. The modified atmospheric wind changes are in favor of the maintenance of the triple SST anomalies.

The Eurasian climate in spring may be connected to that in the following summer. Thus, revealing the source of climate variability and predictability in spring would be of help for understanding summer climate anomalies. This study highlights the role of the North Atlantic triple SST anomaly pattern in sustaining the atmospheric anomalies over the mid-to-high-latitude Eurasia continent from boreal winter to the following spring after the early 1990s. This may provide a potential source of the seasonal predictability of the Eurasian spring climate.

It should be mentioned that the formation of the winter North Atlantic triple SST anomaly pattern may also be related to the extratropical atmospheric internal disturbance (e.g., Wu and Liu 2002; Huang et al. 2004; Huang and Shukla 2005; Chen et al. 2015a). Further investigation is needed to explore the relative contributions of ENSO and internal atmospheric variability to the triple SST anomalies. In addition, comparison of Figs. 7 and 8 indicates substantial differences in the evolution of SST anomalies in the midlatitudes of the North Pacific. Whether these differences in SST anomalies in the midlatitudes of the North Pacific have a contribution to the differences in the atmospheric circulation anomalies over the North Atlantic and Eurasia remain to be explored. Furthermore, several previous studies have demonstrated that the Arctic sea ice variation may have an influence on the Eurasian atmospheric circulation anomalies (Honda et al. 2009; Liu et al. 2012; Kug et al. 2015). The possible role of the Arctic sea ice variations in contributing to the interdecadal change in the connection between Eurasian winter and spring dominant modes of atmospheric variability around the early 1990s is still unclear and should be further investigated.

The present study indicates that the connection of ENSO with the North Atlantic triple SST anomaly pattern is weak during 1968–88 (Fig. 15). This may be because the ENSO-induced atmospheric circulation anomalies over the North Atlantic are much weaker and shift westward compared to that after the early 1990s (figure not shown). Several studies suggested that response of the atmospheric circulation anomalies to the SST change may be dependent on the location of the induced atmospheric heating (e.g., Zuo et al. 2013; Chen and Wu 2017). We have examined the atmospheric heating anomalies (represented by precipitation) related to the Niño-3.4 index of $-1$ during 1968–88 and 1994–2014, respectively (figures not shown). The significant negative precipitation anomalies related to the negative phase of the Niño-3.4 index are stronger and located more westward during 1994–2014 than during 1968–88 (not shown). These differences in the precipitation anomalies may be related to changes in the types of ENSO events around the early 1990s. Studies found that the central Pacific (CP)-type ENSO events occur more frequently after the early 1990s (e.g., Ashok et al. 2007; Kao and Yu 2009; Kug et al. 2009; Yeh et al. 2009). In particular, locations of the center of the maximum SST anomalies related to the CP ENSO events shift westward to the central Pacific compared to those related to the traditional eastern Pacific (EP)-type ENSO events. Graf and Zanchettin (2012) have analyzed influences of CP and EP El Niño events on the North Atlantic–European climate. They found that the connection of the EP El Niño with the North Atlantic–European...
atmospheric circulation and surface temperature variations is weak. By contrast, CP El Niño could induce a negative phase of the NAO, and lead to significant cooling over Europe. They further suggested that the change in the connection of CP and EP El Niño with the North Atlantic atmospheric circulation is mainly attributed to the change in the locations of convective heating over the tropical Pacific. Based on the above evidence, it is speculated that the change in the position of atmospheric heating may be one of the possible factors for the change in the ENSO-related atmospheric circulation over the North Atlantic. Note that other factors, such as a change in the mean circulation (Kug and Jin 2009; Zuo et al. 2013; Chen and Wu 2017), may also play a role in contributing to this interdecadal change, which should be further investigated.

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