

Impact of Heating Anomalies Associated with Rainfall Variations over the Indo-Western Pacific on Asian Atmospheric Circulation in Winter

Jian Zheng · Qinyu Liu · Chunzai Wang ·
Xiao-Tong Zheng

Received: 11 February 2012 / Accepted: 30 July 2012 / Published online: 12 August 2012
© Springer-Verlag 2012

Abstract Observational data show that the dominant mode of the boreal winter rainfall anomalies in the tropical Indo-Western Pacific (IWP) is a west-east dipolar pattern, which is called the Indo-Western Pacific Dipole (IWPD) mode and is related to El Niño-Southern Oscillation. It is found that corresponded to the IWPD mode is a new atmospheric teleconnection pattern—a wave train pattern emitted from the IWP toward Asia and the northwest Pacific in winter. During the positive (negative) phase of the IWPD, the teleconnection pattern features the negative (positive) anomalies of 200-hPa geopotential height (H200) centered at 30°N, 110°E and the positive (negative) anomalies of H200 centered at 45°N, 140°E. The teleconnection pattern represents the dominant mode of the boreal winter H200 anomaly over Asia. A series of simple atmospheric model experiments are performed to confirm that this winter teleconnection pattern is induced by the heating anomalies associated with the IWPD, and the heating anomalies over the equatorial central Pacific are not important to this teleconnection pattern from the IWP toward Asia and the northeast Pacific. The IWPD is strengthened after the climate regime shift of the 1970s, which leads to a stronger teleconnection pattern.

Keywords Tropical Indo-Western Pacific · Rainfall · Teleconnection · East Asia · Decadal change

1 Introduction

El Niño-Southern Oscillation (ENSO) is a dominant mode of climate variability in Earth's climate system. ENSO has profound influences not only on the tropics but also on the midlatitude (e.g. Klein et al. 1999; Alexander et al. 2002). The diabatic heating anomaly over the central equatorial Pacific during ENSO can excite a stationary barotropic Rossby wave train propagating into extratropical regions. This teleconnection is known as the Pacific-North American (PNA) pattern (Wallace and Gutzler 1981) in the Northern Hemisphere and the Pacific-South American (PSA) pattern (Robertson and Mechoso 2003) in the Southern Hemisphere. For El Niño, the PNA-like pattern is usually characterized as negative height anomalies over the North Pacific, positive anomalies over western Canada, and negative anomalies over the southeastern United States (Horel and Wallace 1981). The theoretical work of Hoskins and Karoly (1981) indicates a ray path connecting the anomaly centers of the PNA wave train, which is directed first poleward, then curved eastward, and finally directed back equatorward.

Zhang et al. (1996) pointed out that when El Niño matures in boreal winter, the anomalous circulation is characterized by an anomalous subsidence over the maritime continent, an anomalous anticyclone over the tropical western North Pacific (Wang et al. 1999), and a reversed Hadley circulation in the western Pacific. The atmospheric circulation in the Pacific during the different phases of ENSO is shown in Wang (2002). The convective activity and atmospheric circulation in the western

J. Zheng (✉) · Q. Liu · X.-T. Zheng
Physical Oceanography Lab., Key Lab. of Ocean-Atmosphere
Interaction and Climate in Universities of Shandong, College
of Physical and Environmental Oceanography, Ocean University
of China, Qingdao 266100, China
e-mail: jian.zheng@colorado.edu

C. Wang
Physical Oceanography Division, NOAA Atlantic
Oceanographic and Meteorological Laboratory,
Miami, FL 33149, USA

Pacific are related to ENSO events in the tropical eastern Pacific.

The teleconnection between the tropics and extratropics also occurs over the tropical western Pacific and Asia. Nitta (1987) found an association between abnormal convective activity over the tropical western North Pacific and atmospheric circulation anomalies over the midlatitude of East Asia in summer (June to August; JJA). This teleconnection was named as the Pacific-Japan (PJ) pattern. Associated with the PJ pattern are a gain of kinetic energy and available potential energy through barotropic and baroclinic energy conversions from the climatological mean state (Kosaka and Nakamura 2010), in addition to available potential energy supplied by anomalous diabatic heating in the tropical western North Pacific. The impact of the tropical western Pacific on the East Asian summer monsoon is also studied by Huang and Sun (1992). They pointed out that the summer monsoon rainfall might be weak in the Yangtze River and the Huaihe River valley, when the convective activity around the Philippines is intensified.

These previous studies mainly focused on the remote influence of heating anomaly over the tropical western Pacific on Asian summer climate. What is the influence of the heating anomaly in the tropical Indo-Western Pacific (IWP) during winter? Can the heating anomalies associated with rainfall variations in the IWP affect the winter climate in East Asia? The rainfall anomalies are observed to show two large variations centered in the central equatorial Pacific and in the tropical IWP (Fig. 1) in boreal winter (December to February; DJF). The large rainfall anomalies over the central equatorial Pacific are accompanied with the PNA-like pattern and have been extensively studied previously. The present paper emphasizes the second large rainfall variation over the IWP, with its rainfall variability in winter comparable with that in summer (Fig. 1). The purpose of this paper is to investigate whether and how the winter rainfall variability over the IWP can trigger atmospheric teleconnection to the midlatitudes of Asia and the northwest Pacific. Furthermore, Indo-Pacific climate displays noticeable changes across the 1976/1977 climate regime shift (e.g. Trenberth and Hurrell 1994), and the influence of the tropical Indian Ocean on the western Pacific and East Asian climate is intensified after the late 1970s (Xie et al. 2010). It is still unknown whether similar decadal change exists in the IWP rainfall variability and its teleconnected influence.

The rest of the paper is organized as follows. Section 2 describes the data and model used in this study. Sections 3 and 4 present the leading mode of rainfall variability over the IWP and its impact on Asia and the northwest Pacific based on observational analysis and model experiments. Section 5 provides a summary and discussion.

2 Data and model

2.1 Data

Two sets of rainfall data are used in this study. One is from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997) from 1979 to 2008. The other one is NOAA's Precipitation Reconstruction Dataset (PREC) (Chen et al. 2002; Arkin and Smith 2003), available on a 2.5° grid for 1948–2009. The climatology is based on the time periods of 1951–1990 over land and 1979–1998 over oceans. The other atmospheric data set is from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996). SST data set is from the monthly National Oceanic and Atmospheric Administration Extended Reconstructed Sea Surface Temperature V3b (NOAA ERSST V3b) (Smith et al. 2008).

First, we examine the rainfall interannual variability for the period of 1979–2008 based on the CMAP data. Then the interdecadal modulations in the interannual variability before and after the 1976/1977 climate shift are examined by using the PREC data. To focus on interannual variations, the detrending and 7-year high pass filtering are performed to monthly anomalies in this study.

2.2 Atmospheric model

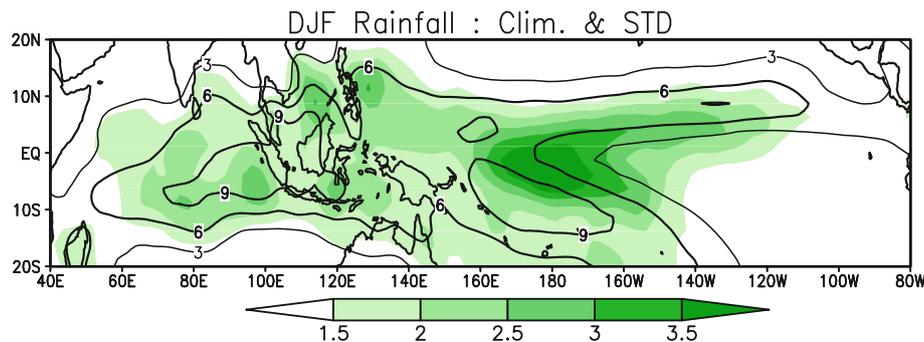
A simple atmospheric model, which is developed by Lee et al. (2009), is used to gain insight into the atmospheric response to heating forcing over the IWP. The model is a steady-state two-level (centered at 250 and 750 hPa) spherical-coordinate primitive equation model, linearized about a specified background flow. The model uses triangular 18 truncations for horizontal grid. The atmospheric mean flow states are climatological mean for boreal winter and vary zonally for the experiments in this study. The detailed description of the simple model is given by Lee et al. (2009).

Some model results of both the local and remote responses of the atmosphere to tropical heating anomalies over the equatorial central Pacific are demonstrated by Lee et al. (2009). Besides, the model successfully simulated the interhemispheric connection between the Atlantic warm pool and the southeastern tropical Pacific (Wang et al. 2010).

3 Leading mode of rainfall variability over the IWP

In order to obtain the spatio-temporal rainfall variability over the IWP in winter, we perform an empirical orthogonal function (EOF) analysis of CMAP winter rainfall anomalies in the domain of 20°S–20°N, 40–160°E. As shown in Fig. 2,

Fig. 1 Climatological mean rainfall (*contours*) and standard deviation of rainfall anomalies (*shading*) in DJF. The rainfall data is from the CMAP from 1979 to 2008. Unit is mm/day



the first EOF mode (EOF1) explains about 36.9 % of the total variance for the period of 1979–2008. For the positive (negative) phase of EOF1, the spatial patterns show a west–east dipolar pattern with more (less) rainfall over the tropical western Indian Ocean (WIO) and less (more) over the tropical eastern Indian Ocean–western Pacific (EI-WP) (Fig. 2a). Over the tropical Indian Ocean, the spatial pattern is similar to the rainfall anomaly pattern for the Indian Ocean Dipole (IOD) mode (Saji et al. 1999). Hereafter, the EOF1 is called the Indo-Western Pacific Dipole (IWP) mode, and the positive phase of the IWP is defined as more (less) rainfall over the WIO (EI-WP). As shown in the PC1 time series, the extreme events of the IWP usually appear in ENSO years, such as 1982 and 1997 (Fig. 2b). The simultaneous correlation between PC1 and the Niño3.4 index is 0.87 (exceeding the 99 % confidence level). The IWP is also correlated with the IOD. The correlation between PC1 and the IOD index in prior months of September–November is 0.68, which is statistically significant at the 99 % level. It is noted that 1982 and 1997 are also IOD years. An extreme IOD event also appears in fall of 1994 (Saji et al. 1999), but there is no significant rainfall variation associated with the IWP in the winter of 1994 (Fig. 2b). The partial correlation between PC1 and the IOD index by removing the Niño3.4 influence is 0.49, and the partial correlation between PC1 and Niño3.4 by removing the IOD influence is 0.70. Thus, the IWP is closely related to the IOD events which occur concomitantly with ENSO.

The spatial patterns of EOF2 show more rainfall over the southern tropical Indian Ocean and less rainfall north of Australia and east to New Guinea (not shown). EOF2 accounts for only 11 % of the total variance. In this study, we therefore focus only on the teleconnected impact of the IWP on Asian atmospheric circulation.

4 Influence of IWP on Asia and the Northwest Pacific

4.1 Observational results

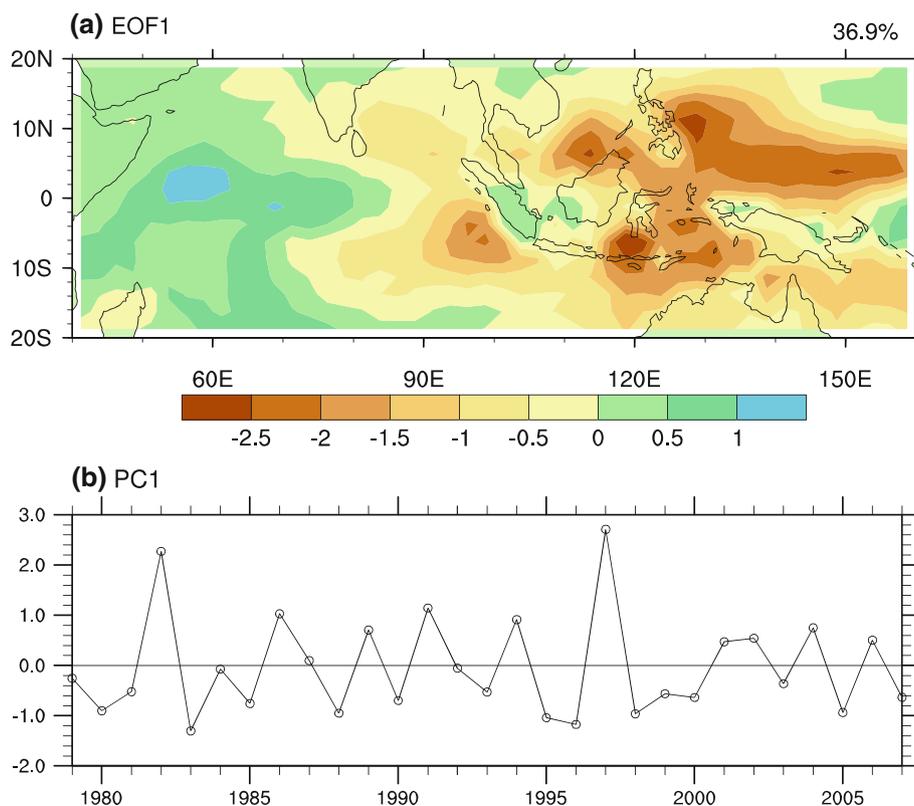
We use the linear regression analysis to examine the possible impact of the IWP on Asian and northwest Pacific

climate in winter. Figure 3 shows the regression maps of oceanic and atmospheric variables onto PC1 time series. The SST anomalies show an El Niño-like pattern in the tropical Pacific, indicating that the IWP is related to ENSO (Fig. 3a). The tropical SST anomalies appear as a zonal tripole, with warm anomalies in the tropical Indian Ocean and eastern equatorial Pacific and cold anomalies in the Maritime Continent/tropical western Pacific.

For the anomalies of 200-hPa geopotential height (H200), there are two anticyclones straddled in the equatorial eastern Pacific associated with the positive phase of the IWP. These anomalous anticyclones are the reminiscent pair of Rossby wave in Matsuno–Gill type of response to the heating anomaly over the equatorial central Pacific (Matsuno 1966; Gill 1980). Over the IWP, the H200 anomaly shows a wedge-shape, which is associated with Kelvin wave produced by the rainfall anomaly in the tropical Indian Ocean. In extratropical regions, the H200 anomaly shows two wave trains (Fig. 3a). One is the PNA-like pattern, which is induced by the heating associated with the warm SST anomalies in the equatorial eastern Pacific. The other wave train is emitted from the IWP toward Asia and the northwest Pacific, with reduced H200 over Asia (centered in 30°N, 110°E) and enhanced H200 over the northeast part of China/Korean Peninsula/Japan (centered at 45°N, 140°E) during the IWP positive phase. At lower level, the circulation anomalies associated with the IWP show low sea level pressure (SLP) anomalies over midlatitude/subpolar Asia and high SLP anomalies over Japan and South Asia. Besides, consistent with the SST anomalies, easterly surface wind anomalies appear over the equatorial Indian Ocean and the far western equatorial Pacific (Wang et al. 1999), and westerly anomalies over the equatorial central Pacific. These indicate that the Walker circulation weakens both in the tropical Indian Ocean and in the tropical Pacific in association with the positive phase of the IWP.

The relationships of the IWP with SST and surface wind in the season of September–November (SON) are shown by the regression map of SON SST and sea surface wind onto PC1 time series (Fig. 3c). The SST pattern is IOD-like in the tropical Indian Ocean and El Niño-like in

Fig. 2 The first EOF mode (a) and the time series (b) of the rainfall anomalies in the Indo-Western Pacific during DJF. The rainfall data is from the CMAP from 1979–2008. The spatial patterns are shown as the rainfall anomaly regressions onto the normalized PC time series. The year number in b indicates December of year



the tropical Pacific. According to previous studies, the feedback effect of SST to atmosphere can be derived by regression of the atmospheric anomalies upon the leading SST anomalies (Frankignoul et al. 1998; Liu et al. 2006; Fan et al. 2011). Thus, the DJF rainfall dipolar pattern over the IWP could be the union response to the SST anomalies in the tropical Indian Ocean and tropical Pacific during SON (Fig. 3c). The positive SST anomaly over the tropical Indian Ocean east to 80°E in DJF (Fig. 3a) could be result of local reduced cloud cover and wind anomaly as responses to the El Niño (Klein et al. 1999; Reason et al. 2000; Yang et al. 2007). However, we should keep in mind that large rainfall anomalies do not necessarily follow high SST anomalies, similar to the rainfall-SST distribution in the tropical Pacific during El Niño (the positive rainfall anomalies are located in the equatorial central Pacific, but high SST anomalies are normally resided in the equatorial eastern Pacific). During SON, associated with the SST anomalies are easterly wind anomalies in the equatorial Indian Ocean and westerly wind anomalies in the equatorial central and eastern Pacific. As in the case during winter, these variations indicate that the Walker circulation weakens both in the tropical Indian Ocean and in the tropical Pacific. Previous studies have shown El Niño could reduce the rainfall and suppress the convection over the western Pacific through the change of the Walker circulation (e.g. Zhang et al. 1996; Wang 2002; Alexander et al.

2004). Yang et al. (2010) also showed the dipolar rainfall pattern over the IWP during November–January is related to the preceding IOD-like SST anomalies during October–December (their Fig. 8). Thus, the developing El Niño in SON and the matured IOD will lead to the IWPD mode, which emits the wave train from the IWP to East Asia and the northwest Pacific in the following winter.

To determine whether the wave train is the primary mode of interannual variability of atmospheric circulation, an EOF analysis is performed using the H200 anomaly in the domain of 20°S–70°N, 0°–180° during winter (DJF). The spatial pattern and time series of the first mode are shown in Fig. 4. EOF1 accounts for 23.3 % of the total variance for the period of 1979–2008. The spatial structure of EOF1 features below-normal height over East China and above-normal height extending from Balkhash to the northwestern Pacific. The pattern resembles the wave train in the regression map. The PC1 time series shows that the extreme events of EOF1 often occur in ENSO years, being similar to the IWPD. In fact, the correlation between the PC1 time series of H200 EOF1 (Fig. 4b) and the PC1 time series of the IWPD (Fig. 2b) is 0.88, which is statistically significant at the 99 % level. In view of these resemblances to the wave train in the regression map, it is suggested that the wave train triggered by the IWPD represents the dominant mode of the interannual variability of the Asian atmospheric circulation during winter.

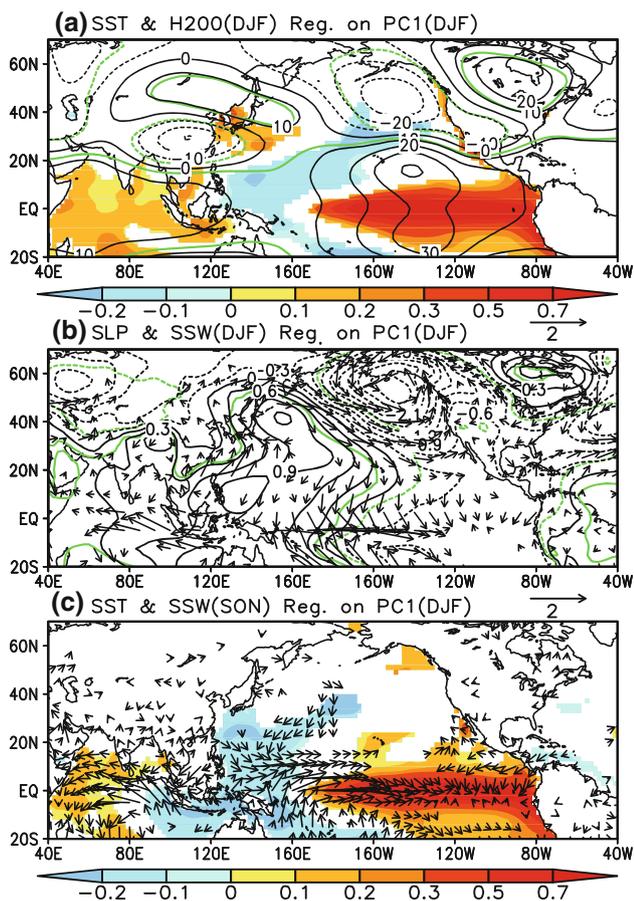


Fig. 3 Regressions of SST anomalies (*shading*) and 200 hPa geopotential height (H200; *contours*) in **a**, of sea level pressure (SLP; *contours*) and sea surface wind (SSW; *vectors*) in **b** onto PC1 time series in DJF. **c** Regressions of SON SST anomalies (*shading*) and SSW (*vectors*) onto PC1 time series in DJF. In **a** and **b**, the *green contours* denote the 90 % confidence level for H200 and SLP, respectively. For SST and SSW, only those exceed the 90 % confidence level are plotted

To further confirm that the teleconnection pattern is a Rossby wave train forced in the tropics, the associated wave-activity flux is shown in Fig. 4c. The wave-activity flux is defined and formulated by Takaya and Nakamura (2001). Consistent with the teleconnection pattern from the IWP to East Asia, the wave-activity flux emits from the tropics towards the Sea of Japan. This indicates that the poleward Rossby wave propagation is related to the teleconnection pattern.

In summary, our analyses of observational data show that there is a new atmospheric teleconnection pattern from the IWP to East Asia and the teleconnection pattern may be a Rossby wave train forced by the heating anomalies associated with the IWP. In next section, we will use a simple atmospheric model to investigate atmospheric responses to various heating distributions.

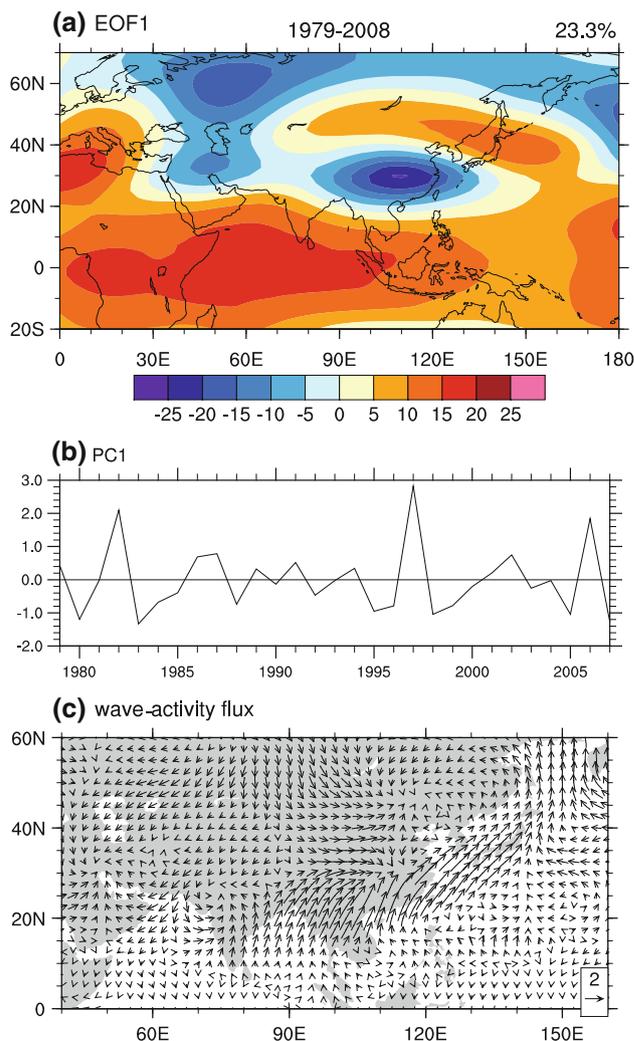


Fig. 4 **a** First EOF mode of DJF H200 anomalies and **b** its time series for the period of 1979–2008, **c** the associated wave-activity flux defined by Takaya and Nakamura (2001) for stationary Rossby waves. The spatial patterns are shown as the H200 anomaly regressions onto the normalized PC time series

4.2 Results from simple atmospheric model experiments

In this subsection, we use simple atmospheric model experiments to test and confirm the observed results in the last section. According to the spatial pattern of the IWP, a dipolar heating pattern as observed in the IWP is used to represent the dominant forcing of the winter rainfall anomaly, with a positive heating centered in 0°, 60°E and a negative heating centered in 5°N, 125°E. Each center of the heating has a Gaussian-shape:

$$Q = Q_0 \exp \left[-\left(\frac{\theta - \theta_0}{L_\theta} \right)^2 \right] \exp \left[-\left(\frac{\lambda - \lambda_0}{L_\lambda} \right)^2 \right], \quad (1)$$

Table 1 The parameter values of Eq. (1) and background mean flows used in the simple model experiments

Expt	Heat forcing			Period for calculating background mean flow
	WIO	EI-WP	EP	
Case 1	$Q_0 = 0.4Q_{0EP}^*$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	$Q_0 = -0.5Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1979–2008
Case 2	$Q_0 = 0.4Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	None	None	1979–2008
Case 3	None	$Q_0 = -0.5Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1979–2008
Case 4	None	None	$Q_0 = Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 160^\circ W$ $L_\theta = 5^\circ, L_\lambda = 25^\circ$	1979–2008
Case 5	$Q_0 = 0.4Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	$Q_0 = -0.5Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1979–2008 ($\bar{U} = 0$)
Case 6	$Q_0 = 0.4Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	$Q_0 = -0.5Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1979–2008 ($\bar{U} = 0$)
Case 7	$Q_0 = 0.4Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	$Q_0 = -0.5Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1977–2009
Case 8	$Q_0 = 0.5Q_{0EP}$ $\theta_0 = 0^\circ, \lambda_0 = 60^\circ E$ $L_\theta = 10^\circ, L_\lambda = 15^\circ$	$Q_0 = -0.7Q_{0EP}$ $\theta_0 = 5^\circ N, \lambda_0 = 125^\circ E$ $L_\theta = 10^\circ, L_\lambda = 25^\circ$	None	1977–2009

* $Q_{0EP} = 2.5 \times 10^{-2}$ W/kg (2.15 K/day)

where θ is latitude and λ is longitude. The parameter values of Eq. (1) for each experiment are summarized in Table 1. Similar to other simple models, the heating is applied to the level at 500 hPa. The basic mean states are set to the values in winter (DJF). The basic flow is obtained by using the stream functions at 250 and 750 hPa from the monthly NCEP-NCAR reanalysis. The stream functions at 750 hPa are constructed by linear interpolation, as 750 hPa is not a standard pressure level. For all cases in the study, the basic flow varies zonally rather than the zonally averaged as in Wang et al. (2010). Since the model atmosphere consists of two levels (centered at 250 and 750 hPa), the variables can be separated into the barotropic and baroclinic components:

$$\bar{Y} = 0.5(Y_1 + Y_2), \quad (2)$$

$$\hat{Y} = 0.5(Y_1 - Y_2). \quad (3)$$

Here Y stands for any variable with subscripts 1 and 2 denoting values at the upper (250 hPa) and lower (750 hPa) levels, respectively.

Figure 5a shows the model response of the barotropic stream function in the case of the dipolar heating forcing of the IWP. The response shows a pattern of alternating high and low centers from the IWP to East Asia and the northwest Pacific. The wave train of barotropic stream function is similar to that of observations (Figs. 3, 4). But in the model, the wave train induced by the IWP heating anomalies can even propagate to North America (Fig. 5a). Two other experiments (Cases 2 and 3) are run with the single polar heat forcing over the WIO and over the EI-WP, respectively. In the case forced by the heating over the WIO, the wave train is very weak (Fig. 5b). But the wave train is strong in the case forced by the cooling over the EI-WP (Fig. 5c), which is comparable with the combined case (Fig. 5a). This indicates that the cooling over the EI-WP is the major contributor to the wave train.

The SST anomaly in the tropical eastern Pacific and the corresponding rainfall anomaly in the equatorial central

Fig. 5 Barotropic stream function anomalies (contours at $1 \times 10^6 \text{ m}^2\text{s}^{-1}$ intervals; dash and blue lines are for negative values; solid and red lines are for positive values) and heating anomalies (dark shading and solid lines represent positive heating anomalies; light shading and dotted lines represent negative heating anomalies) in the simple model experiments for basic mean flows in DJF. Shown are the model experiments of Case 1 (a), Case 2 (b), Case 3 (c) and Case 4 (d). For the detail of these experiments, see Table 1

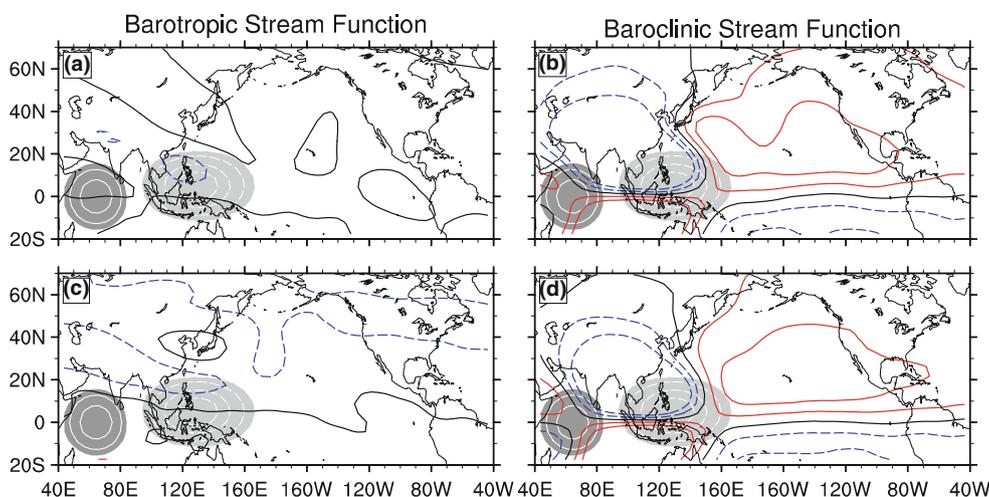
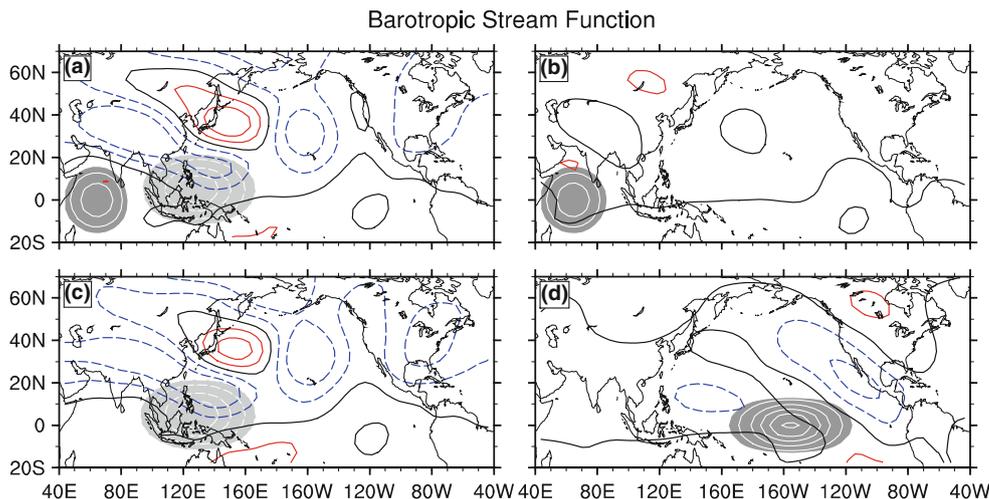


Fig. 6 Stream functions (contours at $1 \times 10^6 \text{ m}^2\text{s}^{-1}$ intervals; dash and blue lines are for negative values; solid and red lines are for positive values) of the barotropic (left panels) and baroclinic (right panels) components from the simple model experiments. a, b Are from Case 5 without the background vertical wind shear over the Northern Hemisphere ($\bar{U} = 0$); c and d are from Case 6 without the

background vertical mean wind over the Northern Hemisphere ($\bar{U} = 0$). Dark shading and solid lines represent positive heating anomalies, and light shading and dotted lines represent negative heating anomalies

Pacific are closely related to the IWP. A natural question is whether the heating anomaly in the equatorial central Pacific can make any contribution to the wave train emitted from the IWP to East Asia and the northwest Pacific. The model experiment with single heating in the tropical central or eastern Pacific (EP) is performed (Case 4 in Table 1). In this case, the response of barotropic stream function shows a wave train over the eastern Pacific and North America, which lies east of the location of classical PNA pattern. Note that the wave train is more similar to the classical PNA pattern if the basic mean flow is replaced with the zonally averaged one in the model (not shown). The signal over the IWP-East Asia is very weak (Fig. 5d). This indicates that the heating in the equatorial central Pacific has

little direct influence on the wave train from the IWP to East Asia and the northwest Pacific.

In brief, the results of the simple model experiments verify and confirm that the IWP atmospheric heating can induce a wave train from the IWP to East Asia and the northwest Pacific. Either of the isolated anomalous heating or cooling associated with the IWP can trigger the wave train, but their combination produces the wave train much closer to the observed one.

As shown in Lee et al. (2009) and Wang et al. (2010), the background mean zonal winds are important for the propagation of heating-induced signals from the tropics to high latitudes in the Atlantic. The background vertical wind shear is important in converting energy from the

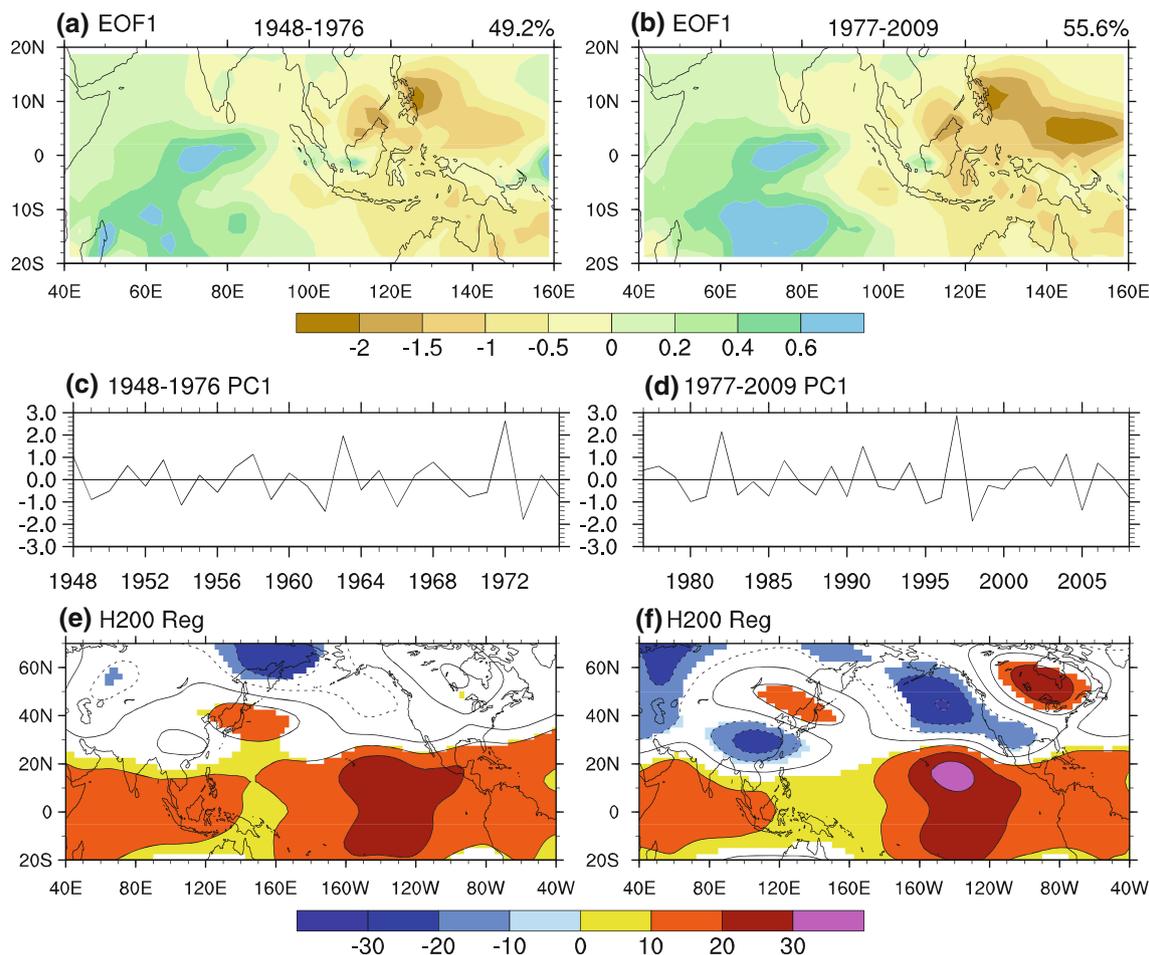


Fig. 7 Top First EOF mode of DJF rainfall anomalies (shown as regressions onto the corresponding time series), middle the corresponding time series and bottom the regression of H200 (contours at intervals of 10 m and dash lines for negative values; shading exceeds

the 90 % confidence level) on the rainfall EOF time series for the period of (left) 1948–1976 and (right) 1977–2009. The EOF analyses of rainfall are based on PREC data

heating-induced baroclinic flow anomalies into barotropic motions near the heating source. Then the barotropic flow anomalies in turn interact with the background vertical mean wind to transmit the barotropic signals to the high latitudes. To examine the effect of the background mean flow on the wave train induced by IWP heating, two model experiments (Cases 5 and 6) are performed with purely barotropic or baroclinic background flows (i.e. $\hat{U} = 0$, $\bar{U} = 0$) over the Northern Hemisphere (Fig. 6). The baroclinic responses to IWP heating in the two experiments are similar (Fig. 6b, d), and resemble those with the realistic background flow experiment (not shown). It is indicated that the baroclinic response to tropical heating is largely independent of the background mean state. But the barotropic stream function response is very different. With no background vertical wind shear (Fig. 6a), the barotropic stream function anomalies are nearly zero in the Northern Hemisphere because the heating-induced baroclinic anomalies cannot interact with the

background vertical wind shear to produce barotropic anomalies in the Northern Hemisphere. Besides, with no background vertical mean wind, the barotropic stream function anomalies are too weak to propagate into high latitudes (also see Fig. 6c). Therefore, the model experiments show that the background mean flow is important for the IWP heating-induced signals to propagate to East Asia and the northwest Pacific.

4.3 Decadal change of IWP and the teleconnection pattern

In the preceding sections, it is shown that a new teleconnection pattern to Asia and the northeast Pacific is induced by the IWP based on the CMAP rainfall data during 1979–2008. Does the teleconnection pattern exist only in this period? In this subsection, we use the PREC data covering from 1948 to 2009 to examine whether the IWP and the teleconnection pattern display changes across the

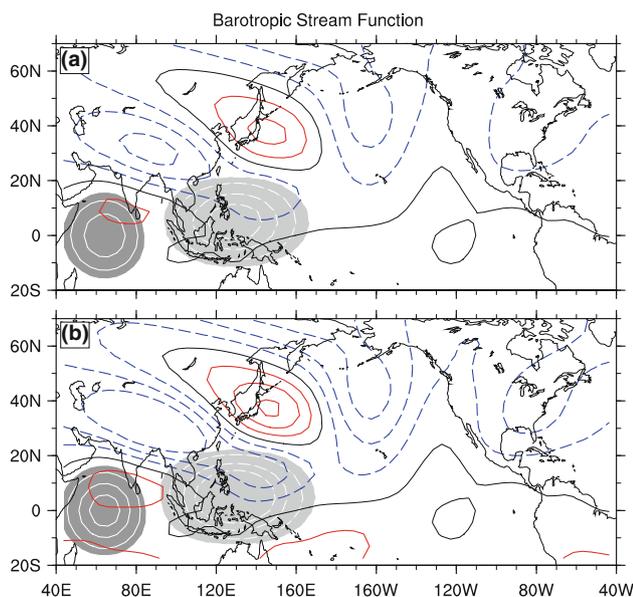


Fig. 8 Barotropic stream function anomalies (contours at $1 \times 10^6 \text{ m}^2 \text{ s}^{-1}$ intervals) for the model experiments of Case 7 (a) and Case 8 (b). Dash and blue lines are for negative values, and solid and red lines are for positive values. Dark shading and solid lines represent positive heating anomalies, and light shading and dotted lines represent negative heating anomalies

1976/1977 climate regime shift. For convenience, the period of 1948–1976 and 1977–2009 are called the PRE and POST epoch, respectively.

The top two panels of Fig. 7 show the EOF1 and the corresponding PC time series of DJF precipitation anomalies over the IWP for the PRE (left panels) and POST (right panels) epoch, respectively. In each period, EOF1 features a dipole pattern with enhanced rainfall over the Indian Ocean and reduced rainfall over the EI-WP, resembling the pattern obtained from the CMAP data of the period 1979–2008. By comparing the patterns in these two epochs, one can see that the IWPDP is stronger in POST than that in PRE, with an additional rainfall anomaly center ($< -1.5 \text{ mm/day}$) to the north of New Guinea in POST (Fig. 7a, b). The IWPDP also accounts for a greater fraction of variance in POST than in PRE (55.6 versus 49.2 %). In the regression maps of H200 anomalies on the time series of the IWPDP (Fig. 7e, f), both the PNA-like pattern and the new teleconnection pattern emitted from the IWP towards Asia are more obvious in POST than those in PRE, corresponded to the IWPDP strength.

We perform an extra pair of model experiments (Cases 7 and 8) to confirm that the decadal change of the IWPDP is responsible for the teleconnection pattern decadal change. The difference between these two experiments is that the heating forcing has stronger amplitude in Case 8 than in Case 7 (Table 1). The model results show that the barotropic stream function anomalies are stronger over Asia,

Sea of Japan and the northern Pacific in Case 8 (Fig. 8b), indicating a stronger teleconnection pattern with the stronger heat forcing in the EI-WP region.

As indicated in the last subsection, the background mean flow is important for the IWPDP heating-induced signals to propagate to East Asia and the northwest Pacific. We also examine whether or not the background mean flow change contributes to the teleconnection pattern decadal change. The result shows that the difference of background mean flows in the PRE epoch and POST epoch is very small, thus has little influence on the decadal change of the teleconnection pattern (not shown). Thus, we can conclude that the decadal change of the teleconnection pattern is mainly due to the intensification of the IWPDP after the climate shift of the 1970s.

5 Summary and discussion

In this study, the interannual variability of rainfall in the tropical Indo-Western Pacific (IWP) and its influence on Asian and northwest Pacific atmospheric circulation during boreal winter are analyzed. It is found the leading mode of rainfall variability is a dipolar pattern mode, with more rainfall in the tropical western Indian Ocean and less rainfall in the tropical eastern Indian Ocean and Maritime Continent/tropical western Pacific. This pattern is called the tropical Indo-Western Pacific Dipole (IWPDP) mode.

The analysis of observational data reveals that the convective heating associated with the IWPDP pattern can trigger a wave train emitted from the IWP to East Asia and the northwest Pacific, with reduced H200 over Asia (centered at 30°N , 110°E) and enhanced H200 over the northeast part of China/Korean peninsula/Japan (centered at 45°N , 140°E). The wave train represents the teleconnected pattern induced by the IWPDP heating, influencing the Asian atmospheric circulation in winter. At lower level of the troposphere, associated with this pattern is an anomalous anticyclone east to Japan.

A simple model forced by heating anomalies confirms and supports the IWPDP-induced teleconnection. The wave train from the IWP to East Asia and the northwest Pacific is triggered by the IWPDP heating anomalies. The model experiments also show that the direct effect of heating in the tropical central Pacific (associated with ENSO heating) on the wave train is very small. The model results indicate that the background mean flow is important for the tropical signals to transmit from the IWP to East Asia and the northwest Pacific. Both the IWPDP and the wave train display a change across the 1976/1977 climate regime shift. Because the IWPDP is intensified after the climate shift, the wave train emitted from the IWP to East Asia and the northwest Pacific is stronger during the POST epoch.

The SST anomaly pattern associated with the IWPD is the IOD and ENSO SST anomaly distribution (Fig. 3a, c). Although the numerical experiment shows small effect of the heating associated rainfall anomaly in equatorial central Pacific on the wave train in East Asia, the SST anomalies over equatorial eastern Pacific are still important to the western Pacific. Previous studies have shown El Niño could reduce the rainfall and suppress the convection over the western Pacific through the change of the Walker circulation (e.g. Zhang et al. 1996; Wang 2002; Alexander et al. 2004). The observational analyses show that the IWPD and the wave train usually occur in El Niño years (Figs. 2b, 4b). A natural question is: Can the IOD and/or ENSO lead to the IWPD and then the wave train from the IWP to East Asia and the northwest Pacific? The study by Cai et al. (2011) pointed out that the IOD can trigger a wave train propagating into the extratropics in the Southern Hemisphere during SON. Can/how we predict the IWPD and the wave train based on the preceding SST anomaly in the tropics? Why does the IWPD exhibit a change across the 1976/1977 climate regime shift? All these issues need be addressed by further studies.

Acknowledgments We thank Dr. Sang-Ki Lee for sharing the model code and helping to run the model. We also thank two anonymous reviewers for their constructive comments and suggestions. This work is supported by Ministry of Science and Technology of China (National Basic Research Program of China 2012CB955602 and National Key Program for Developing Basic Science 2010CB428904), the Changjiang Scholar Program and Natural Science Foundation of China (40830106, 40921004, 41176006 and 41106010).

References

- Alexander MA, Blade I, Newman M, Lanzante JR, Lau N-C, Scott JD (2002) The Atmospheric Bridge: the influence of ENSO teleconnections on air-sea interaction over the global oceans. *J Clim* 15:2205–2231
- Alexander MA, Lau N-C, Scott JD (2004) Broadening the atmospheric bridge paradigm: ENSO teleconnections to the North Pacific in summer and to the tropical west Pacific-Indian Oceans over the seasonal cycle. In: Wang C, Xie S-P, Carton J (eds) *Earth climate: the ocean-atmosphere interaction*. AGU Monograph, pp 85–104
- Arkin, Smith TM (2003) Reconstruction of the oceanic precipitation from 1948 to the present. In: *The AMS 14th symposium on global change and climate variations*, Long Beach, CA, 2003
- Cai W, Rensch PV, Cowan T, Hendon HH (2011) Teleconnection pathways of ENSO and the IOD and the mechanisms for impacts on Australian rainfall. *J Clim* 24:3910–3923
- Chen M, Xie P, Janowiak JE, Arkin PA (2002) Global land precipitation: a 50-yr monthly analysis based on gauge observations. *J Hydrometeorol* 3:249–266
- Fan L, Liu Z, Liu Q (2011) Robust GEFA assessment of climate feedback to SST EOF modes. *Adv Atmos Sci* 28(4):907–912
- Frankignoul C, Czaja A, L'Heveder B (1998) Air-sea feedback in the North Atlantic and surface boundary conditions for ocean models. *J Clim* 11:2310–2324
- Gill AE (1980) Some simple solutions for heat-induced tropical circulation. *Q J R Meteorol Soc* 106:447–462
- Horel JD, Wallace JM (1981) Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Mon Wea Rev* 109:813–829
- Hoskins BJ, Karoly DJ (1981) The steady linear response of a spherical atmosphere to thermal and orographic forcing. *J Atmos Sci* 38:1179–1196
- Huang R, Sun F (1992) Impacts of the tropical western Pacific on the East Asian summer monsoon. *J Meteorol Soc Jpn* 70:243–256
- Kalnay E et al (1996) The NCEP/NCAR 40-year reanalysis project. *Bull Am Meteorol Soc* 77:437–471
- Klein SA, Sode BJ, Lau N-C (1999) Remote sea surface temperature variations during ENSO: evidence for a tropical atmospheric bridge. *J Clim* 12:917–932
- Kosaka Y, Nakamura H (2010) Mechanisms of meridional teleconnection observed between a summer monsoon system and a subtropical anticyclone. Part I: the Pacific–Japan pattern. *J Clim* 23:5085–5108
- Lee S-K, Wang C, Mapes BE (2009) A simple atmospheric model of the local and teleconnection responses to tropical heating anomalies. *J Clim* 22:272–284
- Liu Q, Wen N, Liu Z (2006) An observational study of the impact of the North Pacific SST on the atmosphere. *Geophys Res Lett* 33:L18611. doi:10.1029/2006GL026082
- Matsuno T (1966) Quasi-geostrophic motions in the equatorial area. *J Meteorol Soc Jpn* 44:25–43
- Nitta T (1987) Convective activities in the tropical western Pacific and their impact on the northern hemisphere summer circulation. *J Meteorol Soc Jpn* 65:373–390
- Reason CJC, Allan RJ, Lindesay JA, Ansell TJ (2000) ENSO and climatic signals across the Indian Ocean basin in the global context: Part I. Interannual composite patterns. *Int J Climatol* 20:1285–1327
- Robertson AW, Mechoso CR (2003) Circulation regimes and low-frequency oscillations in the South Pacific sector. *Mon Wea Rev* 131:1566–1576
- Saji NH, Goswami BN, Vinayachandran PN, Yamagata T (1999) A dipole mode in the tropical Indian Ocean. *Nature* 401:360–363
- Smith TM, Reynolds RW, Peterson TC, Lawrimore J (2008) Improvements to NOAA's historical merged land–ocean surface temperature analysis (1880–2006). *J Clim* 21:2283–2296
- Takaya K, Nakamura H (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
- Trenberth KE, Hurrell JW (1994) Decadal atmospheric-ocean variations in the Pacific. *Clim Dyn* 9:303–319
- Wallace JM, Gutzler DS (1981) Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon Wea Rev* 109:784–812
- Wang C (2002) Atmospheric circulation cells associated with the El Niño–Southern Oscillation. *J Clim* 15:399–419
- Wang C, Weisberg RH, Virmani J (1999) Western Pacific interannual variability associated with the El Niño–Southern Oscillation. *J Geophys Res* 104:5131–5149
- Wang C, Lee S-K, Mechoso CR (2010) Inter-hemispheric influence of the Atlantic warm pool on the southeastern Pacific. *J Clim* 23:404–418
- Xie P, Arkin PA (1997) Global precipitation: a 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull Amer Meteorol Soc* 78:2539–2558
- Xie S-P, Du Y, Huang G, Zheng X-T, Tokinaga H, Hu K, Liu Q (2010) Decadal shift in El Niño influences on Indo-western Pacific and East Asian climate in the 1970s. *J Clim* 23:3352–3368

- Yang J, Liu Q, Xie S-P, Liu Z, Wu L (2007) Impact of the Indian Ocean SST basin mode on the Asian summer monsoon. *Geophys Res Lett* 34:L02708. doi:[10.1029/2006GL028571](https://doi.org/10.1029/2006GL028571)
- Yang J, Liu Q, Liu Z (2010) Linking Observations of the Asian Monsoon to the Indian Ocean SST: possible Roles of Indian Ocean Basin Mode and Dipole Mode. *J Clim* 23:5889–5902
- Zhang R, Sumi A, Kimoto M (1996) Impact of El Niño on the East Asia Monsoon: a diagnostic study of the '86/87 and '91/92 events. *J Meteor Soc Jpn* 74:49–62