Interannual Variability of the North Pacific Mixed Layer Associated with the Spring Tibetan Plateau Thermal Forcing

RUIZAO SUN

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, and University of Chinese Academy of Sciences, Beijing, China

ANMIN DUAN

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, and College of Earth Science, University of Chinese Academy of Sciences, Beijing, and Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters, Nanjing University of Information Science and Technology, Nanjing, China

LILAN CHEN

School of Atmospheric Sciences, Nanjing University, Nanjing, China

YANJIE LI

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

ZHIANG XIE AND YU ZHAO

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, and University of Chinese Academy of Sciences, Beijing, China

(Manuscript received 4 September 2018, in final form 7 March 2019)

ABSTRACT

By using multisourced data and two sets of sensitivity runs from the coupled general circulation model CESM1.2.0, we investigated the effects of the spring [March, April, and May (MAM)] surface sensible heating over the Tibetan Plateau (SHTP) on the interannual variability of the North Pacific Ocean sea surface temperature (SST) and mixed layer. The results indicated that an above-normal MAM SHTP can generate a Rossby wave downstream and form an anomalous equivalent barotropic anticyclone over the North Pacific, inducing anticyclonic wind stress anomalies. As a result of Ekman transport and Ekman pumping, sea currents converge near 40°N, accompanied by weak downwelling motion. The mixed layer heat budget diagnosis indicates that the net heat fluxes, together with meridional advection anomalies, contributed significantly to changes in the mixed layer temperature (MLT). As a result, the SST anomalies (SSTAs) and MLT anomalies both present a horseshoelike pattern. In addition, the significant warm SSTAs show a maximum in the late spring, but the significant warm MLT anomalies centered under the sea surface (25-m depth) could be sustained until summer, acting like a signal storage for the anomalous spring SHTP. Moreover, the midlatitude ocean–atmosphere interaction provides a positive feedback on the development of the anomalous anticyclone over the North Pacific, since the SSTa pattern could strengthen the oceanic front and induce more active transient eddy activities. The eddy vorticity forcing that is dominant among the total atmospheric forcings tends to produce an equivalent barotropic atmospheric high pressure, which in turn intensifies the initial anomalous anticyclone.

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0577.s1>.

Corresponding author: Anmin Duan, amduan@lasg.iap.ac.cn

DOI: 10.1175/JCLI-D-18-0577.1

© 2019 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the [AMS Copyright Policy](#) (www.ametsoc.org/PUBSReuseLicenses).

1. Introduction

The Tibetan Plateau (TP) is a huge terrain in the subtropical central and eastern Eurasian continent that acts as an intense, large, and elevated heat source that exerts both thermal and dynamic effects on the climate of the Northern Hemisphere (Flohn 1957; Ye and Gao 1979; Yanai et al. 1992; Boos and Kuang 2010; Wu et al. 2012). In the boreal spring [March, April, and May (MAM)], the air column over the TP changes from an atmospheric heat sink to a source of thermal conditions that is represented mainly by surface sensible heating over the TP (SHTP) (Ye and Gao 1979; Luo and Yanai 1984; Wu et al. 1997; Duan et al. 2017). SHTP maintains a daily heat rate of 2°–4°C over the entire troposphere during the transitional season (Ye and Gao 1979; Zhang and Wu 1999; Duan and Wu 2005) and serves as an efficient air pump (Wu et al. 1997). The impacts of the SHTP on large-scale atmospheric circulation, the Asian summer monsoon and the weather systems over eastern China have been well recognized (e.g., Yanai et al. 1992; Ose 1996; Wu et al. 1997; Zhang and Wu 1999; Hsu and Liu 2003; Duan and Wu 2005; Rajagopalan and Molnar 2013).

Moreover, many previous studies have addressed the effects of TP elevated heating anomalies on the tropical and extratropical sea surface temperatures (SST) over the Pacific Ocean and associated physical processes. On one hand, the TP can affect the atmosphere–ocean interactions in the tropics and midlatitudes of the Pacific through the atmospheric circulation over the North Pacific by stimulating large-scale teleconnections, such as the Asian–Pacific oscillation, an extratropical zonal–vertical atmospheric pattern that connects Asia and the Pacific (Nan et al. 2009; Zhou et al. 2009; Zhao et al. 2011; Liu et al. 2017). On the other hand, anomalous TP heating could emanate wave energy downstream along the westerly jet stream and induce lower-layer geopotential height anomalies, further changing the meridional winds away from the equator and the zonal winds over the tropical eastern Pacific (Chen et al. 2001; Zhao and Chen 2001; Zhao et al. 2007, 2009; Wu and Li 2012). The two ways mentioned above further influence the North Pacific SST and the development of the ENSO.

However, the above studies mainly focused on the response of the Pacific SST to TP-heating anomalies. In our research, we pay more attention on the changes in the mixed layer of the North Pacific. One reason for doing so is that the changes over the tropical Pacific are mainly attributed to dynamic processes in the internal ocean, such as Ekman pumping, vertical displacement of the thermocline, advections by surface ocean

currents, and entrainment of deep ocean water (Lau and Nath 1996; Jin et al. 1999). Thus, it is challenging to discuss the influence of anomalous TP thermal conditions on the tropical Pacific that are induced by air–sea interaction. For extratropical oceans, for example, the North Pacific, it is widely agreed that midlatitude SST anomalies are mainly caused by the atmosphere through air–sea heat fluxes and upper-ocean current response (Cayan 1992; Di Lorenzo et al. 2008; Fang and Yang 2016; Na et al. 2018). Understanding the climate variations over the North Pacific Ocean is essential for discerning external influences on climate variability and for improving large-scale and regional climate predictability (Latif and Barnett 1994; Englehart and Douglas 2003; Yi et al. 2015).

In addition, the oceanic mixed layer, which is in close contact with the atmosphere, is the manifestation of vigorous turbulent mixing processes that are active in the upper ocean. The transfer of mass, momentum, and energy across the mixed layer provides the source of almost all oceanic motions, and the thickness of the mixed layer determines the heat content and mechanical inertia of the layer that directly interacts with the atmosphere (Chen et al. 1994; Levitus et al. 2001; Montégut et al. 2004; Huang et al. 2010). To make accurate estimates of oceanic variability, a better understanding of mixed layer dynamics may help to improve the distribution of SST and heat fluxes, thereby reducing the climate drift that is observed in coupled models (Sterl and Kattenberg 1994).

This study seeks to answer the following questions: What is the relationship between the spring SHTP and North Pacific mixed layer on an interannual scale? In which months does the connection between the spring SHTP and North Pacific mixed layer become the most significant? How deep do thermal anomalies penetrate and what are the main factors that induce mixed layer temperature (MLT) anomalies? Do the induced MLT anomalies have the feedback on atmospheric circulation?

The remainder of the paper is arranged as follows. Section 2 introduces the data, methods, and model used in this study. Section 3 shows the interannual variability of the North Pacific SST anomalies (SSTAs) associated with MAM SHTP by partial correlation analysis. Section 4 presents the influences of the spring SHTP on the North Pacific mixed layer in coupled general circulation model (CGCM) experiments in comparison with data analysis. Section 5 discusses the mechanisms of the formation of the anomalous barotropic anticyclone and oceanic feedback over the North Pacific. A summary and discussion are presented in section 6.

2. Data, methods, and model

a. Data

The data used in this study include the following:

- 1) The surface air temperature T_a , ground surface temperature T_s , and wind speed at 10 m above the surface V_{10} derived from regular surface meteorological observations at 73 stations over the TP, provided by the China Meteorological Administration (CMA): Data were gathered four times per day (0000, 0600, 1200, and 1800 UTC). The CMA routinely applied quality control procedures to eliminate erroneous data and ensure homogeneity. SHTP was calculated according to the bulk aerodynamic method as $SH = C_p \rho C_{DH} V_{10} (T_s - T_a)$ [for details, see [Duan et al. \(2011\)](#)]. Note that, for the choice of C_{DH} , a recent study ([Zhao et al. 2018](#)) suggested a lower C_{DH} value over the TP, but the interannual variability of SHTP has little change for the selection of the C_{DH} values in our study. For convenience, therefore, we still take that of the original calculation scheme in [Duan et al. \(2011\)](#).
- 2) The monthly geopotential height, zonal winds, and meridional winds in the isobaric coordinate system, whose horizontal resolution is $1.5^\circ \times 1.5^\circ$, with 37 vertical pressure levels and wind speed at 10 m above the surface, were gathered from the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim; [Dee et al. 2011](#)).
- 3) The monthly SST data were derived from the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST1) dataset at a horizontal resolution of $1^\circ \times 1^\circ$ ([Rayner et al. 2003](#)).
- 4) The monthly vertical diffusion heating rates were obtained from the National Centers for Environmental Prediction (NCEP)–U.S. Department of Energy AMIP II reanalysis (Reanalysis 2) at a horizontal resolution of $1.875^\circ \times 1.904^\circ$ with 28 vertical sigma levels ([Kanamitsu et al. 2002](#)).
- 5) The pentad of potential temperatures, zonal currents, meridional currents, vertical velocities, and mixed layer depths were derived from the NCEP Global Ocean Data Assimilation System (GODAS), whose domain extends from 75°S to 65°N . The data have an average $1^\circ \times 1^\circ$ horizontal resolution and 40-level vertical resolution with meridional resolution enhanced to $\frac{1}{3}^\circ$ within 10°S – 10°N ([Behringer and Xue 2004](#)).
- 6) The monthly evaporation data and the daily atmospheric and surface heat flux data were from the Japanese 55-yr Reanalysis (JRA-55; [Ebata et al. 2011](#)),

with a horizontal resolution of $1.25^\circ \times 1.25^\circ$. The pentad variables of GODAS and daily variables of JRA-55 are interpolated to a horizontal resolution of $1^\circ \times 1^\circ$ for convenient analysis in the MLT tendency equation.

All of the datasets range from 1980 to 2014, except for the vertical diffusion heating rate data (which range from 1980 to 2006 because of the time limit). This study is based primarily on monthly mean data. To focus on the interannual variability of the system, the linear trend is excluded from the datasets along with the extraction of an interannual signal by a Lanczos 2–9-yr bandpass filter.

In addition, the NCEP–National Center for Atmospheric Research (NCAR) daily reanalysis data are used to calculate spring-mean (MAM) transient eddy thermal and dynamical forcing based on the definitions presented in [section 5](#). MAM diabatic heating is diagnosed as a residual in the thermodynamic equation ([Fang and Yang 2016](#)).

b. Methods

Atmospheric forcing may induce MLT anomalies through altered wind-driven horizontal and vertical motions, vertical turbulent mixing, and surface energy fluxes. To understand the relative roles of ocean advection and surface heat flux terms in causing the North Pacific MLT tendencies, the oceanic mixed layer heat budget needs to be diagnosed. The MLT tendency equation is written as follows ([Lin et al. 2007](#)):

$$\frac{\partial T}{\partial t} = -\underbrace{u \frac{\partial T}{\partial x}}_A - \underbrace{v \frac{\partial T}{\partial y}}_B - \underbrace{w \frac{\partial T}{\partial z}}_C + \underbrace{\frac{Q_{\text{net}} - Q_{\text{pen}}}{\rho_0 c_p H}}_D + \underbrace{\varepsilon}_E \quad (1)$$

A B C D E F,

In Eq. (1), $\overline{(\quad)} = \frac{1}{H} \int_{-H}^0 dz$ represents the vertical means of the oceanic terms over the mixed layer, where H is the mixed layer depth, T is the ocean temperature, u is the zonal current, v is the meridional current, and w is the vertical current speed. Flux Q_{net} is the net downward (down direction defined as positive) surface heat flux (including the surface sensible, latent heat fluxes, net downward shortwave radiation at the ocean surface, and net downward longwave radiation), and Q_{pen} is the downward shortwave penetration through the mixed layer depth. The seawater density is defined as $\rho_0 = 1.029 \times 10^3 \text{ kg m}^{-3}$, with a specific heat of $C_p = 3996 \text{ J kg}^{-1} \text{ K}^{-1}$. The calculation of Q_{pen} follows the solar radiation penetration parameterization scheme introduced by [Paulson and Simpson \(1977\)](#):

$$Q_{\text{pen}} = \text{SW} \times [R \times e^{-H/L_1} + (1 - R)e^{-H/L_2}], \quad (2)$$

where SW is the downward shortwave radiation at the sea surface, $R = 0.58$, $L_1 = 0.35$ m, and $L_2 = 23$ m. The ε from Eq. (1) represents the residual term, which is a relatively small term.

c. Model

In this study, we used the Community Earth System Model, version 1.2.0 (CESM1.2.0), developed by NCAR (for more details, see <http://www.cesm.ucar.edu/models/cesm1.2/>). The atmospheric component used in this study is the Community Atmospheric Model, version 4.0 (CAM4.0; Neale et al. 2010); the land component is the Community Land Model, version 4.0 (CLM4.0; Oleson et al. 2010); the oceanic component is the Parallel Ocean Program, version 2 (POP2; Smith et al. 2010); the river-runoff component is the River Transport Model; and the sea ice component is the Los Alamos sea ice model, version 4 (CICE; Hunke and Lipscomb 2010).

1) EXPERIMENTAL DESIGNS

CESM is a fully coupled, global climate model that supports numerous resolutions and component configurations. In this study, we used the f19_g16 configuration with approximately 2° horizontal resolution for the atmospheric model and approximately 1° for the ocean model with the “B_2000” component selected (for more details, see <http://www.cesm.ucar.edu/models/cesm1.2/cesm/doc/modelnl/compsets.html>). To investigate the influences of the spring SHTP on the North Pacific SSTA, we conducted three experiments using the fully coupled CESM. The design of the experiments is shown in Table 1. In the first experiment, referred to as the control run (CNTL), the atmosphere and the ocean were freely coupled, with the external forcing (such as from greenhouse gases, aerosols, volcanic activity, and solar activity) fixed at the year-2000 level. This experiment was integrated for 120 yr, and we used outputs for the last 40 yr. The final two experiments were TP-heating experiments. One of the TP experiments represented an above-normal SHTP, referred to as “Strong” (Table 1), which was the same as the CNTL but with a vertical profile of sensible heating (vertical diffusion heating) at the grid points in the TP region (70° – 110°E ; 20° – 45°N) (Duan et al. 2017) with an elevation above 1500 m manually set to be 150% of normal during the spring months (MAM). We initialized the simulation in “warm start” mode, which meant using the restart files provided by the first year of CNTL and integrated for 40 yr. The second TP experiment represented the below-normal SHTP, referred to as “Weak” (Table 1), which was designed the same as the Strong

TABLE 1. Design of the CGCM experiments.

Expt	Design
CNTL	The atmosphere and ocean are freely coupled, with the external forcing fixed at the year-2000 level. This experiment was integrated for 120 yr. Outputs for the last 40 yr were used.
TP heating: Strong	The vertical profile of vertical diffusion heating at the grid points in the TP region (70° – 110°E ; 20° – 45°N) with elevation above 1500 m manually set to be 150% of the CNTL during the spring months (MAM). The simulations were initialized in “warm start” mode, meaning that we used the restart files provided by the 81st year of CNTL and integrated for 40 yr.
TP heating: Weak	As in the Strong experiment, but with the vertical profile of vertical diffusion heating set to be 50% of the CNTL.

experiment but with the vertical profile of sensible heating set to be 50% of normal. This design closely followed those presented by Wang et al. (2014) and Hu and Duan (2015).

2) MODEL EVALUATION

Previous studies have verified that CESM could capture the relatively large SST anomaly variance in the tropical central and eastern Pacific, as well as in the North Pacific (Capotondi 2012; Capotondi 2013; Jha et al. 2014). After checking the ENSO representation in the model, generally, the spatial pattern of climatological mean SST and SSTA variance, the ENSO seasonality, and ENSO spectral characteristics diagnosed from Niño-3.4 SST anomalies were reproducible (Fig. 1). Modeling overestimated the intensity of ENSO events, and the cold tongue over the eastern tropical Pacific extended more westward in the model compared to observations (Fig. 1a). In addition, there were two peaks in the power spectrum of the Niño-3.4 index with periods of approximately 2–4 and 5–6 yr in the observation (Fig. 1c), but the model did not capture the two peaks, particularly for the 5–6-yr-period peak. These differences between the model and observations were consistent with previous studies (e.g., Capotondi et al. 2006; Deser et al. 2006; Bellenger et al. 2014; Taschetto et al. 2014; Tang et al. 2016). Next, we plotted the observed and simulated 850-hPa horizontal wind vectors, SST, and mixed layer depth, as well as their differences in the spring (Fig. 2). The position and intensity of the large-scale circulation simulated by the control run (Fig. 2b) presented a general resemblance of the observations (Fig. 2a), including the western Pacific subtropical high and the westerlies over the midlatitude.

FIG. 1. (a) Spatial pattern of climatological mean SST (contours; K) and SST anomaly (SSTA) variance (shading) in the observations (top panel) and the CESM control run (bottom panel). (b) ENSO seasonality diagnosed from monthly average standard deviations of the Niño-3.4 SST anomalies in the observations (red bars) and the CESM control run (blue bars). (c) ENSO spectral characteristics diagnosed from Niño-3.4 SST anomalies in the observations (black line) and the CESM control run (red line).

In addition, the model (Fig. 2b) could reasonably reproduce the SST and mixed layer depth distribution and magnitude compared to the observations (Fig. 2a). The differences between the simulation and the observation slightly overestimated SST and slightly underestimated mixed layer depth in the model over the central North Pacific (Fig. 2c). We also obtained similar results for the middle and upper troposphere (figures not shown). These results indicated that the CESM can reproduce the spring mean SST pattern, mixed layer structure, and atmospheric circulation in the Pacific.

The ability of the model to capture the basic features of SHTP in terms of intensity and annual cycle was critical to estimate its impact on the oceanic response in the North Pacific. To verify this point, we interpolated

the simulated SHTP to the 73 stations included in the datasets and plotted the annual cycle (Fig. 3a). The model was able to simulate the SHTP's seasonal variability in the observations, including the annual peak in spring. SHTP depends primarily on the change in surface wind speed and the temperature difference between the ground surface and surface air according to the bulk formula (Duan et al. 2011), but for seasonal and annual mean surface air temperatures, the model always shows substantial cold biases over the TP, especially during cold seasons (Chen and Frauenfeld 2014; Duan et al. 2014). Thus, the simulated values of SHTP were smaller than the stations' values in the spring. Next, we compared the vertical diffusion heating averaged over the TP region (70°–110°E; 20°–40°N),

FIG. 2. The spring mean SST (shading; $^{\circ}\text{C}$), 850-hPa wind vectors (arrows; m s^{-1}), and mixed layer depth (m; contour interval is 20 m) in the (a) observations and (b) the coupled CESM, along with (c) the difference between (b) and (a).

which could represent the sensible heating conditions between the reanalysis data and the simulation. The black, red, and blue solid lines in Fig. 3b represent the climatology ($\sim 4 \text{ K day}^{-1}$), strongest ($\sim 6 \text{ K day}^{-1}$), and weakest ($\sim 2 \text{ K day}^{-1}$) vertical profiles of the sensible heating in the spring from 1980 to 2014. The noted range was approximately 0.5–1.5 times that of the climate mean state. This range provided a reference for the sensitivity run design of the vertical diffusion heating. The simulation result is shown in Fig. 3c, demonstrating that the model was able to capture the vertical profile of the sensible heating well, even though it overestimated the magnitude by approximately 2 K day^{-1} when compared with the reanalysis data.

Overall, despite these discrepancies, the consistency in the large-scale circulation, mixed layer condition, and surface sensible heating over the TP between the models and observations provided confidence for the following experimental results.

3. Interannual variability of the North Pacific SSTAs associated with the spring SHTP

This section examines the spatial pattern of the North Pacific SSTAs connected with the spring SHTP by correlation and partial correlation analysis. Here, we define the SHTP index as 73-station-averaged spring (MAM) SH over the TP (Duan et al. 2017) and the North Pacific domain as areas over 20° – 60°N and 110°E – 110°W (Frankignoul and Sennéchaël 2007). For clarity, unless

otherwise noted, the SHTP mentioned in the following text refers to the MAM SHTP and ENSO is defined by the Niño-3.4 SST anomalies in the previous winter [December, January, and February (DJF)]. The correlation and partial correlation fields associated with MAM North Pacific SST are shown in Figs. 4a–c. When the SHTP was above normal (Fig. 4a), the SSTAs over the North Pacific featured a horseshoelike pattern, with basinwide warm anomalies in the central North Pacific surrounded by cold anomalies along the west coast of the North America; this SSTA pattern associated with the SHTP in spring is similar to that associated with the summer Asian tropospheric temperature (Zhao et al. 2011). In contrast to the SHTP, the warm ENSO events seem to have the opposite effect on the SSTAs in the central and eastern North Pacific (Fig. 4b). Weare et al. (1976) noted that the above-normal SST conditions in the equatorial Pacific are accompanied by anomalous cold water in the central North Pacific, as well as warm waters off the western North American coasts. As previously mentioned, the ENSO is the dominant signal of the climate system in terms of the interannual variability; it may prevent us from judging the influence of SHTP on North Pacific SSTAs to a large extent. Thus, we eliminated the ENSO signal by partial correlation (Fig. 4c). The horseshoelike SSTAs proved to be even more significant than the data presented in Fig. 4a, with the maximum amplitude of SSTAs along 40°N in the central and eastern North Pacific.

Next, we inspected the accompanying large-scale circulation pattern related to the SSTAs. Figures 4d–f display the simultaneous partial correlation fields between SHTP and the geopotential height, as well as winds at different tropospheric levels (i.e., 200, 500, and 850 hPa) in spring. The above-normal SHTP connected with an anomalous anticyclone and positive geopotential height anomalies over the North Pacific exhibited an equivalent vertical barotropic structure. From midlevels to high levels (Figs. 4d,e), the anomalous anticyclone and corresponding positive geopotential height anomalies were oriented southwest–northeast and extended to the south of the TP. Such a North Pacific anticyclone was also detected in association with thermal conditions over the TP in previous studies (Nan et al. 2009; Zhao and Chen 2001; Zhao et al. 2009, 2011).

How will the SSTA evolve once it appears? We conducted a lagging partial correlation between SST and MAM SHTP for three successive months (months are denoted by the first letter of each month; e.g., April, May, and June is AMJ). The results are shown in Fig. 5. Compared to the simultaneous SSTA pattern (Fig. 4c), we found that the SSTAs reached a maximum in AMJ (Fig. 5a), which lagged behind the anomalous MAM

FIG. 3. (a) Annual cycle of the surface sensible heating averaged at the 73 stations over the TP (W m^{-2}). The black line represents the values of the station, and the green line represents the values of the CNTL run. Also shown are the spring mean vertical profile of the vertical diffusion heating (K day^{-1}) averaged over the TP (70° – 110°E ; 20° – 40°N) in the (b) NCEP2 reanalysis (the black line represents the climatic mean value, the red line represents the strongest value, and the blue line represents the weakest value during 1980–2006) and (c) model simulation (the black line represents the CNTL value, the red line represents the Strong sensitivity experiment value, and the blue line represents the Weak sensitivity experiment value).

SHTP by 1 month, indicating that the atmospheric force dominated the air–sea interaction (Frankignoul and Sennéchaël 2007; Frankignoul and Hasselmann 1977).

4. Comparison between CGCM experimental results and data analysis on the changes of mixed layer over the North Pacific

To further explore the results (Figs. 4 and 5), this section shows the influence of the spring SHTP on the North Pacific atmospheric circulation and mixed layer by comparing the CGCM experiment results with their counterpart in the data analysis.

First, we selected cases to do the composite analysis. For the observation, Fig. 6a displays the time series of normalized MAM SHTP from 1980 to 2014. We chose the strong and weak SHTP years with standard

deviations >0.7 or <-0.7 (Table 2). To be consistent with the partial correlation, we also excluded the ENSO-decaying years before composite analysis. We obtained the classification of ENSO events from <http://ggweather.com/enso/oni.htm>. The index for identifying the ENSO is the running 3-month mean SSTAs for the Niño-3.4 region (i.e., 5°N – 5°S ; 120° – 170°W). Events are defined as five consecutive, overlapping, 3-month periods at or above the $+0.5^{\circ}\text{C}$ anomaly for warm (El Niño) events and at or below the -0.5°C anomaly for cold (La Niña) events. The threshold is further divided into weak (with a 0.5 – 0.9 SST anomaly), moderate (1.0 – 1.4), strong (1.5 – 1.9), and very strong (≥ 2.0) events. According to the criteria, we categorized the years in Table 2. Then, we removed the moderate and strong event years. Then, seven strong SHTP cases (1980, 1982, 1986, 1991, 2004, 2012, and 2014) and eight weak

FIG. 4. Spatial distribution of the correlation coefficient (shading) between MAM North Pacific SST and (a) MAM SHTP or (b) DJF ENSO. (c) As in (a), but for partial correlation coefficient. Also shown is the simultaneous spatial distribution of partial correlation coefficient between MAM SHTP and geopotential height (shading) or winds (vectors) at (d) 200, (e) 500, and (f) 850 hPa. In all panels, stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test. The black vectors are significant at the 90% confidence level, and the gray vectors are not significant. The black box in (c) indicates the area of North Pacific. In some panels, the irregular black solid line, sometimes with gray shading, indicates the TP.

SHTP cases (1981, 1985, 1990, 1997, 2001, 2002, 2005, and 2013) were selected (boldfaced years in Table 2).

For the experiments, we also selected years without ENSO events in the experiments to obtain reliable results. The index for identifying the ENSO was the same as above, with the running 3-month mean SSTAs for the Niño-3.4 region in the CNTL run. Figure 6e displays the time series of the running 3-month mean Niño-3.4 index values during the last 40 years in the CNTL run. According to the criteria, the years between the two dashed lines whose mean SSTAs were below the $+0.5^{\circ}\text{C}$ or above the -0.5°C as the no ENSO events years were chosen and are presented in Fig. 6e. We selected years 90, 92, 93, 95, 100, 101, 103, 108, 111, 113, and 115. Next, we used these years to obtain the difference fields between the Strong and Weak experiments.

a. Comparison of atmospheric circulation patterns over the North Pacific

Figure 6 displays the spring anomalous atmospheric circulation patterns at different tropospheric levels (i.e., 200, 500, and 850 hPa) in the observation composite analysis (Figs. 6b–d) and the composite results of the CGCM-sensitivity experiments (Figs. 6f–h). Similar to Figs. 4d–f, when the SHTP was above normal, anomalous anticyclone and positive geopotential height

anomalies appeared over the North Pacific and exhibited an equivalent vertical barotropic structure. Note that, in our previous study, similar results (Fig. 8 in Duan et al. 2017) were obtained from the CGCM experiments simulated in the Flexible Global Ocean–Atmosphere–Land System Model, spectral version 2 (FGOALS-s2), so the response of the large-scale circulation to the anomalous spring SHTP was not sensitive to the chosen model.

b. Comparison of mixed layer changes over the North Pacific

From Fig. 5 it is apparent that the SSTA is featured by a tripole pattern in the longitudinal direction, so we take the average over 150°E – 150°W for further analysis of the vertical structure and evolution of the mixed layer in both observations and modeling.

Figures 7a–d display the observation composite differences in the latitude–depth profile of the North Pacific mixed layer between strong and weak SHTP cases. For the evolution of SSTA (Figs. S1a–d in the online supplemental material), it is apparent that it is very similar to the partial correlation analysis results shown in Fig. 5. The signal of warm SSTA reaches the maximum in AMJ (Fig. S1b) and gradually disappears afterward. However, for the MLT anomalies (Figs. 7a–d), when the signal of a significant warm SSTA disappears,

FIG. 5. Spatial distribution of partial correlation coefficient between MAM SHTP and (a) AMJ, (b) MJJ, and (c) JJA North Pacific SST (shading) or 850-hPa winds (vectors). Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test. The black vectors are significant at the 90% confidence level, and the gray vectors are not significant. Shown in (a) are the A1, A2, and A3 averaging distances that are used in the tripole index.

the maximum anomalies gradually move to a depth of approximately 25 m below the sea surface (Fig. 7c) and dampen slowly after summer (supplemental Figs. S2a–d). Accompanying the MLT anomalies, anomalous northward sea currents from the low latitudes and southward sea currents from the high latitudes converge and then sink near 40°N.

According to the Ekman pumping theory, in which the wind stress curl generates ocean-surface divergence or convergence, water upwelling will occur in the center of low-pressure areas and downwelling will occur in high-pressure areas (Pickett and Paduan 2003). As indicated in Fig. 6d, the positive geopotential height anomalies center is around 40°N, so the above-sea current anomalies match well with the anomalous atmospheric circulation pattern. If we further compare the changes between SSTA and MLT, it is obvious that, even though the significant SSTA signal disappears after late spring, the weak downwelling motion (noting that the magnitude of vertical velocity is 10^{-5} of the meridional velocity) does bring heat to the subsurface, and the heat can be stored for several months (until summer) because of the large heat capacity of the ocean. That is, considering only the evolution of SSTA may underestimate the duration of the anomalies over

the ocean, and the ocean mixed layer helps to maintain the signal of the anomalous spring SHTP longer. To verify this point, the simulation results from the coupled model are compared with the counterpart of the data diagnosis in the following section.

The differences in successive 3-month mean MLT and sea currents between the Strong and Weak experiments are plotted in Figs. 7e–h. In general, the simulation can reproduce the observational results well; that is, the warm SSTA shows a maximum in AMJ (Figs. S1e–h) and then the warm core gradually moves downward (Fig. 7g), accompanied by the convergence and sinking of sea currents near 40°N (Fig. 7e). Moreover, the cold anomalies near 60°N (Figs. 7e–h) persist even longer than the warm anomalies. This phenomenon may be associated with the change in mixed layer depth (MLD) over the North Pacific. So, we checked the MLD variation next.

At the southern edge of the anomalous anticyclone (Figs. 6d,h), the anomalous wind direction was opposite to the climatologic wind fields (Figs. 2a,b), and hence, the MLD decreased due to the weakened wind fields (blue contours in Figs. 6d,h). In contrast, at the east and north edges of the anomalous anticyclone (Figs. 6d,h), the anomalous wind direction was consistent with the climatologic wind fields (Figs. 2a,b), and the MLD increased (red contours in Figs. 6d,h) for the strengthened wind fields. That is, the cold anomalies corresponded to the deeper MLD and the warm anomalies corresponded to the shallower MLD. Indeed, Bathen (1972) noted that lower heat storage in the shallow MLD causes the MLT to be sensitive to high-frequency processes such as clouds and fronts. For the deep MLD, such high-frequency processes are considerably less effective. This can explain why cold anomalies last longer than warm anomalies.

We display the spatial distributions of the MLT and sea currents at different depths to show the three-dimensional features of the mixed layer. In general, the mixed layer depth was approximately 60–100 m (Figs. 2a,b) in the North Pacific during the boreal spring (Kara et al. 2000). We selected three depths as typical layers: 5 m for the near surface, 40 m for the middle of the mixed layer, and 100 m for the intersection of the mixed layer and thermocline. From both composite results of the data and model (left and right panels of Fig. 8, respectively), it is apparent that the anomalous MLT patterns were similar to the SSTA distribution (Fig. 5), with basin-wide warm anomalies in the central North Pacific surrounded by the cold anomalies and MLT anomalies that gradually decay with depth.

An overlay of the surface wind stress (the green vectors) on the ocean currents at a depth of 5 m is shown in Figs. 8a and 8d. It is apparent that the pattern of

FIG. 6. (a) Time series of normalized MAM SHTP from 1980 to 2014. The solid horizontal lines represent ± 0.7 standard deviations. Also shown is the composite difference of geopotential height (shading; gpm) and winds (vectors; m s^{-1}), at (b) 200 (c) 500, and (d) 850 hPa between the strong and weak SHTP cases. The mixed layer depth difference (m; contour interval is 8 m, with red and blue indicating positive and negative values and blue, respectively) is overlain on (d). (e) Time series of running 3-month mean SSTAs for the Niño-3.4 region in the last 40 yr of the CNTL run. The solid horizontal lines represent $\pm 0.5^\circ\text{C}$ of SSTA. (f)–(h) As in (b)–(d), but for the difference between the Strong and Weak sensitivity experiments. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student’s test.

anomalous wind stress was similar to the anomalous atmospheric circulation shown in Figs. 6d and 6h, featured by the anomalous anticyclone over the central North Pacific. The anomalous ocean currents were most significant at the near-surface (Figs. 8a,d; black vectors) pointing to the right side of wind stress (green vectors), with the strength of currents gradually decaying with depth. This process is called Ekman transport, in which a steady wind stress acts together with the Coriolis

force to cause the transport of water to the right of the wind (Price et al. 1987). This theory also explains why the sea currents could converge near 40°N , as mentioned above (Fig. 7).

c. Comparison of energy budget of the mixed layer over the North Pacific

What are the main factors that induce MLT anomalies? By using the MLT tendency equation mentioned

TABLE 2. The years with standard deviations greater than 0.7 or less than -0.7 in Fig. 6a. Here, W represents weak ENSO events, M represents moderate ENSO events, and S represents strong ENSO events. POS represents strong SHTP cases; NEG represents weak SHTP cases. The boldfaced years are selected for the composite analysis.

ENSO	W			M		M	S	M		M		
POS	1980	1982	1986	1987	1991	1995	1998	2003	2004	2010	2012	2014
ENSO		W	S			W		W				
NEG	1981	1985	1988	1990	1997	2001	2002	2005	2013

FIG. 7. The composite difference of the longitude–depth profile averaged between 150°E and 150°W for (a) MAM, (b) AMJ, (c) MJJ, and (d) JJA ocean potential temperature (shading; K) and ocean currents (vectors; v : m s^{-1} and w : 10^{-5} m s^{-1}) between the strong and weak SHTP cases. (e)–(h) As in (a)–(d), but for the difference between the Strong and Weak sensitivity experiments. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test. The black vectors are significant at the 90% confidence level, and the gray vectors are not significant.

above, we calculate each term of Eq. (1). In Eq. (1), term A represents the MLT tendency; terms B–E are the zonal advection, meridional advection, vertical advection, and net surface heat fluxes, respectively. Term F represents either model errors or processes other than heat fluxes and temperature advection (Lin et al. 2007).

We analyzed their relative contributions to temperature changes in the mixed layer for both observations and experiments. Figure 9 shows the differences between strong and weak SHTP cases for each term in Eq. (1) in data composite analysis. As shown in Fig. 9a, the anomalous MLT tendency also showed a horseshoelike pattern, with significant warming at $0.24\text{--}0.36 \text{ K month}^{-1}$ over the central North Pacific and cooling of from -0.24 to $-0.36 \text{ K month}^{-1}$ along the west coast of North America. The sum of terms B, C, D, and E from the right side of Eq. (1) is shown in Fig. 9f. The pattern of

Fig. 9f is similar to that of Fig. 9a, suggesting that the calculation on both sides of Eq. (1) was generally balanced. Next, we analyzed the relative contribution to the MLT tendency changes term by term. The differences from term B–E in Eq. (1) correspond to Figs. 9b–e. The zonal advection (Fig. 9b) and vertical advection (Fig. 9d) terms contributed only slightly to the MLT tendency changes. The net heat fluxes (Fig. 9e) significantly contributed to the positive MLT tendency changes of $0.24\text{--}0.36 \text{ K month}^{-1}$ increasing over the central North Pacific and the negative MLT tendency changes of from -0.24 to $-0.36 \text{ K month}^{-1}$ decreasing over the southeast North Pacific. The meridional advective (Fig. 9c) were the another important factor for the overall MLT tendency changes, which favored positive MLT tendency changes of $0.06\text{--}0.24 \text{ K month}^{-1}$ over the southeastern North Pacific because of Ekman transport from south to north (Fig. 8a) and negative MLT tendency changes of

FIG. 8. The composite difference of MAM ocean potential temperature (shading; K) and ocean currents (vectors; m s^{-1}) at depths of (a) 5, (b) 40, and (c) 100 m between the strong and weak SHTP cases. The green arrows in (a) denote the surface wind stress (dyn cm^{-2}). (d)–(f) As in (a)–(c), but for the difference between the Strong and Weak sensitivity experiments. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test. The black vectors are significant at the 90% confidence level, and the gray vectors are not significant.

from -0.06 to $-0.18 \text{ K month}^{-1}$ over the northeastern North Pacific because of Ekman transport from north to south (Fig. 8a). However, its contribution was much weaker than the net heat fluxes.

Figure 10 is similar to Fig. 9, but presents the differences between two sensitivity experiments. The simulation also reproduced similar results when calculated with the reanalyzed data. Correspondingly, the net heat fluxes (Fig. 10e) significantly contributed to the positive MLT tendency changes with 0.18 – $0.36 \text{ K month}^{-1}$ increasing over the central North Pacific and the negative MLT tendency changes with from -0.12 to $-0.18 \text{ K month}^{-1}$ decreasing along the western coast of the North America. Additionally, the meridional advective (Fig. 10c) were another noticeable factor for the overall MLT tendency changes that favored the positive MLT tendency changes at 0.06 – $0.18 \text{ K month}^{-1}$ over the southeast North Pacific and negative MLT tendency changes at from -0.06 to $-0.12 \text{ K month}^{-1}$ over the northeastern North Pacific.

Furthermore, we split the net heat fluxes (Figs. 9e, 10e) into sensible heat fluxes, latent heat fluxes, shortwave radiation, and longwave radiation, as displayed in Fig. 11. Obviously, the change pattern in the net heat fluxes was mainly attributed to the anomalous latent heat fluxes (Figs. 11b,f). Sensible heating, together with

shortwave radiation, also contributed to the anomalies to a limited degree (Figs. 11a,c,e,g). So, what are the main factors causing the anomalous latent heat fluxes? Figures 12a and 12b show the anomalous surface wind stress and evaporation. When compared with the climatologic wind fields shown in Figs. 2a and 2b, we found that the anomalous easterlies along 30° – 40°N were opposite to the climatologic wind fields (Figs. 2a,b), causing the surface evaporation to decrease and further warming the SST. The anomalous northwesterlies and northeasterlies along the west coast of North America were consistent with the climatologic wind fields and, thus, they facilitated surface evaporation, leading to further SST cooling. The above results were similar to those of Fang and Yang (2016), who noted that the initial midlatitude surface westerly anomaly accompanied by an intensified Aleutian low tended to force a negative SST anomaly by increasing upward surface heat fluxes and driving a southward Ekman current anomaly.

The evolution of the MLT and MLT tendency averaged at the area of warm anomalies (30° – 45°N , 150°E – 150°W , and 0 – 50 m ; the green box in Figs. 7a,e) are illustrated in Figs. 12c and 12d. It is obvious that, because of the strong spring SHTP, the MLT over the

FIG. 9. The composite difference between strong and weak SHTP cases in the MAM mean mixed layer heat budget (K month^{-1}) calculated from Eq. (2) for (a) term A; (b) term B; (c) term C; (d) term D; (e) term E; (f) the sum of terms B, C, D, and E. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test.

central North Pacific can warm up by approximately 0.2 K in 2 months for observations and 1 month for the model. Prior research on extratropical atmosphere–ocean interactions has examined the lag correlations between monthly atmospheric and SSTa fields, finding that large-scale air–sea coupling is strongest when the atmosphere leads the ocean by 1 month at midlatitudes (Davis 1976; Lanzante 1984; Wallace and Jiang 1987). In addition, the MLT tendency continued to fall as the value changed from positive to negative. Frankignoul and Sennéchaël (2007) noted that, once the SSTAs were generated, the associated MLT tendency would gradually change sign as a result of a negative turbulent heat flux feedback, and therefore the SSTAs damp slowly.

On the basis of above analysis over the North Pacific, the influences of MAM SHTP on the North Pacific SSTAs and mixed layer can be summarized as follows. When the SHTP is above normal, large-scale circulation is characterized by an anomalous anticyclone over the entire North Pacific in the lower troposphere. Therefore, changes in the near sea surface wind fields lead to variations in net surface heat flux, which are mainly caused by latent heating, along with meridional advections caused by Ekman transport. Then, the SSTAs and MLT anomalies over the North Pacific

present horseshoelike patterns in 1–2 months. In addition, the MLT anomalies damp more slowly than the SSTa, reflecting the good heat storage capacity of the mixed layer.

5. Mechanisms of the formation of an anomalous equivalent barotropic anticyclone over the North Pacific

From the above analysis, we found that changes in the North Pacific could be attributed mainly to the anomalous equivalent vertical barotropic anticyclone over the North Pacific. Our remaining questions are: How was the signal that originated from the anomalous SHTP conveyed to the North Pacific? Why is there an equivalent vertical barotropic anticyclonic anomaly over the North Pacific? Does the horseshoelike SSTa pattern have feedback with the equivalent vertical barotropic structure of the anomalous anticyclone? To answer these questions, we do further analyses to explain the mechanisms.

a. Rossby wave and vorticity budget diagnosis

We used the wave activity flux formulated by Takaya and Nakamura (2001) to confirm the propagation of

FIG. 10. As in Fig. 9, but for the difference between the Strong and Weak sensitivity experiments.

a stationary Rossby wave. Additionally, we applied the Rossby wave ray tracing method (Li and Li 2012; Li et al. 2015; Zhao et al. 2015, 2019) to characterize the pathway of the influence of spring TP heating.

The wave activity flux is calculated from geopotential height anomalies between two sensitive experiments because these anomalies are planetary in theory (Takaya and Nakamura 2001). We calculated the wave ray tracing with the smoothed climatological MAM basic flow by setting the start point at the TP region and the initial zonal wavenumber 4. The details of the calculation can be found in Li et al. (2015). We also tested the sensitivity of the ray trajectories to the initial zonal wavenumber. The results were generally consistent for the initial zonal wavenumbers 3–5. The wave activity flux pattern (Fig. 13a) suggested that strong southeastward-propagating waves were entering the North Pacific. The wave rays (Fig. 13a) sourced from the TP regions showed significant eastward propagation in the midlatitudes, passing through the North Pacific. The Rossby wave diagnosis indicated that the interannual changes in the spring SHTP could alter wave activity and circulation over the North Pacific section.

Vorticity analysis is clearly of value to better understand the anomalous subtropical general circulation related to the TP heating. After neglecting

vertical advection, twisting, and part of the divergence term ζD , the vorticity equation can be approximated as (Lu 2001)

$$\varepsilon \zeta = -\mathbf{V} \cdot \Delta \zeta - \beta v - fD, \quad (3)$$

where ε is a coefficient of Rayleigh friction; ζ and D are relative vorticity and divergence, respectively; and $\mathbf{V} = (u, v)$ is horizontal wind. The three terms on the right side of Eq. (3) respectively represent the advection of relative vorticity, the advection of planetary vorticity, and divergence.

Figures 13b–d show the differences in these three terms for two sensitivity experiments. The negative advection of relative vorticity (Fig. 13b), together with the positive advection of planetary vorticity east of 180° and negative advection of planetary vorticity west of 180° (Fig. 13c) over the North Pacific, both of these conditions were favorable for the formation of anomalous anticyclones. Additionally, the sign of divergence (Fig. 13d) was opposite to the relative vorticity advection, suggesting that the divergence tended to compensate for the advection of perturbation vorticity in waves with various frequencies and, finally, only the stationary wave train without frequency dispersion could exist. For the vertical barotropic structure, the “advection limit” theory of Smagorinsky (1953) suggests that

FIG. 11. The composite differences between the strong and weak SHTP cases in MAM mean (a) sensible heating, (b) latent heating, (c) shortwave radiation, and (d) longwave radiation (K month^{-1}). (e)–(f) As in (a)–(d), but for the difference between the Strong and Weak sensitivity experiments. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test.

when there is local heating then wind direction changes with height. In areas outside the heating zone, the wind direction does not vary with height, showing an equivalent barotropic structure.

b. Positive feedback induced by the ocean–atmosphere interaction on the development of an equivalent barotropic anticyclone

When displaying the vertical structure of anomalous atmospheric circulation, while the atmospheric temperature anomalies over the North Pacific show a vertical baroclinic structure with warm anomalies over the low level and cold anomalies over the high level, the geopotential height shows an equivalent barotropic structure in terms of the hydrostatic balance (Figs. 14a,b); such a phenomenon was called an equivalent barotropic warm/ridge structure (Fang and Yang 2016).

Unlike those in the tropics where thermal forcing is dominant, the midlatitude atmospheric seasonal-mean anomalies are generally linked to transient eddy forcing, especially that from synoptic eddies. Thus, causative and maintenance mechanisms of the equivalent barotropic warm/ridge structure and the associated processes of midlatitude ocean–atmosphere interactions must be examined.

Recent studies by Fang and Yang (2016) and Wang et al. (2017, 2019) proposed a dynamical framework with a quasigeostrophic potential vorticity (QGPV) equation in which the diabatic heating and transient eddy forcing are both considered to diagnose the role of transient eddy forcing in generating the midlatitude seasonal-mean atmospheric anomalies. From those studies, an atmospheric transient eddy dynamical (vorticity) forcing mechanism is proposed to explain the

FIG. 12. The composite difference of evaporation (shading; mm day^{-1}) and surface wind stress (vectors; dyn cm^{-2}) (a) between the strong and weak SHTP case and (b) between the Strong and Weak sensitivity experiments. Also shown are composite differences in time evolution of MLT (K) and MLT tendency (K month^{-1}) averaged at the area of $30^{\circ}\text{--}45^{\circ}\text{N}$, $150^{\circ}\text{E--}150^{\circ}\text{W}$, and 0–50 m (the green box in Figs. 7a,e) (c) between the strong and weak SHTP cases and (d) between the Strong and Weak sensitivity experiments. In (a) and (b), stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test, the black vectors are significant at the 90% confidence level, and the gray vectors are not significant.

dynamics and structure of midlatitude climate anomalies, and an unstable midlatitude air–sea interaction mechanism is hypothesized, in which the oceanic front (meridional SST gradient) and the atmospheric synoptic transient eddy are two key players.

When the underlying SSTA shows a horseshoelike pattern (Figs. 5a, 14c), the meridional gradient of SST increases near 40°N (Fig. 14d), thus enhancing the subarctic frontal zone (Wang et al. 2017, 2019). The transient eddy activities are closely related to the oceanic front zones. Correspondingly, as an adjustment of the atmospheric boundary layer to the enhanced oceanic

front, the meridional gradient of low-level atmospheric temperature and the atmospheric baroclinicity

$$(\sigma_{\text{BI}} = 0.31f|\partial\mathbf{V}/\partial z|/N)$$

both increase north of 40°N , in which baroclinic eddies can be generated and become more active (Figs. 14e,f). The geopotential height anomaly tendency induced by the diabatic heating and transient eddy forcing anomalies can be determined by a QGPV anomaly equation proposed by Fang and Yang (2016):

$$\left(\frac{1}{f}\nabla^2 + \frac{f}{\sigma_1} \frac{\partial^2}{\partial p^2}\right) \left(\frac{\partial\Delta\bar{\phi}}{\partial t}\right) = -f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \frac{\Delta\bar{Q}_d}{T}\right) - f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \frac{\Delta\bar{Q}_{\text{eddy}}}{T}\right) + \Delta\bar{F}_{\text{eddy}} + R_v \quad (4)$$

F1
F2
F3.

The details of Eq. (4) are provided in Fang and Yang (2016). The first three terms on the right side of Eq. (4) respectively represent the vertical gradient of diabatic heating forcing (F1), the vertical gradient of transient eddy heating forcing (F2), and transient eddy vorticity forcing (F3).

To detect oceanic feedback to the atmosphere, here we define a new index (tripole index = $A1 - A2 - A3$; Fig. 5a) to perform a correlation analysis with the above

three terms, in which A1 represents the spring mean SST averaged over $170^{\circ}\text{E--}170^{\circ}\text{W}$ along the 38°N and A2 and A3 are the same as A1 but are along 22° and 55°N , respectively. Noting that the SSTA pattern (Fig. 5a) is similar to the PDO-like SSTA, to reflect the SHTP influence to a maximum extent and avoid the PDO influence (Fang and Yang 2016), we regress the spring PDO index on the tripole index by using the least squares technique. Then, the estimated PDO-related

FIG. 13. (a) Stationary wave activity flux (vectors; $\text{m}^2 \text{s}^{-2}$) and stationary Rossby wave ray trajectories (curves) with four zonal wavenumbers for TP sources (black dots) at 200 hPa. Also shown are composite differences in the (b) advection of relative vorticity term, (c) advection of planetary vorticity term, and (d) divergence term (shading; 10^{-11}s^{-2}) between the Strong and Weak sensitivity experiments at 200 hPa.

components are subtracted from the tripole index time series; the residuals are regarded as “PDO free” parts and are subjected to the analysis (Gong et al. 2011). The correlation coefficient between the tripole index and spring SHTP is increased from 0.58 to 0.63, significant at the 99% confidence level. For the spring PDO index, the correlation coefficient with the tripole index is decreased from -0.41 (significant at the 95% confidence level) to -0.10 , which is not significant. The results indicated that this new tripole index almost completely removed the signal of PDO.

The correlation fields between the new tripole index and the three forcing terms on the right side of the QGPV equation [Eq. (4)] are plotted in Fig. 15. The response of atmospheric circulation to both diabatic heating (Fig. 15a) and transient eddy heating (Fig. 15b) tends to be baroclinic over the North Pacific, whereas that to the transient eddy vorticity forcing is barotropic (Fig. 15c). Moreover, the atmospheric response to all three forcing sources ($F1 + F2 + F3$) tends to display an equivalent barotropic structure (Fig. 15d) with the geopotential high centered at the upper troposphere over the North Pacific near 40°N , reflecting the dominance of transient eddy vorticity forcing among the three forcing

terms. Therefore, the initial anomalous equivalent barotropic anticyclone over the North Pacific can be further intensified, indicating a positive air–sea feedback in the midlatitude North Pacific which was found in the observations by Fang and Yang (2016).

6. Summary and discussion

a. Summary

This study presents the impacts of spring SHTP on the interannual variability in SSTAs and the mixed layer of North Pacific as determined through data analysis and model simulations. Our main findings are as follows:

- 1) When the MAM SHTP was above normal, large-scale circulation over the North Pacific exhibited an anomalous anticyclone with equivalent vertical barotropic structures. The SSTAs over the North Pacific presented horseshoelike patterns, with basinwide warm anomalies over the central North Pacific that were surrounded by cold anomalies along the western coast of the North America. Such anomalous SSTA patterns were detected more clearly after removing the ENSO signal from the previous winter.
- 2) The anomalous anticyclone over the North Pacific induced changes in wind stress, after which sea currents converged and experienced weak downwelling near 40°N due to Ekman transport and Ekman pumping. The significant warm SSTAs showed a maximum in late spring, but the significant warm MLT anomalies that were centered under the sea surface (25-m depth) could be sustained until summer, reflecting the strong heat storage capacity of the mixed layer. As such, the ocean mixed layer could seem as a signal storage for the anomalous spring SHTP.
- 3) Through the mixed layer heat budget diagnosis, we find that the anomalous MLT tendency distribution showed a horseshoelike pattern similar to SSTAs. The net heat fluxes, together with the meridional advection anomalies, contributed significantly to the MLT tendency. Moreover, the anomalous latent heating that resulted from changes in evaporation were the largest in anomalous net heat fluxes.
- 4) For the formation and maintenance of the anomalous equivalent barotropic anticyclone, we present two possible explanations. One is a direct way in which TP heating could generate a stationary Rossby wave train that passes through the North Pacific. The advection of relative vorticity and advection of planetary vorticity over the North Pacific facilitated

FIG. 14. Longitude–altitude section along 35°N of the composite difference of geopotential height (contours; gpm) and air temperature (shading; K) (a) between the strong and weak SHTP cases and (b) between the Strong and Weak sensitivity experiments. Also shown are the meridional distribution of the composite difference of (c) SST (K) and (d) SST meridional gradient $[\text{K} (110 \text{ km})^{-1}]$ and latitude–altitude sections of (e) meridional air temperature gradient $[\text{K} (110 \text{ km})^{-1}]$ and (f) Eady growth rate (day^{-1}), averaged over 150°E–150°W between the strong and weak SHTP cases. Stippling denotes regions where anomalies are significant at the 90% confidence level according to the Student's test.

the existence of the anomalous anticyclone. According to the advection-limit theory (Smagorinsky 1953; Wu et al. 2015), the Rossby wave has an equivalent barotropic structure when leaving the heating area. The second explanation is indirect, through ocean–atmosphere interactions from which the tri-pole SSTA pattern in the longitudinal direction induced by the anomalous spring SHTP, strengthening the oceanic front. The low-level atmospheric meridional temperature gradient and the low-level atmospheric baroclinicity would be enhanced as

an adjustment to the oceanic front, inducing more active transient eddy activities. According to the diagnosis of the QGPV equation, the eddy vorticity forcing that plays a leading role among all atmospheric forcings tends to produce an equivalent barotropic atmospheric high pressure near 40°N, which could intensify the anomalous anticyclone over the North Pacific. Hence, the midlatitude ocean–atmosphere interaction can provide a positive feedback for the development of an initial anomaly.

FIG. 15. The correlation fields between the new tripole index and latitude–altitude sections of (a) diabatic heat forcing (F1), (b) transient eddy heat forcing (F2), (c) transient eddy vorticity forcing (F3), and (d) total forcing ($F1 + F2 + F3$), averaged over 150°E – 150°W .

b. Discussion

As mentioned in our previous study (Duan et al. 2017), stronger spring SHTP is usually followed by an enhanced and westward extension of the western Pacific subtropical high (WPSH) in summer, and vice versa. In addition, the anomalous spring SHTP could store its signal over the equatorial Pacific Ocean, featuring warm SSTAs in the equatorial central Pacific. Subsequently, an atmospheric Rossby wave is stimulated to the northwest of the warm SSTAs, which in turn enhances the atmospheric dipole anomalies over the western Pacific. Thus, the air–sea interactions tend to reinforce the effect of SHTP on the summer WPSH anomaly. Therefore, could the signal stored in the North Pacific mixed layer that is induced by the spring SHTP exert its influence on the climate system in later seasons? And how does this work? These questions should be addressed in future studies.

Acknowledgments. We thank Prof. Jiabei Fang for providing constructive suggestions regarding the modification of our article. This work was supported by the National Natural Science Foundation of China (Grants 41725018 and 91637312). Author Yanjie Li is supported by the National Natural Science Foundation of

China (Grant 41575060). The authors also appreciate three anonymous reviewers for their constructive comments and suggestions to improve the paper.

REFERENCES

- Bathen, K. H., 1972: On the seasonal changes in the depth of the mixed layer in the north Pacific Ocean. *J. Geophys. Res.*, **77**, 7138–7150, <https://doi.org/10.1029/JC077i036p07138>.
- Behringer, D., and Y. Xue, 2004: Evaluation of the global ocean data assimilation system at NCEP: The Pacific Ocean. *Proc. Eighth Symp. on Integrated Observing and Assimilation Systems for Atmosphere, Oceans, and Land Surface*, Seattle, WA, Amer. Meteor. Soc., 2.3, https://ams.confex.com/ams/84Annual/techprogram/paper_70720.htm.
- Bellenger, H., É. Guilyardi, J. Leloup, M. Lengaigne, and J. Vialard, 2014: ENSO representation in climate models: From CMIP3 to CMIP5. *Climate Dyn.*, **42**, 1999–2018, <https://doi.org/10.1007/s00382-013-1783-z>.
- Boos, W. R., and Z. Kuang, 2010: Dominant control of the South Asian monsoon by orographic insulation versus plateau heating. *Nature*, **463**, 218, <https://doi.org/10.1038/nature08707>.
- Capotondi, A., 2012: El Niño–Southern Oscillation ocean dynamics: Simulation by coupled general circulation models. *Climate Dynamics: Why Does Climate Vary? Geophys. Monogr.*, Vol. 189, Amer. Geophys. Union, 105–122, <https://doi.org/10.1029/2008GM000796>.
- , 2013: ENSO diversity in the NCAR CCSM4 climate model. *J. Geophys. Res. Oceans*, **118**, 4755–4770, <https://doi.org/10.1002/jgrc.20335>.

- , A. Wittenberg, and S. Masina, 2006: Spatial and temporal structure of tropical Pacific interannual variability in 20th century coupled simulations. *Ocean Modell.*, **15**, 274–298, <https://doi.org/10.1016/j.ocemod.2006.02.004>.
- Cayan, D. R., 1992: Latent and sensible heat flux anomalies over the northern oceans: The connection to monthly atmospheric circulation. *J. Climate*, **5**, 354–369, [https://doi.org/10.1175/1520-0442\(1992\)005<0354:LASHFA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1992)005<0354:LASHFA>2.0.CO;2).
- Chen, D., A. J. Busalacchi, and L. M. Rothstein, 1994: The roles of vertical mixing, solar radiation, and wind stress in a model simulation of the sea surface temperature seasonal cycle in the tropical Pacific Ocean. *J. Geophys. Res.*, **99**, 20 345–20 359, <https://doi.org/10.1029/94JC01621>.
- Chen, L., and O. W. Frauenfeld, 2014: Surface air temperature changes over the twentieth and twenty-first centuries in China simulated by 20 CMIP5 models. *J. Climate*, **27**, 3920–3937, <https://doi.org/10.1175/JCLI-D-13-00465.1>.
- Chen, L. X., W. Li, and P. Zhao, 2001: Impact of winter thermal condition of the Tibetan Plateau on zonal wind anomaly over equatorial Pacific. *Sci. China*, **44D**, 400–409, <https://doi.org/10.1007/BF02912012>.
- Davis, R. E., 1976: Predictability of sea surface temperature and sea level pressure anomalies over the North Pacific Ocean. *J. Phys. Oceanogr.*, **6**, 249–266, [https://doi.org/10.1175/1520-0485\(1976\)006<0249:POSSTA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1976)006<0249:POSSTA>2.0.CO;2).
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Deser, C., A. Capotondi, R. Saravanan, and A. S. Phillips, 2006: Tropical Pacific and Atlantic climate variability in CCSM3. *J. Climate*, **19**, 2451–2481, <https://doi.org/10.1175/JCLI3759.1>.
- Di Lorenzo, E., and Coauthors, 2008: North Pacific Gyre Oscillation links ocean climate and ecosystem change. *Geophys. Res. Lett.*, **35**, L08607, <https://doi.org/10.1029/2007GL032838>.
- Duan, A. M., and G. X. Wu, 2005: Role of the Tibetan Plateau thermal forcing in the summer climate patterns over subtropical Asia. *Climate Dyn.*, **24**, 793–807, <https://doi.org/10.1007/s00382-004-0488-8>.
- , F. Li, M. Wang, and G. Wu, 2011: Persistent weakening trend in the spring sensible heat source over the Tibetan Plateau and its impact on the Asian summer monsoon. *J. Climate*, **24**, 5671–5682, <https://doi.org/10.1175/JCLI-D-11-00052.1>.
- , Z. X. Xiao, and J. Hu, 2014: Can Current AGCMs Reproduce Historical Changes in the Atmospheric Diabatic Heating over the Tibetan Plateau? *Atmos. Ocean. Sci. Lett.*, **7**, 143–148, <https://doi.org/10.3878/j.issn.1674-2834.13.0084>.
- , R. Sun, and J. He, 2017: Impact of surface sensible heating over the Tibetan Plateau on the western Pacific subtropical high: A land–air–sea interaction perspective. *Adv. Atmos. Sci.*, **34**, 157–168, <https://doi.org/10.1007/s00376-016-6008-z>.
- Ebita, A., and Coauthors, 2011: The Japanese 55-year Reanalysis “JRA-55”: An Interim Report. *SOLA*, **7**, 149–152, <https://doi.org/10.2151/sola.2011-038>.
- Englehart, P. J., and A. V. Douglas, 2003: Assessing warm season drought episodes in the central United States. *J. Climate*, **16**, 1831–1842, [https://doi.org/10.1175/1520-0442\(2003\)016<1831:AWSDEI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<1831:AWSDEI>2.0.CO;2).
- Fang, J., and X. Yang, 2016: Structure and dynamics of decadal anomalies in the wintertime midlatitude North Pacific ocean–atmosphere system. *Climate Dyn.*, **47**, 1989–2007, <https://doi.org/10.1007/s00382-015-2946-x>.
- Flohn, H., 1957: Large-scale aspects of the “summer monsoon” in South and East Asia. *J. Meteor. Soc. Japan Ser. II*, **35**, 180–186, https://doi.org/10.2151/jmsj1923.35a.0_180.
- Frankignoul, C., and K. Hasselmann, 1977: Stochastic climate models. Part II: Application to SST anomalies and thermocline variability. *Tellus*, **29**, 289–305, <https://doi.org/10.3402/tellusa.v29i4.11362>.
- , and N. Sennéchal, 2007: Observed influence of North Pacific SST anomalies on the atmospheric circulation. *J. Climate*, **20**, 592–606, <https://doi.org/10.1175/JCLI4021.1>.
- Gong, D. Y., J. Yang, S. J. Kim, Y. Q. Gao, D. Guo, T. J. Zhou, and M. Hu, 2011: Spring Arctic Oscillation–East Asian summer monsoon connection through circulation changes over the western North Pacific. *Climate Dyn.*, **37**, 2199–2216, <https://doi.org/10.1007/s00382-011-1041-1>.
- Hsu, H. H., and X. Liu, 2003: Relationship between the Tibetan Plateau heating and East Asian summer monsoon rainfall. *Geophys. Res. Lett.*, **30**, 2066, <https://doi.org/10.1029/2003GL017909>.
- Hu, J., and A. Duan, 2015: Relative contributions of the Tibetan Plateau thermal forcing and the Indian Ocean Sea surface temperature basin mode to the interannual variability of the East Asian summer monsoon. *Climate Dyn.*, **45**, 2697–2711, <https://doi.org/10.1007/s00382-015-2503-7>.
- Huang, B. Y., Y. Xue, D. X. Zhang, A. Kumar, and M. J. McPhaden, 2010: The NCEP GODAS ocean analysis of the tropical Pacific mixed layer heat budget on seasonal to interannual time scales. *J. Climate*, **23**, 4901–4925, <https://doi.org/10.1175/2010JCLI3373.1>.
- Hunke, E., and W. Lipscomb, 2010: CICE: The Los Alamos sea ice model documentation and software user’s manual, version 4.1. Los Alamos National Laboratory Tech. Rep. LA-CC-06-012, 76 pp.
- Jha, B., Z. Z. Hu, and A. Kumar, 2014: SST and ENSO variability and change simulated in historical experiments of CMIP5 models. *Climate Dyn.*, **42**, 2113–2124, <https://doi.org/10.1007/s00382-013-1803-z>.
- Jin, X., X. Zhang, and T. Zhou, 1999: Fundamental framework and experiments of the third generation of IAP/LASG world ocean general circulation model. *Adv. Atmos. Sci.*, **16**, 197–215, <https://doi.org/10.1007/BF02973082>.
- Kanamitsu, M., W. Ebisuzaki, J. Woollen, S. K. Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter, 2002: NCEP–DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc.*, **83**, 1631–1643, <https://doi.org/10.1175/BAMS-83-11-1631>.
- Kara, A. B., P. A. Rochford, and H. E. Hurlburt, 2000: Mixed layer depth variability and barrier layer formation over the North Pacific Ocean. *J. Geophys. Res.*, **105**, 16 783–16 801, <https://doi.org/10.1029/2000JC900071>.
- Lanzante, J. R., 1984: A rotated eigenanalysis of the correlation between 700 mb heights and sea surface temperatures in the Pacific and Atlantic. *Mon. Wea. Rev.*, **112**, 2270–2280, [https://doi.org/10.1175/1520-0493\(1984\)112<2270:AREOTC>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<2270:AREOTC>2.0.CO;2).
- Latif, M., and T. P. Barnett, 1994: Causes of decadal climate variability over the North Pacific and North America. *Science*, **266**, 634–637, <https://doi.org/10.1126/science.266.5185.634>.
- Lau, N. C., and M. J. Nath, 1996: The role of the “atmospheric bridge” in linking tropical Pacific ENSO events to extratropical SST anomalies. *J. Climate*, **9**, 2036–2057, [https://doi.org/10.1175/1520-0442\(1996\)009<2036:TROBTBI>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<2036:TROBTBI>2.0.CO;2).
- Levitus, S., J. I. Antonov, J. Wang, T. L. Delworth, K. W. Dixon, and A. J. Broccoli, 2001: Anthropogenic warming of Earth’s climate system. *Science*, **292**, 267–270, <https://doi.org/10.1126/science.1058154>.

- Li, Y., and J. P. Li, 2012: Propagation of planetary waves in the horizontal non-uniform basic flow (in Chinese). *Chin. J. Geophys.*, **55**, 361–371.
- , —, F. F. Jin, and S. Zhao, 2015: Interhemispheric propagation of stationary Rossby waves in a horizontally nonuniform background flow. *J. Atmos. Sci.*, **72**, 3233–3256, <https://doi.org/10.1175/JAS-D-14-0239.1>.
- Lin, P., H. Liu, and X. Zhang, 2007: Sensitivity of the upper ocean temperature and circulation in the equatorial Pacific to solar radiation penetration due to phytoplankton. *Adv. Atmos. Sci.*, **24**, 765–780, <https://doi.org/10.1007/s00376-007-0765-7>.
- Liu, G., P. Zhao, and J. M. Chen, 2017: Possible effect of the thermal condition of the Tibetan Plateau on the interannual variability of the summer Asian–Pacific oscillation. *J. Climate*, **30**, 9965–9977, <https://doi.org/10.1175/JCLI-D-17-0079.1>.
- Lu, R., 2001: Interannual variability of the summertime North Pacific subtropical high and its relation to atmospheric convection over the warm pool. *J. Oceanogr. Soc. Japan Ser. II*, **79**, 771–783, <https://doi.org/10.2151/jmsj.79.771>.
- Luo, H., and M. Yanai, 1984: The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979. Part II: Heat and moisture budgets. *Mon. Wea. Rev.*, **112**, 966–989, [https://doi.org/10.1175/1520-0493\(1984\)112<0966:TLSCAH>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<0966:TLSCAH>2.0.CO;2).
- Montégut, C. B., G. Madec, A. S. Fischer, A. Lazar, and D. Iudicone, 2004: Mixed layer depth over the global ocean: An examination of profile data and a profile-based climatology. *J. Geophys. Res. Oceans*, **109**, C12003, <https://doi.org/10.1029/2004JC002378>.
- Na, H., K.-Y. Kim, S. Minobe, and Y. N. Sasaki, 2018: Interannual to decadal variability of the upper-ocean heat content in the western North Pacific and its relationship to oceanic and atmospheric variability. *J. Climate*, **31**, 5107–5125, <https://doi.org/10.1175/JCLI-D-17-0506.1>.
- Nan, S., P. Zhao, S. Yang, and J. Chen, 2009: Springtime tropospheric temperature over the Tibetan Plateau and evolutions of the tropical Pacific SST. *J. Geophys. Res.*, **114**, D10104, <https://doi.org/10.1029/2008jd011559>.
- Neale, R. B., and Coauthors, 2010: Description of the NCAR Community Atmosphere Model (CAM5.0). NCAR Tech. Note NCAR/TN-486+STR, 268 pp., www.cesm.ucar.edu/models/cesm1.1/cam/docs/description/cam5_desc.pdf.
- Oleson, K. W., and Coauthors, 2010: Technical description of version 4.0 of the Community Land Model (CLM). NCAR Tech. Note NCAR/TN-478+STR, 257 pp., <https://doi.org/10.5065/D6FB50WZ>.
- Ose, T., 1996: The comparison of the simulated response to the regional snow mass anomalies over Tibet, Eastern Europe, and Siberia. *J. Meteor. Soc. Japan*, **74**, 845–866, https://doi.org/10.2151/jmsj1965.74.6_845.
- Paulson, C. A., and J. J. Simpson, 1977: Irradiance measurements in the upper ocean. *J. Phys. Oceanogr.*, **7**, 952–956, [https://doi.org/10.1175/1520-0485\(1977\)007<0952:IMITUO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1977)007<0952:IMITUO>2.0.CO;2).
- Pickett, M. H., and J. D. Paduan, 2003: Ekman transport and pumping in the California Current based on the U.S. Navy's high-resolution atmospheric model (COAMPS). *J. Geophys. Res.*, **108**, 3327, <https://doi.org/10.1029/2003JC001902>.
- Price, J. F., R. A. Weller, and R. R. Schudlich, 1987: Wind-driven ocean currents and Ekman transport. *Science*, **238**, 1534–1538, <https://doi.org/10.1126/science.238.4833.1534>.
- Rajagopalan, B., and P. Molnar, 2013: Signatures of Tibetan Plateau heating on Indian summer monsoon rainfall variability. *J. Geophys. Res. Atmos.*, **118**, 1170–1178, <https://doi.org/10.1002/jgrd.50124>.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, <https://doi.org/10.1029/2002JD002670>.
- Smagorinsky, J., 1953: The dynamical influence of large-scale heat sources and sinks on the quasi-stationary mean motions of the atmosphere. *Quart. J. Roy. Meteor. Soc.*, **79**, 342–366, <https://doi.org/10.1002/qj.49707934103>.
- Smith, R., and Coauthors, 2010: The Parallel Ocean Program (POP) reference manual. Los Alamos National Lab Tech. Rep. LAUR-10-01853, 141 pp., <http://www.cesm.ucar.edu/models/cesm1.0/pop2/doc/sci/POPRefManual.pdf>.
- Sterl, A., and A. Kattenberg, 1994: Embedding a mixed layer model into an ocean general circulation model of the Atlantic: The importance of surface mixing for heat flux and temperature. *J. Geophys. Res.*, **99**, 14 139–14 157, <https://doi.org/10.1029/94JC00813>.
- Takaya, K., and H. Nakamura, 2001: A formulation of a phase-independent wave-activity flux for stationary and migratory quasigeostrophic eddies on a zonally varying basic flow. *J. Atmos. Sci.*, **58**, 608–627, [https://doi.org/10.1175/1520-0469\(2001\)058<0608:AFOAPI>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<0608:AFOAPI>2.0.CO;2).
- Tang, Y., L. Li, W. Dong, and B. Wang, 2016: Tracing the source of ENSO simulation differences to the atmospheric component of two CGCMs. *Atmos. Sci. Lett.*, **17**, 155–161, <https://doi.org/10.1002/asl.637>.
- Taschetto, A. S., A. S. Gupta, N. C. Jourdain, A. Santoso, C. C. Ummenhofer, and M. H. England, 2014: Cold tongue and warm pool ENSO events in CMIP5: Mean state and future projections. *J. Climate*, **27**, 2861–2885, <https://doi.org/10.1175/JCLI-D-13-00437.1>.
- Wallace, J., and Q. R. Jiang, 1987: On the observed structure of the interannual variability of the atmosphere/ocean climate system. *Atmospheric and Oceanic Variability*, H. Cattle, Ed., Royal Meteorological Society, 17–43.
- Wang, L. Y., X. Q. Yang, D. J. Yang, Q. Xie, J. B. Fang, and X. G. Sun, 2017: Two typical modes in the variabilities of wintertime North Pacific basin-scale oceanic fronts and associated atmospheric eddy-driven jet. *Atmos. Sci. Lett.*, **18**, 373–380, <https://doi.org/10.1002/asl.766>.
- , H. B. Hu, and X. Q. Yang, 2019: The atmospheric responses to the intensity variability of subtropical front in the wintertime North Pacific. *Climate Dyn.*, **52**, 5623–5639, <https://doi.org/10.1007/s00382-018-4468-9>.
- Wang, Z., A. Duan, and G. Wu, 2014: Time-lagged impact of spring sensible heat over the Tibetan Plateau on the summer rainfall anomaly in East China: Case studies using the WRF model. *Climate Dyn.*, **42**, 2885–2898, <https://doi.org/10.1007/s00382-013-1800-2>.
- Weare, B. C., A. R. Navato, and R. E. Newell, 1976: Empirical orthogonal analysis of Pacific sea surface temperatures. *J. Phys. Oceanogr.*, **6**, 671–678, [https://doi.org/10.1175/1520-0485\(1976\)006<0671:EOAOPS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1976)006<0671:EOAOPS>2.0.CO;2).
- Wu, G. X., W. P. Li, and H. Liu, 1997: Sensible heating-driving air pump of the Tibetan Plateau and the Asian summer monsoon. *Memorial Volume of Prof. Zhao JZ*, D. Z. Ye, Ed., Science Press, 116–126.
- , Y. Liu, B. Dong, X. Liang, A. Duan, Q. Bao, and J. Yu, 2012: Revisiting Asian monsoon formation and change associated

- with Tibetan Plateau forcing: I. Formation. *Climate Dyn.*, **39**, 1169–1181, <https://doi.org/10.1007/s00382-012-1334-z>.
- , B. He, Y. M. Liu, Q. Bao, and R. C. Ren, 2015: Location and variation of the summertime upper-troposphere temperature maximum over South Asia. *Climate Dyn.*, **45**, 2757–2774, <https://doi.org/10.1007/s00382-015-2506-4>.
- Wu, Z., and J. Li, 2012: Modulation of the Tibetan Plateau snow cover on the ENSO teleconnections: From the East Asian summer monsoon perspective. *J. Climate*, **25**, 2481–2488, <https://doi.org/10.1175/JCLI-D-11-00135.1>.
- Yanai, M., C. Li, and Z. Song, 1992: Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. *J. Meteor. Soc. Japan Ser. II*, **70**, 319–351, <https://doi.org/10.2151/jmsj1965.70.1b.319>.
- Ye, D., and Y. Gao, 1979: *Meteorology of the Qinghai-Xizang Plateau*. Science Press, 278 pp.
- Yi, D. L., L. Zhang, and L. Wu, 2015: On the mechanisms of decadal variability of the North Pacific Gyre Oscillation over the 20th century. *J. Geophys. Res. Oceans*, **120**, 6114–6129, <https://doi.org/10.1002/2014JC010660>.
- Zhang, Y. S., and G. X. Wu, 1999: Diagnostic investigations on the mechanism of the onset of Asian summer monsoon and abrupt seasonal transitions over the Northern Hemisphere. Part II: The role of surface sensible heating over Tibetan Plateau and surrounding regions. *Acta Meteor. Sin.*, **1**, 56–73.
- Zhao, P., and L. X. Chen, 2001: Role of atmospheric heat source/sink over the Qinghai-Xizang Plateau in the quasi-4-years oscillation of atmosphere-land-ocean interaction. *Chin. Sci. Bull.*, **46**, 241–245, <https://doi.org/10.1007/BF03187178>.
- , Z. J. Zhou, and J. P. Liu, 2007: 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, <https://doi.org/10.1175/JCLI4205.1>.
- , X. D. Zhang, Y. F. Li, and J. M. Chen, 2009: Remotely modulated tropical-North Pacific ocean-atmosphere interactions by the South Asian high. *Atmos. Res.*, **94**, 45–60, <https://doi.org/10.1016/j.atmosres.2009.01.018>.
- , S. Yang, M. Q. Jian, and J. M. Chen, 2011: Relative controls of Asian-Pacific summer climate by Asian land and tropical-North Pacific sea surface temperature. *J. Climate*, **24**, 4165–4188, <https://doi.org/10.1175/2011JCLI3915.1>.
- , and Coauthors, 2018: The third atmospheric scientific experiment for understanding the Earth-atmosphere coupled system over the Tibetan Plateau and its effects. *Bull. Amer. Meteor. Soc.*, **99**, 757–776, <https://doi.org/10.1175/BAMS-D-16-0050.1>.
- Zhao, S., J. Li, and Y. Li, 2015: Dynamics of an interhemispheric teleconnection across the critical latitude through a southerly duct during boreal winter. *J. Climate*, **28**, 7437–7456, <https://doi.org/10.1175/JCLI-D-14-00425.1>.
- , —, —, F.-F. Jin, and J. Zheng, 2019: Interhemispheric influence of Indo-Pacific convection oscillation on Southern Hemisphere rainfall through southward propagation of Rossby waves. *Climate Dyn.*, **52**, 3203–3221, <https://doi.org/10.1007/s00382-018-4324-y>.
- Zhou, X. J., P. Zhao, J. M. Chen, L. X. Chen, and W. L. Li, 2009: Impacts of thermodynamic processes over the Tibetan Plateau on the Northern Hemispheric climate. *Sci. China.*, **52D**, 1679–1693, <https://doi.org/10.1007/s11430-009-0194-9>.