Possible Effect of the Thermal Condition of the Tibetan Plateau on the Interannual Variability of the Summer Asian–Pacific Oscillation

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ABSTRACT

The summer (June–August) Asian–Pacific Oscillation (APO), a large-scale atmospheric teleconnection pattern, is closely associated with climate anomalies over the Northern Hemisphere. Using the NOAA/CIRES twentieth-century reanalysis, the ECMWF twentieth-century atmospheric reanalysis, and the NCEP reanalysis, this study investigates the variability of the summer APO on the interannual time scale and its relationship with the thermal condition over the Tibetan Plateau (TP). The results show that the interannual variability of the APO is steadily related to the summer TP surface air temperature during the last 100 years.

Observation and simulation further show that a positive heating anomaly over the TP can increase the upper-tropospheric temperature and upward motion over Asia. This anomalous upward flow moves northward in the upper troposphere, and then turns and moves eastward, before finally descending over the mid- to high latitudes of the central-eastern North Pacific, concurrently accompanied by anomalous upward motion over the lower latitudes of the central-eastern North Pacific. The anomalous downward and upward motions over the central-eastern North Pacific reduce the in situ mid- and upper-tropospheric temperature, mainly through modulating condensation latent heat from precipitation and/or dry adiabatic heat, which ultimately leads to the interannual variability of the summer APO. In this process, the zonal vertical circulation over the extratropical Asian–North Pacific sector plays an important bridging role.

1. Introduction

The intrinsic relationships between the atmospheric circulation systems of Asia and the North Pacific have been extensively studied by many researchers (e.g., Kutzbach 1970; Kidson 1975; Wallace and Gutzler 1981; Barnston and Livezey 1987; Nitta 1987). Consequently, several atmospheric teleconnections, such as the East Asia–North America pattern (Lau 1992; Lau and Weng 2002), the western North Pacific–North America teleconnection (Wang et al. 2001), the circumglobal teleconnection (Ding and Wang 2005), the Asian–Pacific–American climate teleconnection (Zhang et al. 2005), and the upper-level Asia–North America teleconnection (Zhu and Li 2016), have been found to link the summer climate anomalies over Asia, the Pacific, and North America. These teleconnection patterns have deepened our understanding of the intrinsic relationships among climate anomalies over different regions of the Northern Hemisphere (NH). Among them, there are two dominant tropical–extratropical teleconnections (i.e., the circumglobal teleconnection and the western Pacific–North America teleconnection) that modulate the summer climate variability over the extratropical NH; furthermore, they even have a global impact by connecting anomalous convective heating over the Asian monsoon area with the circulation of the NH and related climate anomalies (Ding and Wang 2005; Ding et al. 2011; Lee et al. 2011; Lee and Wang 2014; Lee and Ha 2015). Moreover, these two teleconnections can be reproduced well by several coupled climate models (Lee et al. 2011; Lee and Wang 2014), providing significant signal sources for climate predictability over the NH.

In addition to the above two dominant teleconnections, there is another extratropical large-scale teleconnection pattern that links atmospheric circulation anomalies on the hemispheric scale (Zhou and
and measures the variability of summer monsoon rainfall over South Asia, East Asia, and extratropical North America (Zhao et al. 2012b), and this is the Asian–Pacific Oscillation (APO). The APO and associated climate anomalies have been successfully captured by many models (Zhao et al. 2012a; Chen et al. 2013; Y. Huang et al. 2013, 2014; Zhang et al. 2016; Zhou 2016; Wu and Chen 2016; Zhou et al. 2017; Zhou and Xu 2017a,b). Evidently, the APO is an important factor affecting the monthly-scale predictability of East Asian precipitation (Chen et al. 2016).

Several studies have shown that complicated interactions among the atmosphere, ocean, and land modify the behaviors of the Asian and American monsoons (Mechl 1994; Douville and Royer 1996; Yang and Lau 1998; Chen and Yoon 2000; Douville 2003; Misra 2008). During boreal summer, the elevated heating of the Tibetan Plateau (TP), one of the most important heat sources in the NH, plays a prominent role in atmosphere–ocean–land interactions. For example, the heating anomalies of the TP affect atmospheric circulation and precipitation over Asia (e.g., Zhao and Chen 2001a,b; Duan and Wu 2005; Wang et al. 2008; Liu et al. 2012; Wu and Liu 2016) and even the larger-scale atmospheric circulation and climate over the NH (e.g., Ose 1996; Zhao et al. 2007; Nan et al. 2009; Zhou et al. 2009).

Some studies have shown that the interannual variability of the summer APO can be modulated by the TP’s heating anomalies (Nan et al. 2009; Zhou et al. 2009; Liu et al. 2015). However, these studies have not clarified in detail the physical processes involved in the effect of TP heating on the APO. Moreover, the interannual relationships between monsoon climates and ocean/land forcing are not stable (Bamzai and Shukla 1999; Kumar et al. 1999; Wang 2002; Wu et al. 2014). For example, the relationship between ENSO and the Asian summer monsoons is clearly unstable on the decadal time scale (Kumar et al. 1999; Wang 2002). Decadal change has also been found in the interannual relationship between the summer air temperature anomaly over northeast China, the spring snow-cover anomaly in situ, and the spring North Atlantic sea surface temperature (SST) anomaly (Wu et al. 2014). Also, an inverse correlation between ENSO and the Indian summer monsoon has broken down during recent decades (Wang et al. 2015). Therefore, it is necessary to investigate the interannual variability of the summer APO and associated mechanisms using long-term data.

The present study investigates the interannual relationship between the summer APO and the thermodynamic condition of the TP, along with the associated physical mechanisms, by using long-term data analyses and climate simulations. The datasets and methods are described in section 2, and then the interannual variability of the summer APO and its relationship with the thermal condition of the TP is examined in section 3. A physical explanation for this relationship is given in section 4, followed by a summary and discussion in section 5.

2. Data, methods, and models

This study utilizes the monthly National Oceanic and Atmospheric Administration (NOAA)/Cooperative Institute for Research in Environmental Sciences (CIRES) Twentieth Century Reanalysis V2 product (NOAA-20C; Compo et al. 2011) during 1871–2012. Several studies have demonstrated that this reanalysis dataset can be successfully applied to studying atmospheric circulation and climate over Asia and the Pacific (e.g., Cao et al. 2012; Chowdary et al. 2012; Krishnamurthy and Krishnamurthy 2014), as well as the long-term variability of the APO (Zhao et al. 2012b). Additionally, we use the European Centre for Medium-Range Weather Forecasts (ECMWF) twentieth-century atmospheric reanalysis (ERA-20C) during 1900–2010 (Poli et al. 2013), the National Centers for Environmental Prediction (NCEP) reanalysis during 1948–2015 (Kalnay et al. 1996), and phase 5 of the Coupled Model Intercomparison Project (CMIP5) output from the NCAR Community Climate System Model, version 4 (CCSM4), to assess the robustness of the results from NOAA-20C. The output of this model used in this study is from its historical simulation, based on the observational anthropogenic and natural forcing from 1870 to 2006 (see https://esgf-node.llnl.gov/search/cmip5/ for further details). Because our focus here is on the interannual time scale, we extract the interannual components from the original variables via 10-yr high-pass Lanczos filtering (Duchon 1979). The empirical orthogonal function (EOF) (Kundu and Allen 1976), correlation, composite, and regression analysis methods are also employed and, unless otherwise stated, the statistical significance of correlation coefficients is assessed using the Student’s t test.

The NCAR Community Climate System Model, version 3 (CCSM3), which includes an atmospheric component, an oceanic component (Smith and Gent 2002), a land model (Bonan et al. 2002), and a sea ice model (Bitz et al. 2001), is employed to investigate the impact of the TP elevated heating on the APO during summer. A previous study showed that CCSM3 can successfully simulate the impacts of Asian land heating on the Asian–Pacific climate (Zhao et al. 2011). In this study, we perform a “free run” experiment using the
original CCSM3, referred to as CCSM3-Control, and a sensitivity experiment to examine the possible impact of TP elevated heating on the APO during summer (referred to as CCSM3-TP). The configuration in CCSM3-TP is the same as that in CCSM3-Control, except with a 50% reduction in the topographic height over the TP region above 3000 m. For each experiment, the CCSM3 model is integrated for 100 years and the outputs for the last 50 years are analyzed. Furthermore, to isolate the individual effect of the TP thermal forcing, we use the NCAR Community Atmosphere Model, version 3 (CAM3), with prescribed monthly SST (Collins et al. 2004), to perform two numerical experiments with altered surface vegetation (CAM3-Tree and CAM3-Bare, respectively). Following Zhao et al. (2012a, 2016), in CAM3-Tree, the surface vegetation type in each grid with a topographic height above 1500 m over the TP region is set to “broadleaf evergreen tropical tree,” which has a low surface albedo. In CAM3-Bare, the surface vegetation type over the same region is set to “bare soil,’’ which has a higher surface albedo than the configuration in CAM3-Tree. It should be noted that the feedback of ocean–atmosphere interactions is excluded because of prescribed monthly SST in both experiments. The aim in conducting these two experiments is that the difference between their results will reveal the individual effect of the TP thermal forcing. In both experiments, CAM3 is integrated for 20 years, and the values from the last 15 years are used for analysis.

3. Interannual variability of the summer APO and its relationship with the thermal condition of the TP

Following Zhao et al. (2012b), we perform an EOF analysis of the interannual component of the summer (June–August) upper-tropospheric (300–200 hPa) eddy temperature ($T'$) anomaly over the entire NH using NOAA-20C for the period 1871–2012, in which $T'$ is defined as the deviation of temperature from the global zonal mean. The leading EOF mode (EOF1) accounts for 17% of the total variance for the interannual component of the reanalysis dataset, significant at the 99% confidence level through 1000 Monte Carlo simulations (Besag and Clifford 1989). Meanwhile, EOF1 is found to be significantly distinct from the other modes, based on the criterion of North et al. (1982). EOF1 displays an extratropical out-of-phase variation between Eurasia and the Pacific–Atlantic region, indicating a clear APO pattern (Zhao et al. 2012b) on the interannual time scale during the last 100 years. For convenience, the regional mean values of the $T'$ interannual components over Eurasia and the western North Pacific (30°–55°N, 30°–135°E) and the central-eastern North Pacific (30°–55°N, 165°E–120°W) (indicated by two red boxes in Fig. 1a) are referred to as the Eurasian and Pacific $T'$ indices, respectively.
respectively. The APO index is calculated simply through the arithmetic difference between the Eurasian and Pacific $T^0$ indices. This APO index shows a consistent variation with the principal component of EOF1 (shown in Fig. 1b), with a correlation coefficient of 0.95, significant at the 99.9% confidence level. Thus, the difference between the Eurasian and Pacific $T^0$ indices can represent the interannual variability of the summer APO well, and is defined as the APO index.

Figure 2a shows the distribution of correlation coefficients between the summer APO index and simultaneous surface air temperature during 1871–2012, in which the correlation is based on NOAA-20C reanalysis data. (b),(c) As in (a), but for ERA-20C during 1900–2010 and NCEP during 1948–2015, respectively. Yellow and orange (light and dark blue) shading denotes positive (negative) correlations significant at the 95% and 99% confidence level, respectively. The black rectangle in (a) is the TP region. Red dashed contours indicate the area over 1500 m.

Fig. 2. (a) Distribution of correlation coefficients between the summer APO index and simultaneous surface air temperature during 1871–2012, in which the correlation is based on NOAA-20C reanalysis data. (b),(c) As in (a), but for ERA-20C during 1900–2010 and NCEP during 1948–2015, respectively. Yellow and orange (light and dark blue) shading denotes positive (negative) correlations significant at the 95% and 99% confidence level, respectively. The black rectangle in (a) is the TP region. Red dashed contours indicate the area over 1500 m.

Fig. 3. (a) Distribution of correlation coefficients between the summer TP thermal index and simultaneous upper-tropospheric $T^0$ during 1871–2012, in which the correlation is based on NOAA-20C reanalysis data. (b),(c) As in (a), but for ERA-20C during 1900–2010 and NCEP during 1948–2015, respectively. Yellow and orange (light and dark blue) shading denotes positive (negative) correlations significant at the 95% and 99% confidence level, respectively. The black rectangle in (a) is the TP region. Red dashed contours indicate the area over 1500 m.

To measure the variability of the TP thermal condition, referring to Fig. 2a, we define the regional mean SAT over the TP ($31^\circ$–$40^\circ$N, $69^\circ$–$93^\circ$E and $27^\circ$–$40^\circ$N, $93^\circ$–$105^\circ$E) as a TP thermal index. Figures 3a and 3b show the correlation between the summer TP thermal index and the simultaneous upper-tropospheric (300–200 hPa) $T^0$ according to the NOAA-20C and ERA-20C reanalysis datasets, respectively. It is apparent that there are positive correlations over most of the extratropical Eurasian continent and negative correlations over the extratropical central-eastern North Pacific, which constitutes a clear APO pattern. Correlation analysis reveals a correlation coefficient of 0.37 (significant at the 99.9% confidence level) between the summer TP and APO indices during 1871–2012 for NOAA-20C, and 0.39 (significant at the 99.9% confidence level) during 1900–2010 for ERA-20C. These results support the link between the APO and TP SAT. To further verify the robustness of this link, we calculate the 50-yr running
correlation coefficient between the summer TP and APO indices for the NOAA-20C and ERA-20C re-analysis datasets (Figs. 4a,b). All correlation coefficients are significant at the 95% confidence level during the last 100 years, demonstrating the robustness of the relationship between the APO and TP SAT. Because NOAA-20C only assimilates observations of surface synoptic pressure, monthly SST, and sea ice distribution (Compo et al. 2011), and ERA-20C only assimilates surface pressure and surface marine wind observations (Poli et al. 2013), we further examine the link between the APO and TP SAT on the interannual time scale (1948–2015) using the NCEP reanalysis, which assimilates both surface and radiosonde observations. The correlation pattern between the summer APO index and the simultaneous NCEP SAT (Fig. 2c) shows a significant positive correlation over the TP region. Meanwhile, the correlation distribution between the summer TP thermal index (mean SAT over the region 31°–39°N, 75°–105°E) and the simultaneous upper-tropospheric $T'$ from NCEP (Fig. 3c) also clearly displays the APO pattern. These correlation patterns are similar to those in Figs. 2a,b and 3a,b. The consistency among the NCEP, NOAA-20C, and ERA-20C reanalysis datasets further supports the close interannual relationship between the APO and TP SAT, implying the reliability of both NOAA-20C and ERA-20C in reflecting the link between the summer APO and TPSAT on the interannual time scale.

Additionally, we investigate the relationship between the summer APO and TP SAT using the CMIP5 outputs from the CCSM4 model for the period 1871–2005. It can be seen from the results that the APO index is positively correlated to the TP SAT (Fig. 5a) and that the pattern of anomalous upper-tropospheric $T'$ associated with the summer TP thermal index shows the APO pattern (Fig. 5b), with a correlation coefficient of 0.54 between the APO and TP thermal indices during 1871–2005, significant at the 99.9% confidence level. This result indicates that CCSM4 captures the observed interannual relationship between the summer APO and TP SAT reasonably well, lending further support to the robustness of this relationship.

4. Possible physical explanation

Previous studies have shown that anomalous TP heating can affect the APO anomalies during spring and summer (Nan et al. 2009; Zhou et al. 2009). In this section, we analyze in detail the possible physical processes involved in the impact of anomalous TP heating on the APO on the interannual time scale during summer.

a. Observational results

Figure 6a shows the spatial structure of the regressed vertical circulation and $T'$ against the summer TP thermal index based on NOAA-20C data over the period 1871–2012. When the summer TP SAT index is high (corresponding to a stronger TP thermal condition), tropospheric temperature increases over Asia, with upward-motion anomalies over the northern TP and its adjacent areas (between 35° and 50°N) (see the cross section along 85°E in Fig. 6a). The anomalous upward
air moves northward in the upper troposphere and then turns to move eastward at the mid- to high latitudes along anomalous upper-tropospheric westerly winds (see the cross section along 55°N in Fig. 6a). These anomalous westerly winds derive from a northeastward extension of the strengthened East Asian subtropical jet indicated by D.-Q. Huang et al. (2014). The anomalous eastward airflow turns to move southward to the north of 37°N and, concurrently, descends over the central-eastern North Pacific (see the cross section along 165°W in Fig. 6a). Meanwhile, anomalous upward motion appears to the south of the anomalous downward motion. In the cross section along 165°W in Fig. 6a, negative temperature anomalies appear in the mid- and upper
troposphere over the central-eastern North Pacific, corresponding to the anomalous downward (upward) motion over the high (low) latitudes of the central-eastern North Pacific. Similar features are also observed based on NCEP data (Fig. 6b), thus supporting the result from NOAA-20C.

Therefore, the interannual variability of the TP thermal condition is closely associated with the summer APO anomaly through the zonal and vertical circulations over East Asia and the central-eastern North Pacific. Similar features are also observed based on NCEP data (Fig. 6b), thus supporting the result from NOAA-20C.

b. Model results

In this subsection, we examine the response of the summer APO to an elevated TP heating anomaly and the related physical processes using the results from experiments CCSM3-Control and CCSM3-TP. Figure 7a shows the composite difference in 500-hPa air temperature between CCSM3-Control and CCSM3-TP (the former minus the latter). In this figure, positive anomalies of 500-hPa air temperature above 1°C appear over the TP region, with a central value exceeding 3°C, which indicates a decrease in the local 500-hPa air temperature with a topographic reduction over the TP in CCSM3-TP (i.e., an effect of elevated heating over the TP on local SAT). Positive temperature anomalies also appear in the upper troposphere over the TP and adjacent areas, with central values near the TP (Fig. 7b), which is similar to the observational results shown in Fig. 3. Meanwhile, significant negative anomalies of upper-tropospheric \( T' \) appear over the Pacific–Atlantic extratropical regions. This anomaly pattern in Fig. 7b generally exhibits an APO-like pattern. Also of note is that negative anomalies appear over the high latitudes of East Asia, which is inconsistent with the observational result (Fig. 3). Despite this inconsistency, the APO-like pattern can be clearly identified in the simulation, which demonstrates the reasonability of our model experiments and an effect of elevated TP heating on the APO pattern.

Figure 8 shows the composite differences in vertical circulation and \( T' \) between the experiments CCSM3-Control and CCSM3-TP. Corresponding to the TP elevated heating anomaly, significant positive anomalies of air temperature extend from the surface to the upper troposphere over the TP and adjacent areas, with upward-motion anomalies over the northern TP (see the cross section along 100°E in Fig. 8). These upward-motion anomalies then turn eastward along 40°N and gradually descend southward over the midlatitudes (to the north of 30°N) of the central-eastern North Pacific along 160°W, with upward-motion anomalies over the lower latitudes (to the south of 30°N). The anomalies are broadly similar to the observational ones in Fig. 6, although the simulated features shift systematically southward. Actually, the simulated southward-shifted vertical circulation anomalies are in good agreement with the systematic southward APO pattern in the CCSM3 results (Fig. 7b). These results reveal that the observed vertical circulation anomalies associated with the summer TP thermal index, shown in Fig. 6, may be forced by the TP elevated heating anomaly.

It is important to note that the reduction of the TP topography also includes a mechanical forcing effect. To isolate the individual effect of TP thermal forcing, we next examine the results from CAM3-Tree and CAM3-Bare. It is apparent that similar anomalous features are simulated solely by the TP thermal forcing. Positive temperature anomalies above 1°C again appear at the 500-hPa level over the TP (Fig. 9a). Correspondingly, significant positive upper-tropospheric \( T' \) anomalies occur.
over the Asian continent, while negative upper-tropospheric $T'$ anomalies occur over the extratropical North Pacific, which also exhibits an APO-like pattern (Fig. 9b). Figure 10 shows the composite differences in vertical circulation and $T'$ between CAM3-Tree and CAM3-Bare. Generally speaking, the tropospheric temperature and vertical circulation anomalies over the TP–eastern Pacific region are similar to those produced by the CCSM3 simulation (Fig. 8) and those observed (Fig. 6). For example, positive and negative temperature anomalies in the troposphere appear around the TP and over the extratropical North Pacific, respectively (Figs. 9b and 10), which indicates an APO pattern. Meanwhile, the upward-motion anomalies appear over the TP region, turn eastward, and descend southward over the midlatitudes of the North Pacific (Fig. 10). This similarity between the CAM3 and CCSM3 simulations further reveals that the atmospheric circulation variations simulated by changing the TP elevated heating (in Figs. 7 and 8) may mainly be triggered by an individual thermal forcing effect over the TP, and that the mechanical forcing effect, involving a change in the topographic height, is relatively weaker during summer. This is also supported by the conclusion of Liu et al. (2007), who showed that the TP's dynamic function is evidently weaker than its thermal function during summer. Thus, the CCSM3 simulations successfully reproduce the observed effect of the TP’s thermal condition on the APO. In the following analysis, we examine the mechanism underlying the impact of the TP elevated heating anomaly on the APO during summer.

Over the midlatitudes of the central-eastern North Pacific, the downward-motion anomalies may cause a reduction in local convective rainfall and condensation latent heat produced by precipitation. Figure 11a shows evidence of this reduced latent heat. The net longwave and shortwave radiation absorbed by the atmosphere and the sensible heat at the surface (figure not shown) are smaller relative to the latent heat over these latitudes. As such, mainly owing to the decrease in the latent heat, the atmospheric apparent heat source...
The composite differences in vertical circulation (red vector lines) and $T^*$ (°C; contours and shading) between CAM3-Tree and CAM3-Bare (CAM3-Tree minus CAM3-Bare). The cross sections are along 80°E, 45°N, and 155°W. Positive (negative) values significant at the 95% confidence level are shaded yellow (blue). The black shaded area denotes the terrain.

The composite differences in (a) condensation latent heat, (b) atmospheric apparent heat source ($Q_1$), (c) dry adiabatic heat between 500 and 100 hPa, and (d) the sum of ($Q_1$) and dry adiabatic heat (W m$^{-2}$) between CCSM3-Control and CCSM3-TP (CCSM3-Control minus CCSM3-TP). Yellow and orange (light and dark blue) shading denotes positive (negative) values significant at the 95% and 99% confidence level in (a)–(c) but indicates positive (negative) values greater (smaller) than 10 and 20 W m$^{-2}$ ($-10$ and $-20$ W m$^{-2}$) in (d). The upper and lower red boxes indicate the mid- ($30°$–$40°N, 170°–130°W$) and lower ($20°$–$30°N, 170°–130°W$) latitudes of the North Pacific key region of the APO, respectively.
Figure 11c shows the composite difference in dry adiabatic heat between CCSM3-Control and CCSM3-TP. As can be seen, there is only a very small area with significant positive anomalies of dry adiabatic heat, near 30°N, 170°W, while large-scale negative anomalies of dry adiabatic heat appear over the northern key area of the APO’s North Pacific sector (30°–40°N, 170°–130°W; upper red box in Fig. 11c). Therefore, there are clear negative anomalies in the sum of dry adiabatic heat and \( Q_i \) over this sector (Fig. 11d) and, evidently, the increase in dry adiabatic heat caused by the downward-motion anomalies does not balance the decrease in the local \( Q_i \). Therefore, the downward-motion anomalies forced by the TP elevated heating generally cause a decrease in the upper-tropospheric atmospheric temperature over the midlatitudes of the central-eastern North Pacific.

Over the lower latitudes of the central-eastern North Pacific, corresponding to the upward-motion anomalies, there are negative values of dry adiabatic heat (Fig. 11c). Although the upward-motion anomalies also strengthen local convective activity and increase the local \( Q_i \) to some extent (Fig. 11b), the positive anomalies of \( Q_i \) are weak (nonsignificant). Because the positive anomalies of \( Q_i \) cannot balance the reduction in dry adiabatic heat caused by the upward-motion anomalies, the sum of dry adiabatic heat and \( Q_i \) is still negative over the southern key area of the APO’s North Pacific sector (20°–30°N, 170°–130°W; lower red box in Fig. 11d). Thus, the upward-motion anomalies over the lower latitudes of the central-eastern North Pacific also lower the tropospheric temperature in situ.

5. Summary and discussion

In the present study, using NOAA-20C and ERA-20C reanalysis data, we investigate the variability of the summer APO and its relationship with the thermal condition of the TP on the interannual time scale. The results show that the interannual variation of the summer APO correlates significantly with the TP’s SAT in summer, and that this relationship is stable during the past 100 years. The relationship is also supported by NCEP reanalysis data and by CMIP5 model output.

Observation and simulation results reveal that the relationship between the TP thermal condition and the APO reflects an impact of the former on the latter. During summer, the strengthened heating over the TP can raise the mid- and upper-tropospheric temperature over Eurasia and strengthen the upward motion over the northern TP and adjacent areas. This anomalous upward flow turns eastward and then sinks southward over the mid- and high latitudes of the central-eastern North Pacific, accompanied by upward-motion anomalies over the lower latitudes of the central-eastern North Pacific. The downward-motion anomalies over the mid- and high latitudes lower the in situ mid- and upper-tropospheric temperature, mainly through weakening the condensation latent heat from precipitation and, hence, the atmospheric apparent heat source \( Q_i \). Meanwhile, the upward-motion anomalies over the lower latitudes reduce the in situ mid- and upper-tropospheric temperature, mainly through dry adiabatic cooling from the expansion of rising air masses. Accordingly, negative temperature anomalies appear in the mid- and upper troposphere over the central-eastern North Pacific, which leads to the APO pattern. In this process, the zonal vertical circulations over the Asian–North Pacific sector act as a “bridge” linking the TP climate with the North Pacific atmospheric circulation. Moreover, during summer, the TP’s thermal forcing, rather than its dynamic forcing, plays the more important role in modulating the variation of the summer Asian–Pacific climate.

In this study, we emphasize the importance of the TP thermal condition to the interannual variability of the summer APO. However, it remains unclear what factors are responsible for the variation of the thermal condition of the TP on the interannual time scale. Previous studies have documented the effects of snow cover over the TP on the local atmospheric temperature (Zhao et al. 2007; Liu et al. 2014a,b), and the effects of the previous winter’s SST anomaly in the tropical central-eastern Pacific on the subsequent summer TP temperature (Zhao et al. 2009; Liu et al. 2015). We intend to further investigate the potential effects of snow cover over the TP, and the North Pacific SST, on the variability of the summer APO in future work.

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