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RESEARCH ARTICLE

Inter-decadal change of the spring North Atlantic Oscillation impact on the summer Pamir–Tienshan snow cover

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Known as the "water tower of central Asia" and a crucial component of the Tibetan Plateau (TP), the Pamir–Tienshan snow cover (PTSC) exerts profound impacts on regional and global climate extremes during summer. However, researches on origins of the summer PTSC variability have not attracted adequate attention. In recent works, the spring North Atlantic Oscillation (NAO) has been shown to "prolong" its influence through the North Atlantic tri-pole sea surface temperature anomalies (Tri-SSTAs) and affect remote Asian climate in the subsequent summer. The present study discovers that the spring NAO has a significant positive correlation with the summer PTSC in the period 1967–1981, but such connection weakens after 1981. The North Atlantic Tri-SSTAs induced by the spring NAO during these two epochs can both persist into ensuing summer but excite significantly different downstream atmospheric teleconnections. The Rossby wave train triggered by positive (or negative) Tri-SSTAs during the 1967–1981 period exhibits an anomalous cyclonic (or anticyclonic) centre over the Pamir–Tienshan region, favouring an excessive (or reduced) PTSC. By contrast, such atmospheric anomaly centre is not evident during the 1982–2014 period. Numerical experiments suggest that westwards shift of the spring NAO northern centre can modulate the Tri-SSTA pattern via the wind–evaporation–SST feedback, which could further result in the displacement of the atmospheric anomaly centres in the Rossby wave train in summer. Therefore, such change in the spring NAO itself is responsible for the inter-decadal change of the spring NAO impact on the summer PTSC.

KEYWORDS
inter-decadal change, spring North Atlantic Oscillation, summer Pamir–Tienshan snow cover

1 | INTRODUCTION

The Tibetan Plateau (TP) is generally recognized as "The Third Pole" and the highest elevated land in the world with the averaged elevation of more than 4,000 m a.s.l. (Wu *et al*., 2016). It is an enormous cooling (or heating) source in the cold (or warm) seasons in the mid-troposphere of Asia. Numerous researches have suggested that the TP thermal anomaly can not only affect surrounding regional climate variability, such as the East Asian summer monsoon (EASM; e.g., Zhao and Chen, 2001; Hsu and Liu, 2003; Fujinami and Yasunari, 2009; Duan and Wu, 2006; Zhao *et al*., 2007; Seal and Hong, 2009; Duan *et al*., 2011) and the South Asian summer monsoon (SASM; e.g., Hahn and Manabe, 1975; Wu and Zhang, 1998; Boos and Kuang, 2010; Abe *et al*., 2013), but also modulate global climate through teleconnections (e.g., Liu *et al*., 2002; Lin and Wu, 2011; 2012).

In summer, the Pamir–Tienshan snow cover (PTSC) is known as the "water tower of central Asia" (Xu *et al*., 2008) and a crucial component of the TP. It can provide considerably freshwater to the surrounding densely populated, arid lowland areas and bring the atmosphere sufficient water vapour (Sorg *et al*., 2012). Its characteristic of persistence through the warm seasons over high altitudes (Pu *et al*., 2007; Wu *et al*., 2012a; 2012b) makes it a potential predictability source for summer climate. Although the summer TP snow cover (TPSC) exists at a relatively small area (Pamir–Tienshan) compared to that of winter, its possible climatic influence cannot be overlooked and has already been studied by many researchers (e.g., Tao and Ding, 1980; Ye, 1981; Zhang *et al*., 2004; Wang *et al*., 2008). The PTSC was found to have a close relationship with the summer East Asian climate (Wu and Kirtman, 2007) and could explain more than 30% of the total variances of heat wave variability in the southern Europe and northeastern Asia (SENA) region through a teleconnection pattern across the Eurasian continent (Wu *et al*., 2016). Another study also found that the decadal to inter-decadal variations of northern China heatwave frequency were associated with the summer PTSC (Wu *et al*., 2012a). Additionally, El Niño–Southern Oscillation (ENSO) could extend its influence to the eastern Asia during the reduced PTSC summers, which enhanced its connection with the EASM (Wu *et al*., 2012b).

A considerable amount of studies had paid attention to the summer PTSC as a climatic prediction source (e.g., Wu *et al*., 2012a; 2012b; 2016). However, what factors may influence the summer PTSC variability? So far, the discussion about the origin of the PTSC variability is relatively few. Yuan *et al*. (2009; 2012) presented that some slow boundary forcing, such as ENSO and Indian Ocean Dipole (IOD), might modulate the TPSC during the wintertime. Similarly, limited studies found that ENSO could influence the winter TP snow depth (Shaman and Tziperman, 2005). During the winter of 2009/2010, the negative North Atlantic Oscillation (NAO) and El Niño event was demonstrated to be responsible for the Northern Hemisphere (NH) snow anomalies (Seager *et al*., 2010). However, in summer counterpart, what can influence the PTSC is still quite unclear. Hence, considering the importance of the summer PTSC, it is essential to explore the potential extra-forcing factors that may contribute to the summer PTSC variability.

The NAO is a major circulation pattern over the middle-to-high latitudes of NH (Hurrell, 1995), which features as a large-scale seesaw atmospheric pressure at sea level (SLP) between the Icelandic low and the Azores high (Walker and Bliss, 1932; Bjerknes, 1964; van Loon and Rogers, 1978). Through fluctuations in the strength of the Icelandic low and the Azores high, it controls the strength and direction of westerly winds and location of storm tracks across the North Atlantic, which can further influence weather and climate from the eastern North America to the Europe (Hurrell, 1995; Rodwell *et al*., 1999; Cullen *et al*., 2002; Riviere and Orlanski, 2007). Besides, the extreme weathers were also coincided with the NAO, such as the coldest winter in Europe and the warmest winter in Canada during 2010/2011 (Seager *et al*., 2010; Lin and Wu, 2011). Recently, researchers find that the NAO is able to influence climate anomalies in downstream remote areas. For example, on the inter-annual timescale, the wintertime NAO could affect the subsequent spring climate over the European region (Herceg and Kucharski, 2013). The February NAO could extend its signals to the East Asia and the North Pacific (Watanabe, 2003) and the spring NAO had also been identified to influence the EASM variations through subpolar atmospheric teleconnections (Wu *et al*., 2009; Wu *et al*., 2012b). The NAO and the Asian–Pacific Oscillation (APO) correlated very well for the period of May–August (Zhou and Wang, 2015). In addition, the summer NH high-latitude climate is influenced by the previous winter NAO (Ogi *et al*., 2003). Moreover, on the inter-decadal timescale, the strengthened NAO could weaken the relationship between ENSO and Indian summer monsoon (Chang *et al*., 2001). The winter NAO could affect the precipitation over southern China (SCP) in the following spring (March–May), but this relationship was weakened after the early 1980s (Zhou, 2013). The relationship between the NAO and the tropical cyclone frequency over the western North Pacific in summer appears to have an inter-decadal change from weak connection to strong connection around 1980 (Zhou and Cui, 2014).

Up to now, some researches have found that the spring NAO could influence downstream mid-latitude climate anomalies through Eurasian teleconnections (Watanabe, 2003; Wu *et al*., 2009). Besides, such climatic effect undertook significant inter-decadal change (Wu *et al*., 2012b). These raise the new questions of whether the spring NAO can affect variations of the summer PTSC, which is located at the high altitude in Central Asia, and whether their connection is stable. This work will attempt to discuss these issues.

This paper is organized as follows. In section 2, the data sets, methodology and model used in this manuscript are described. Section 3 shows the relationship between the NAO and the summer PTSC. Section 4 presents how the spring NAO can or cannot influence the summer PTSC through the North Atlantic tri-pole sea surface temperature anomalies (Tri-SSTAs) and subpolar atmospheric teleconnections during the two epochs. The possible reason for such inter-decadal change of the spring NAO impact on the summer PTSC is explained in section 5.

2 | DATA, METHODOLOGY AND MODEL

The principle data sets employed in this study include (a) 1967–2014 monthly snow cover from the Global Snow Lab (Rutgers University, http://climate.rutgers.edu/snowcover), which are obtained using polar stereographic projection of NH and each monthly matrix having 89 × 89 grids (Robinson *et al*., 1993; Robinson and Frei, 2000); (b) 1967–2013 monthly atmospheric circulation fields, which are taken from Japanese 55-year Reanalysis (JRA-55, http://jra.kishou.go.jp); (c) 1967–2014 monthly SST data from the NOAA Extended Reconstructed Sea Surface Temperature V3b (ERSST V3b; Smith *et al*., 2008); (d) the NAO index

(NAOI) used in this study is defined as the difference in the normalized monthly sea level pressure zonal-averaged over the North Atlantic sector from 80°W to 30°E between 35°N and 65°N (Li *et al.*, 2013).

Since this work focuses on inter-decadal change of inter-annual variability, the 1967–2014 period is divided into two epochs: 1967–1981 and 1982–2014. The month by month calculation method is applied in this study. Annual cycles are removed from each month and linear trends are, respectively, removed during the two epochs. During the whole article, the summer refers to the months of June, July and August (JJA) and spring refers to the late spring with 1-month lead.

A variety of statistical methods, including linear correlation, one-point correlation, sliding correlation, regression, composite and empirical orthogonal function (EOF) analysis, are employed in this article. In particular, the plumb wave activity flux (WAF; Plumb, 1985) is calculated to discover the teleconnection pattern, which is derived for linear, quasi-geostrophic disturbances on a zonal flow and is specific to the stationary component of atmospheric wave motions.

To verify the teleconnection change related to the Tri-SSTAs between the two periods, three numerical experiments are performed based on the GFDL AM2.1, with a horizontal resolution of 2.5° longitude × 2° latitude with specified SST boundary conditions and 32 levels in the vertical dimension (Geng and Zhang, 2017). These experiments are referred to as the "Ctrl," "Exp1," "Exp2," which are short for "control experiment," "sensitive experiment 1," "sensitive experiment 2." The control experiment is driven by the historical SST. To simply mimic the diabatic heating influence of Tri-SSTAs in the sensitive experiments, the observational summer Tri-SSTAs at the area of Tri-SSTI1/2 based on the composite analysis are used as the summer forcing sources (Figure 11a,b). All these three experiments are integrated for 20 years from 1982 to 2001, and the ensemble mean of the last 5 years are employed for eliminating the model spin-up.

3 | RELATIONSHIP BETWEEN NORTH ATLANTIC OSCILLATION AND JJA PAMIR-TIENSHAN SNOW COVER

Figure 1a presents the PT and its adjacent areas, which are located over the western and northern TP (as shown by the

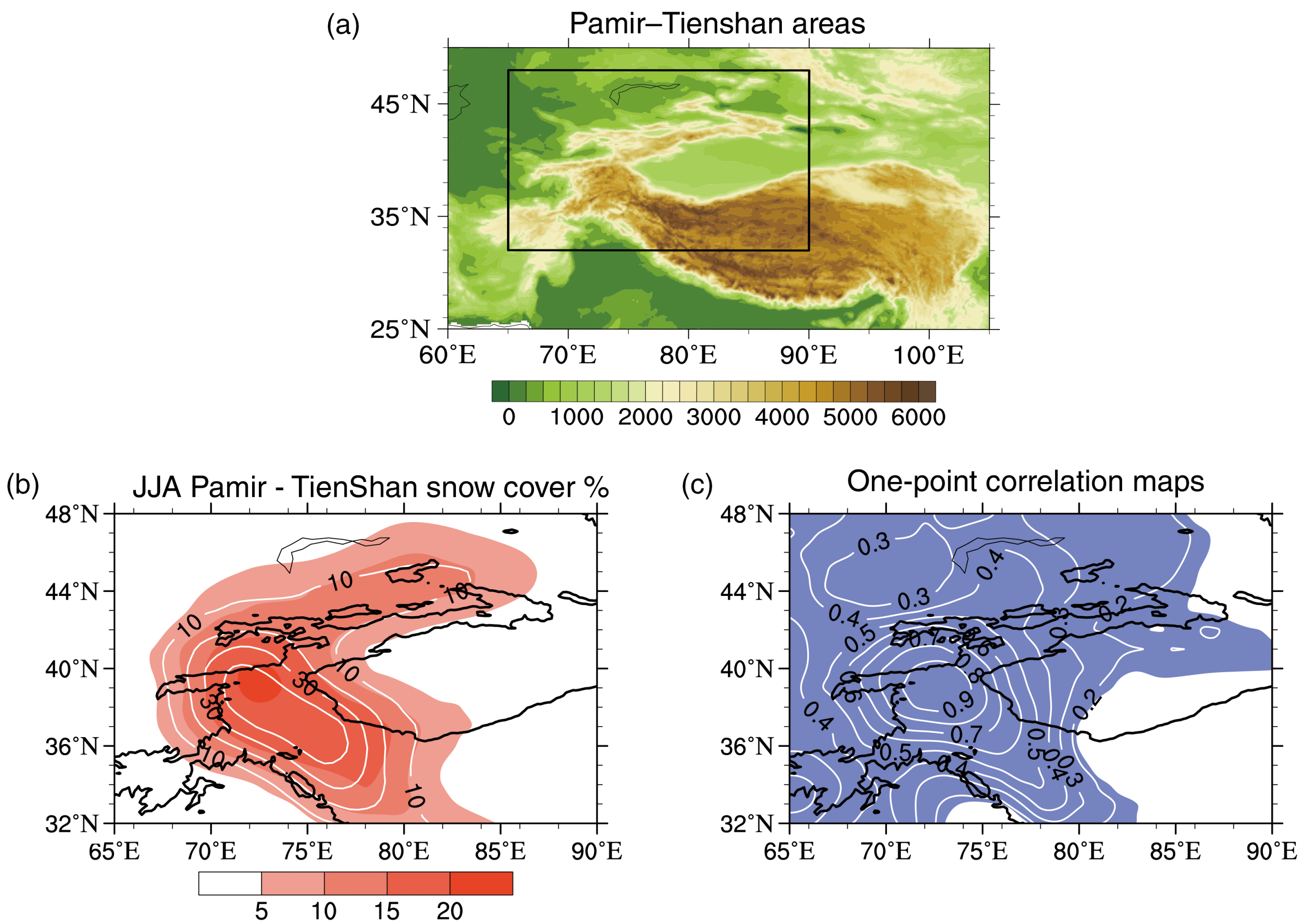

FIGURE 1 (a) Altitude (colour shadings, m) of the Pamir–Tienshan (black box) and its adjacent areas. (b) Long-term mean (contours, %) and standard deviations (colour shadings) of the PTSC during boreal summer in the period 1967–2014. The areas included by the black curves are 2,500 m a.s.l. (c) One-point (39°N, 73°E) correlation maps of the summer snow cover. The summer PTSC anomalies averaged in the plot areas and passed 99% confidence level based on a Student's *t* test (colour shading areas) is defined as a PTSCI [Colour figure can be viewed at wileyonlinelibrary.com]

black box: 32°–48°N, 65°–90°E). The climatology and standard deviation of the summer PTSC are shown in Figure 1b, which presents that the PT areas exhibit relatively a large amount of snow cover, and the snow cover here experiences strong inter-annual variability. Such notable snow cover variations have the ability to "memorize" and influence middle-tropospheric atmosphere directly and lead to anomalous atmospheric diabatic heating.

In order to quantify the variations of the summer PTSC, the point (39°N, 73°E), which has the largest inter-annual variability, is selected to make the one-point correlation distribution with the PTSC (Figure 1c). The detrended summer PTSC anomalies, where the shaded areas with correlation coefficients exceeding 99% confidence level, are averaged as a PTSC index (PTSCI). Figure 2a shows the temporal evolution of the normalized PTSCI (blue line) and spring (red dash line)/summer (red solid line) NAOI. Their correlation coefficients both fail to pass the significance test during the 1967–2014 period. To further investigate their potential inter-decadal variations, we carry out the 11-year sliding correlation (linear trends are removed during each 11 years) between the NAOIs and the PTSCI (Figure 2b). Interestingly enough, the relationship between the spring/summer NAO and the summer PTSC dramatically changes around 1982. During the 1967–1981 period, the correlation coefficients reach 0.67 in preceding spring and 0.80 in simultaneous summer, both exceeding the 99% confidence level. But after 1981, this connection breaks down, with their correlation coefficients decreasing to −0.24 and −0.18.

To further check whether such inter-decadal change does exist or not, Figure 3 shows the lead–lag correlation coefficients between the JJA PTSCI and the NAOI from January, February and March (JFM) to October, November and December (OND). The results show that only the MJJ and JJA NAOI are significantly correlated with the JJA PTSCI in the earlier period, both exceeding the 95% confidence level (black horizontal dash line), which imply the NAO can influence the PTSC by leading 1-month during the 1967–1981 epoch. For the reason that there are many NAOI definition methods, different NAOIs are adopted to verify their relationships with the PTSCI (Table 1). It shows that the inter-decadal relationship between the spring $NAOI_{NOAA}$/$NAOI_{Jones}$ and PTSCI is consistent with the $NAOI_{LJP}$'s, which proves that the inter-decadal change of relationship between the NAO and PTSC exists robustly. The spatial correlation distributions are also calculated between the spring/summer NAOI and the summer PTSC during these two periods (Figure 4). A large area of significant positive values is discerned over the PT during the earlier period (Figure 4a,c), which implies that the positive (negative) spring/summer NAO anomalies usually correspond to the excessive (reduced) summer PTSC. During the 1982–2014 period, no significant anomalous snow cover signal is found over the PT areas (Figure 4b,d).

All the analysis above indicates that the preceding spring NAO can influence the summer PTSC in the period 1967–1981, but this relevance weakens after 1981. So the following research will focus on the spring NAO. Then what

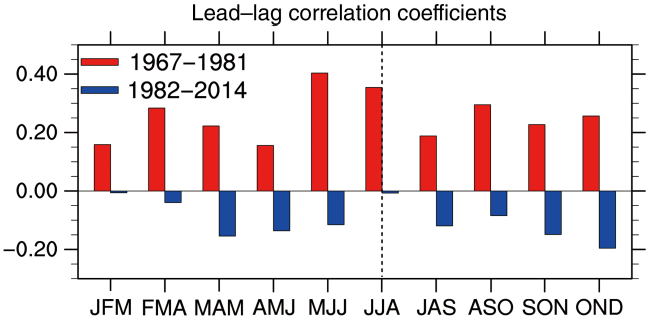

FIGURE 3 Lead–lag correlation coefficients between the PTSCI and the NAOI from JFM to OND during these two epochs. The bars above and below zero line, respectively, represent the 1967–1981 period and 1982–2014 period. The dash horizontal lines represent the 95% confidence level during the two epochs, respectively [Colour figure can be viewed at wileyonlinelibrary.com]

TABLE 1 Correlation coefficients between the PTSCI and different NAOIs in different periods ("**" and "*" respectively represent the Student's *t* test exceeding 95% and 90%). $NAOI_{NOAA}$: The loading pattern of the NAO is defined as the first leading mode of rotated empirical orthogonal function (REOF) analysis of monthly mean 500 mb height (Barnston and Livezey, 1987). $NAOI_{Jones}$: P_G–P_S. P = normalized SLP anomaly; S = Stykkisholmur (Iceland); G = Gibraltar (Jones *et al.*, 1997)

MJJ(JJA) Indexes	1967–1981 PTSC index	1982–2014	1967–2014
$NAOI_{LJP}$	0.40**(0.35**)	−0.12(−0.01)	0.05(0.10)
$NAOI_{NOAA}$	0.29*(0.31**)	−0.15(0.05)	−0.03(0.14)
$NAOI_{Jones}$	0.33**(0.22*)	−0.22 (−0.02)	−0.01(0.06)

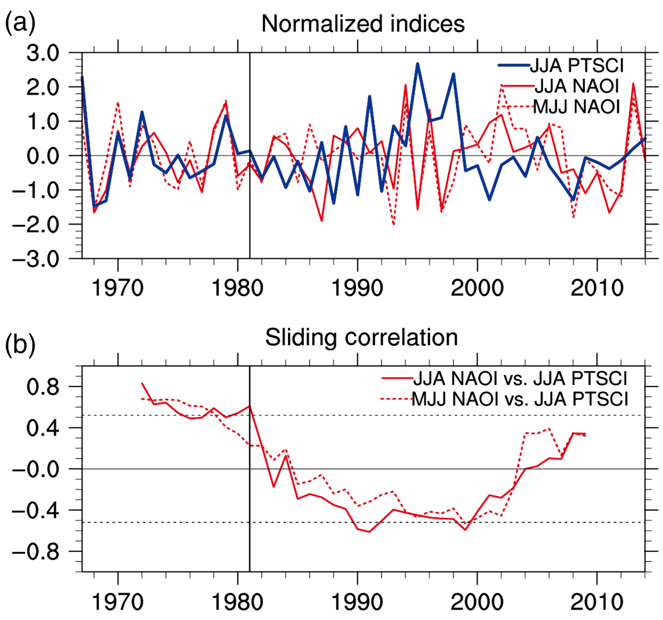

FIGURE 2 (a) Normalized time series of the PTSCI (bold solid line), the summer NAOI (solid line) and the spring NAOI (dash line). (b) 11-year sliding correlation between the NAOI and the PTSCI [Colour figure can be viewed at wileyonlinelibrary.com]

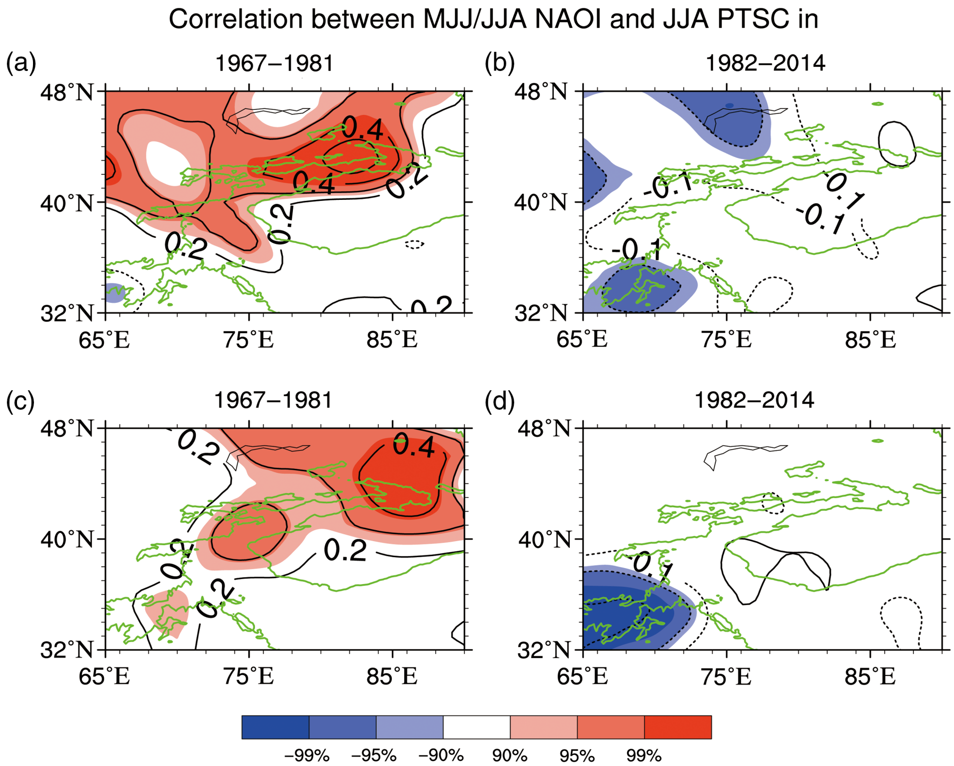

FIGURE 4 Correlation between the NAOI in the spring (a, 1967–1981; b, 1982–2014)/summer (c, 1967–1981; d, 1982–2014) and the summer PTSC. The shaded areas exceeding 90, 95, 99% confidence level are based on a Student's *t* test. The areas included by green curves are 2,500 m a.s.l [Colour figure can be viewed at wileyonlinelibrary.com]

may be the physical mechanism of NAO influencing PTSC and existent inter-decadal change?

4 | PHYSICAL MECHANISMS

As the major atmospheric circulation mode, the NAO signal changes quickly. The researches on the NAO usually concern coupled mechanisms that involve slowly varying low boundary forcing, which can "prolong" the NAO impact. For example, the North Atlantic Tri-SSTAs can store the NAO signal via air–sea interactions and then, in the subsequent months, influence remote regional climate (Cayan, 1992; Shukla, 1998; Eden and Willebrand, 2001; Wu *et al.*, 2009; Czaja and Frankignoul, 2013). The similar method will be adopted here to interpret that how the spring NAO can or cannot affect the downstream summer PTSC.

Figures 5 and 6, respectively, show the spring/summer SSTA patterns regressed to the spring NAOI and the PTSCI during the 1967–1981 and 1982–2014 periods. As can be seen clearly, the SSTAs related to the spring NAO can both persist from spring to summer during the two periods, but the SSTAs undertake significant spatial transformation. During the earlier period, the SSTAs are the zonal distribution structure with the Tri-SSTAs in the extratropical North Atlantic and the significant positive values in the tropical areas. But during the later period, the tropical positive SSTAs disappear and the Tri-SSTAs expand to the Tropic. These centres move southwards (the northern centre approximately 5° and the others 10°) and the areas of centre enlarge, which is almost like a "horseshoe" pattern in the North Atlantic. The SSTAs induced by the PTSC are also computed, which show exactly the similar Tri-SSTAs pattern for 1967–1981 (but without the tropical SSTAs) and apparently different after 1981. To further quantitatively contrast such change of the Tri-SSTAs pattern, the differences between the sum of averaged SSTAs in positive correlation box and averaged SSTAs in two negative correlation boxes are defined as the simple tri-pole SST indices during these two periods (Tri-SSTI1/2), respectively (positive domain minus negative domain). The correlation coefficients between the Tri-SSTI1 and the NAOI/PTSCI in the period 1967–1981 reach 0.51/0.54, exceeding the 99% confidence level. But in the period 1982–2014, no significant correlation is found.

Based on the above statistical analysis, the transformation of the Tri-SSTAs may be responsible for this inter-decadal change. Nevertheless, we still have not clarified the causality between atmosphere and ocean. To illustrate this issue, we take the Tri-SSTI1 as a reference and compute the lead–lag regression with SLP field in the earlier period (Figure 7). It shows a NAO-like pattern from −2 to −1 month (Figure 7a, b), which implies that the NAO-like atmospheric anomalies are driving the Tri-SSTAs in the period 1967–1981. Furthermore, such NAO-like signal reaches its peak at −1-month lead (Figure 7b) and dies away rapidly after the simultaneous summer (Figure 7d), indicating that the memory effect of ocean may be responsible for the persistence of the Tri-

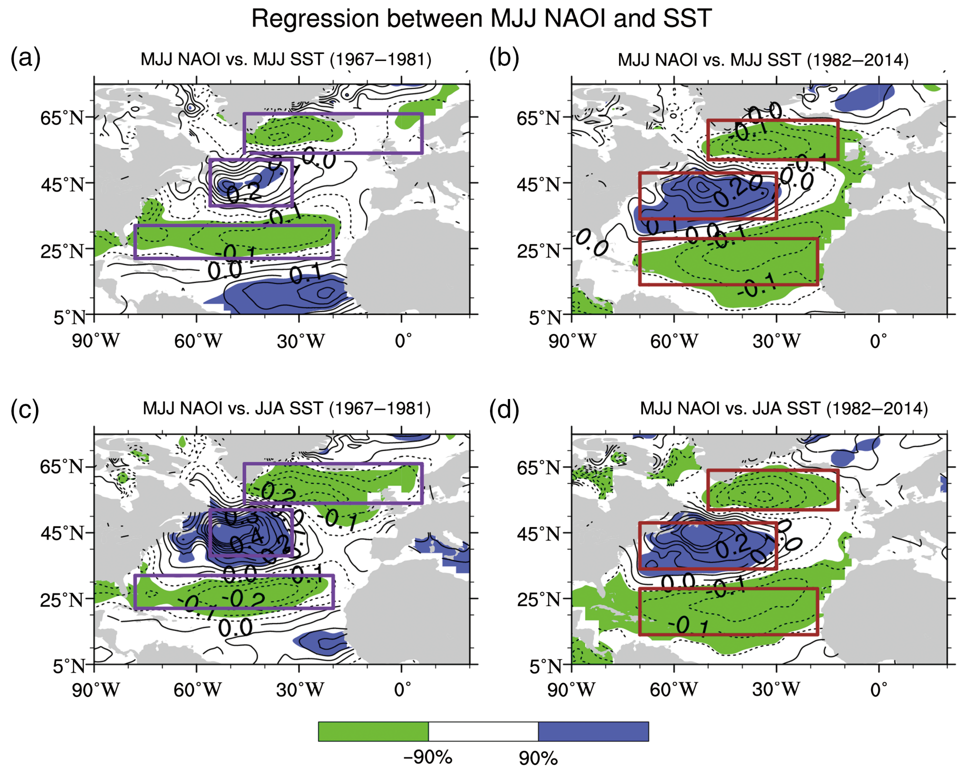

FIGURE 5 Regression maps between the spring NAOI and the spring/summer SST (contours, K). The shaded areas exceeding 90% confidence level are based on a Student's *t* test. Difference of summer SSTAs during the 1967–1981 period between the box with positive values (32°–56°W, 38°–52°N) and the two boxes with negative values [(20°–78°W, 22°–32°N) and (46°W–6°E, 54°–66°N)] is defined as a tri-pole sea surface temperature index (Tri-SSTI1); and difference of summer SSTAs during the 1982–2014 period between the box with positive values (30°–70°W, 34°–48°N) and the two boxes with negative values [(18°–70°W, 14°–28°N) and (12°–50°W, 52°–64°N)] is also defined as a tri-pole sea surface temperature index (Tri-SSTI2) [Colour figure can be viewed at wileyonlinelibrary.com]

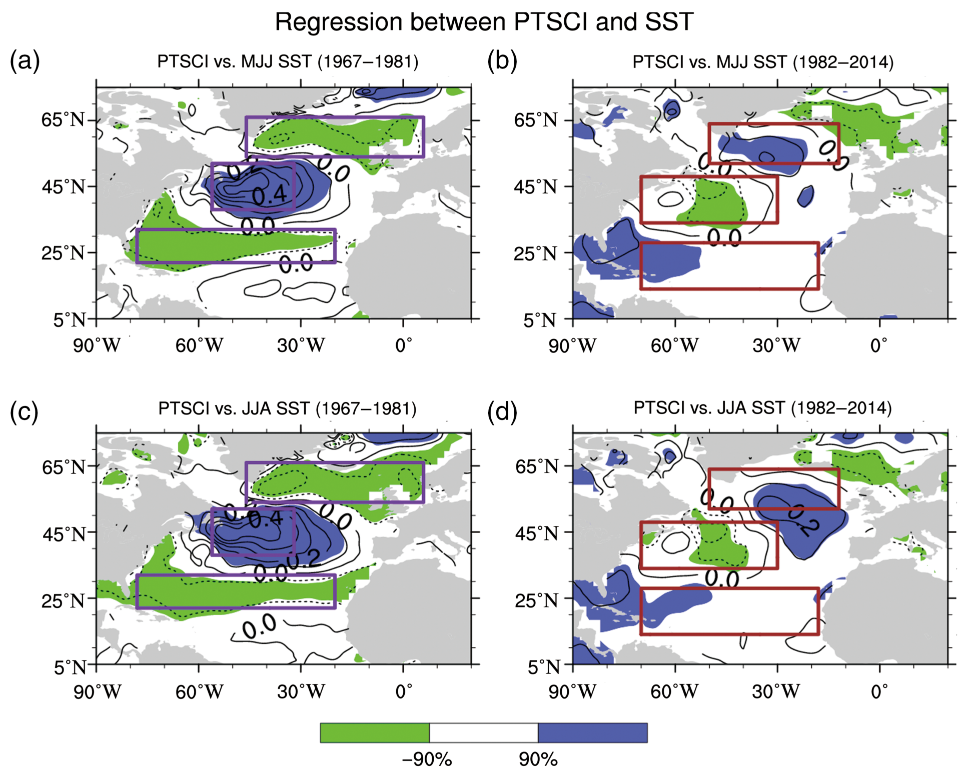

FIGURE 6 Same as Figure 5 but for the regression to the PTSCI [Colour figure can be viewed at wileyonlinelibrary.com]

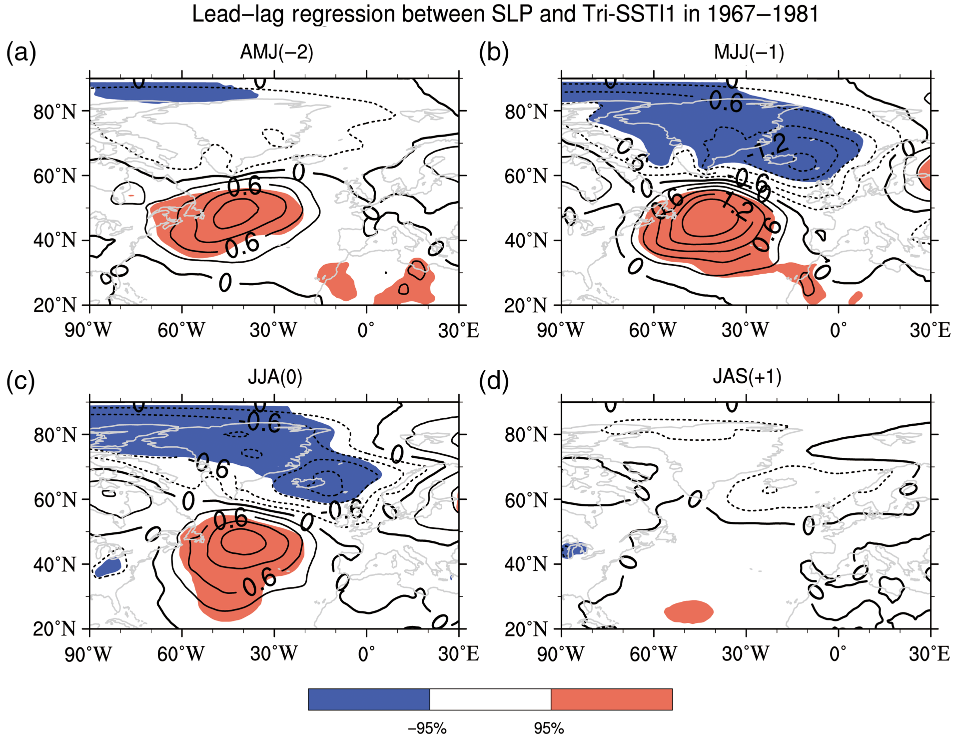

FIGURE 7 Lead–lag regression maps between the summer Tri-SSTI1 and the SLP (contours, hPa) from April, May and June (AMJ) to July, August and September (JAS) in the period 1967–1981. The shaded areas exceeding the 99% confidence level are based on a Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

SSTAs from spring to summer. The similar results are shown in the latter period (Figure 8), which are all consistent with previous studies (Deser and Timlin, 1997; Pan, 2005; Wu *et al*., 2009; Wu *et al*., 2012a; 2012b; 2016).

So how can such Tri-SSTAs affect the summer atmospheric circulation? Figure 9 shows the WAF and stream function (SF) at 200 hPa regressed to the Tri-SSTIs during these two periods, respectively. In the earlier period, the

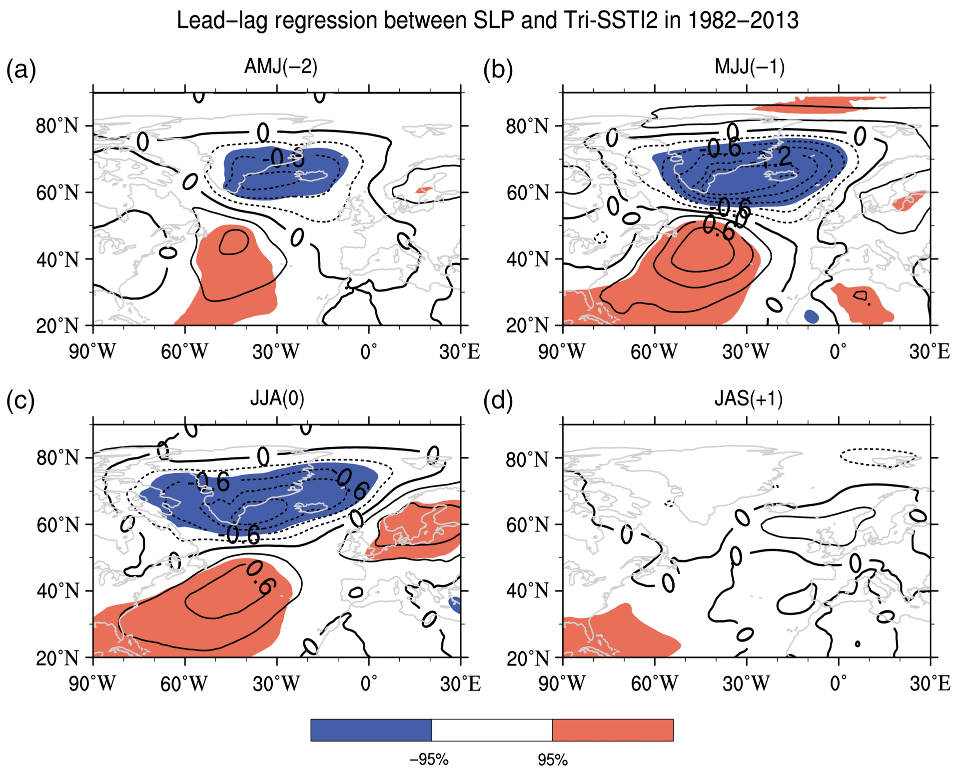

FIGURE 8 Same as Figure 7, but in the period 1982–2013 [Colour figure can be viewed at wileyonlinelibrary.com]

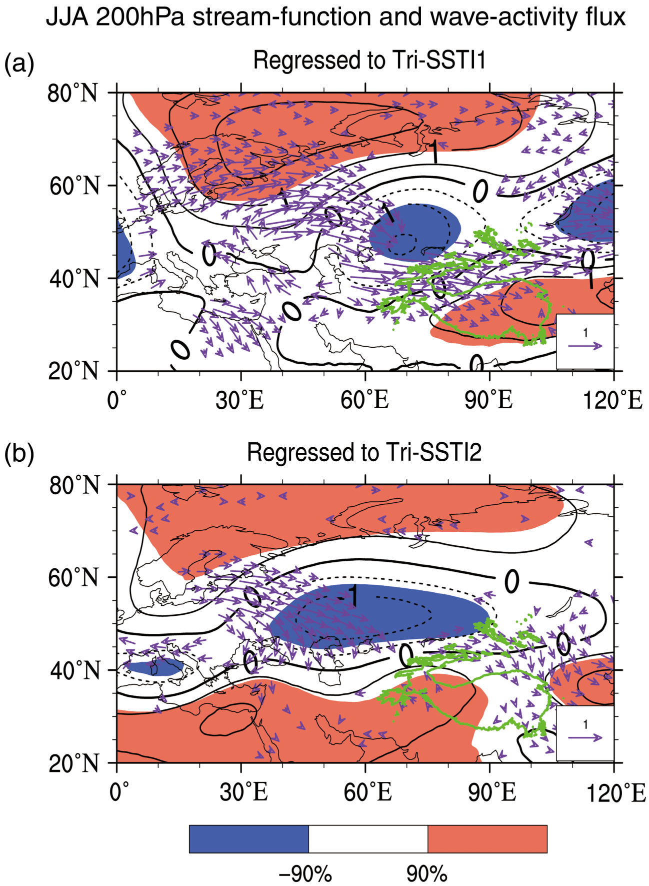

FIGURE 9 200 hPa summer SF (contours, 10^7 m^2/s) and WAF (purple vectors, 10^2 m^2/s^2) regression for the Tri-SSTI1 (a) during the 1967–1981 period and the Tri-SSTI2 (b) during the 1982–2013 period. The shaded areas represent the regression coefficient exceeding the 90% confidence level based on a Student's *t* test. And the vectors only show the parts of WAF exceeding the 90% confidence level [Colour figure can be viewed at wileyonlinelibrary.com]

significant feature is that a Rossby wave train, southeastwards propagation of the WAF from the Norwegian Sea/Scandinavian Peninsula to the Balkhash Lake, PT and East Asia, which is also notable and more clearly in the regression between the PTSCI and geopotential height fields (Figure 10a,b). That is to say, the positive (negative) Tri-SSTAs usually correspond to anticyclone (cyclone) centres at the East European Plain and East Asia, and cyclone (anticyclone) centres at the Iceland and PT, favouring an excessive (reduced) PTSC. By contrast, during the latter period, the corresponding WAF only can propagate to the Black Sea–Caspian Sea area and such anomalous atmospheric centre is not evident over the PT (Figure 9b). There is also no remarkable feature in the 200/500 hPa circulation fields regressed to the PTSCI (Figure 10c,d).

To further examine such change of the teleconnection pattern forced by the Tri-SSTAs transformation, three numerical experiments (one was the control run driven by the historical SST, the other two were forced by the different North Atlantic Tri-SSTAs forcing, respectively) were executed with the GFDL AM2.1 model described in section 2. Under the two Tri-SSTAs forcing (Figure 11a,b), the simulations both can capture the downstream teleconnections, but they are significantly different (Figure 11c,d). The notable differences are similar as the observational analyses, which shows the displacement of teleconnection centres (Figure 11e,f). Although the numerical experiments cannot perfectly simulate the atmospheric circulation change over the PT areas, such simulation can further verify that the change of the North Atlantic Tri-SSTAs related to the NAO can modulate downstream teleconnections.

Therefore, based on the analysis above, the North Atlantic Tri-SSTAs induced by the spring NAO can both persist into ensuing summer, but excite significantly different downstream atmospheric teleconnections, which may be the direct reason for the inter-decadal change. However, what causes such different North Atlantic Tri-SSTAs during the two periods is still a key problem to be solved.

5 | REASONS FOR INTER-DECADAL CHANGES

The inter-annual variability of the winter NAO underwent a remarkable change after 1978, especially the eastwards shift of its northern centre (Ulbrich and Christoph, 1999; Hilmer and Jung, 2000). The plausible explanation for this shift may be the increase in strength of mean westerly winds and eastwards shift of Atlantic storm track activity (Luo and Gong, 2006; Luo *et al*., 2010). And the corresponding climate change in North America, North Atlantic and Europe has already been discussed (Jung *et al*., 2003). Nevertheless, the research on the NAO cannot only be confined in its strong phase (winter). Whether the inter-decadal change of the NAO itself exists in other season, such as spring, is an interesting topic. If so, can such spring NAO own change be the reason for its inter-decadal relationship with the summer PTSC? These issues are worthwhile exploring.

Figure 12, respectively, shows the spring SLP regressed to the simultaneous NAOI in the two periods and their differences. The NAO pattern exhibits a northeast–southwest tilted dipole anomaly for 1967–1981 (Figure 12a), with the northern centre located at the southeast of Island and the southern centre located at approximately 35°–50°N, 35°–55°W. But the observable feature exhibits a north–south meridional dipole anomaly for 1982–2014 (Figure 12b), with the northern centre located between the Greenland and Island and the southern centre located at approximately 35°–45°N, 20°–60°W. The spring NAO northern centre is found to experience an observable westwards shift after 1981, which is opposite to that of the winter NAO in previous studies (Ulbrich and Christoph, 1999; Hilmer and Jung, 2000; Jung *et al*., 2003). Considering the limitation of using NAOI to carry out the analysis above, the first EOF (EOF1) mode of spring North Atlantic SLP anomalies during the two periods are shown in Figure 12. It is not difficult to find

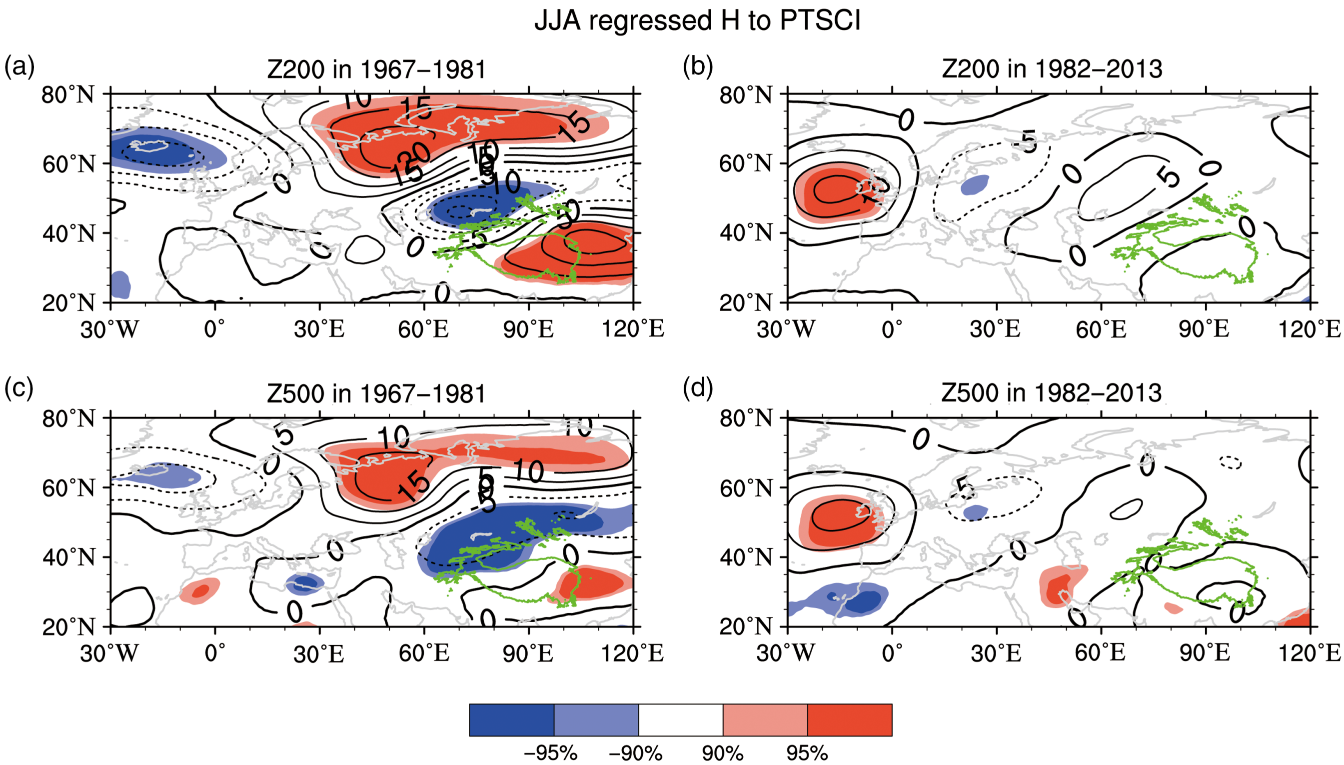

FIGURE 10 Two-hundred/five-hundred hectopascal summer geopotential height (contours, gpm) regression for the PTSCI during (a, c) the 1967–1981 period and (b, d) the 1982–2013 period. The shaded areas represent the regression coefficient exceeding the 90%, 95% confidence level based on a Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

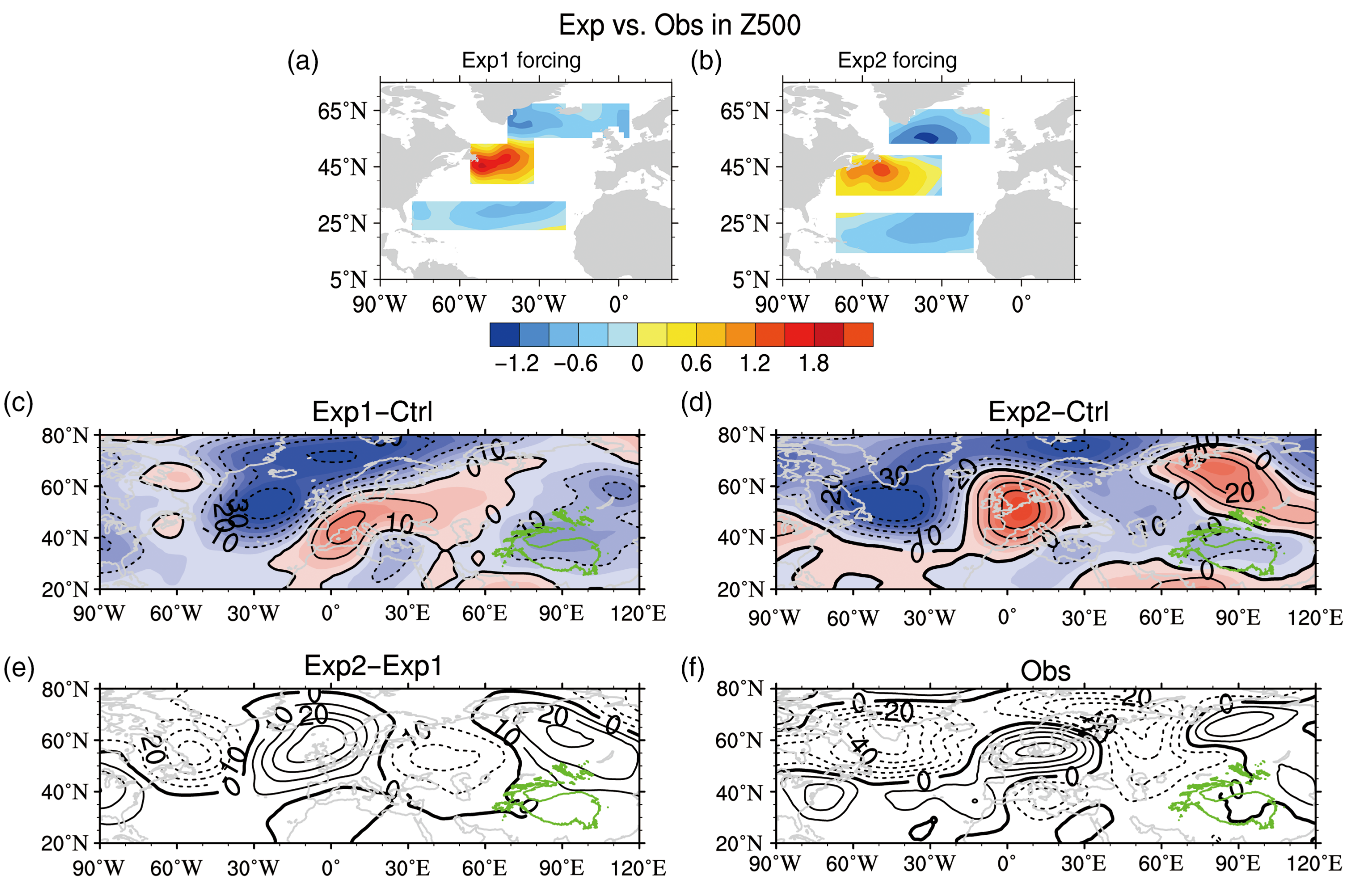

FIGURE 11 The numerical experiments based on the GFDL AM2.1. (a, b) Forcing sources (summer SSTAs, K) based on the observational composite analysis of the North Atlantic Tri-SSTAs in the high-low NAOI years. (c, d) Five-hundred hectopascal summer geopotential height (contours, gpm) differences between Exp1/Exp2 and Ctrl. (e, f) Comparison of the 500 hPa summer geopotential height (contours, gpm) differences between two periods in the numerical experiments and observations [Colour figure can be viewed at wileyonlinelibrary.com]

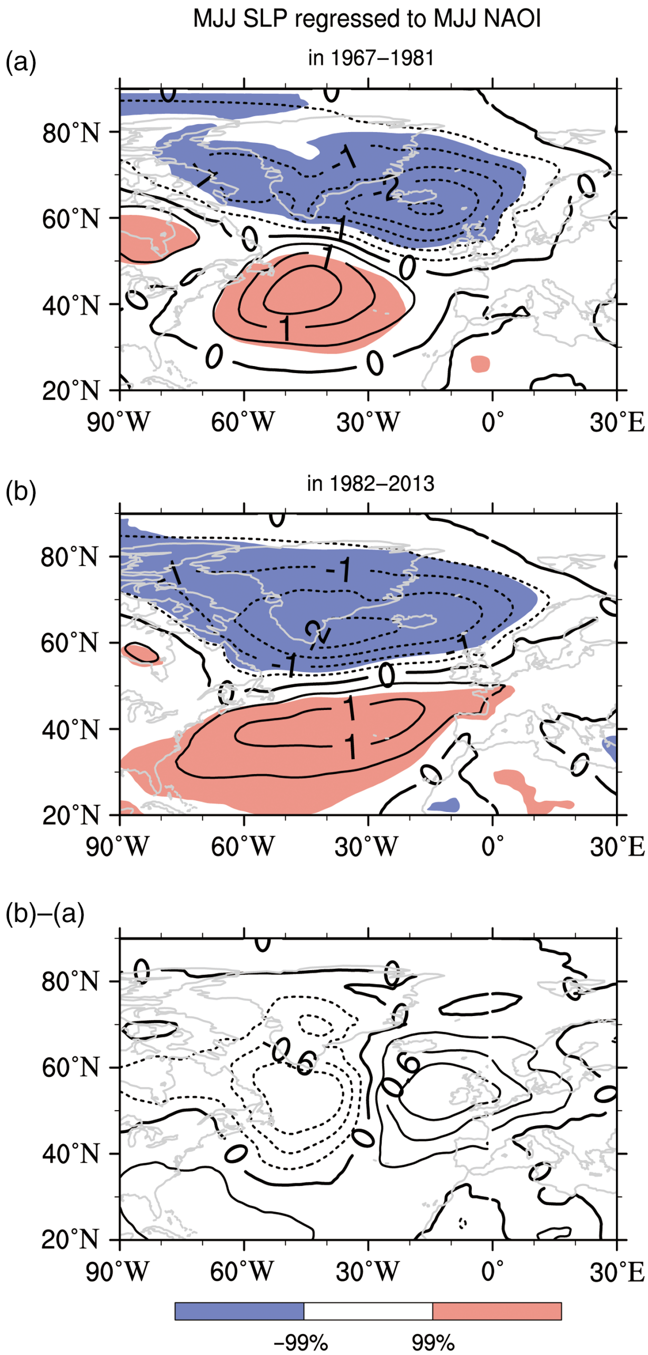

FIGURE 12 The spring SLP (contours, hPa) regression for the spring NAOI during (a) the 1967–1981 period and (b) the 1982–2013 period. The shaded areas represent the regression coefficient exceeding the 99% confidence level based on a Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

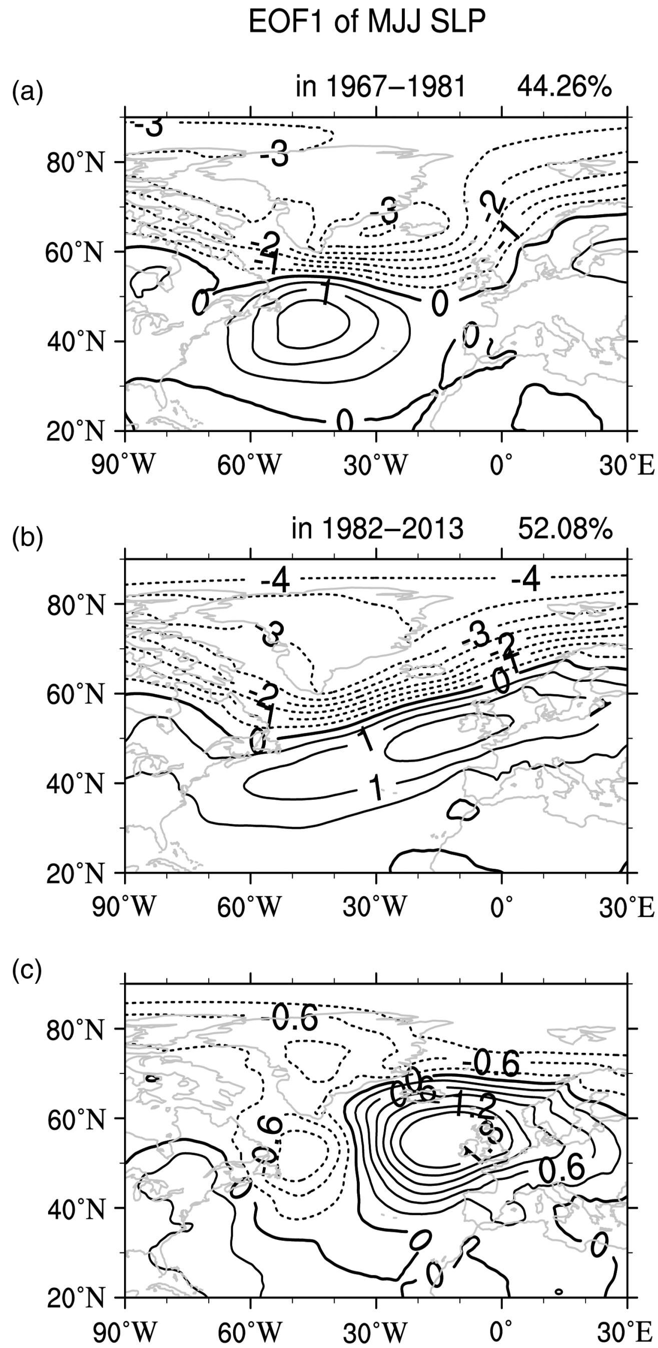

FIGURE 13 EOF1 of the spring SLP during (a) the 1967–1981 period and (b) the 1982–2013 period in the North Atlantic basin. The variance explanations are listed in the figures

that the differences of EOF1 between the two periods are the same as the regression analysis (Figures 12c and 13c), which can further confirm the shift of NAO northern centre.

How can such inter-decadal change of the spring NAO itself affect its relationship with the downstream summer PTSC? In order to understand how ocean can restore such variations of atmospheric mode, we firstly choose 0.85 standard deviation as a threshold for the spring NAOI to obtain the high-NAOI years and low-NAOI years during the two periods, respectively. The same method is taken to select the extreme PTSC years for comparison (Table 2). In the earlier period, the high-PTSCI years are all consistent with the high-NAOI years (1967, 1972, 1979), and the low-PTSCI years also have the same year with the low-NAOI years (1981). However, in the latter period, they almost have no sharing high/low years, which is in accordance with the analyses above.

Previous researchers have pointed out that a positive (negative) NAO is always coupled with the northwards (southwards) displacement of the North Atlantic storm track (Thompson and Wallace, 2000a; 2000b; Trenberth *et al.*,

TABLE 2 High and low value years of the spring NAOI and the PTSCI based on 0.85 standard deviation during the two epochs. The years of bold font represent the sharing years between the NAOI and the PTSCI

		MJJ NAOI	PTSCI
1967–1981	High	**1967**, 1970, **1972**, **1979**	**1967**, **1972**, **1979**
	Low	**1968**, 1971, 1975, 1980	**1968**, 1969
1982–2014	High	1986, 1994, 1999, 2002, 2006,2013	1991, 1995, 1996, 1997, 1998
	Low	1993, 1995, 1997, **2008**, 2011, 2012	1984, 1986, 1988, 1990, 1992, 2001, **2008**

2005) but the surface wind speed varies in phase with the high-level jet stream. Such abnormal surface wind speed can not only modulate the exchanges of sensible/latent heat flux at the ocean–atmosphere interface but also stir up surface meridional Ekman transport (Thompson and Wallace, 2000a; 2000b; Trenberth *et al.*, 2005). Consequently, surface wind speed related to the NAO can trigger the North Atlantic Tri-SSTAs (Wu *et al.*, 2009). As can be seen in the composite analyses of SLP/surface 10-m zonal wind speed (Uwnd10m)/SST in the high-NAOI years (Figure 14), during the 1967–1981 period, the strong spring NAO is associated with the high surface wind speed within (50°–70°N, 0°–70°W) and the low surface wind speed within (50°–70°N, 30°–50°E) (Figure 14a, middle). But during the 1982–2014 period, the significant westwards motions are found in both high and low wind speed regions (Figure 14b, middle) due to the westwards shift of the NAO northern centre (Figure 14a,b, left). It is known that high (low) wind speed usually leads to northwards (southwards) Ekman transport of cold water and rise (decline) in ocean heat release, resulting in negative (positive) SSTAs (Wu *et al.*, 2009; Dou and Wu, 2016). Therefore, the northern two centres of the Tri-SSTAs in the high-NAOI years exhibit notable southwestwards shift and the southern centre exhibits southeastwards shift via such wind–evaporation–SST feedback and restriction of sea–land boundary in North Atlantic basin (Figure 14, left). The similar composite results of the low-NAOI years are shown in Figure 15.

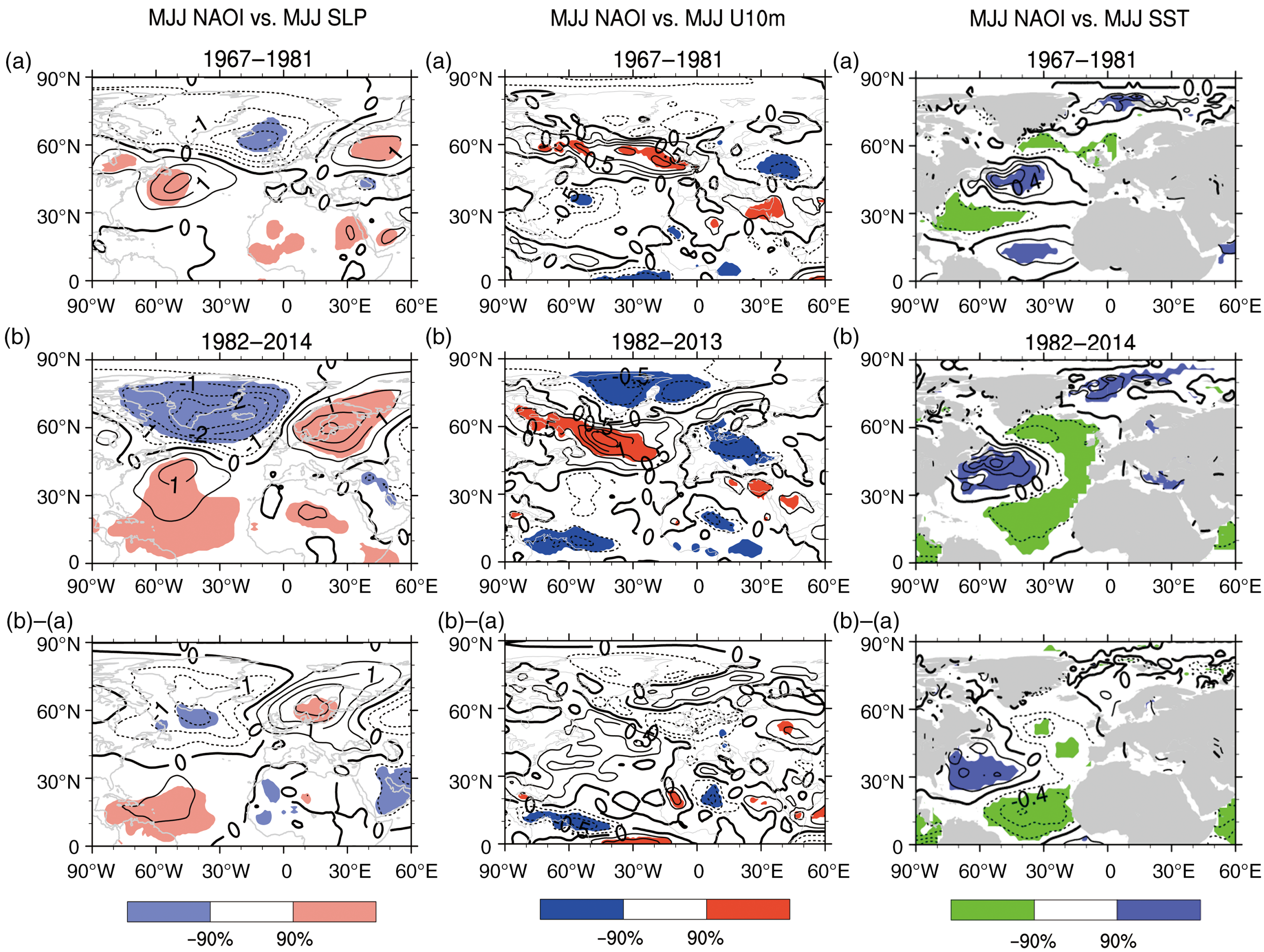

FIGURE 14 Composite analysis of the spring SLP (left, contours, hPa), the spring Uwnd10m (middle, contours, m/s), the spring SST (right, contours, K) in the *high* NAOI years. (a, b), respectively, represent the 1967–1981 period and the 1982–2014 period. The shaded areas exceeding the 90% confidence level are based on a Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

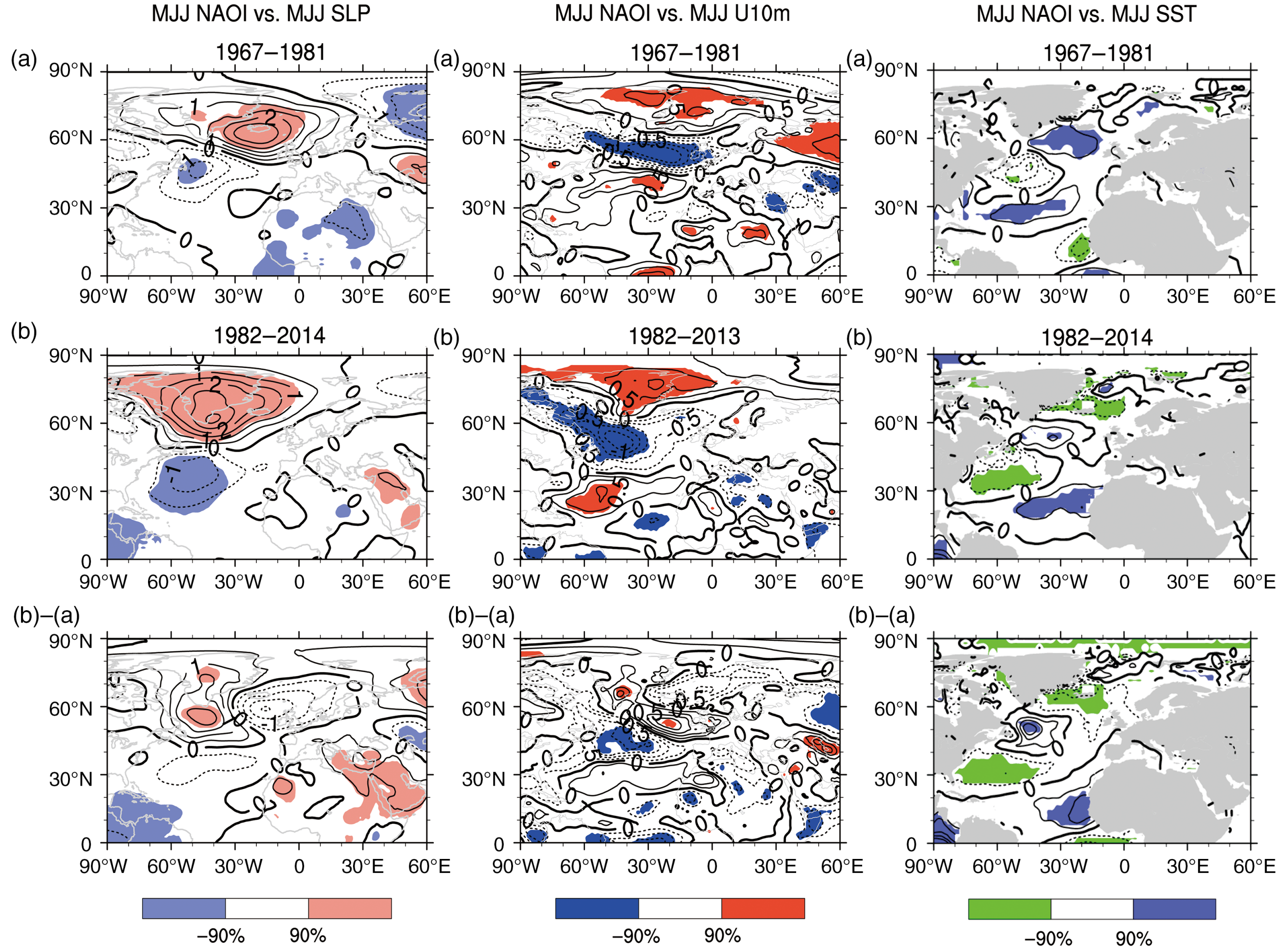

FIGURE 15 Same as Figure 14, but for the *low* NAOI years [Colour figure can be viewed at wileyonlinelibrary.com]

These changes of the spring Tri-SSTAs in the high-NAOI and low-NAOI years next can persist into ensuing summer (not shown, but similar to the regression analysis in Figure 5) and then modulate the downstream atmospheric teleconnections, finally resulting in the inter-decadal change of the spring NAO impact on the summer PTSC.

6 | CONCLUSION AND DISCUSSION

As an indispensable part of central Asia and TP, the PTSC can exert its direct effect on tropospheric atmosphere and persist through the whole summer. Its climatic effects have been investigated by many studies (Wu and Kirtman, 2007; Wu *et al.*, 2012a; Wu *et al.*, 2016). However, limited researches focus on what may explain the summer PTSC variability. Therefore, exploring the origin of the PTSC variations becomes an impending issue. NAO, the major atmospheric circulation pattern over mid–high latitudes of NH (Walker and Bliss, 1932; Bjerknes, 1964; van Loon and Rogers, 1978; Hurrell, 1995), can motivate downstream teleconnections to affect Asian climate (Shukla, 1998; Chang *et al.*, 2001; Wu *et al.*, 2009). Now that the PT is located at the central Asia, there might be a potential linkage between the NAO and the PTSC.

The current paper indicates that the spring (MJJ) NAO can influence the summer (JJA) PTSC during the 1967–1981 period, but this teleconnection weakens after 1981. Comprehending the reason for such inter-decadal change should be the main difficulty in our study. According to the analyses, the North Atlantic Tri-SSTAs can “prolong” the influence of the spring NAO and regulate the downstream atmospheric teleconnections in both periods. Moreover, the westwards shift of the spring NAO northern centre can result in the displacement of the North Atlantic Tri-SSTA centres via wind–evaporation–SST feedback. Numerical results show that the atmospheric teleconnections triggered by the Tri-SSTAs, with anomalous atmospheric centre over the PT, can influence the PTSC in the earlier period. But in the latter period, such atmospheric anomaly centre is not evident over the PT. Briefly, the westwards shift of the NAO northern centre can account for the inter-decadal change of the spring NAO impact on the summer PTSC.

In addition, as shown in Figure 5, the positive values at the southern of 15°N are present in the 1967–1981. But the significant negative values are present at the tropical eastern

Pacific in the 1982–2014 (not shown). The recent studies found that the tropical SSTAs may have the ability to affect the atmospheric circulation in the middle and high latitude. For example, the synchronous behaviours in winter between the mega-ENSO and the NAO are tied to the anomalous tropical North Atlantic SST (Zhang *et al.*, 2017). And the tropical central Pacific SSTAs can influence East Asian climate through the Philippine Sea anticyclone, which is called the Pacific–East Asian teleconnection (Wang, 2002). Therefore, what are the roles of such tropical SSTAs on the NAO–PTSC connection? Besides, Figure 9 partly shows the different atmospheric teleconnection in East Asia during the two periods. Hence what other effects such teleconnection change has on the downstream areas, such as East Asia and Pacific? These are interesting issues and can be discussed in future work.

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