Progress in the Study of the Dynamics of Extratropical Atmospheric Teleconnection Patterns and Their Impacts on East Asian Climate

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ABSTRACT

In the extratropics of the Northern Hemisphere, there exist many kinds of atmospheric teleconnection patterns. According to their spatial structure, these teleconnection patterns are generally divided into two groups. One group comprises north-south dipole patterns, such as the North Atlantic Oscillation and the North Pacific Oscillation, which have two anomalous centers of opposite signs in the north-south direction. The other group includes the wave train-like patterns, which have several anomalous centers of opposite signs distributed mainly in the zonal direction, such as the Pacific/North American and Eurasian Patterns. These teleconnection patterns greatly impact weather and climate not only in the regions where the teleconnection patterns are active, but also in the regions thousands of kilometers away. Studying and understanding the formation mechanisms of these teleconnection patterns form the basis for the short-term climate prediction. This paper reviews advances in the study of the dynamics of these teleconnection patterns, with particular attention paid to the teleconnection patterns that significantly influence the weather and climate of East Asia.

Key words: teleconnection pattern, Arctic Oscillation, North Pacific Oscillation, Pacific/North American Pattern, climate anomaly over East Asia

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1. Introduction

The middle and high latitudes of the Northern Hemisphere have a high population density, so changes in weather and climate have been studied extensively as they heavily influence quality of life. There exist, particularly in winter, various low-frequency climate variabilities, with time periods on seasonal, interannual, decadal, or longer scales. These anomalies have characteristic spatial structures, and have been named "teleconnection patterns" by Wallace and Guzler (1981). These teleconnection patterns are classified into two groups, according to their spatial structure. The first group has two anomalous centers of different polarity running north-south, and is called "dipole-like pattern." Examples of such dipole-like teleconnection patterns are the North Atlantic Oscillation (NAO; van Loon and Rogers, 1978), the Arctic Oscillation (AO; Thompson and Wallace, 1998), and the North Pacific Oscillation (NPO; Rogers, 1981). The other group has many anomalous centers running east-west, is called "wave train-like pattern." Wave train-like teleconnection patterns include the Eurasian Pattern (EU; Wallace and Guzler, 1981), and Pacific/North America Pattern (PNA; Wallace and Guzler, 1981). The growth and development of these teleconnection patterns are believed to be closely related to local wave-mean flow interaction. These teleconnection patterns influence not only the weather and climate in their immediate vicinity, but also the atmospheric systems located thousands of kilometers away, such as the El Niño–Southern Oscillation, the

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Madden–Julian Oscillation (MJO), or the stratospheric polar vortex. Recently, the mid-high latitude teleconnections have been the focus of study for many meteorologists. This article reviews the progress so far achieved in this field, with special attention paid to those teleconnection patterns that impact weather and climate in China during winter.

The paper is organized as follows. Section 2 reviews the impacts of the AO/NAO and the NPO on weather and climate over East Asia. Section 3 reviews the Eurasian teleconnection patterns and their climate impacts. The interaction between the AO/NAO and the tropical phenomena, such as ENSO, is discussed in Section 4. Stationary waves and their relation to teleconnection patterns are described in Section 5, and some basic questions related to teleconnection patterns are answered in Section 6. Finally, Section 7 summarizes the present state of research in this field, and discusses the questions that remain to be addressed by future studies.

2. Influence of the AO/NAO and NPO on the East Asian climate

2.1 Relationship between the AO/NAO and climate anomalies over East Asia and the influence of solar cycle

The AO is a dominant pattern of mid-high latitude climate variability in the Northern Hemisphere, present on the interannual timescale. Associated with the strength variation of the AO, there is a "seesaw" in the pressure and atmospheric mass between the Arctic region and midlatitudes, manifested as the NAO over the North Atlantic region. The AO has a significant impact on the climate not only in the Arctic region, but also in low to mid latitudes (Thompson and Wallace, 1998, 2000). Several studies have reported that the East Asian winter monsoon (EAWM) tends to be weak during the positive phase of the AO (Gong et al., 2001; Wu and Wang, 2002; Ju et al., 2004; Chen et al., 2005; Suo et al., 2009). Many studies have demonstrated the interdecadal variations of the East Asian climate, and most of them attribute this to the sea surface temperature (SST) anomalies or the global warming effect (e.g., Zhou et al., 2006; Wang et al., 2008; Wei et al., 2011). Solar radiation is the primary energy source for the motion of the atmosphere, and the most important interdecadal timescale is the 11-yr solar cycle. Several studies have found that the influence of the solar cycle was enhanced through ocean-atmosphere coupling in the tropics (van Loon et al., 2007; Meehl et al., 2008; Zhou and Chen, 2012; Zhou et al., 2013). Possible solar influences on the AO or NAO have also been reported, with a confined structure in the Atlantic sector during low solar activity, and a hemispherical structure during high solar activity (Kodera, 2003).

The recent work of Chen and Zhou (2012) further investigated the modulation by the 11-yr solar cycle of the relationship between the AO and the East Asian winter climate. They found that robust warming appeared in northern Asia in response to a positive AO phase during winters with high solar activity. This corresponded to an enhanced anticyclonic circulation at 850 hPa over northeastern Asia and a weakened East Asian trough at 500 hPa, implying that the cold waves affecting East Asia were relatively inactive. However, during winters with low solar activity, both the surface warming and the intensities of the anticyclonic flow and the East Asian trough were much less pronounced in the presence of a positive AO phase. Chen and Zhou (2012) proposed that the possible mechanism for this 11-yr solar cycle modulation may be the indirect influence of solar activity on the structure of the AO. Their reasoning was that during high solar activity winters, the sea level pressure (SLP) oscillation between the polar region and the midlatitudes associated with the AO became stronger, with the significant influence of AO extending to East Asia. Meanwhile, the AO-related zonal-mean zonal winds tended to extend further into the stratosphere during high solar activity winters, leading to a stronger coupling of the troposphere to the stratosphere. These trends might have led to an enhanced AO phase difference, and thus, the associated East Asian climate anomalies became larger and more significant. The situation tended to reverse during low solar activity winters. Chen and Zhou (2012) also revealed that the relationship between the winter AO and surface climate anomalies in the following spring was also modulated by the 11-yr solar cycle, with significant signals appearing only during high solar activity phases. Thus, solar cycle variation should be taken into account when the AO is used to predict winter and spring climate anomalies over East Asia.

Several studies demonstrated the impact of the spring AO/NAO on the East Asian summer monsoon (Gong and Ho, 2003; Wu et al., 2009). Wu et al. (2009) suggested that the North Atlantic triple SST anomaly pattern, and associated subpolar teleconnections, played a dominant role during this process. Gu et al. (2009) further argued that this relationship between the AO/NAO and the summer rainfall in eastern China was not stable, and had an interdecadal variability. The recent work of Zhou (2013) revealed that the 11-yr solar cycle significantly modulated the relationship between the SST and the NAO in spring (May–June average). A typical triple pattern was evident for the correlations between the NAO and the North Atlantic SST anomalies in spring, with a center of positive values around 40°N, and negative values to both the northern and southern sides, consistent with previous study of Wu et al. (2009). Grouping the data into high and low solar activity phases, this triple feature associated with the SST anomalies was only evident in the low solar activity phase, and both the domain and the correlation value became larger. During the high solar activity phase, although there were larger areas of positive correlations between 30° and 45°N, the negative correlations to both the southern and northern sides of the 30°–45°N zone were nearly absent. Hence, the triple SST anomalies in North Atlantic tended to be much weaker from spring to summer in the high solar activity phase, which may have affected the development of subpolar teleconnection in northern Eurasian continent. In this case, there was no close relationship between the spring NAO and the summer climate anomalies in East Asia. In contrast, the spring NAO may have a robust impact on the East Asian summer climate in the low solar activity phase (Zhou, 2013). However, the issue of how the 11-yr solar cycle influences the persistence of NAO-related triple SST anomalies in North Atlantic is still open, and may benefit from further investigations including diagnostic and numerical simulations.

2.2 Influence of the NPO on the East Asian winter climate

The NPO manifests as the second eigenvector of the SLP field over North Pacific on interannual timescales, with a teleconnection mode of opposite pressure variations over the Aleutian Islands and Hawaii (Wallace and Gutzler, 1981). When the NPO is in positive phase, the north-south dipole pattern in the SLP shows an anomalous cyclone around the Bering Strait and an anomalous anticyclone in the whole of the subtropical Pacific, with a barotropic structure extending from the lower troposphere to the tropopause region (Wang et al., 2011). It is generally recognized that in winter, the NPO influences the climate over the "downstream" region of North America (Linkin and Nigam, 2008) and also the strength of the EAWM. During boreal winter, the subtropical Pacific anticyclone associated with the NPO can extend westward to the coast of East Asia and weaken the northwesterlies of the EAWM. Consequently, significant warming is observed over East Asia and more precipitation appears in the south of the Huai River area (Wang et al., 2011).

In the mid 1970s, the atmospheric circulation underwent a significant change over North Pacific, as the Aleutian low (AL) both deepened and shifted eastward (Trenberth and Hurrell, 1994). Wang et al. (2007) also found a corresponding change of the typical period of the NPO in the mid 1970s, and a related interdecadal change in the relationship between the NPO and East Asian winter climate. Before 1976, the influences of the NPO on the air temperature over East Asia were significant along and close to the coast, but very weak north of 40°N inland. However, this situation reversed after 1976. This interdecadal change was suggested to be related to the weakening of the remote forcing from the tropical eastern Pacific SST anomaly, associated with the ENSO to the NPO (Wang et al., 2007). The ENSO had a dominant influence on the NPO via the stationary wave propagation from the equatorial central-eastern Pacific to North Pacific before 1976. At the same time, the NPO mainly influenced the winter air temperature over southern East Asia, indicating a close link to the southern path of EAWM. After 1976, the atmospheric circulation related to the NPO was quite different, and exhibited a circumglobal wave train pattern over the extratropical regions in the Northern Hemisphere. There were enhanced northward wave propagations, particularly over the extratropics of East Asia. Hence, the influence of the equatorial central-eastern Pacific SST on the NPO was very weak after 1976. The NPO mainly influenced the winter air temperature over northern East Asia, thus indicating a close link to the northern path of the EAWM (Wang et al., 2009a).

The change around 1976/77 in North Pacific also appears in the ocean itself, referred to as the Pacific Decadal Oscillation (PDO; Mantua et al., 1997). In the warm phase of the PDO, the SST anomalies are cold in central North Pacific, and warm in the inshore waters along the west coast of the Americas. The situation is roughly opposite in the cold phase of PDO. The PDO is a strong signal of climate variability on the interdecadal timescale. On the one hand, the PDO is a disturbance superimposed on the long-term climate trend, which can directly lead to interdecadal changes of climate in the Pacific and its surrounding area. On the other hand, the PDO is an important background for the interannual variability, significantly modulating the interannual variation, such as the ENSO and its impacts (Yang et al., 2004; Zhu et al., 2008; Feng et al., 2014). The typical influence of ENSO on the North American climate was found to be modulated by the PDO (Gershunov and Barnett, 1998). The ENSO-Australian precipitation relationship was found to be robust only during the low phase of the PDO (Power et al., 1999). Several recent studies also indicated that the relationship of East Asian winter climate to the tropical and extratropical systems underwent a significant change around 1976 (Wang et al., 2007, 2008, 2009b, 2010). They found that the impacts of the atmospheric internal signals over mid-high latitudes such as the Ural blocking, the NPO, etc., on the East Asian winter climate were enhanced, whereas the impact from ENSO weakened. The climate over East Asia was warmer during El Niño winters, while the EAWM was usually weaker (Zhang et al., 1996; Chen et al., 2000). However, this relationship was also subject to interdecadal change, particularly related to the phase of the PDO. The close relationship between the ENSO and the winter climate over East Asia was robust only in the low phase of the PDO. This modulation may be accounted for by the difference in the strength and location of the anticyclone over the Philippine Sea during the different phases of the PDO (Wang et al., 2008; Chen et al., 2013a). Since the PDO was in its high phase from 1976 to the beginning of 21st century, the impact of ENSO on the East Asian winter climate was weaker than that before 1976.

The surface air temperature (SAT) over East Asia can be significantly influenced by the second singular value decomposition (SVD2) mode, a metric obtained by applying an SVD analysis on 500-hPa geopotential height in the Northern Hemisphere and SST in the tropical Pacific Ocean, with both observations and output of seasonal forecasts from atmospheric general circulation models (Jia and Lin, 2011). The atmospheric component of SVD2 has been found to share many spatial similarities with the AO. The time variation of SVD2, however, was found to be more closely correlated to the variation of SAT over China than the AO. When the SVD2 was in its positive phase, the SAT over China tended to be warmer than normal. Furthermore, the seasonal forecasts of SAT and precipitation over China were post-processed by using a statistical approach (Jia et al., 2010, 2014), based on the relationship mentioned above. The results present a significant improvement. The mechanisms of the influence of this mode on East Asian climate, and the relationship of this mode to the tropical SST and NPO require further investigation.

3. Teleconnection patterns over the Eurasian continent and their climate impacts

3.1 Ural blocking and the East Asian winter climate

During winter, the Ural Mountains region has one of the highest blocking frequencies in the Northern Hemisphere. This blocking is of great importance for the weather and climate of downstream regions such as East Asia (Ding and Krishnamurti, 1987; Takaya and Nakamura, 2005). Most previous studies have focused on cases of blocking based on the definition of blocking by Liu (1994). They first suggested that the persistence of a blocking pattern could be expressed in terms of the duration of the corresponding anomaly patterns, and defined a blocking index to measure the resemblance of a particular circulation pattern to the blocking regime. Based on this index, Barriopedro et al. (2006) developed a modified version to capture the winter mean regional anomaly signature of the Ural blocking regime. The blocking frequency in the region was defined as high when the Ural blocking index (UBI) was positive, and vice versa. Using the UBI, Wang et al. (2010) investigated the interannual variation of winter Ural blocking activity and its impact on the EAWM. They suggested that the Ural blocking exerted its impact via a quasi-barotropic wave train over the Eurasian continent. When the Ural blocking activity was strong, this wave train propagated both upward and eastward, and negative geopotential height anomalies were induced over East Asia. Hence, the East Asian trough deepened and the EAWM strengthened.

However, deciphering the causes of interannual variation of the Ural blocking is difficult, since the variability from internal dynamics is much larger than that from external factors. The previous study of Li (2004) suggested that the North Atlantic SST anomalies could affect the interannual variation of Ural blocking in early winter. However, this influence became weak in mid winter. In mid winter, the NAO had a close relationship with the blocking over North Atlantic, whereas its relation to the Ural blocking was weak (Huang J. P. et al., 2006). There are increasing pieces of evidence suggesting that the stratospheric circulation anomaly can influence the weather and climate in the troposphere (Baldwin and Dunkerton, 2001; Chen and Kang, 2006; Huang et al., 2007; Chen et al., 2013c). Recently, Wang et al. (2010) showed that the stratospheric polar jet had a close relationship with the UBI, especially when the stratosphere led the troposphere by one month. Further analysis indicated that the Ural blocking in winter had become weak in recent decades, leading to a decreased blocking frequency. However, the relationship between the Ural blocking activity and the East Asian trough and Siberian high–both may represent the EAWM–became much closer. Particularly after 1976/1977, strengthened stratospheric polar night jet and polar vortex prevented the planetary waves from propagating into the stratosphere. Therefore, the Ural blocking signal was found to propagate further eastward in the troposphere to East Asia, thus exerting more influences on the East Asian winter climate. Evidently, further study on how the stratospheric polar night jet influences the Ural blocking is needed.

3.2 The Eurasian teleconnection (EU) patterns and their impact on climate

China has a wide meridional coverage, so climate-linked disasters such as drought, flood, and winter freezing rain/snow storms are related to both the tropical SST and the mid-high latitude atmospheric circulation. For example, Huang Ronghui et al. (2006) found an obvious interdecadal change around 1976, with persistent drought in North China and increased rainfall in the Yangtze and Huai River valleys. They further suggested that this change was associated with an EU anomaly. Several further studies also suggested that a persistent positive EU-like anomaly played a dominant role in the record-breaking, long-lasting cold episode over southern China in January 2008 (Wen et al., 2009; Zhou et al., 2009). However, there were three EU patterns identified in previous studies by Wallace and Gutzler (1981) and Barnston and Livezey (1987), so it was not clear which EU pattern played the most important role. Hence, Liu et al. (2014) conducted a comprehensive side-by-side comparison among the three EU patterns, including their temporal variability, three-dimensional structure, and possible mechanisms.

The conventional EU pattern was first identified by Wallace and Gutzler (1981). Later, Barnston and Livezey (1987) identified two different EU-like patterns, using the rotated empirical orthogonal function (REOF) method to decompose the anomalies of monthly geopotential height at 700 hPa in the subtropics of the Northern Hemisphere. They were named the EU1 and EU2 by Barnston and Livezey (1987), and were later called the Scandinavia (SCAND) and the East Atlantic/West Russia (EATL/WRUS) patterns by the US Climate Prediction Center/National Oceanic and Atmospheric Administration (CPC/NOAA). Liu (2013) identified the teleconnection patterns by applying the REOF method to the monthly mean 500-hPa geopotential height anomalies in the extratropics of the Northern Hemisphere. Figure 1 presents the obtained REOF spatial distributions and their time series. The REOF5 had four centers of action, with two negative centers around the Scandinavian Peninsula (50°–70°N, 0°–20°E) and East Asia (40°–50°N, 110°–130°E) and two positive centers over the Atlantic Ocean to the west of Europe (40°–55°N, 30°–10°W) and Siberia (55°–75°N, 60°–90°E) (Fig. 1a). This mode somewhat resembled the conventional EU pattern (Wallace and Gutzler, 1981). Liu (2013) calculated the monthly EU index according to the definition of Wallace and Gutzler (1981) from January 1948 to December 2009. The temporal and spatial correlation coefficients between the REOF5 and the conventional EU pattern were 0.64 and 0.88, respectively. Hence, the REOF5 was identified as the conventional EU pattern. The mode shown in Fig. 1c had three centers of action, with two negative centers over western Europe (40°–55°N, 30°W–10°E) and Siberia (45°–60°N, 80°–105°E) and a positive one around the Scandinavian Peninsula (60°–70°N, 20°–40°E), resembling the EU1 (Barnston and Livezey, 1987) or SCAND pattern. The correlation coefficient between the time series of REOF3 and the monthly mean SCAND index provided by CPC/NOAA was 0.76. Therefore, the REOF3 was the SCAND pattern identified in previous studies. The REOF10 had three centers of action, with a positive center over western Europe (55°–60°N, 15°W–0°E) and two negative centers north of the Black and Caspian seas (45°–60°N, 30°–45°E) and North Atlantic (40°–50°N, 60°–40°W) (Fig. 1e), resembling the EU2 (Barnston and Livezey, 1987) or EATL/WRUS pattern. The correlation coefficient be-

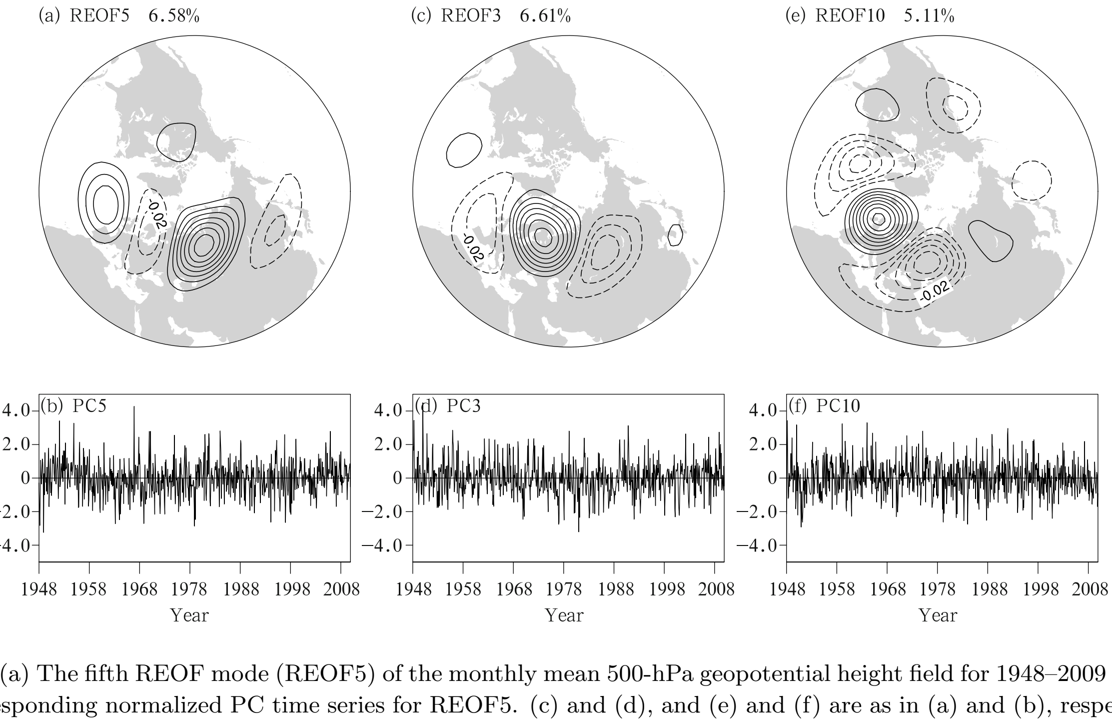

Fig. 1. (a) The fifth REOF mode (REOF5) of the monthly mean 500-hPa geopotential height field for 1948–2009 and (b) the corresponding normalized PC time series for REOF5. (c) and (d), and (e) and (f) are as in (a) and (b), respectively, except for the third and tenth REOF modes. The percentages at the top right indicate the variance explained by each REOF mode. The contour interval in (a), (c), and (e) is 0.01 m. The zero contour lines are omitted.

tween the REOF10 time series and the monthly mean EATL/WRUS index provided by the CPC/NOAA was 0.62. Therefore, the REOF10 reflected the EATL/WRUS pattern identified in previous studies.

Figure 2 shows the monthly evolution of the variance of the EU, SCAND, and EATL/WRUS indices. The variances of the three teleconnection patterns were larger during January–March and October–December, but smaller during May–August. Among the distributions of the three index variances there were also differences. The EU had a maximum in December, its secondary extreme value in February, and was low during May–August with the minimum in July. The SCAND had a maximum in November, its secondary extreme value in January, and was low during May–August with the minimum in July. The EATL/WRUS had a maximum in January, its secondary extreme value in February, and its minimum in July. Generally, the three teleconnection patterns had larger variances during autumn (October and November) and winter (December–February) and smaller variances during summer (June and July). Their different climate impacts were compared in the Northern Hemisphere winter (Liu and Chen, 2012; Liu, 2013). The research on the winter EU indicated that in winter, during a positive EU phase, cold anomalies appeared in most of China, with significant cooling in the south to the Yangtze River, and the region from Inner Mongolia to southern Northeast China (Liu and Chen, 2012). In a winter with positive SCAND phase, significant cooling anomalies occurred in most of eastern China and Xinjiang Region, and significant warming anomalies occurred to the east of the Tibetan Plateau. The influences of the EATL/WRUS pattern on the climate in China were weak, with the SAT anomalies mainly to the north of 40°N. In addition, Liu and Wang (2014) found that the SCAND pattern experienced a clear and abrupt change in 1979. Further analysis indicated that the centers of both the SCAND patterns over Europe and Siberia at 500 hPa extended further southeastward after 1979. The center over the Scandinavian Peninsula, however, showed little change. Accompanied with the changes in spatial pattern after 1979, the influences of the SCAND pattern on the wintertime SAT in the Northern Hemisphere were also enhanced. In particular, there were enlarged regional temperature anomalies in the polar zone associated with the positive (negative) phase of

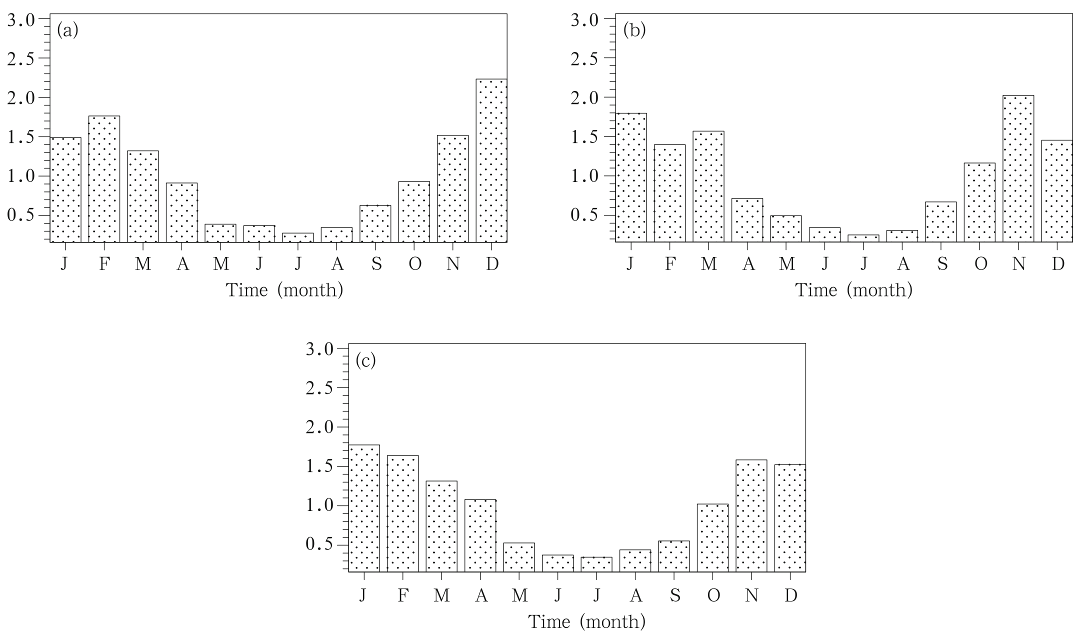

Fig. 2. Monthly evolution of the variance of the (a) conventional EU index, (b) SCAND index, and (c) EATL/WRUS index.

the SCAND phase after 1979. In the meantime, negative (positive) SAT anomalies associated with the positive (negative) phase of the SCAND pattern over the northern Eurasian continent extended further southeastward, even reaching the Yangtze River valley and Japan.

3.3 Cyclones and anticyclones over China in winter

Cyclones and anticyclones are main weather systems influencing mid-high latitudes, and are very active during the EAWM, so an investigation of wintertime cyclones and anticyclones is helpful for understanding the variation of the EAWM.

3.3.1 Climatological features

Most past cyclone and anticyclone studies over China have been based on poor-quality datasets, spanning short time periods, or subjective methods using weather charts (e.g., Chen et al., 1991; Zhu et al., 2000). At the same time, objective methods for detecting and tracking cyclones and anticyclones based on high-quality reanalysis datasets encompassing longer time periods and automatic numerical algorithms have been rapidly developing, and applied to cyclone and anticyclone studies (Murry and Simmonds, 1991; Hodges, 1994; Simmonds and Keay, 2000). Wang X. et al. (2009) studied cyclones over East Asia based on the objective detecting and tracking method and the ECMWF reanalysis dataset. They found that the highest cyclone frequency was in spring, and the lowest in winter. Moreover, winter cyclone frequency was low from 1958 to the mid 1970s, increased until the mid 1980s, and subsequently decreased again. However, the annual mean cyclone intensity (defined as the minimum pressure in each cyclone center) had a decreasing trend for the time period they considered. Based on the NCEP/NCAR reanalysis data, Chen L. et al. (2014) studied the wintertime cyclone and anticyclone activities over China for the time period 1948–2007. They found that among the cyclones and anticyclones affecting China, half were generated out of China (mainly in Inner Mongolia) and half within China. The cyclones and anticyclones assumed significant asymmetry in their source, path, and lysis. For the cyclones, their primary sources were, in order, Inner Mongolia, Northeast China, and the Jianghuai region. A secondary source was the zone from Xinjiang to North China to the northeast of Southwest China. For the anticyclones, Inner Mongolia and its surrounding, particularly northeastern Inner Mongolia, was a primary source region. A secondary source region was the zone from eastern Northwest China across Shanxi-Henan-Hebei to the Jianghuai region. The island of New Siberia was also a main source of anticyclones.

Away from their source regions, the cyclones tended to move eastward or northeastward across East Asia. The entire North Pacific, and the Northeast China–Jeju Island and further eastward, were a lysis region. The anticyclones moved largely southwestward, and died out most frequently over East Asia and its neighboring West Pacific. The asymmetrical cyclone and anticyclone path may result from the steering by the stationary East Asian trough. Typically, cyclones (anticyclones) were located east (west) of the upper wave trough axis. Upper waves fully developed and became stationary when they moved upstream to the climatological location of the East Asian trough due to land-sea thermal forcing. Under the steering of the East Asian trough, the low-level anticyclones moved along the northwesterlies of the East Asian trough and diminished at lower latitudes. In contrast, the low-level cyclones were steered along the southwesterlies of the East Asian trough and travelled long distances, occasionally reaching even Northeast Pacific (Chen L. et al., 2014).

3.3.2 Relation to upper-level jets

In the past decades, significant changes in the Northern Hemisphere wintertime cyclone and anticyclone activities have been detected. The cyclone frequency increased in high latitudes but decreased at midlatitudes during 1957–1997, while the cyclone intensity increased for both latitude bands (McCabe et al., 2001). Such changes were most significant during the mid 1970s. The North Atlantic showed the same trend as the whole Northern Hemisphere. Over North Pacific, the cyclone path was more southward, and its intensity increased at the same time (Chang and Fu, 2002; Nakamura and Izumi, 2002; Nie et al., 2008;

Zhang et al., 2012).

Since the mid 1980s, the EAWM has weakened significantly (Xu et al., 2006; Gao, 2007; Wang et al., 2009b; Wang and Chen, 2010). This was associated with the increasing frequency of the cyclones and anticyclones, i.e., being lower (higher) before (after) the mid 1980s (Chen L. et al., 2014). Responding to this change, the East Asian polar front jet was weaker (stronger) before (after) the mid 1980s. This means that the increase in strength of the polar-front jet was likely partially responsible for the increase in the cyclonic and anticyclonic activities (Chen L. et al., 2014).

4. Interaction between the AO/NAO and tropical atmosphere

4.1 Influence of the spring AO on ENSO

ENSO is the strongest interannual variability mode of the climate system, with signals in the tropical Pacific SST, and associated SLP and atmospheric and oceanic circulations. ENSO may directly induce weather and climate anomalies in the tropical Pacific, but it also indirectly exerts substantial influences on the global weather and climate via teleconnection (e.g., Huang and Wu, 1989; Zhang et al., 1996; Wang et al., 2000; Huang et al., 2004). Recently, several studies have indicated that the AO had a significant impact not only on the northern extratropical weather and climate, but also on the tropical climate variability (e.g., Zhou and Miller, 2005; Choi et al., 2012). In addition, the spring AO influenced the following ENSO occurrence. Nakamura et al. (2006, 2007) demonstrated that the anomalous westerly winds in the tropical western Pacific associated with a positive AO phase in spring tended to propagate eastward during the following summer and autumn, triggering an ENSO event. They suggested that dry and cold surges occurred more frequently during the spring with positive AO phase. The inflow of surge associated cold advections would strengthen the convective activity in the tropical western Pacific, leading to the Gill type response in the atmosphere (Gill, 1980). A pair of anomalous cyclonic cells appeared off-equator on both sides of the equatorial western Pacific. Correspondingly, the anomalous westerly winds were seen in the equatorial western Pacific. However, this interpretation contradicts the results of Jeong and Ho (2005), who found that the frequency of the wintertime cold surge occurrence was higher in negative than in positive and neutral AO phases. Chen et al. (2014a) further explored the physical process of the influence of AO in spring on the ENSO in the following winter. They suggested that the formation of the westerly wind burst in the tropical western Pacific was not caused by cold surge activity, but via the interaction between synoptic-scale eddies and the mean flow over North Pacific. The equatorial westerly anomalies excited downwelling equatorial Kelvin waves, which propagated subsequently to the equatorial central-eastern Pacific, leading to SST warming there in summer through to fall. The tropical SST, atmospheric heating, and atmospheric circulation anomalies were sustained and developed through the Bjerknes positive feedback mechanism, which resulted in an El Niño event in the tropical eastern Pacific in fall and winter. Chen et al. (2014b) also found that the AO–ENSO relationship experienced a pronounced interdecadal change at the beginning of the 1970s. The spring AO influence on the subsequent ENSO was weak before 1970. In contrast, the spring AO exerted a substantial influence on the following winter ENSO after 1970.

Moreover, several studies have demonstrated that the wintertime NPO was able to affect the occurrence of ENSO in the following winter via the seasonal footprinting mechanism (SFM; Vimont et al., 2001, 2003). Figures 3a–c present the first leading SVD mode of the wintertime (November–March average) SLP over North Pacific and the SST in the following winter (October–February average) over the tropical Pacific, and its time series. The SVD1 accounted for 62% of the total covariance. The anomalous SLP pattern in Fig. 3a bears close resemblance to the NPO structure obtained from the Empirical Orthogonal Function (EOF) analysis (Linkin and Nigam, 2008). The anomalous SST pattern in Fig. 3b illustrates a typical El Niño peak phase during the following winter. The regression patterns of the SLP in winter and the SST during the following winter upon the normalized SLP-PC1 are further displayed in Figs. 3d and 3e to confirm that the wintertime extratropical NPO was significantly linked to the ENSO in the tropical Pacific in the following winter. However, studies also found that NPO-like atmospheric forcing from midlatitudes did not always trigger an ENSO (Alexander et al., 2010; Park et al., 2013). The spring AO can influence the following winter SST in the equatorial central-eastern Pacific. Further, the springtime SST anomalies over the subtropical North Pacific associated with the spring AO bore close resemblance to those associated with the preceding wintertime NPO. Motivated by these qualitatively similar structures of SST anomalies, Chen S. et al. (2013) hypothesized that the spring AO related SST anomalies could strengthen or weaken the SST "footprint" in the upper subtropical North Pacific generated by the preceding winter NPO. This would then influence the formation of the westerly wind anomalies generated by the air-sea interaction in the following summer and finally influenced the

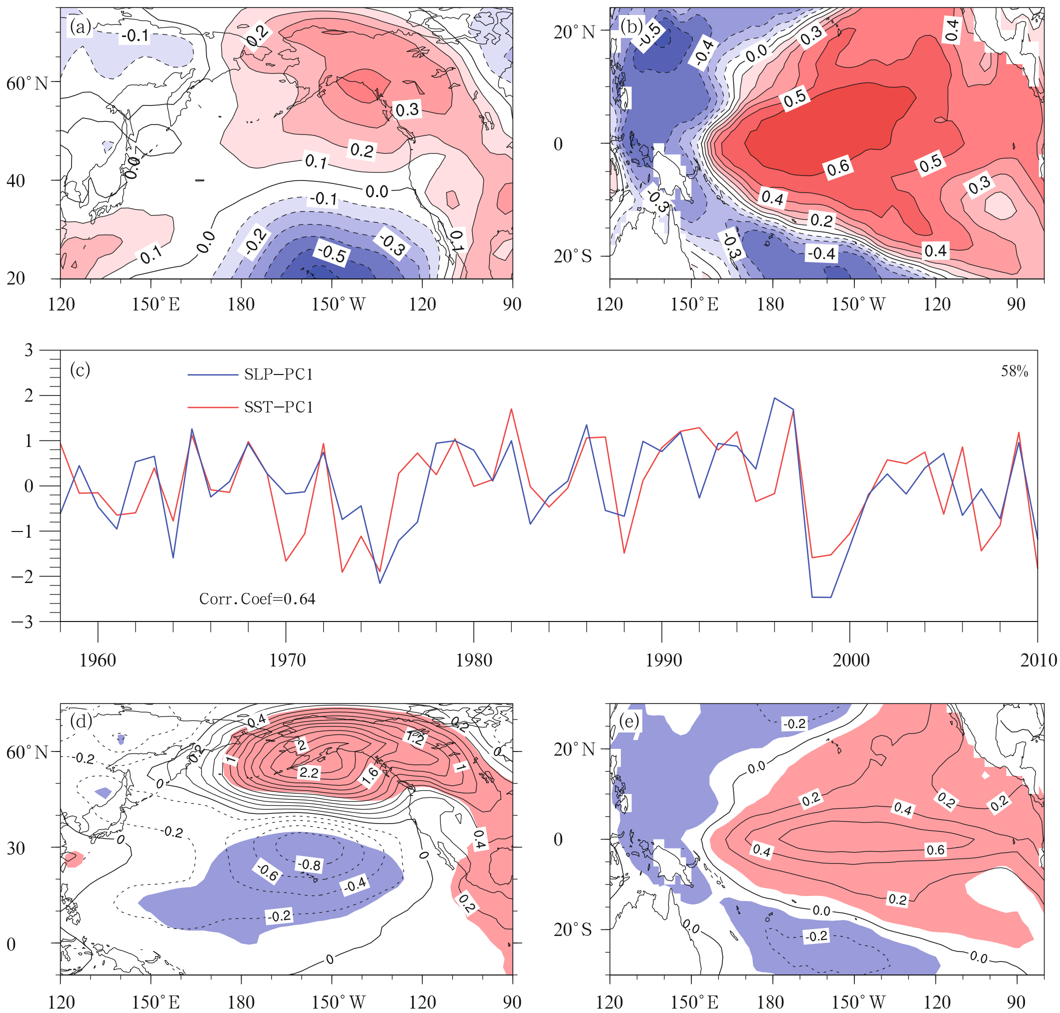

Fig. 3. The first leading SVD homogeneous modes of (a) wintertime (i.e., November through next March) SLP over North Pacific (20°–80°N, 120°E–90°W) and (b) SST in the following winter (i.e., Ocotober through next February) over the tropical Pacific (25°S–25°N, 120°E–80°W), and (c) the time series of the first leading SVD mode. Anomalies of (d) SLP as for (a), and (e) SST as for (b), regressed onto the normalized time series for SLP in (c). The shadings in (d) and (e) indicate the anomalies different from zero at the 5% significance level. Contour intervals are 0.2 hPa in (d) and 0.2℃ in (e). The zero contour lines are omitted and negative values are dashed.

SFM-driven SST anomalies in the equatorial centraleastern Pacific. With diagnostics of the data, Chen S. et al. (2013) presented evidence that the spring AO exerted a robust modulation on the connection between the NPO and ENSO. Only when the spring AO was in positive phase, could a positive (negative) phase of wintertime NPO lead to significant El Niño-like warming (La Niña-like cold) anomalies in the tropical central-eastern Pacific during the following winter via the SFM. However, the connection between the NPO and ENSO was weak when the spring AO was in its negative phase.

4.2 Relationship between the North Atlantic jet and tropical convective heating

The NAO has been shown to exhibit variability on a wide range of timescales: from its intrinsic timescale of about 10 days to decades and even centuries (Hurrell and Loon, 1997; Feldstein, 2000; Semenov et al., 2008). The NAO variability may be caused by the internal dynamics of the extratropics (Robinson, 1993; Yu and Hartmann, 1993; Lee and Feldstein, 1996) or by tropical heating. The NAO may in turn affect, through exciting a wave train in the area downstream, even the tropical Indian Ocean and West Pacific (Palmer, 1988). On the decadal timescale, Hoerling et al. (2001) found by numerical modeling that a warming of the Indian and Pacific Ocean surface was able to force a circulation pattern for later half of the 20th century, very similar to the positive phase of the NAO. The same result was obtained by later numerical studies (Hoerling et al., 2004; Selten et al., 2004; Bader and Latif, 2005; Zhou and Miller, 2005). These findings contrast with those of Cohen and Barlow (2005), who argued that the SST trend in the Indo–western Pacific warm pool region was unrelated to the NAO trend.

On the intraseasonal timescale, there was also evidence that tropical convection can modify the wintertime circulation over North Atlantic. For example, phase 3 of the tropical Madden–Julian Oscillation (MJO) index, characterized by enhanced convection over the Indian Ocean and reduced convection over the western Pacific Ocean, was associated with positive polarity of the NAO. When the MJO was observed to be in phase 6, which exhibits opposite convection characteristics to that of phase 3, the NAO index tended to coincide with its negative polarity (Zhou and Miller, 2005; L' Heureux and Higgins, 2007; Cassou, 2008; Lin et al., 2009; Lin and Brunet, 2011).

Considering the NAO's downstream influence, Hsu et al. (1990) found that a subtropical Rossby wave train excited by the NAO and propagating southeastward from the North Atlantic to the Indian Ocean, was a trigger for convection over the Indian Ocean. Similar high-frequency Rossby wave trains were also observed to propagate southeastward from the Asian-Pacific jet and forced convection over the central tropical Pacific (Matthews and Kiladis, 1999; Palmer, 1988).

Bader and Latif (2005) suggested that tropical convection over the Indian Ocean and the NAO may be linked through a circumglobal teleconnection pattern (CTP). Based on the same idea, Yuan et al. (2011) investigated further the relationship between convection over the tropical Indian and West Pacific oceans. They found that on the subseasonal scale, negative (positive) NAO events led to enhanced precipitation over the tropical Indian (West Pacific) Ocean. They also found that convection over the tropical Indian (West Pacific) Ocean led to a positive (negative) NAO. On the decadal scale, convection over the tropical Indian Ocean, compared with the tropical West Pacific, enhanced significantly from 1958–1979 to 1980–2001, possibly indicating a positive phase trend of NAO during 1980–2001.

5. Stationary waves and teleconnections over North Pacific in winter

Stationary waves are mostly active during winter in the Northern Hemisphere. At the surface, stationary waves assume a structure with three semipermanent atmospheric centers of action: the Siberian high, AL, and the Icelandic low (IL). In the mid and upper troposphere, stationary waves manifest in the height field as a negative anomalous center over North Pacific, a positive center over western North America, a negative center over eastern North America, and a positive center over eastern North Atlantic and downstream. These stationary waves have apparent variability on seasonal to interannual and decadal timescales, heavily affecting the circulation and climate over the wintertime Northern Hemisphere.

The three-dimensional propagation of stationary waves is well described by the wave activity flux (Plumb, 1985). The flux is parallel to the local group velocity and thus provides a good indicator of wave energy propagation. In the lower and mid troposphere, there exist two centers of wave activity over East Asia and western North Pacific, and North Atlantic, respectively. There also exists a center over the eastern North Pacific and North America. The first two centers are called the East Asia/West Pacific and North Atlantic wave trains, respectively, while the last center is called the eastern North Pacific wave train (EPW), which is weaker than the previous two (Plumb, 1985; Zhou et al., 2012).

5.1 EPW and PNA

Though it is weaker in magnitude, the variance of EPW is as large as the other wave trains, which means that the EPW has apparent variability that impacts the circulation and climate over the EPW area, its surroundings, and even beyond. Zhou et al. (2012) defined an EPW index, in which the vertical wave flux is chosen, before being averaged over the lower to mid troposphere over the eastern North Pacific for each winter. The results showed that the index was in its second maximum during 1958–1964, weakest in 1965–1975, and reached its maximum in 1976–1987. After 1987, the EPW weakened again (Zhou et al., 2012).

To ascertain the EPW's structure in the geopotential height field for its weak and active phases, the wave index was standardized to enable a winter to be labeled as an active (a non-active) winter when the standardized index of the winter was larger than 1 (less than −1). The composite of geopotential height shows that at the surface, the AL was very active, with its center beyond the International Date Line (IDL) for the active winters. In contrast, the AL was weak, with its center located west of the IDL for weak winters. In the upper troposphere and for active winters, the negative anomalous center over North Pacific was strong and extended from the East Asian coast to the southeast of the Aleutian Islands; the positive height anomaly over western North America was also strong. For weak winters, the negative height anomaly over North Pacific was weak and located over West Pacific to the west of the IDL, while the positive height anomaly over western North America was very weak with its center shifted westward to the eastern North Pacific. Very interestingly, the composite height difference of the active minus weak winters assumed a PNA-like pattern over the North Pacific-North America sector. The EPW resembled the PNA not only in spatial structure, but also in time variability. The correlation was as high as 0.7 for the EPW and PNA indices (Zhou et al., 2012).

Therefore, EPW and PNA were the same phenomenon, but in different physical fields. PNA described the anomalous pattern in the geopotential field and the EPW was a wave train in a wave flux field. Physically, when the atmospheric centers of action such as the AL and the upper-layer anomalous centers over North Pacific and North America varied in their location or strength, low-frequency waves were excited and propagated downstream and vertically from the lower troposphere to the stratosphere. In this way, the low-frequency PNA-like wave train formed. During active (weak) winters, the EPW was strong (weak), and assumed a positive (negative) phase PNA (Zhou et al., 2012). Both the horizontal and vertical propagations of the EPW were of fundamental importance to atmospheric teleconnections discussed in the next section.

5.2 EPW, AL–IL seesaw, and AO

The AO is a strong signal of climate variability in the wintertime extratropics (Thompson and Wallace, 1998, 2000). Since the work of Thompson and Wallace (1998), however, there has been an ongoing debate over the question of whether the AO pattern is a real physical mode or if it is just a consequence of the EOF methodology used in its definition (Deser, 2000; Ambaum et al., 2001; Wallace and Thompson, 2002). On the one hand, due to the fact that AO is strongly correlated with NAO, it was believed that AO is actually the NAO, not a new mode. However, if this is true, the existence of the Pacific center cannot be explained. On the other hand, Thompson and

Wallace (2000) proposed that the AO was a fundamental mode of climate variability, describing the seesaw of mass between the Arctic and midlatitudes. If this were the case, the three centers of action of the AO, i.e., the Arctic center, Atlantic center, and Pacific center, should be closely correlated. Deser (2000), however, found that among the three centers, only the Arctic and Atlantic centers were strongly correlated, exhibiting a negative value, indicative of the well-known NAO, while the Pacific center was only weakly correlated with the other two centers. Consequently, the debate on the AO mechanism has lasted for a long time.

The Arctic–Pacific link is generally weak, though some studies have shown that the AL and IL do not fluctuate independently; rather, the two lows exhibit a strong negative correlation from one winter to the next (Kutzbach, 1970; van Loon and Rogers, 1978; Wallace and Gutzler, 1981; van Loon and Madden, 1983). This negative correlation is called a seesaw oscillation, and its formation and relationship to the AO formation are very interesting.

Honda et al. (2001) found that the AL–IL seesaw showed a seasonal dependence, with the most significant seesaw observed in late winter (February to mid March) during 1973–1994. They further demonstrated that the AL–IL seesaw may be excited by a PNA-like wave train propagating across North Pacific and North America, and then to North Atlantic. Castanheira and Graf (2003) presented another interesting finding that the AL–IL seesaw may be impacted by the stratospheric polar vortex. For a strong (weak) polar vortex, the seesaw was remarkably strong (weak). This means that the polar vortex may have a significant impact on the AL–IL seesaw formation by its flexion of the upward propagation of planetary waves from the troposphere.

Recently, Sun and Tan (2013) further investigated the seasonal and interannual variation and mechanism of the AL–IL seesaw and proposed that both the horizontal and vertical propagation and the reflection of the polar vortex may contribute to the AL–IL seesaw formation. They showed that the AL–IL seesaw exhibited apparent seasonal, interannual, and decadal variability. The AL and IL correlation was –0.33 and –0.35 for December and January, respectively. For February the correlation was –0.26, smaller than in December and January, while in March the two lows showed no statistically significant correlation. The correlation for the entire winter was –0.26. Interdecadal variability in the AL–IL correlation was also apparent. In December, the two lows were significantly negatively correlated only for the sub-period 1995–2009 (the correlation was –0.55), while for the other two periods, 1948–1972 and 1973–1994, the AL had no significant relation to the IL. In January, the AL–IL correlation was –0.4 or so for the three sub-periods, however, only for period 1948–1972 was the AL–IL correlation statistically significant; while in February the AL–IL correlation reached its highest value of –0.73 for 1973–1994, and there was no statistical significance for the other two sub-periods. Therefore, the conclusion of Honda et al. (2001) that the AL–IL seesaw was most significant in late winter was only apparent for the sub-period 1973–1994, not for the entire period of 1948–2009. For March, AL and IL still showed no statistically significant correlation for any of the three sub-periods.

Sun and Tan (2013) considered the effects of the EPW and the polar vortex on the AL–IL seesaw, separately and combined. Without considering the strength of the EPW, the AL and IL had no significant correlation in the case of a weak polar vortex, while for the strong polar vortex case, the AL–IL correlation reached –0.36. Without considering the strength of the polar vortex, the AL and IL had no relation for a weak EPW (strength below average) and the AL–IL correlation was –0.42 when EPW was strong (strength above average). The stronger the EPW, the higher the AL–IL correlation. Therefore, it was clear that both EPW and polar vortex contributed to the AL–IL seesaw, and were able to work cooperatively. For weak EPW and polar vortex, the AL and IL fluctuated independently. For weak EPW, the AL–IL correlation was as high as –0.34 under strong polar vortex conditions, so the polar vortex had a strong impact on the seesaw for the weak EPW case. When the EPW was strong, the AL–IL correlation was strong, independent of the polar vortex condition. This indicates that for a strong

EPW case, the AL–IL seesaw was mainly caused by horizontal propagation of the EPW from East Pacific through North America, before turning toward Iceland. Sun and Tan (2013) provided evidence of the horizontal and vertical propagation of EPW, supporting the above statistical results.

Sun and Tan (2013) argued that the high inverse correlation between the AL and IL actually depicted the NAO, while the Pacific center in the AO may hint at the presence of the AL–IL seesaw. To confirm their argument, Sun and Tan (2013) performed EOF analyses with the months of strong AL–IL seesaw and the remaining months. The first EOF showed a stronger Pacific center than the climatology for the former case, and the Pacific center disappeared from the EOF pattern for the latter case. This implied that the AO was really an imprint left by both the AO and the AL–IL seesaw.

6. Further consideration of the factors associated with atmospheric teleconnection

6.1 Formation of temperature anomaly associated with AO

In the geopotential height field, the AO has a pattern of three anomalous centers: the Artic center, the North Atlantic Center, and the North Pacific center; while in the lower-level temperature field, the AO pattern has three anomalous cooling centers over the Arctic Ocean, Greenland, and northeastern Canada, respectively, and two warming centers over Eurasia and the United States of America. Thompson and Wallace (1998) showed that the temperature anomalies may be caused by temperature advection induced by zonal mean anomalous zonal wind. Enhanced westerlies associated with the positive AO phase bring more warm marine air into the interior of continents, warming the air there, and pushing cold continental air into the western oceans, cooling the air there. Their explanation emphasized the role of land-sea thermal contrast. Some weakness, however, still exists in their explanation. First, they used the zonal mean anomalous zonal wind, not the anomalous zonal wind itself, to calculate the anomalous temperature advection, which cannot reflect adequately the local characteristics of the temperature anomalies. Second and more importantly, they did not consider the role of temperature advection induced by the anomalous meridional wind and the temperature contrast between the higher and lower latitudes, which is the most dominant temperature gradient. In view of this argument, and based on NCEP/NCAR reanalysis data, Suo et al. (2009) recalculated the temperature anomalies for a positive AO, and obtained a better result. The temperature advection by the anomalous zonal wind north of 40° brought warm ocean air over the continents and cold land air over the oceans, dominantly contributing to the cooling over Greenland, its surroundings, and the Bering Sea. The anomalous meridional wind assumed a wave-like form with wavenumber 3, which brought warm air northward and cold air southward, and caused warming in the United States and cooling in southern Europe and Canada. The temperature advection by both the zonal and meridional wind anomalies caused the warming in northern Eurasian continent.

6.2 Mechanism of poleward propagation of the anomalous westerlies over North Atlantic

Wave-mean flow interaction is coherently a mechanism responsible for low-frequency variability on various timescales. The poleward propagation of the anomalous westerlies is a direct consequence of wave-mean flow interaction. Examination of this poleward propagation aids understanding of the mechanism of the AO and Antarctic oscillation. Evidence of poleward propagation was first shown in the observational study of Riehl et al. (1950). James et al. (1994) found that similar poleward propagation could also occur in an idealized model. With a barotropic weakly nonlinear model, James and Dodd (1996) suggested that the poleward propagation originated from the interaction between the mean flow and equatorward-propagating Rossby waves from mid and high latitudes. The eddy momentum flux convergence at the poleward flank of a positive zonal wind anomaly caused the positive zonal wind anomaly to shift poleward. Robinson (2000) further argued that a poleward shift in low-level baroclinicity may in fact play an important role in driving the poleward propagation. The above two mechanisms were confirmed by the observational study of Feldstein

(1998).

The modeling study of Lee et al. (2007) showed that the poleward propagation was initiated in the tropics by the breaking of Rossby waves generated at midlatitudes. This wave breaking deposits negative zonal momentum and decelerates the ambient zonal wind. This results in a poleward shift in the location of the critical latitude and a reduction in the meridional potential vorticity gradient. As a result, subsequent midlatitudinal waves break poleward of the previous critical latitude. In this manner, the zonal-mean flow anomalies propagate poleward. Recently, Yuan et al. (2013) further confirmed with the ERA40 data the modeling result of Lee et al. (2007).

6.3 Spatial and temporal scales of atmospheric teleconnection

Most past studies on low-frequency climate variability are based on monthly or seasonal mean data (Wallace and Gutzler, 1981; Barnston and Livezey, 1987). This time average is sufficient for understanding anomalies that have a timescale longer than two months, such as seasonal, interannual, or longer timescales. However, daily unfiltered data have lately been used to study the intraseasonal variability of the teleconnections (Feldstein, 2002; Athanasiadis et al., 2010; Johnson and Feldstein, 2010). Feldstein (2002) and Johnson and Feldstein (2010) found some teleconnections, such as the PNA and NAO, have timescales of around 10 days, in good agreement with the timescale obtained with other approaches such as numerical modeling (Cash and Lee, 2001; Feldstein, 2002).

As for the spatial shapes of the teleconnections, the results obtained with different approaches (e.g., point correlation and EOF analysis) do not agree, which means that the shapes of the teleconnections may not be unique. Johnson and Feldstein (2010) applied a cluster analysis and a self-organizing map (SOM) analysis to the study of teleconnections based on unfiltered daily data. They found that there existed a number of PNA-like patterns in the sea-level pressure field. These PNA-like SOMs possessed two polarities, each corresponding to one phase of the PNA. The timescale of the PNA-like SOMs was also around 10 days. Using this method, they tried to relate the PNA to convective heating over the tropics. Similarly, Johnson et al. (2008) provided an explanation on the formation of the decadal variation of NAO, and Lee and Feldstein (2013) detected the relationship between ozone depletion, forcing of greenhouse gases, and the southern annular mode.

6.4 Spatial relation of jets and storm tracks

Jets and storm tracks are closely related, and always interact with each other. Jets can trigger synoptic waves through barotropic or baroclinic instability, forming storm tracks, while the synoptic wave can, in turn, supply energy to the planetary wave or the mean flow for their development. Therefore, to study the relation of storm tracks with jets benefits our understanding of teleconnection formation mechanisms.

It has long been known that in both the Northern and Southern Hemispheres, the storm tracks and their associated jets do not coincide completely, with the peak of storm tracks shifted poleward about 10° latitude apart. Yang et al. (2007) studied the baroclinic wave-mean flow interactions in an extended quasi-geostrophic two-layer model in which two kinds of nonlinearity, nonlinear advection of vorticity and vorticity-divergence production, were included. They found that the former kind of nonlinearity led to downstream development of the baroclinic wave packets, while the second kind of nonlinearity was responsible for the asymmetry in the north-south flanks of the upstream part of the wave packets, which caused the poleward shift of the center of the storm track.

6.5 Energetics and variability of storm tracks

Storm tracks vary in position and intensity on timescales ranging from seasonal (Lau, 1988; Nakamura, 1992; Christophy et al., 1997) to interannual (Trenberth and Hurrell, 1994; Straus and Shukla, 1997; Zhang and Held, 1999) and decadal (Ebisuzaki and Chelliash, 1998; Nakamura and Izumi, 1999; Geng and Sugi, 2001; Chang and Fu, 2002). Chang et al. (2002) found that storm tracks were strongly modulated by ENSO on the interannual timescale. During El Niño years, enhanced Hadley cell drove the storm track over North Pacific to shift equatorward and eastward. During La Niña, the reverse was true. However, the ENSO impact on the storm track over North Atlantic was weak. It is well known that AO is the strongest signal in the extratropical regions, so it seems reasonable to assume that AO has a significant influence on the Northern Hemisphere storm tracks. Nie et al. (2008) studied this assumption and found that the North Atlantic storm track shifted northward and eastward during strong positive AO winters, in agreement with McCabe et al. (2001). A similar shift was also detected in the zone of baroclinic energy conversion, caused by planetary wave-baroclinic wave interaction. In contrast, the North Atlantic storm track retreated westward during strong negative AO winters. They also found that the baroclinic waves moved eastward along middle Atlantic, while during strong negative AO winters, the storm track split into northern and southern branches. The impact associated with this bifurcation on weather and climate over Eurasia and China needs further study.

7. Summary and discussion

This paper has reviewed the progress made in recent years in the study of the dynamics of atmospheric teleconnection, particularly the teleconnection patterns that have had a significant impact on China. The key points are as follows.

(1) The impact of solar activity on the East Asian climate was documented. The solar cycle was shown to lead to changes in the spatial structure of the AO/NAO, via its influence on the temperature and winds in the stratosphere, and induce anomalous East Asian climate in winter or spring. Moreover, the 11-yr solar cycle also modulated the relationship between the spring AO/NAO and the following East Asian summer monsoon. The possible mechanism of this modulation was suggested to be solar activity impacting the air-sea interaction in North Atlantic.

(2) The relationship between the stratospheric polar night jet and the circulation over East Asia was established. After 1976, the polar night jet strengthened, suppressing vertical propagation of the signal associated with the anomalous circulation over the Urals toward the stratosphere. More signals propagated in the troposphere, influencing the circulation over East Asia. Associated with this change, the winter cyclone and anticyclone activity over East Asia increased (decreased) after (before) the mid 1970s. Particularly after the mid 1980s, the upper polar-front jet was significantly enhanced and the frequency of cyclone and anticyclone activities also reached a peak.

(3) Three Eurasian teleconnections (EU, SCAND, and EATL/WRUS) were defined, all with a clear seasonal variation. They were weak during summer, but strong during autumn and winter. These three teleconnections were associated with different external forcing sources, and had different impacts on the winter climate in China.

(4) The link between the teleconnections in mid and high latitudes and the tropical atmosphere was discussed. On the seasonal to decadal timescales, the tropical convective heating influenced the NAO through circumglobal wave trains, and the reverse was also true. In addition, anomalous teleconnection patterns in the extratropics, such as the NPO, were shown to influence the occurrence of ENSO in subsequent seasons via the SFM. The spring AO-associated circulation anomalies were supported by the interaction between synoptic-scale eddies and mean flows, and the anomalous westerly wind correspondingly formed in the equatorial western Pacific. These equatorial westerly anomalies excited downwelling equatorial Kelvin waves, leading to SST warming in the central-eastern tropical Pacific in summer. The tropical SST anomalies sustained and developed through the Bjerknes positive feedback mechanism, and thus, an El Niño event occurred in fall and winter. On the other hand, the spring AO-related SST anomalies were shown to either increase or decrease the previous winter NPO-related SST anomalies over the subtropical North Pacific, influencing the westerly wind anomalies over the tropical western Pacific in summer via the air-sea interaction. Hence, the SST anomalies forced by the SFM in the equatorial central-eastern Pacific were affected. The primary influence was on the occurrence of the ENSO.

(5) The relation between the EPW and PNA was discussed, and it was shown that the EPW played an important role in the AL–IL seesaw through its horizontal and vertical propagation and reflection by the polar vortex. Both the NAO and PNA contributed to the AO formation.

(6) The atmospheric teleconnections, such as PNA, WP, and EP, could be resolved into a number of basic structures called SOMs. These SOMs were similar in structure and had two polarities. Each group of the SOMs with the same polarity constituted one phase of the related teleconnection. The SOMs had an intrinsic timescale of approximately 10 days, and the variability on longer timescales such as seasonal, interannual, or decadal, could be obtained by statistical averaging of the SOMs' frequency.

Although progress in the study of the dynamics of the teleconnections in extratropical regions has been achieved, most work has focused on descriptions of the phenomenon. Further studies on mechanisms are still needed, especially those that can address the following questions.

(i) Past studies have focused only on relationships between the ENSO, AO/NAO, and NPO (e.g., Chen et al., 2013b; Chen et al., 2014a). Evidently, further studies on the interaction between different teleconnections in mid and high latitudes, and their integrated impacts on climate, are needed.

(ii) Several studies (e.g., Chen and Li, 2007; Wei et al., 2007; Chen and Wei, 2009) have revealed the link of the AO, EU, and PNA to the stratosphere. The ENSO in particular can influence the stratosphere circulation, which in turn influences the teleconnections in mid-high latitudes in the troposphere. However, the mechanisms are yet to be determined.

(iii) The factors, such as local wave-mean flow interaction, convective heating in the tropics, changes in polar sea ice, strength of the polar vortex, and activity of solar spots, which determine the frequency of the SOMs, are still unclear.

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