A summer teleconnection pattern over the extratropical Northern Hemisphere and associated mechanisms

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Abstract Using monthly data from the European Center for Medium-Range Weather Forecast 40-year reanalysis (ERA-40), we have revealed a teleconnection pattern over the extratropical Northern Hemisphere through the empirical orthogonal function analysis of summer upper-tropospheric eddy temperature. When temperature is higher (lower) over the Eastern Hemisphere (EH), it is lower (higher) over the Western Hemisphere (WH). The teleconnection manifested by this out-of-phase relationship is referred to as the Asian–Pacific oscillation (APO). The values of an index measuring the teleconnection are high before 1976 and low afterwards, showing a downward trend of the stationary wave at a rate of 4% per year during 1958–2001. The index also exhibits apparent interannual variations. When the APO index is high, anomalous upper-tropospheric highs (lows) appear over EH (WH). The formation of APO is likely associated with a zonal vertical circulation in the troposphere. Unforced control runs of both the NCAR Community Atmospheric Model version 3 and the Community Climate System Model version 3 capture

the major characteristics of the teleconnection pattern and its associated vertical structure. The APO variability is closely associated with sea surface temperature (SST) in the Pacific, with a significantly positive correlation between APO and SST in the extratropical North Pacific and a significantly negative correlation in the tropical eastern Pacific. Sensitivity experiments show that the anomalies of SST over these two regions influence the APO intensity, but their effects are opposite to each other. Compared to the observation, the positive and negative anomalous centers of the extratropical tropospheric temperature triggered by the SST anomalies have a smaller spatial scale.

Keywords Summer teleconnection · Extratropical Northern Hemisphere · Observation and simulation · Atmospheric circulation · Mechanism

1 Introduction

Documenting the intrinsic variability within the atmosphere and understanding the associated physical processes have been the major challenges in studying the variability of the climate system. For winter seasons, Wallace and Gutzler (1981) identified the eastern Atlantic, Pacific/North American, western Atlantic, western Pacific, and Eurasian patterns. The North Atlantic oscillation during winter was addressed by Hurrell (1995) and others. Moreover, Yang et al. (2002) particularly targeted the teleconnection pattern linking the upper-tropospheric jet stream and the Asian–Pacific–American climate. These teleconnection patterns have been applied to explain the anomalies of winter climate over the Northern Hemisphere.

The summer teleconnection patterns over the Asian–Pacific–North American sector have also been investigated.

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For example, Kutzbach (1970) showed a zonal teleconnection pattern in July sea-level pressure between Asia and the North Pacific. Barnston and Livezey (1987) discussed a teleconnection pattern involving the Pacific, the Caribbean Sea, and the western Atlantic in the warm season. Nitta (1987) suggested that the meridional Pacific–Japan (PJ) pattern links the summer convective activity from East Asia to North America through sea surface temperature (SST) of the tropical western Pacific. Lau (1992) and Lau and Weng (2002) found that the East Asian–North American teleconnection pattern might be used to explain a close relationship in rainfall variability between East Asia and North America. Wang et al. (2001) found that a teleconnection between the Indian and East Asian summer monsoons may be part of a global-scale wave train linking Asia and North America, explaining a suppressed western North Pacific monsoon and the deficient rainfall over the Great Plains of the United States. According to Rodwell and Hoskins (2001), the climate over central North America might be a hemispheric response to the Asian monsoon heating. Moreover, Lau et al. (2004) considered the North Pacific as a regulator of summertime climates over Eurasia and North America. Zhang et al. (2005) further showed that the variability of the summer upper-tropospheric South Asian high is linked to the changes in the subtropical western Pacific high, the mid-Pacific trough, and the Mexican high, causing variability of surface temperature and precipitation in the Asian–Pacific–American region. More recently, Zhao et al. (2007a) identified a zonal teleconnection of summer tropospheric temperature at the mid-latitudes of the Asian–Pacific sector, namely the Asian–Pacific oscillation (APO), and discussed the associated Asian monsoon climates on the inter-decadal time scale.

Previous studies have also examined summer teleconnection patterns between Asia and Africa. For example, Rodwell and Hoskins (1996) suggested a monsoon-desert mechanism for desertification whereby remote diabatic heating in the Asian monsoon region can induce a Rossby-wave pattern to the west and exert an influence on the summertime climate in the eastern Sahara/Mediterranean and Arabian Sea regions. The South Asian monsoon circulation has a profound effect on the North African monsoon rainfall through the mid-upper tropospheric easterly jet, which extends from the eastern Indian Ocean and South Asia to as far as the west coast of North Africa (Webster and Fasullo 2003).

The larger-scale summer teleconnection patterns over the entire Northern Hemisphere have also been investigated. Zhao and Chen (2001) revealed a lower-stratospheric circumglobal wave train at the mid-high latitudes that exhibits a zonal wavenumber-3 structure and is associated with the summer Tibetan heating. Some studies noted that a disturbance generated over the southern Tibetan Plateau can

reach widely separated points within the jet stream waveguide (Branstator 2002; Zhao et al. 2007b). Using the 56-year National Centers for Environmental Prediction–NCAR reanalysis, Ding and Wang (2005) found a recurrent teleconnection pattern in summertime atmospheric circulation at the mid-latitudes of the Northern Hemisphere. This pattern has a zonal wavenumber-5 structure and is accompanied by the significant anomalies of rainfall and surface temperature in the continental regions of West Europe, European Russia, India, East Asia, and North America.

Although the previous studies have made great progress in understanding the intrinsic links between summer atmospheric circulation patterns over the Northern Hemisphere, some questions still remain unanswered. For example, the empirical orthogonal function (EOF) analysis of summer tropospheric temperature has showed that when positive values of the first EOF mode appear over the Eastern Hemisphere (EH, including the Eurasian and African lands), negative values appear over the Western Hemisphere (WH, including the North Pacific, North America, and the North Atlantic) (Zhao et al. 2007a). Does this feature mean a larger-scale out-of-phase relationship between EH and WH? If yes, how is it associated with the variations of summer atmospheric circulation over the Northern Hemisphere? What factors may affect it? With these questions in mind, we conduct this study to further examine the relationship of tropospheric temperature between EH and WH, as well as the associated atmospheric circulation patterns and physical mechanisms.

The rest of this paper is organized as follows. In Sect. 2, we describe the main features of the datasets, models, and analysis methods applied in the study. In Sect. 3, we examine the extratropical zonal teleconnection pattern over the Northern Hemisphere and its relationship with APO and discuss the atmospheric circulation patterns associated with this teleconnection on the interannual time scale. Results simulated by climate models are discussed in Sect. 3. In Sect. 4, we discuss the factors associated with the teleconnection. Finally, a summary and further discussion are provided in Sect. 5.

2 Data, models, and analysis methods

This study uses the monthly ERA-40 reanalysis with a horizontal resolution of 2.5° in both latitude and longitude (Uppala et al. 2005) and the monthly SST from the Had-ISST dataset with a horizontal resolution of 1° in both latitude and longitude (Rayner et al. 2003) during 1958–2001. An EOF analysis with area-weighting is carried out to detect the teleconnection pattern over the Northern Hemisphere. Correlation and composite analyses are used to examine the relationships between pairs of variables. The

statistical significance of correlation coefficients, composite differences, and non-zero trends is assessed using the Student's t test. The significance discussed in this study is at the 95% confidence level unless otherwise stated.

The NCAR Community Atmospheric Model version 3 (CAM3) with prescribed SST (Collins et al. 2004) and the Community Climate System Model version 3 (CCSM3) are also used. The CCSM3 model includes CAM3 as the atmospheric component, an ocean component (Smith and Gent 2002), a land model (Bonan et al. 2002), and a sea ice model (Bitz et al. 2001).

3 Extratropical teleconnection and associated atmospheric circulation

3.1 Extratropical teleconnection and its association with APO

Following Zhao et al. (2007a), we use eddy air temperature (T') to perform an EOF analysis of the anomaly of summer (June–July–August, JJA) mean upper-tropospheric (300–200-mb) T' over the Northern Hemisphere (0°N – 90°N , 180°E – 180°W) during 1958–2001, in which $T' = T - \bar{T}$. T is the air temperature and $\bar{T}$ is the zonal mean of T . Results show that the first EOF mode (hereafter EOF1) accounts for 27% of the total variance and the second EOF mode (hereafter EOF2) accounts for 11%.

Figure 1a shows EOF1 of JJA upper-tropospheric T' . In the figure, positive values exceeding 0.2 appear mainly over the mid-latitudes of Eurasia, northern Africa, and the tropical Atlantic, with their central values exceeding 0.6. Negative values below -0.2 emerge mainly over the extratropics of the North Pacific, North America, and the North Atlantic, with their central values below -0.5 . Correlation analysis further shows that the East Asian positive center (over 40°N – $45^{\circ}\text{N}/110^{\circ}\text{E}$ – 120°E) has a correlation of -0.78 with the North American negative center (over 37.5°N – $42.5^{\circ}\text{N}/95^{\circ}\text{W}$ – 85°W) and has a correlation of -0.71 with the Atlantic negative center (over 37.5°N – $40^{\circ}\text{N}/50^{\circ}\text{W}$ – 45°W). Thus both the EOF and correlation analyses show an out-of-phase relationship over the extratropics between EH and WH during summer.

The standardized time series of EOF1 shows a positive phase before 1976 and a negative one afterwards (Fig. 1b), which indicates a decrease at a rate of 4% per year during 1958–2001 (significant at the 99% confidence level). This feature suggests a negative trend of the stationary wave over the extratropical Northern Hemisphere. To test whether the structure of EOF1 (shown in Fig. 1a) is due to the linear trend, we perform a similar analysis of the anomaly of detrended upper-tropospheric T' . It is found that the EOF1 and EOF2 without the linear trend account for 24 and 11% of the total variance, respectively. Consistent patterns are observed between the original (Fig. 1a) and detrended (Fig. 2) results for EOF1. Moreover, we perform an EOF

Fig. 1 **a** EOF1 ($\times 0.1$) of the anomaly of JJA upper-tropospheric (300–200-mb) mean T' over the Northern Hemisphere (0°N – 90°N , 180°E – 180°W) (the shaded areas are greater than zero). **b** Standardized time series (bars) of EOF1 during 1958–2001 and its linear trend (solid line)

Fig. 2 Same as in Fig. 1a but for JJA EOF1 after removing the linear trend during 1958–2001

Fig. 3 **a** Climatology of JJA 300–200-mb T' ($^{\circ}\text{C}$) during 1958–2001. **b** Same as in **a** but for longitude-height cross section of JJA T' ($^{\circ}\text{C}$) along 35°N (the black shaded areas denote mountains)

analysis of the anomaly of JJA upper-tropospheric T' during two periods of 1958–1975 and 1976–2001 (figures not shown), respectively, and obtain consistent results with Fig. 1a. Clearly, this tropospheric teleconnection occurs on both the interannual and inter-decadal time scales.

To understand the physical implications of this summer teleconnection, we analyze the climatology of upper-tropospheric T' . The upper-tropospheric mean temperature is warmer over EH than over WH, with positive T' values over most land areas of EH and negative T' values over most oceanic areas from the central Pacific to the eastern Atlantic (Fig. 3a). As seen from a vertical cross section (Fig. 3b), except for the small-scale positive T' values in the lower troposphere over the North American mountains, large-scale positive and negative T' values generally appear

in the entire troposphere over EH and WH, respectively, with their centers in the upper troposphere. The EH positive (WH negative) anomalies of EOF1 in Fig. 1a generally correspond to the positive (negative) values in Fig. 3a. Thus, this extratropical teleconnection reflects an out-of-phase variation between the upper-tropospheric warm center over the EH land and the upper-tropospheric cold centers over the WH oceans. When the time series of EOF1 is higher (lower), the tropospheric temperature is higher (lower) over land and lower (higher) over oceans, indicating a stronger (weaker) zonal thermal contrast between the EH land and the WH oceans.

Comparing the pattern in Fig. 1a with APO (Zhao et al. 2007a), it is found that the structure of this Northern Hemispheric extratropical teleconnection over the Asian–

Fig. 4 Detrended standardized time series of EOF1 shown in Fig. 1a

Pacific region is similar to that of APO. Based on the definition of APO (Zhao et al. 2007a), we calculate the summer APO index using the ERA-40 reanalysis. The time series of EOF1 in Fig. 1a is highly correlated with the APO index in summer, with a correlation coefficient of 0.91 for 1958–2001, which further verifies the consistency between APO and the Northern Hemispheric extratropical teleconnection pattern. For convenience, this teleconnection pattern is still referred to as APO and the time series shown in Fig. 1b is used as an APO index in the following analyses.

3.2 Interannual variability of atmospheric circulation associated with APO

To detect the variations of the extratropical atmospheric circulation over the Northern Hemisphere associated with the interannual variability of the summer APO, we select several highest and lowest APO-index years based on the detrended JJA APO index (displayed in Fig. 4). The high-index (HI) years (i.e. strong APO years) are 1961, 1962, 1967, 1984, 1994, 1996, and 2000 and the low-index (LI) years (weak APO years) are 1965, 1969, 1980, 1981, 1983, 1987, and 1992. The APO-index anomalies in both HI and LI cases are beyond one standard deviation (0.85).

Figure 5 shows the composite detrended JJA tropospheric T' in the HI and LI cases. In the HI case (Fig. 5a), positive anomalies of T' appear mainly over EH and negative anomalies occur largely over WH. In the LI case (Fig. 5b), opposite features are observed. This result shows that a high (low) APO index, or a strong (weak) APO, reflects a strong (weak) summer thermal contrast between the EH continents and the WH oceans.

Figure 6 shows a longitude-height cross section of the composite difference in detrended JJA T' between the HI and LI cases along 35°N. Significantly positive and negative anomalies are mainly seen in the middle-upper troposphere between 0°E and 140°E and between 140°E and 20°W, respectively, with central values of 2 and -1°C . According to the equation of static equilibrium, an increase (a decrease) of temperature in an air column is associated

with an increase (a decrease) of geopotential height at the top of the column and a decrease (an increase) of geopotential height at the bottom due to expansion (contraction) of the air column. Thus, over EH, corresponding to the positive anomalies of T' , there are positive anomalies of eddy geopotential height (H') in the upper troposphere and negative anomalies in the lower troposphere (Fig. 7a), where $H' = H - \bar{H}$; H is geopotential height and $\bar{H}$ is the zonal mean of H . Over WH, corresponding to the cooling troposphere, there are negative anomalies of H' in the upper troposphere and positive anomalies in the lower troposphere.

At 100 mb (Fig. 7b), positive anomalies of H' cover most of the EH land areas, with central values exceeding 30 m, while negative anomalies appear over 15°N–60°N of WH, with central values below -30 m. Compared to the climatology of summer 100-mb geopotential height (figure not shown), the positive anomalies of H' at the mid–low latitudes of Eurasia reflect a stronger and more northward South Asian high under the high APO-index conditions and the negative anomalies at these latitudes of WH reflect a stronger trough over the North Pacific and Atlantic Oceans and a weaker ridge over North America.

3.3 Results in CCSM3 and CAM3

In order to examine whether APO arises from the natural variability of the atmosphere and can be captured in climate models, we carry out several experiments using the NCAR models: CCSM3 and CAM3. We conduct an unforced run for 100 years using the CCSM3 coupled model and an unforced run for 50 years using the CAM3 model with the climatological SST from the original model (Collins et al. 2004). These experiments are referred to as CCSM3_Control and CAM3_Control, respectively. Our analyses show that CAM3 rapidly reaches quasi-equilibrium and CCSM3 reaches quasi-equilibrium over the last 50 years. The outputs over the last 30 years in CAM3 and over the last 50 years in CCSM3 are analyzed. Similar to Sect. 3.1, we perform an EOF analysis of the anomaly of

Fig. 5 Composite JJA 300–200-mb mean T' ($\times 0.1^\circ\text{C}$) in the HI (a) and LI (b) cases

Fig. 6 Longitude-height cross section of the composite difference in JJA T' ($^\circ\text{C}$) between the HI and LI cases along 35°N . The light shaded areas are at the 95% confidence level and the black shaded areas denote mountains

JJA upper-tropospheric (300–200-mb) mean T' from the model output.

In experiment CCSM3_Control, EOF1 and EOF2 account for 27 and 13% of the total variance, respectively. Figure 8a shows the simulated JJA EOF1. In the figure, an out-of-phase pattern between EH and WH is seen, with positive and negative values mainly over EH and WH, respectively. Compared to the teleconnection pattern in the ERA-40 reanalysis (Fig. 1a), the strongest positive and negative centers in CCSM3 are moved slightly westward, appearing over Europe and the central North Pacific respectively. Based on the standardized time series of the simulated JJA EOF1 (figure not shown), we select seven

years of the high and low model APO indices, respectively, to perform a composite analysis. Figure 8b shows a longitude-height cross section of the composite difference of T' along 35°N in CCSM3. Positive and negative anomalies appear in the middle-upper troposphere over EH and WH, respectively, with central values of 3 and -1.5°C . Thus, the extratropical teleconnection pattern also appears in the natural variability of a coupled climate system.

The simulation from CAM3 further suggests that the teleconnection pattern is part of the atmospheric intrinsic variability. In CAM3, EOF1 and EOF2 of JJA upper-tropospheric mean T' account for 25 and 14% of the total variance, respectively. In Fig. 9a, which shows the

Fig. 7 **a** Longitude-height cross section of the composite difference in JJA H' ($\times 10$ m) between the HI and LI cases along 35°N . **b** Composite difference in JJA 100-mb H' ($\times 10$ m) between the HI and LI cases. The *light shaded* areas are at the 95% confidence level and the *black shaded* areas denote mountains in **a**

Fig. 8 **a** Same as in Fig. 1a but for JJA EOF1 in CCSM3. **b** Same as in Fig. 6 but for JJA T' along 35°N in CCSM3

Fig. 9 **a** Same as in Fig. 1a but for JJA EOF1 in CAM3. **b** Same as in Fig. 6 but for JJA T' along 35°N in CAM3

simulated JJA EOF1 in CAM3, positive values emerge over Eurasia and northern Africa and negative values appear at the mid-latitudes of the central-eastern North Pacific and over North America and the North Atlantic Ocean. Figure 9b shows a longitude-height cross section of the composite difference of JJA T' along 35°N in CAM3. Positive and negative anomalies of T' occur in the middle–upper troposphere over EH and WH, respectively, with central values of 2 and -1°C . These features are also similar to those from the ERA-40 reanalysis and the CCSM3 output.

It is evident that CCSM3 and CAM3 can capture the summer upper-tropospheric teleconnection pattern between EH and WH. This consistency between the observation and the CAM3 simulation also implies that APO is part of the intrinsic variability of the troposphere.

4 Factors associated with the summer extratropical teleconnection

4.1 Mean zonal vertical circulation

In this section, we try to provide a possible explanation for the formation of the summer APO. Figure 10 shows a

longitude-height cross section of the climatology of JJA zonal vertical circulation along 15°N–50°N from the ERA-40 reanalysis. There exists an anticlockwise vertical cell between 100 and 200 mb over Eurasia and its center appears over the Tibetan Plateau. Meanwhile, there is also a large-scale clockwise zonal vertical cell in the troposphere, with its two centers in the lower troposphere over the North Pacific and the North America–Atlantic region, respectively, in which the Asian–Pacific part of this tropospheric cell has been noted by Ye (1981). The upward motion of the tropospheric zonal cell appears over the Asian-central Pacific sector, with the strongest upward motion over 70°E–110°E where the Tibetan Plateau is located. The downward motion dominates the eastern Pacific–European sector. Diagnosing the CCSM3 and CAM3 outputs, we observe similar zonal vertical cells. Compared with Fig. 3b, the upward motion corresponds to the warm troposphere over the EH land and the downward motion corresponds to the cold troposphere over the WH oceans. Thus the tropospheric zonal vertical cell is likely associated with the thermal contrast between the EH land and WH oceans.

Figure 10 also shows the composite difference in JJA vertical velocity between the HI and LI cases. Under higher APO-index conditions, negative anomalies of vertical

Fig. 10 Longitude-height cross section of the climatology of vertical circulation (*streamlines*) of the ERA-40 reanalysis along 15°N–50°N and the composite difference in JJA vertical p -velocity (*color*; $\times 0.001 \text{ Pa s}^{-1}$) between the HI and LI cases

velocity cover most of 70°E–180°E where the upward motion of the mean zonal circulation appears, indicating a strong upward motion of the tropospheric zonal cell over this region. Meanwhile, positive anomalies appear mainly over the eastern Pacific, North America, the Atlantic, and Europe, indicating a strong downward motion. This result shows that a strengthened upward motion of the tropospheric vertical cell is linked to the strengthening of the associated downward motion; and vice versa.

From the foregoing analysis, when the tropospheric temperature increases over EH, both the upper-tropospheric high and the low-level trough strengthen over EH and the local upward motion strengthens, leading to the strong downward motion over WH. The strengthened downward motion over WH corresponds to a decrease of the atmospheric heating over WH (Zhao et al. 2007a), accompanying the cooling of the local troposphere. Accordingly, the lower-tropospheric subtropical high and the upper-tropospheric trough over WH strengthen. These variations in atmospheric circulation have been shown in Figs. 6, 7, and 10. Therefore, the extratropical zonal vertical circulation over the Northern Hemisphere may be used to explain the formation of the summer APO pattern.

In the following section, we describe further numerical simulations performed to examine an impact of ocean surface heating anomalies over the Pacific on APO.

4.2 Extratropical North Pacific and Tropical Pacific SSTs

Because the temperature over the WH troposphere may be affected by ocean surface heating, it is useful to examine a link between APO and oceans. Figure 11 shows the composite difference in detrended summer HadISST SST between the HI and LI cases. It is seen that large-scale significant positive anomalies of SST occur in the mid-latitudes of the North Pacific, with a central value exceeding 1°C near 40°N of the central North Pacific. Meanwhile, large-scale significant negative anomalies appear over the tropical eastern Pacific, with a central value below -1°C , which suggests a relationship between APO and the El Niño–Southern oscillation (ENSO). The correlation between the APO index and the summer regional mean SST over 35°N–50°N/140°E–160°W is 0.68 for 1958–2001. Moreover, the APO index has also a significant negative correlation of 0.52 with the synchronous Niño 1 + 2 index.

Fig. 11 Composite difference in detrended JJA SST ($^\circ\text{C}$) from the HadISST dataset between the HI and LI cases. The two boxes are used to represent the regions for the extratropical North Pacific and the tropical eastern Pacific, respectively. The shaded areas are at the 95% confidence level

In order to further discuss potential impacts of the Pacific SST anomalies on APO during summer, we focus on the positive and negative anomalous centers of the Pacific SST (shown in Fig. 11) and design three experiments. Experiment Extratrop_SST is for testing an impact of the extratropical North Pacific SST; experiment Trop_SST is for testing an impact of the tropical eastern Pacific SST; and experiment Pacific_SST is a combination of experiments Extratrop_SST and Trop_SST. Experiment Extratrop_SST is the same as experiment CAM3_Control but with an increase of 1°C over the region 35°N – 50°N / 140°E – 160°W , and experiment Trop_SST is the same as CAM3_Control except a decrease of 1°C over 10°S – 10°N / 130°W – 80°W (shown in Fig. 11). The extratropical and tropical Pacific SST in experiment Pacific_SST includes

the changes made in both experiments Extratrop_SST and Trop_SST. In these experiments, CAM3 is run for 50 years and the mean values over the last 30 years are used. The 30-year mean summer values are equivalent to those from an ensemble of 30 sensitivity experiments changing the extratropical or/tropical eastern Pacific SST using different initial atmospheric and land surface conditions.

Corresponding to the increase in JJA SST over the extratropical North Pacific between experiments Extratrop_SST and CAM3_Control, positive anomalies of T' appear in the entire troposphere over the extratropical North Pacific and Atlantic Oceans, with a central value of 0.8°C near 180° , and negative anomalies appear in the troposphere over Eurasia, and North America, with a negative center of -0.6°C over Europe (Fig. 12a). This

Fig. 12 Longitude-height cross section of composite differences in JJA T' ($^{\circ}\text{C}$) along 35°N between experiments Extratrop_SST and CAM3_Control (a), between experiments Trop_SST and CAM3_Control (b), and between Pacific_SST and CAM3_Control (c). The shaded areas are at the 95% confidence level

feature indicates an increase in tropospheric temperature over the North Pacific and a decrease over Eurasia, as well as a weak thermal contrast between them.

When SST over the tropical eastern Pacific decreases in experiment Trop_SST relative to experiment CAM3_Control, negative anomalies of JJA T' occur over the tropical central-eastern Pacific and stretch northward to the extratropical central-western North Pacific, indicating a decrease in tropospheric temperature over these regions (figure not shown). Figure 12b exhibits a longitude-height cross section of the composite difference in JJA T' between experiments Trop_SST and CAM3_Control along 35°N. In the figure, a pronounced wave train is seen in the troposphere, with negative anomalies over the central-western Pacific and the Atlantic Ocean and positive anomalies over North America and Eurasia, indicating a strong thermal contrast between Eurasia and the North Pacific.

Clearly, the anomalous tropospheric temperature pattern triggered by SST anomalies over the extratropical North Pacific or the tropical eastern Pacific shows a zonal wavenumber-2 feature over the extratropical Northern Hemisphere, with an opposite variation between Eurasia and the North Pacific. Compared with the APO pattern in the ERA-40 reanalysis (in Fig. 6), however, the positive and negative anomalous centers of T' in Fig. 12a, b have a smaller spatial scale.

Because the anomalies of T' over the Asian–Pacific region forced by tropical and extratropical SSTs (in Fig. 12a, b) are generally opposite in phase, one may wonder how the combination of the SST anomalies over these two regions exerts influences on the summer APO. Figure 12c shows the composite difference between experiments Pacific_SST and Control_SST. The figure shows that the large-scale significant anomalies of T' between Asia and the North Pacific shown in Fig. 12a, b weaken pronouncedly, implying a weaker impact of the anomalous Pacific SST distribution shown in Fig. 11 on the summer APO.

5 Summary and discussion

Using the monthly ERA-40 reanalysis and the empirical orthogonal function analysis, we identified an extratropical teleconnection over the Northern Hemisphere during the summer. This teleconnection pattern shows an out-of-phase relationship in tropospheric temperature between EH and WH, reflecting a link between the upper-tropospheric warm center over the Eurasian and African continents and the upper-tropospheric cold centers over the North Pacific and Atlantic Oceans. When temperature is higher (lower) over EH, it is lower (higher) over WH. Because this

teleconnection pattern is highly consistent with the summer APO shown by Zhao et al. (2007a), it is still referred to as APO. Corresponding to a high (low) APO index, the thermal contrast between EH and WH is stronger (weaker).

Over the extratropical Northern Hemisphere, the summer APO index exhibits a pronounced decadal variation, tending to be positive before 1976 and negative afterwards, showing a decrease at the rate of 4% per year in 44 years and a weakening of the stationary wave. This feature also suggests a reducing trend in the summer tropospheric thermal contrast between EH and WH during 1958–2001. Moreover, the APO index also shows substantial interannual variability. In a higher APO-index year, anomalous highs appear in the EH upper troposphere, indicating a stronger South Asian high, while anomalous lows appear in the WH upper troposphere, indicating a stronger trough over the North Pacific and Atlantic Oceans and a weaker ridge over North America.

The summer extratropical teleconnection pattern and its vertical structure can be captured by both NCAR CCSM3 and CAM3. This consistency between the observation and the CAM3 simulation shows that APO is part of the intrinsic variability of the troposphere. The formation of the summer APO pattern is likely associated with an extratropical zonal vertical circulation. This zonal circulation consists of the upward motion from Tibet to the western Pacific and downward motion over the eastern Pacific and the Atlantic. When tropospheric temperature increases over EH and decreases over WH, the upward motion strengthens over the Asian-central Pacific sector and the downward motion intensifies over the eastern Pacific-European sector; and vice versa.

The APO variability is related to summer SST over the extratropical North Pacific and tropical eastern Pacific. When the APO index is higher, significantly positive and negative SST anomalies appear at the mid-latitudes of the North Pacific and in the tropical eastern Pacific, respectively. The sensitivity experiments with the CAM3 model further show that the SST anomalies over the extratropical and tropical Pacific regions affect the teleconnection pattern between Eurasia and the extratropical North Pacific. However, the impacts of SST over these two regions are opposite to each other. Compared to the observation, the extratropical tropospheric temperature anomalies triggered by the SST anomalies have a smaller spatial scale, showing a zonal wavenumber-2 feature.

The APO pattern establishes a link in atmospheric circulation between EH and WH, which also provides a way to explore interactions between the Asian and Pacific atmospheric circulations. Through this extratropical teleconnection, ocean–atmosphere interactions over the Pacific possibly affect the Asian atmospheric circulation and climate. On the other hand, it is noted from Figs. 6 and 10

that the positive anomalies of the tropospheric temperature and the strongest upward motion appear over the Tibetan Plateau and its adjacent regions. The previous modeling studies have shown that the temperature and flow anomalies near the Tibetan plateau (shown in Figs. 6 and 10) may result from its thermal and dynamical functions (e.g., Ye 1981; Zhao et al. 2009; Nan et al. 2009). Thus we want to know impacts of the Tibetan elevated heating anomaly on the APO intensity and then climates over the extratropical Northern Hemisphere. More research may help understand the relationships of Tibetan heating with the extratropical atmospheric circulation and climate over the entire Northern Hemisphere. This should be addressed in the future work.

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References

- Barnston AG, Livezey RE (1987) Classification, seasonality and persistence of low-frequency atmospheric circulation patterns. *Mon Weather Rev* 115:1083–1126
- Bitz CM, Holland MM, Eby M, Weaver AJ (2001) Simulating the ice-thickness distribution in a coupled climate model. *J Geophys Res* 106:2441–2463
- Bonan GB, Oleson KW, Vertenstein M, Levis S, Zeng X, Dai Y, Dickson RE, Yang ZL (2002) The land surface climatology of the Community Land Model coupled to the NCAR Community Climate Model. *J Climate* 15:3123–3149
- Branstator G (2002) Circumglobal teleconnections, the jet stream waveguide, and the North Atlantic oscillation. *J Climate* 15:1893–1910
- Collins WD et al (2004) Description of the NCAR community atmosphere model (CAM3). Tech. Rep. NCAR/TN-464_STR, National Center for Atmospheric Research, Boulder, p 226
- Ding QH, Wang B (2005) Circumglobal teleconnection in the Northern Hemisphere summer. *J Climate* 18:3484–3505
- Hurrell JW (1995) Decadal trends in the North Atlantic oscillation: regional temperature and precipitation. *Science* 269:676–679
- Kutzbach JE (1970) Large-scale features of monthly mean Northern Hemisphere anomaly maps of sea-level pressure. *Mon Weather Rev* 98:708–716
- Lau KM (1992) East Asian summer monsoon rainfall variability and climate teleconnection. *J Meteorol Soc Jpn* 70:211–241
- Lau KM, Weng HY (2002) Recurrent teleconnection patterns linking summertime precipitation variability over East Asia and North America. *J Meteorol Soc Jpn* 80:1309–1324
- Lau KM, Lee JY, Kim KM, Kang IS (2004) The North Pacific as a regulator of summertime climate over Eurasia and North America. *J Climate* 17:819–833
- Nan SL, Zhao P, Yang S (2009) Springtime tropospheric temperature over the Tibetan Plateau and tropical Pacific sea surface temperature. *J Geophys Res* 114:D10104. doi:10.1029/2008JD011559
- Nitta T (1987) Convective activities in the tropical western Pacific and their impacts on the Northern Hemisphere summer circulation. *J Meteorol Soc Jpn* 65:373–390
- Rayner NA, Parker DE, Horton EB, Folland CK, Alexander LV, Rowell DP, Kent EC, Kaplan A (2003) A Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J Geophys Res* 108(D14):4407. doi:10.1029/2002JD002670
- Rodwell MJ, Hoskins BJ (1996) Monsoons and the dynamics of deserts. *Q J R Meteorol Soc* 122:1385–1404
- Rodwell MJ, Hoskins BJ (2001) Subtropical anticyclones and summer monsoons. *J Climate* 14:3192–3211
- Smith R, Gent P (2002) Reference manual for the parallel Ocean Program ocean component of the Community Climate System Model (CCSM2.0). <http://www.cesm.ucar.edu/models>
- Uppala SM, Kållberg PW, Simmons AJ et al (2005) The ERA-40 reanalysis. *Q J R Meteorol Soc* 131:2961–3012. doi:10.1256/qj.04.176
- Wallace JM, Gutzler DS (1981) Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon Weather Rev* 109:784–812
- Wang B, Wu RG, Lau KM (2001) Interannual variability of the Asian summer monsoon: contrasts between the Indian and the western North Pacific–east Asian monsoons. *J Climate* 14:4073–4090
- Webster PJ, Fasullo J (2003) Monsoon: dynamical theory. In: Holton J, Curry JA (eds) *Encyclopedia of atmospheric sciences*. Academic Press, London, pp 1370–1386
- Yang S, Lau KM, Kim KM (2002) Variations of the east Asian jet stream and Asian–Pacific–American winter climate. *J Climate* 15:306–325
- Ye DZ (1981) Some characteristics of the summer circulation over the Qinghai–Xizang (Tibet) Plateau and its neighborhood. *Bull Am Meteorol Soc* 62(1):14–19
- Zhang PQ, Yang S, Kousky VE (2005) South Asian high and Asian–Pacific–American climate teleconnection. *Adv Atmos Sci* 22:915–923
- Zhao P, Chen LX (2001) Interannual variability of atmospheric heat source/sink over the Qinghai–Xizang (Tibetan) Plateau and its relation to circulation. *Adv Atmos Sci* 18:106–116
- Zhao P, Zhu YN, Zhang RH (2007a) An Asian–Pacific teleconnection in summer tropospheric temperature and associated Asian climate variability. *Clim Dyn* 29:293–303
- Zhao P, Zhou ZJ, Liu JP (2007b) Variability of Tibetan spring snow and its associations with the hemispheric extratropical circulation and East Asian summer monsoon rainfall: an observational investigation. *J Climate* 20:3942–3955
- Zhao P, Zhang XD, Li YF, Chen JM (2009) Remotely modulated tropical–North Pacific ocean–atmosphere interactions by the South Asian high. *Atmos Res* 94:45–60