Role of Latent Heating over the Tropical Western Pacific in Surface Temperature Change over North America during Boreal Spring

BIAN HE

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

SONG YANG

School of Atmospheric Sciences, and Guangdong Province Key Laboratory for Climate Change and Natural Disaster Studies, and Institute of Earth Climate and Environment System, Sun Yat-sen University, Guangzhou, China

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ABSTRACT

Based on observational data, a linear baroclinic model, and an atmospheric general circulation model (AGCM), the major modes of spring precipitation over the tropical Asian and Pacific regions are identified, and the influence of latent heating over the tropical western Pacific (TWP) on global climate is investigated. Results show that the first mode of empirical orthogonal function analysis explains 20% of the total variance in March, the largest in spring, with the maximum center located over the TWP. The precipitation is highly positively correlated with local sea surface temperature (SST) in March, which suggests that the warming SST is the trigger for the precipitation over the TWP. Further analysis suggests that an increase in latent heating over the TWP, especially in March, can produce Rossby waves along the westerly jet, which causes an increase in surface temperature over North America. The propagation intensity decreases from March to May. The changes in location and strength of the westerly jet stream in the Northern Hemisphere are responsible for this decrease. Experiments with both a linear baroclinic model and an AGCM verify the above hypothesis. The study highlights that the spatial distributions of latent heating and westerly jet stream are the two key factors for the formation of teleconnection patterns from eastern Asia to North America.

1. Introduction

Latent heating is an important forcing on atmospheric circulation as it adjusts air temperature and moisture content (Webster 1972; Gill 1980; Hoskins and Karoly 1981; Yang and Webster 1990; Ting 1996; Zhang et al. 2006). In the subtropics, the vertically inhomogeneous latent heating triggers northerly wind above the heating center and produces negative vorticity in the upper

troposphere, because of the so-called Sverdrup balance (Liu et al. 1999; Wu and Liu 2003). This relationship is often used to explain the influence of East Asian precipitation on the formation and interannual variation of the South Asian high and the western Pacific high in summer (Liu et al. 2001; Li et al. 2012; Zhang et al. 2016). Wu et al. (2015) revealed that when the latent heating (usually at 200–400 hPa) over eastern Asia coincided with the subtropical high, a secondary zonal circulation anomaly would be triggered, shifting the upper-tropospheric temperature maximum (at 200–400 hPa) to the west of the heating maximum. The authors emphasized that the response of atmospheric circulation was sensitive to the location of the latent heating. In response to global warming, precipitation has shown a significant increasing trend over the South China Sea (SCS) and the Philippine Sea during spring and summer (Alexander et al. 2013). He et al. (2016) revealed that the increased summertime heating caused an east–west feedback between the tropical western

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Corresponding author: Prof. Song Yang, yangsong3@mail.sysu.edu.cn

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Pacific (TWP) and the Indian Ocean (IO), which contributed to the weakening of the South Asian summer monsoon and the warming of sea surface temperature (SST) in the IO.

Besides local and regional influences, latent heating also acts as a Rossby wave source and affects global climate change. Extensive studies (Ferranti et al. 1994; Graham et al. 1994; Kumar et al. 1994; Lau and Nath 1994; Kharin 1995) have been carried out during the Tropical Ocean and Global Atmosphere (TOGA) program to investigate the influences of tropical SST and associated condensational heating on extratropical atmospheric circulation. As summarized in Trenberth et al. (1998), the distribution of deep convection varies along with the associated changes in heating, low-level convergence, and upper-level divergence, thereby altering the generation of atmospheric vorticity and forcing large-scale atmospheric Rossby waves that could propagate into higher latitudes. There exist several well-known teleconnection patterns in the observations, which are related to tropical heating. Bjerknes (1969) found that the deepening of the Aleutian low pressure system in the North Pacific in winter was associated with El Niño events, and this feature was later recognized as the Pacific–North American (PNA) teleconnection pattern. The Aleutian and Icelandic lows are also a manifestation of atmospheric teleconnection during winter months, and are related to the anomalies of surface temperature and sea level pressure (Honda and Nakamura 2001; Honda et al. 2001; Shi and Nakamura 2014). The tropical–Northern Hemisphere pattern (Mo and Livezey 1986; Barnston et al. 1991) is often observed during wintertime, and its positive phase is associated with above-average precipitation over the central and eastern subtropical North Pacific but below-average precipitation over the North American continent.

For summer, Nitta (1986, 1987) discovered a meridional dipole pattern of cloudiness between the tropical and midlatitude northwestern Pacific, and named this teleconnection the Pacific–Japan (PJ) pattern, which is also referred to as the East Asia–Pacific pattern proposed in Huang and Sun (1992). The PJ pattern appears to be generated by strong convection over the Philippine Sea area associated with warmer local SSTs, which often occur during La Niña (Gambo and Kudo 1983). Similarly, Lau (1992) found a wave train of four centers stretching from Japan along 45°N to North America in summer, and named it the Asia–North America pattern. This pattern is closely related to the latent heating over the western Pacific near the Philippines and the IO.

The Asian rainfall regime is significantly distinguished from other monsoon regimes for its stepwise evolution within the rainy season (Wu and Wang 2001; LinHo

et al. 2008). From spring to summer, rainfall shifts from tropical oceans and the Maritime Continent (MC) to the subtropical Asian mainland as a result of the complex air–sea interaction associated with the changes in SST and surface heat flux (Wu et al. 2008; Hu et al. 2014; Wu and Hu 2015), affecting billions of people in Asia. Compared to the extensive studies of the influence of latent heating in Asian monsoon regions during the boreal summer, or the heating-induced teleconnections in winter associated with El Niño events, our understanding of the connections between latent heating over the Asia–Pacific monsoon region and global climate during boreal spring is still limited. Observations indicate that rainfall has an increasing trend over the SCS, MC, and TWP regions and a decrease in the IO region during spring–summer, associated with an increase in SST in situ (Alexander et al. 2013). Feng et al. (2013) claimed that the warming trend of SST over the Indo-Pacific warm pool played an essential role in the variation of the Hadley circulation during the boreal spring. Li et al. (2016) found that the increase in heating over the SCS and TWP dried southern China through forcing an anomaly of the meridional circulation during the boreal spring. However, the responses of global climate to latent heating in the Asia–Pacific region during the boreal spring have not been fully addressed. In this study, we seek answers to two scientific questions: 1) What are the major features of spring precipitation over the tropical Asia–Pacific region during recent decades? 2) How do the changes in heating influence global climate? It is necessary to mention that the location and magnitude of subtropical Rossby waves are sensitive to changes in upper-tropospheric vorticity gradient, which is controlled by tropical heating, but the physical mechanism for the feature is still unclear (Trenberth et al. 1998). Therefore, this study is also aimed at providing useful information for understanding the related tropical–extratropical interaction.

The remainder of the paper is structured as follows. In section 2, we introduce datasets, methods, and model configuration. In section 3, we diagnose the changes in observed precipitation over the tropical Asia–Pacific region during the boreal spring, in which the major mode of precipitation in each month is also analyzed. In section 4, we provide a statistical analysis of the relationship between major precipitation mode and the change in global climate. The key regions and physical processes are revealed. In section 5, we present the results from numerical experiments using both a linear baroclinic model (LBM) and an atmospheric general circulation model (AGCM) to reveal the impact of heating on global climate. Finally, discussion and summary of our key findings are provided in section 6.

2. Datasets and models

a. Datasets

The precipitation dataset used is the Global Precipitation Climatology Project (GPCP) monthly dataset (Adler et al. 2003; Huffman et al. 2009). The GPCP precipitation has a resolution of $2.5^\circ \times 2.5^\circ$ and is available from January 1979 to the present. This dataset is provided by NOAA at its website (<https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html>). The monthly mean global SST dataset used is the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST) (Rayner et al. 2003), which includes SST with a resolution of $1^\circ \times 1^\circ$. This dataset is available online (<http://www.metoffice.gov.uk/hadobs/hadisst/data/download.html>). The National Centers for Environmental Prediction (NCEP)–U.S. Department of Energy (DOE) AMIP-II reanalysis (Kanamitsu et al. 2002) is also used. It includes multilevel air temperature, specific humidity, and winds with a resolution of $2.5^\circ \times 2.5^\circ$. This dataset, covering the period from 1979 to the present, is available online (<https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html>).

b. LBM

The LBM is based on primitive equations (Watanabe and Kimoto 2000; Watanabe and Jin 2003) is used to study the propagation of the Rossby wave triggered by latent heating. The horizontal resolution of the model uses spectral triangular truncation at wavenumber 42 (T42). The model has 20 vertical levels in a sigma coordinate system ($\sigma = p/p_s$, where p and p_s are pressure and surface pressure, respectively). In addition to vertical diffusion, biharmonic horizontal diffusion having a time scale of 6 h is also employed. Rayleigh friction and Newtonian damping are applied using time scales of $(0.5 \text{ day})^{-1}$ for $\sigma \geq 0.9$, $(1 \text{ day})^{-1}$ for $\sigma \leq 0.3$, and $(20 \text{ day})^{-1}$ for all other sigma levels. The LBM has been used to investigate Rossby wave propagation under different forcing in many studies (Annamalai and Sperber 2005; Ham et al. 2007; Cui et al. 2015; Zhang et al. 2016).

c. AGCM

The Spectral Atmosphere Model of IAP LASG, version 2 (SAMIL2), is used in this study, which was developed at the State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics of the Institute of Atmospheric Physics (Wu et al. 1996; Bao et al. 2013). The SAMIL2 has a spectral rhomboidal truncation at wavenumber 42 (R42) horizontal resolution (2.81° longitude $\times$ 1.66° latitude) and 26 vertical layers in a σ – p hybrid coordinate, extending from the surface to 2.19 hPa. The mass flux cumulus parameterization of Tiedtke (1989) is used to calculate convective precipitation. The cloud

scheme is a diagnostic method (Slingo 1987; Kiehl et al. 1996), combined with a low-cloud statistical method (Dai et al. 2004). A nonlocal scheme is employed to calculate the eddy-diffusivity profile and turbulent velocity scale, and the model incorporates nonlocal transport effects for heat and moisture (Holtslag and Boville 1993). The radiation scheme employed is an updated Edwards–Slingo scheme (Edwards and Slingo 1996; Sun and Rikus 1999).

3. Major modes of precipitation over the Asia–Pacific region during boreal spring

The region 20°S – 30°N , 40° – 160°E analyzed here mainly covers the IO, SCS, the western Pacific, the MC, and southern Asia. Because of the consistency in spatial and temporal variations, precipitation is often used to infer condensational heating by simply multiplying the latent heating coefficient (Li et al. 2012). For simplicity, we depict the major modes of observed precipitation, instead of the condensational heating, which is difficult to observe directly in the real atmosphere. An empirical orthogonal function (EOF) analysis is applied on monthly GPCP data during March–May from 1979 to 2014 to identify the major modes. The first two modes of spatial pattern are shown in Fig. 1. In March, the first EOF mode (EOF1) accounts for 20% of the total variance (Fig. 1a), whereas the second EOF mode (EOF2) only accounts for 9.5% of the total variance (Fig. 1d) (The associated time series are shown in Fig. S5 of the supplemental material.) The EOF1 pattern shows a positive phase over the MC and TWP, with a maximum over the southern Philippine Sea (Fig. 1a). The EOF2 pattern shows two obvious negative centers over the tropical IO and TWP (Fig. 1d). In April (Figs. 1b,e), the spatial pattern of EOF1 accounts of 15.2% of the total variance (Fig. 1b) and is quite similar to that in March (Fig. 1a), except that the intensity in the Philippine Sea is weaker. The EOF2 (Fig. 1e) shows a negative phase over the MC in the Southern Hemisphere and a warm phase in the TWP. In May, the EOF1 pattern (Fig. 1c) is significantly different from the patterns in March and April, because of the evolution of the Asian summer monsoon. The positive phase covers the Bay of Bengal (BOB), the SCS, and the Philippine Sea; the negative phase covers the entire southern IO. EOF2 (Fig. 1f) shows an overall negative phase in southern Asia, the IO, and the tropics, which accounts for 9.2% of the total variance. As a whole, the precipitation pattern of the first EOF mode is similar to the climatological mean pattern (figures not shown), which represents the major variability of precipitation over the tropical Asia–Pacific region.

The time series corresponding to the EOF1 spatial pattern are shown in Fig. 2 (black lines). To verify whether the EOF1 time series [first principal component

FIG. 1. Spatial patterns of the EOF1 of precipitation in (a) March, (b) April, and (c) May from GPCP during 1979–2014. (d)–(f) As in (a)–(c), but for EOF2. The explained variance is shown on the top right of each panel. The black rectangles denote the region of 15°S–10°N, 120°–155°E in (a),(b) and the region of 5°–20°N, 90°–150°E in (c).

(PC1)] captures the interannual variability of regional precipitation, we show the precipitation averaged over 15°S–10°N, 120°–155°E (see the black rectangle in Fig. 1a) using the blue lines in Fig. 2 for comparison with the PC1. It is clear that the positive phases in Figs. 1a–c capture the major mode of spring precipitation, with their PC1s highly correlated with the regional mean values of precipitation (0.95 for March, 0.89 for April, and 0.95 for May). However, the relationships between precipitation and local SST are different among the spring months. The correlation of local SST and precipitation is 0.58 in March (Fig. 2a), meaning that the convection over the TWP and MC mainly originates from local SST forcing in situ. The correlation coefficient decreases to 0.33 in April (Fig. 2b), but the value is still significant, passing the 95% confidence level (0.325). This suggests that local SST forcing still contributes to the tropical precipitation in April but the relationship is weaker. As a result, the EOF1 in April (Fig. 1b) is weaker than that in March (Fig. 1a). However, precipitation in May (Fig. 2c) shows almost no relationship with the SST over the BOB, the SCS, and the Philippine Sea ($R = -0.04$). The above results indicate that the trigger of the convection over the MC and TWP in March and April is similar as a result of the same origination from local SST anomalies, while the formation of precipitation in May is different because of the

monsoon onset. In spite of different mechanisms, precipitation PC1 all show strong interannual variation during 1979–2014, while their long-term linear trends are weak. Therefore, in the next section we mainly investigate the interannual relationship between major precipitation modes and global climate change.

4. Influences of the precipitation major modes on global climate change during boreal spring

Teleconnection patterns link the variability of extratropical large-scale circulation with tropical heating (Wallace and Gutzler 1981; Simmons 1982; Simmons et al. 1983; Kasahara and da Silva Dias 1986). Figure 3 shows the correlation coefficients between the PC1s (Fig. 2) and the geopotential height fields in the upper, middle, and lower troposphere, respectively. In the upper troposphere (i.e., 200 hPa), significant negative values appear over the tropical IO and the tropical Pacific Ocean in all spring months, indicating that an increase in heating lowers the pressure in the tropics associated with the change in the Walker circulation. In the extratropics, two Rossby wave trains occur in both hemispheres in March, and the feature is more apparent in the Northern Hemisphere, which is characterized by three positive and two negative geopotential height anomalies across the Asian and North American continents (Fig. 3a). However, the correlation (Figs. 3d,g) becomes

FIG. 2. (a) Standardized time series of precipitation EOF1 (black line) and the regional mean (15°S – 10°N , 120° – 155°E ; see the black rectangle in Fig. 1a) precipitation (blue line) and observed SST (red line) in March. (b) As in (a), but for April. (c) Standardized time series of precipitation EOF1 (black line) and the regional mean (5° – 20°N , 90° – 150°E ; see the black rectangle in Fig. 1c) precipitation (blue line) and observed SST (red line) in May. Pearson's correlation coefficient between the time series of EOF1 and precipitation/SST is shown in the legend of each panel.

weaker in April and May, especially in May. We also examined the same correlation patterns using both the ERA-Interim and the JRA-55 datasets, and the results are similar to those using the NCEP–NCAR reanalysis (figures not shown), indicating that the existence of the Rossby wave train structure in March is robust. In the middle troposphere (i.e., 500 hPa), the correlation patterns in the three months (Figs. 3b,e,h) are similar to those at 200 hPa, except that the correlations are weaker in the extratropical regions.

However, in the lower troposphere (i.e., 850 hPa), the correlation patterns are quite different from those at 200 and 500 hPa. The negative correlation over the Asia–Pacific region is the dominant signal in all months (Figs. 3c,f,i), and the signal in March is the strongest.

The above analysis reveals that the relationship between precipitation major mode and global climate change is more solid in March than in April and May. To further understand the responses of atmospheric circulation and temperature to precipitation, especially in March, we show the regressions of geopotential height, sea level pressure, and air temperature to the precipitation PC1 shown in Fig. 4. The regression pattern for 200-hPa geopotential height (Fig. 4a) is accompanied by well-built anticyclonic–cyclonic circulation anomalies and is similar to the correlation pattern at the same level (Fig. 3a). The extratropical response of geopotential height is more apparent than the tropical response at 500 hPa (Fig. 4b), which is in phase with that at 200 hPa, and is consistent with the regressions of sea level pressure (Fig. 4c). In the subtropics, the large-scale atmospheric motion follows the principles of quasigeostrophic balance, hydrostatic balance, and thermal wind balance. Thus, the response of air temperature to latent heating is often in phase with the response of upper-level circulation (Wu et al. 2015). We show the regressions of air temperature at 300, 500, and 925 hPa in Figs. 4d–f, respectively. The warming over southern Asia and the North Pacific is strong at 300 (Fig. 4d) and 500 hPa (Fig. 4e), while the cooling over Alaska and warming over North America are especially strong at 925 hPa (Fig. 4f). It is necessary to note that the spatial distributions of temperature and geopotential height regressions in the upper troposphere are highly consistent with those in the lower troposphere, known as equivalent barotropic structure. This feature is helpful for us to infer that the variation of surface temperature over the extratropical regions is due to tropical heating. In brief, the above analysis suggests that the intensified precipitation in March causes an equivalent barotropic Rossby wave train from southern Asia to North America in the upper troposphere, and thus leads to surface warming over the North American continent.

To further understand the effects of TWP heating on North American surface climate and the associated physical mechanism, we show the composite results based on the precipitation PC1 in March in Fig. 5. The years when the PC1 is above (below) its 0.5 (–0.5) standard deviation are recognized as the years of strong (weak) TWP precipitation (Table 1). The SST differences between the strong and weak years are shown in Fig. 5a. It is clear that the SST anomalies present a La Niña pattern, with strong negative anomalies over the central and eastern tropical Pacific and positive anomalies over the TWP. Negative anomalies also

FIG. 3. Spatial patterns of Pearson's correlation coefficient between the time series of precipitation EOF1 and geopotential height for (a) 200, (b) 500, and (c) 850 hPa in March. (d)–(f) As in (a)–(c), but for April. (g)–(i) As in (a)–(c), but for May. Stippling denotes the value significantly exceeding the 95% confidence level of the Student's t test.

exist over the tropical IO, but the magnitude is smaller. The composite 2-m air temperature (t_{2m}) over land is shown in Fig. 5b. Significant warming mainly appears over the North American continent, while cooling mainly occurs in northern China, western areas of Australia, and northern South America. Meanwhile, the composite 200-hPa geopotential height (Fig. 5c) and 500-hPa temperature (Fig. 5d) present similar patterns, as shown in correlation (Fig. 3a) and regression (Fig. 4a). We also checked the relationship between TWP precipitation and global climate change when the El Niño–Southern Oscillation (ENSO) signal is removed (section 1 in the supplemental material), and the results are consistent; specifically, the Rossby wave features are still clear in the midlatitude. To understand wave propagation qualitatively, we analyze the stationary wave-activity flux $\mathbf{W}$ at 200 hPa and the associated streamfunction anomaly. The calculation of $\mathbf{W}$ follows Takaya and Nakamura (1997, 2001):

$$\mathbf{W} = \frac{p}{2|\mathbf{U}|} \left\{ \begin{array}{l} U(\psi_x'^2 - \psi'_x \psi'_{xx}) + V(\psi'_x \psi'_y - \psi'_y \psi'_{xy}) \\ U(\psi'_x \psi'_y - \psi'_y \psi'_{xy}) + V(\psi_y'^2 - \psi'_y \psi'_{yy}) \\ \frac{f_0^2}{N^2} [U(\psi'_x \psi'_z - \psi'_z \psi'_{xz}) + V(\psi'_y \psi'_z - \psi'_z \psi'_{yz})] \end{array} \right\}. \quad (1)$$

In our calculation, $p = 200/100$; U and V are the climatological means of 200-hPa zonal and meridional winds,

respectively. Also, $|\mathbf{U}|$ is the mode of (U, V) , ψ' is the perturbation streamfunction derived from the differences between strong and weak years, and subscripts x , y , and z denote the ψ' gradient in each direction. Here, only the horizontal propagation component is considered. The distributions of ψ' and $\mathbf{W}$ in March are shown in Fig. 6. There are three obvious positive ψ' anomalies in the Northern Hemisphere, located over southern Asia, the North Pacific, and North America, respectively. The first two centers are associated with the northeastward energy dispersion of the Rossby wave energy ($\mathbf{W}$ vectors). The strongest one is located over the North Pacific. Meanwhile, an obvious negative ψ' anomaly is located over the central-eastern tropical Pacific, associated with southward energy dispersion near 150°W . The pattern of $\mathbf{W}$ indicates that the Rossby wave induced by the TWP heating mainly has one energy dispersion path in the upper troposphere, from southern Asia to North America along the westerly jet waveguide. Thus, composite, regression, and correlation analyses all indicate that the TWP heating is an important Rossby wave source, which can influence the North America climate in the boreal spring, especially in March. Consequently, a question emerges: Why is the teleconnection pattern or the Rossby wave train in the extratropical regions clear only in March, and not in the other spring months of April and May?

Previous theoretical studies (Hoskins et al. 1977; Hoskins and Karoly 1981; Webster 1981, 1982; Hoskins

FIG. 4. Regression patterns of the standardized time series of precipitation EOF1 in March on geopotential height at (a) 200 and (b) 500 hPa. (c) As in (a), but for sea level pressure and 1000 hPa for wind. Regression patterns of the standardized time series of precipitation EOF1 in March on air temperature and wind field at (d) 300, (e) 500, and (f) 925 hPa. Stippling denotes the value significantly exceeding the 95% confidence level of the Student's t test.

and Ambrizzi 1993; Ting and Sardeshmukh 1993) have revealed that the extratropical response to tropical heating can be understood in terms of Rossby wave propagation derived from the vorticity equation. The linearized vorticity equation about a basic state consisted of a zonal mean flow on the β plane can be written as (Trenberth et al. 1998)

$$\left(\frac{\partial}{\partial t} + [\bar{u}] \frac{\partial}{\partial x} \right) \nabla^2 \psi' + \frac{\partial \psi'}{\partial x} \left(\beta - \frac{\partial^2 [\bar{u}]}{\partial y^2} \right) = Q, \quad (2)$$

where $[\bar{u}]$ denotes the climatological zonal mean flow, ψ' is the perturbation streamfunction, and Q is the vorticity forcing (i.e., the heating anomaly). A Wentzel–Kramers–Brillouin (WKB) solution to (2) exists in the form of $\psi' = A \exp[i(kx + ly - \sigma t)]$, with a dispersion relation

$$\sigma = [\bar{u}]k - \frac{\beta_* k}{K^2}, \quad (3)$$

FIG. 5. Differences in (a) SST (K), (b) t2m (K), (c) 200-hPa geopotential height (gpm), and (d) 500-hPa air temperature (K) in March between strong and weak years during 1979–2014. Stippling denotes values significantly exceeding the 95% confidence level.

TABLE 1. Years based on the criterion of 0.5 standard deviation of the standardized PC1 of precipitation in March (Fig. 2a) from 1979 to 2014.

Strong years	1982, 1986, 1988, 1989, 1999, 2000, 2001, 2004, 2006, 2008, 2011, and 2012
Weak years	1981, 1983, 1987, 1991, 1992, 1993, 1997, 1998, 2002, 2010, and 2014

where

$$\beta_* = \beta - \frac{\partial^2 [\bar{u}]}{\partial y^2}. \quad (4)$$

Equation (4) is the meridional gradient of absolute vorticity, and $K = (k^2 + l^2)^{1/2}$ is the total wavenumber. For stationary Rossby waves ($\sigma = 0$),

$$K = \left(\frac{\beta_*}{[\bar{u}]} \right)^{1/2}. \quad (5)$$

Thus, Rossby wave propagation is largely affected by the zonal mean flow, the meridional gradient of vorticity, and the β effect. To investigate the causes of teleconnection pattern differences among March, April, and May, we show the climatological mean 200-hPa zonal wind, divergence, and vorticity, respectively, in Fig. 7. The strong divergence (blue contour) in the upper troposphere is consistent with the strong precipitation over the tropics. The distributions of divergence are similar in March (Fig. 7a), April (Fig. 7b), and May (Fig. 7c). However, the intensity of the westerly jet stream, especially that over eastern Asia, is different among the boreal spring months. The value of $[\bar{u}]$ exceeds 50 m s^{-1} over southern Japan in March, associated with strong vorticity (red dashed contour) to the south of the westerly jet stream where the meridional gradient of $[\bar{u}]$ is strong (Fig. 7a). In April (Fig. 7b), $[\bar{u}]$ weakens, especially over Japan, and vorticity also weakens accordingly. In May (Fig. 7c), the westerly jet stream breaks down over the eastern subtropical Pacific and vorticity forcing is not apparent. We further calculate stationary wavenumber K according to (5) for the three spring months. Figure 8 clearly shows that Rossby wave cannot propagate in the easterlies over the tropical IO and the MC where K is missing in all three months. The major differences of wavenumber K among these months exist in

FIG. 6. Composite streamfunction anomaly (shading; $10^6 \text{ m}^2 \text{ s}^{-1}$) and $\mathbf{W}$ (vector; $\text{m}^2 \text{ s}^{-2}$) at 200 hPa in March.

the subtropical Pacific. The value of K is about 2–8 over the subtropical regions from eastern Asia to North America in March (Fig. 8a). It is similar in April, except that the value reduces over southern Japan and in the latitude band of 20° – 30°N over the North Pacific (Fig. 8b). In May (Fig. 8c), K is about 4–6 throughout 30° – 40°N over the North Pacific but is missing in 20° – 30°N over the subtropical Pacific. Therefore, the differences in the distributions of K during the boreal spring demonstrate that low-frequency waves can propagate along the waveguide (westerly jet stream) from southern China to North America in March.

5. Numerical simulations

The above analysis suggests that the patterns of teleconnection between the major mode of latent heating over the TWP and climate change over North America are sensitive to the change in the intensity of westerly jet stream during the boreal spring. To better understand the related Rossby wave propagation features and associated atmospheric dynamics, we conduct a series of sensitivity experiments using both LBM and AGCM.

a. LBM experiments

We conduct nine sensitivity experiments listed in Table 2. For the scientific issue addressed in this paper,

FIG. 7. Climate mean (1979–2014) 200-hPa zonal wind (shading; m s^{-1}), divergence (blue contours; $2 \times 10^{-6} \text{ s}^{-1}$), and vorticity (red dashed contours; $-3 \times 10^{-5} \text{ s}^{-1}$) in (a) March, (b) April, and (c) May.

FIG. 8. Stationary wavenumber K at 200 hPa in (a) March, (b) April, and (c) May.

we design a heating source of idealized ellipse shape located over the TWP region (Fig. 9a), with its center over 5°N , 140°E (Table 2). The vertical structure of the heating source is an idealized latent heating as shown in Fig. 10, with a maximum heating rate of 9 K day^{-1} in the midtroposphere. We test three basic backgrounds using the heating source over the TWP in the boreal spring, which are named TWP_Mar for March, TWP_Apr for April, and TWP_May for May (Table 2). To compare the effect of the heating over the TWP with the heating over other tropical regions, we use six additional experiments for a heating source over the central tropical Pacific (CP_Mar, CP_Apr, and CP_May) and over the western IO (WIO_Mar, WIO_Apr, and WIO_May), again for March, April, and May.

In the TWP heating experiments (Fig. 11), the 200-hPa geopotential height (HGT) fields from different basic-flow backgrounds (Figs. 11a,d,g) all show a positive anomaly in the forcing region on day 1, because latent heating instantly causes an anticyclone in the upper troposphere. On day 3 (Figs. 11b,e,h), the heating

FIG. 9. Heating forcing (K day^{-1}) at sigma level 0.45 used in different experiments: (a) TWP_Mar, (b) WIO_Mar, and (c) CP_Mar.

effect leads to an expansion of HGT anomalies over the tropical region, similar to the Gill-type response (Gill 1980), and produces a weak Rossby wave propagating to the high latitudes in all three cases. However, on day 7 (Figs. 11c,f,i), the features of Rossby wave propagation become different under various basic flow patterns. In March (Fig. 11c), the Rossby wave propagates eastward and a path is clearly seen from eastern Asia to North America with three pairs of positive–negative HGT anomaly centers, in which the anomalies over the northwestern Pacific are the strongest. In April (Fig. 11f), the Rossby wave propagation becomes weaker in the Northern Hemisphere and stronger in the Southern Hemisphere. In May (Fig. 11i), the HGT anomalies in the Southern Hemisphere are larger and the anomalies over the Northern Hemisphere are smaller than those in April (Fig. 11f). This feature suggests that Rossby wave propagation intensifies in the Southern Hemisphere and weakens in the Northern Hemisphere as the westerly jet stream moves seasonally. In these experiments, because the heating forcing is the same, different Rossby wave propagation features are thus due to the change in the basic flow. The results are consistent with the observational analysis in section 4.

TABLE 2. Experimental design for LBM simulations.

Name	Center of heating	Background field
TWP_Mar	5°N , 140°E	March
TWP_Apr	5°N , 140°E	April
TWP_May	5°N , 140°E	May
CP_Mar	5°N , 200°E	March
CP_Apr	5°N , 200°E	April
CP_May	5°N , 200°E	May
WIO_Mar	5°N , 60°E	March
WIO_Apr	5°N , 60°E	April
WIO_May	5°N , 60°E	May

FIG. 10. Heating profile (K day^{-1}) over 5°N , 140°E in the TWP_Mar, TWP_Apr, and TWP_May experiments. Sigma level is shown on the y axis.

The HGT anomaly pattern in Fig. 11c is similar to the regression pattern of HGT in Fig. 4a. It is thus demonstrated that the location and intensity of the westerly jet stream are crucial for adjusting the teleconnection patterns in the Northern Hemisphere shown in Fig. 7.

In addition to the heating over the TWP, we also test the effects of heating over the central Pacific and

western IO, which are the other two regions of distinguished tropical precipitation. We first show different stages of Rossby wave propagation in CP_Mar, CP_Apr, and CP_May in Fig. 12. It is clear that the evolutions of 200-hPa HGT anomaly patterns on day 1 (Figs. 12a,d,g) and day 3 (Figs. 12b,e,h) are similar as in the TWP experiments, with an idealized positive HGT anomaly on day 1, which propagates to the northeast in the tropics on day 3. It is interesting to find that the HGT anomaly patterns in March, April, and May are similar on day 7 (Figs. 12c,f,i), which is the biggest difference from the TWP experiments. Although the basic backgrounds are different in CP_Mar (Fig. 12c), CP_Apr (Fig. 12f), and CP_May (Fig. 12i), the HGT anomalies show a solid Pacific–North American teleconnection pattern. Finally, we show the results from experiments WIO_Mar, WIO_Apr, and WIO_May in Fig. 13. The evolutions of HGT anomaly patterns on day 1 (Figs. 13a,d,g) and day 3 (Figs. 13b,e,h) are highly similar to those in the TWP and CP experiments. On day 7, a Rossby wave train is triggered clearly under the backgrounds of March (Fig. 13c) and April (Fig. 13f) from tropical Africa to the North Pacific, but it cannot reach North America. In May (Fig. 13i), however, Rossby wave propagation becomes weaker in the Northern Hemisphere and stronger in the Southern Hemisphere, but these responses are both confined to the Eastern Hemisphere.

The different patterns of Rossby wave propagation induced by these three heating sources under the basic flow of boreal spring can be understood by the spatial

FIG. 11. HGT anomaly (gpm) at 200 hPa in the TWP_Mar experiment for days (a) 1, (b) 3, and (c) 7. (d)–(f) As in (a)–(c), but for TWP_Apr. (g)–(i) As in (a)–(c), but for TWP_May.

FIG. 12. As in Fig. 12, but for the central tropical Pacific (CP) experiments.

relation between the heating and the westerly jet stream. The westerly jet streams have two cores, over the northwestern Pacific and North America (Fig. 7), and they weaken gradually from March to May. The heating source over the CP located at the entrance of North American jet stream triggers a similar Rossby wave pattern from March to May. The Rossby wave related to the heating over the western IO is able to propagate along the East Asian jet stream from central Asia to the

western Pacific. Uniquely, the heating source over the TWP located to the south of the jet core triggers a Rossby wave train across the Pacific only in March, because the westerly jet stream is strong with a maximum exceeding 60 m s^{-1} (Fig. 7a). However, from April to May, as the westerly jet stream weakens obviously, Rossby wave propagation weakens and even ceases on the edge of the westerly jet stream. In other words, whether the atmospheric heating over the TWP can

FIG. 13. As in Fig. 12, but for the western IO (WIO) experiments.

exert a remote influence on North America depends on both the location of the heating source and the strength of the westerly jet stream.

b. AGCM experiments

Although the LBM simulations provide a clear physical picture on how the atmospheric heating over the TWP in March triggers a strong Rossby wave that propagates to North America along the westerly jet stream in the upper troposphere, the observed connections between the precipitation over the TWP and the surface temperature over North America have not been verified. In section 3, the precipitation over the TWP in March is found to be closely related to local SST, suggesting that local SST warming (Fig. 5a) is important in triggering variations of precipitation and atmospheric heating, which can influence the North American surface temperature through Rossby wave propagation. To test this hypothesis, we carry out several numerical experiments using an AGCM. The control experiment is named the CON run, which is forced by climatological SST with an annual cycle. The CON run is integrated for 35 model years, and the output from the last 30 years is used. The sensitivity run, named SEN run, has prescribed SST anomaly of an idealized rectangle (0° – 10° N, 130° – 160° E) in the TWP using the climatological values in March only, as shown in Fig. 14a. The SEN run is also integrated for 35 years, and the output of the last 30 years is analyzed.

The effect of TWP SST warming is investigated by analyzing the difference between SEN and CON runs. We show the difference in mean precipitation of March in Fig. 14b. We can see that precipitation not only increases over the TWP but also extends northward to the mid-latitudes outside the prescribed SST forcing region. Moreover, convective motion causes a decrease in precipitation over the SCS and MC. These simulation results are consistent with those of He et al. (2016), which showed the effect of TWP heating on the climate over adjacent regions. He et al. (2016) found that the summer latent heating in the middle troposphere over the TWP generated a local anticyclonic anomaly that led to a strong divergence in the upper troposphere. Consequently, this divergence caused descending anomalies over the surrounding regions, leading to a decrease in precipitation.

In response to the changes in tropical precipitation and the associated latent heating, the difference in 200-hPa HGT (Fig. 15a) exhibits an apparent Rossby wave train from southern China to North America. This pattern is similar to the result from the LBM experiment TWP_Mar (Fig. 11c) and the composite analysis (Fig. 5c), showing negative and positive anomalies over western and central-eastern North America, respectively. Furthermore, since the vertical structure of

FIG. 14. Mean differences between SEN and CON runs during March for (a) SST (K) and (b) precipitation (mm day^{-1}).

Rossby wave is equivalent barotropic, the response of surface temperature (Fig. 15b) also shows a similar pattern as the upper-tropospheric HGT, with cooling over northern North America and warming over central-eastern North America. The surface temperature pattern is also close to the observed (Fig. 5b), which helps us understand the formation of observed surface temperature anomalies in North America. In short, the results of AGCM experiments verify our hypothesis that the atmospheric heating over the TWP triggered by local SST warming contributes to the remote surface warming over North America through Rossby wave propagation under the basic flow background of March.

6. Conclusions and discussion

In this study, we have investigated the major mode of precipitation over the TWP during the boreal spring and examined its connections with global climate anomalies based on observations and results from experiments with LBM and AGCM. The location of latent heating over the TWP can affect the stationary wave anomalies in the upper troposphere, which can then affect North American climate under specific basic flow background. The main results from this study can be concluded as follows.

The precipitation mainly occurs in the MC in March and April, and the main rainbelt shifts northward to Indo-China, the SCS, and the TWP in May after the onset of the Asian summer monsoon. The first EOF mode of precipitation, with a strong positive center over the TWP, explains about 20% of the total variance in March, which is the largest among the spring months. Furthermore, precipitation in March is strongly correlated with local SST, implying that the SST variability is an important trigger for the occurrence of convection and precipitation.

Further observational analysis suggests that the latent heating over the TWP is an important Rossby wave source that causes stationary HGT anomalies in the

FIG. 15. Mean differences between SEN and CON runs during March for (a) 200-hPa HGT (gpm) and (b) surface temperature (K) over land.

upper troposphere in March. The location and strength of the westerly jet stream, which acts as a waveguide, vary significantly from March to May, and the Rossby wave train from the TWP to North America appears clearly in March. The surface temperature over North America increases in phase with the HGT anomalies aloft, as the vertical structure of Rossby wave train is equivalent barotropic.

Experiments using the LBM testify that the location of latent heating under different basic flow backgrounds can affect the Rossby wave train in the upper troposphere. Only the heating prescribed over the TWP, on the south edge of the East Asian westerly jet core, can produce an apparent wave train from southern China to North America. However, when the westerly jet stream becomes weaker in the Northern Hemisphere during April and May, the propagation of Rossby wave is limited and becomes stronger in the Southern Hemisphere. Our AGCM experiments demonstrate a more physically meaningful feature, showing that the warming of TWP SST triggers local convection and produces the Rossby wave train in March, which in turn cause an increase in surface temperature in central-eastern North America. The highly consistent results between models and observations ensure the robustness of our conclusions. Furthermore, we would like to mention that the TWP heating also has an effect on strengthening the Walker circulation, and provides a positive contribution to the western North American surface warming (details are given in section 2 of the supplemental material).

The present study highlights the influence of tropical heating on global climate in the boreal spring and fills a gap in the extensive previous studies focusing on the backgrounds of winter and summer seasons. Previous studies (Nitta 1987; Huang and Sun 1992; Lau 1992) emphasized the important role of TWP heating in producing Rossby waves that propagate northeastward in the boreal summer. The basic PJ pattern (Fig. 1 in Nitta

1987) is a sandwich-type teleconnection with a positive correlation center east of the SCS and two negative centers, located over Japan and New Guinea. The correlation of heating with 500-hPa HGT (Fig. 17 in Nitta 1987) suggests two pairs of HGT anomalies across the whole Pacific. Therefore, North America is dominated by negative correlation that shows an increase in heating over the TWP and a decrease in air temperature over North America in June. Lau (1992) roughly repeated the experiments in Nitta (1987) and found that the heating over 95° – 140° E was important in producing a strong Rossby wave train over the Pacific, while the influence of heating forcing over 140° E– 180° was relatively weaker. The analysis of the streamfunction by Lau (1992) showed paired negative and positive anomalies over North America that were somewhat different from those in Nitta (1987) but highly agree with our current study. However, the heating forcing in our study is smaller than that used in Lau (1992), and the background of basic flow is also different. Since the westerly jet stream is weaker in June than in March, the larger pattern of heating plays an important role in the experiments of Lau (1992). These findings suggest that the location and strength of heating and westerly jet stream are crucial in determine the Rossby wave pattern on the global scale.

We have to admit that the conclusions in the current study are qualitative. We noticed that the features of Rossby wave propagation are slightly different between the LBM and AGCM simulations. In the LBM, when the heating is prescribed in the TWP under the background of March (Fig. 11), the 200-hPa HGT anomalies exhibit a strong positive center over eastern Asia with values above 8 gpm, and the wave train is damped gradually to the east and decreases to about 2 gpm over North America. However, in the AGCM (Fig. 15a), the 200-hPa HGT anomalies induced by prescribed SST forcing exhibit a relatively weaker positive center over eastern Asia. It intensifies when propagating to the east, and finally a positive anomaly exceeding 40 gpm appears over North America. These differences imply that the speed and strength of Rossby wave propagation largely depend on wave–flow or wave–wave interactions as presented in previous studies (Dickinson 1969; Lau and Holopainen 1984; Hayashi and Golder 1987; Bueh et al. 2008; Shi et al. 2016). Quantitative studies of the features of Rossby wave propagation induced by tropical heating should contribute to improvement of weather and climate predictions, and will be carried out in our future studies.

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