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Boreal summer convection oscillation over the Indo-Western Pacific and its relationship with the East Asian summer monsoon

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Abstract

Statistical analysis shows that the Indo-Pacific convection oscillation (IPCO) in boreal summer undergoes quasi-biennial variability, and is closely related to the East Asian summer monsoon (EASM). Negative IPCO phases, which exhibit an enhanced convection over the north Indian Ocean and a suppressed convection over the western North Pacific, favor a weakened EASM and an increase of summer rainfall in the Yangtze River valley with the joint actions of the stronger than normal Ural and Okhotsk blocking highs and the subtropical western Pacific high, and vice versa. Copyright © 2013 Royal Meteorological Society

Keywords: Indo-Pacific convection oscillation; Quasi-biennial variability; East Asian summer monsoon

1. Introduction

Boreal summer is the wettest season in East Asia, to which the East Asian summer monsoon (EASM) contributes enormously. The EASM exhibits a great interannual variability. Numerous studies have linked this variability to the Pacific and Indian Ocean sea surface temperature (Huang and Wu, 1989; Chen *et al*., 1992; Zhang *et al*., 1996; Lau and Weng, 2001). The effects of El Niño-Southern Oscillation (ENSO) (Shen and Lau, 1995; Chang *et al*. 2000; Wang *et al*., 2000) and Indian Ocean Dipole (Li and Mu, 2001; Yuan *et al*., 2008) events on the EASM have been extensively discussed. Convection is a key link in the interaction processes between ocean and atmosphere, especially in the tropics (Huang and Sun, 1992; Chang *et al*. 2000; Wang *et al*., 2000).

Tropical convection is characterized by its variability over multiple timescales. Over the Indo-western Pacific, the intraseasonal variability (ISV) is the most pronounced (Lau and Chan, 1988; Chen *et al*., 1996). The convection ISV always exhibits the eastward propagation of large-scale deep convection in boreal summer (Weickmann and Khalsa, 1990; Rui and Wang, 1990; Wang and Rui, 1990; Lee *et al*., 2012), which results in a 'seesaw' over the tropical Indian Ocean and the western Pacific (Lau and Chan, 1986; Zhu and Wang, 1993). This 'seesaw' is closely related to the active/break of Asian monsoon. Besides the ISV, the interannual variability of boreal summer convection over this sector is also more significant than other regions with the standard deviation (from the 1979–2010 mean climatology) up to 10 W m^{-2}. This issue has not yet been fully discussed.

The tropical western Pacific convection can affect the East Asian summer climate through atmospheric teleconnection, i.e. the Pacific-Japan (PJ) pattern (Nitta, 1987; Huang and Sun, 1992). Meanwhile, the EASM is also influenced by tropical Indian Ocean convection, indicated by its correlation map with precipitation (Figure 4(b) in Li and Zeng, 2002). The above mentioned review raises two questions. Does a dipole pattern exist in the year-to-year variability of summer mean convection over the Indo-western Pacific region, like the ISV 'seesaw'? What about its relationship with the EASM? This study mainly focuses on these two questions.

2. Data and methodology

The analysis is based on boreal summer (June to August, JJA) mean data for the period 1979–2010 derived from monthly datasets that include interpolated outgoing longwave radiation (OLR) data (Liebmann and Smith, 1996), the National Centers for Environmental Prediction and Atmospheric Research (NCEP/NCAR) reanalysis 1 (Kalnay *et al*., 1996) including the zonal wind, meridional wind, vertical velocity, and geopotential height, the Hadley Centre sea surface temperature dataset on a 1° × 1° grid (Rayner *et al*., 2003), and the China 160-station rainfall dataset. The 1979–2009 Climate Prediction Center merged analysis of precipitation (CMAP) (Xie and Arkin, 1997) is also used. The OLR, NCEP/NCAR reanalysis, and CMAP data are on a 2.5° × 2.5° grid.

The EASM index (EASMI) used here, which is one of the best indices to measure the intensity of EASM

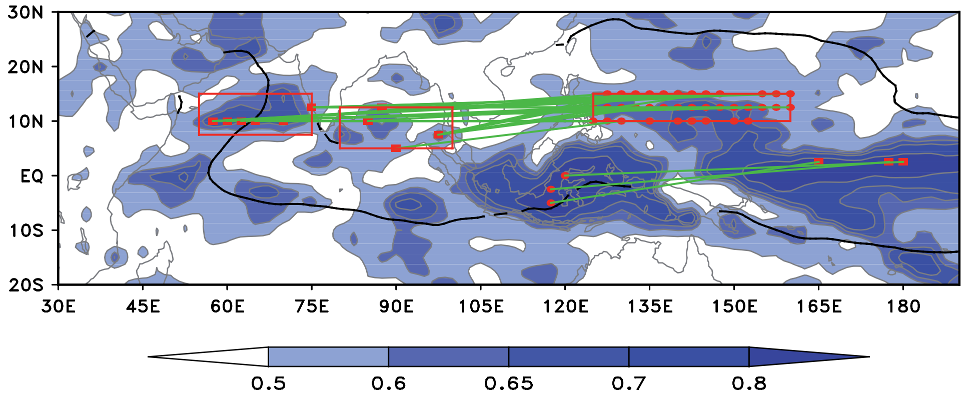

Figure 1. Teleconnectivity map of the JJA anomalies in the OLR field. Only regions with the strongest negative correlation coefficients less than −0.5 are shaded. Green line connects selected point (red solid circle) with that (square) which exhibits the strongest negative correlation on its one-point correlation map. Red rectangles denote selected regions: WNP (10°N–15°N, 125°E–160°E), ARB (7.5°N–15°N, 55°E–75°E), and SBOB (80°E–100°E, 5°N–12.5°N). Black line denotes the 28 °C isoline of climatological (1979–2010) JJA SST.

(Wang *et al*., 2008), is derived from a unified Dynamic Normalized Seasonality (DNS) monsoon index defined by Li and Zeng (2002, 2003). This DNS index is based on intensity of the seasonality of wind field and can be used to depict both the seasonal cycle and interannual variability of monsoons over different areas. Given a pressure level and a gird point (i, j), the DNS index in the mth month of the nth year is given by

$$\delta_{nm}(i,j) = \frac{\|\mathbf{V}_1(i,j) - \mathbf{V}_{nm}(i,j)\|}{\|\mathbf{V}(i,j)\|} - 2 \qquad (1)$$

where $\mathbf{V}_1(i,j)$ (m s^{-1}) is the January climatology wind vector, and $\mathbf{V}_{nm}(i,j)$ (m s^{-1}) is the wind vectors at grid point (i, j) in the mth month of the nth year. $\mathbf{V}(i,j)$ (m s^{-1}) is the mean of January and July climatology wind vectors at grid point (i, j). The norm $||A||$ is defined as $||A|| = (\int\int_S |A|^2 dS)^{1/2}$, where S denotes the domain of integration. Then, we can define a large-scale monsoon index MI_{nm} in the mth month of the nth year as a measure of the averaged DNS over a monsoon domain given by

$$MI_{nm} = \{\delta_{nm}(i,j)\} \qquad (2)$$

where {} denotes the areal average of δ values at grid points within a chosen monsoon domain at a certain pressure level. The EASMI is the JJA mean MI_{nm} of the nth year over the domain (10°–40°N, 110°–140°E). More details on the physical definition are described by Li and Zeng (2000 and 2002), Feng *et al*. (2010) and Li *et al*. (2010).

To determine the dominant modes of interannual convection variation over the Indo-western Pacific region, the teleconnectivity analysis (Wallace and Gutzler, 1981) is employed over the domain (30°E–170°W, 20°S–30°N). Some statistical methods, i.e. correlation, partial correlation, composite and empirical orthogonal function (EOF) analyses are used in this work. Several key regions mentioned frequently in this work are defined as follow: the Arabian Sea (ARB) (7.5°–15°N, 55°–75°E), south of the Bay of Bengal (SBOB) (80°–100°E, 5°–12.5°N), the north Indian Ocean (NIO) (which comprises ARB and SBOB), and the western North Pacific (WNP) region (10°–15°N, 125°–160°E). The normalized time series of OLR anomalies averaged over these regions are denoted as the I_{ARB}, I_{SBOB}, I_{NIO}, and I_{WNP}, respectively. A positive (negative) phase refers to the OLR anomaly greater (less) than +1 (−1) standard deviation.

3. Results

Figure 1 shows the teleconnectivity map of the JJA anomalies in the OLR field. It can be seen the most prominent feature of the opposing variability of convection over the maritime continent and the equatorial western-central Pacific statistically. This dipole pattern is quasi-stationary in all the four seasons, and has been described by Lau and Chan (1983), referred to as the maritime continent-Pacific convection oscillation (MPCO) hereafter. In addition, the significant negative correlations over the ARB, SBOB, and WNP regions (denoted by red rectangles in Figure 1) indicate the out-of-phase fluctuation in convection anomalies over the NIO and WNP, referred to as the Indo-Pacific convection oscillation (IPCO). The MPCO and IPCO represent the dominate traits of the first and second EOF modes (EOF1 and EOF2) (figures not shown) which explain around 23.5 and 13.4% of the variance, respectively. The correlation between the principle component (PC) of EOF1 and EASMI is about 0.39, far less than that between the PC2 and EASMI (0.71, but negative). In this work, we focus on the IPCO.

To highlight the IPCO, the normalized time series of area-averaged OLR anomalies are shown in Figure 2(a). The convection variations over the ARB and the SBOB are in phase, with a significant correlation of 0.71. Therefore, the I_{NIO} can be used to represent the interannual variability of convection

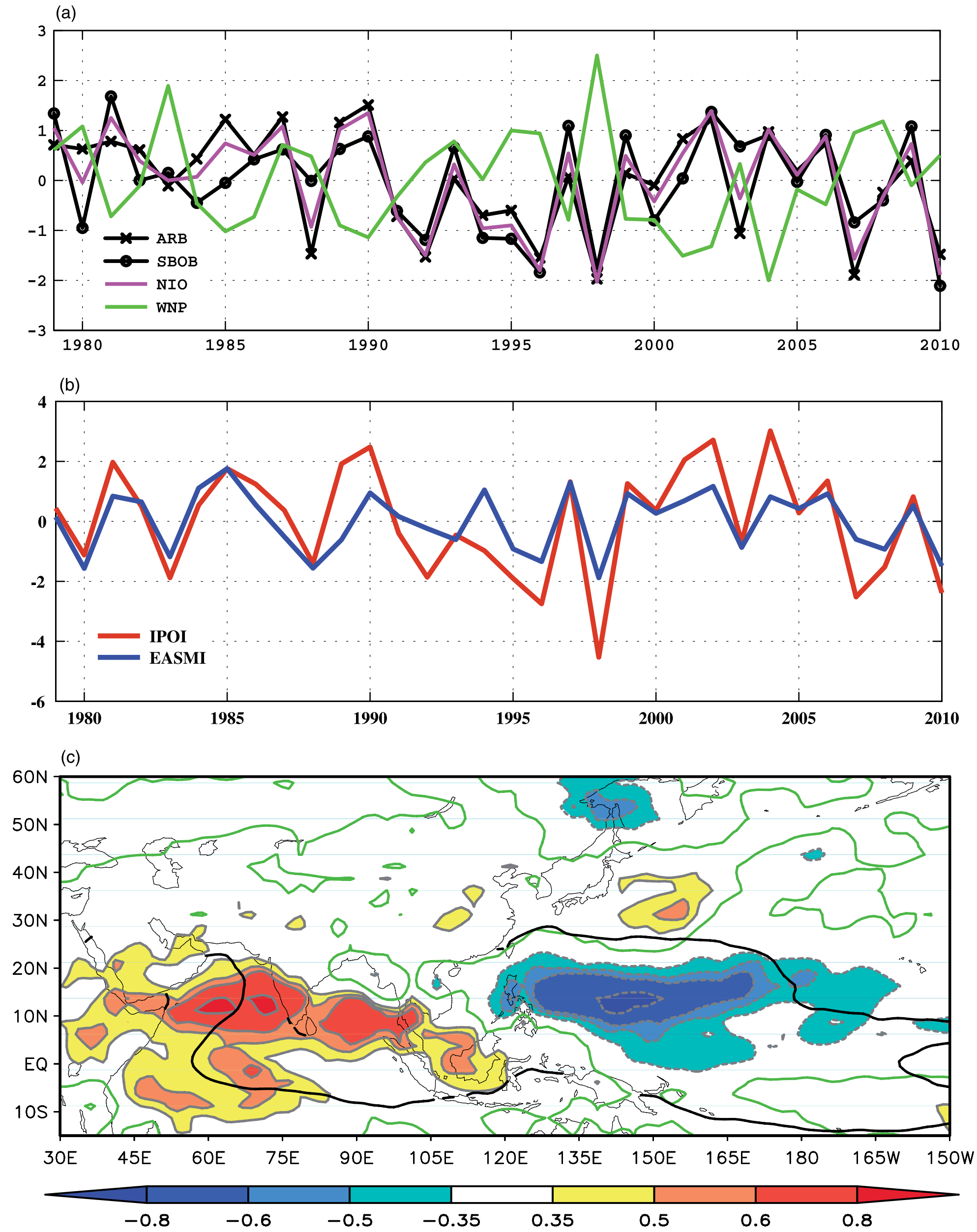

Figure 2. (a) Normalized time series of area-averaged OLR anomalies over the ARB (I_{ARB}, black line with crosses), SBOB (I_{SBOB}, black line with circles), NIO (I_{NIO}, purple line), WNP (I_{WNP}, green line); (b) time series of the IPCOI (red line) and EASMI (blue line); (c) correlation map between the IPCOI and the OLR field. Green line in (c) denotes the zero correlation coefficient, and the black line denotes the 28 °C isoline of climatological (1979–2010) JJA SST. Shading indicates >95% significance by Student's *t*-test.

over the NIO. The correlation between the I_{NIO} and I_{WNP} is negative and significant above the 99% confidence level with the value of 0.62. Thus, when convection over the NIO is enhanced (suppressed), that over the WNP tends to be suppressed (enhanced). An IPCO index (IPCOI) is defined as the difference in the normalized area-averaged OLR anomalies over the NIO and WNP:

$$\mathrm{IPCOI} = I_{\mathrm{NIO}} - I_{\mathrm{WNP}} \quad (3)$$

It can be seen that the positive phases of the IPCO show the convection suppressed over the NIO and enhanced over the WNP, and vice versa. The correlation map of the IPCOI with the OLR field (Figure 2(c)) reveals the spatial pattern of the IPCO. It also can be seen from Figure 2(c) that notable convection anomalies over the regions from eastern China to the western Pacific (east of Japan) and the sea of Okhotsk accompany with the IPCO. The out-of-phase variations in convective activities over the WNP region and the Pacific (east of Japan) are in agreement with the PJ Pattern.

Figure 2(b) shows the interannual variability of the IPCO and the EASM. It is obvious that the IPCO exhibits a quasi-biennial variability. Its positive phases occurred in 1981, 1985, 1989, 1990, 1997, 1999, 2001, 2002, 2004, and 2006, while negative phases occurred in 1980, 1983, 1988, 1992, 1994, 1995, 1996, 1998, 2007, 2008, and 2010. The correlation coefficient

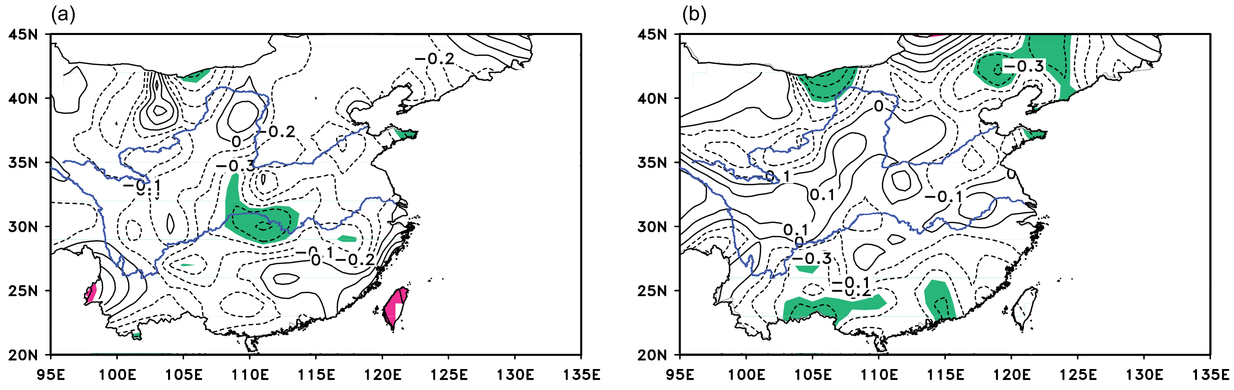

Figure 3. (a) Correlation map between the IPCOI and China rainfall; (b) same as (a), but for partial correlation map with the EASMI signal removed. The shaded areas indicate significance at the 95% confidence level.

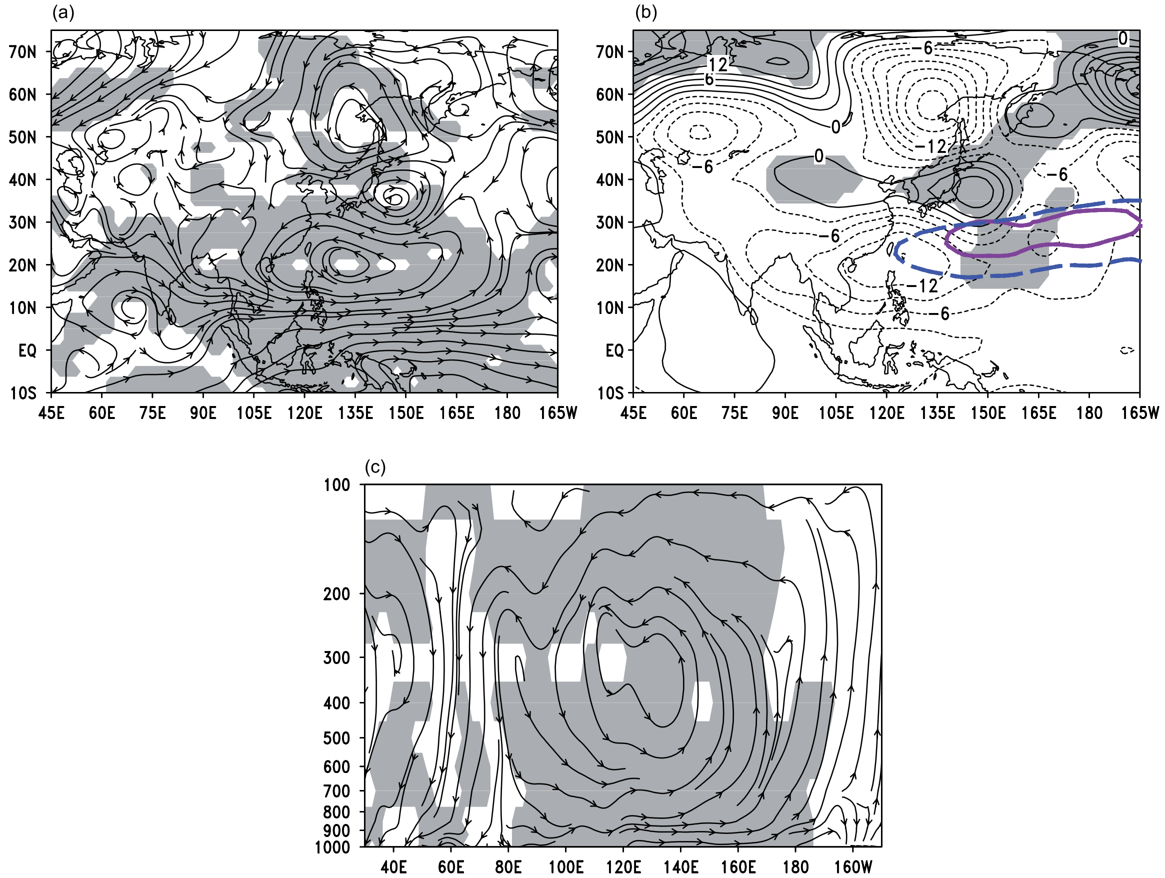

Figure 4. Composite differences between the positive and negative IPCO phases for (a) 850 hPa stream lines, (b) 500 hPa geopotential height (units: gpm, contour interval: 3 gpm), (c) vertical circulation (units of horizontal and vertical components: $m\,s^{-1}$ and $-10^{-2}\,Pa\,s^{-1}$, respectively) along 7.5°N. The purple (blue) line in (b) indicates the 5880 gpm isoline of geopotential height in the positive (negative) IPCO phase. The shaded areas indicate significance at the 95% confidence level.

between the EASMI and IPCOI is 0.8, indicating similarity between interannual variability in the EASM and IPCO (Figure 2(b)). This implies that the quasi-biennial variability of the IPCO must be closely related to that of the EASM.

On the basis of above analysis, it is expected that the IPCO is connected with East Asian summer precipitation. The correlation between the IPCOI and summer precipitation over the Indo-western Pacific sector (figure not shown) exhibits a contrary pattern with Figure 2(c), suggesting the existence of the IPCO in the interannual variability of precipitation. Figure 3 shows the correlation between the IPCOI and summer rainfall in China. Significant negative correlations are observed in the middle reach of the Yangtze River valley. However, the partial correlation suggests these significant correlations become insignificant after removing the EASMI signal. It implies that the IPCO influences summer rainfall in the middle reach of the Yangtze River valley via the bridge of the EASM.

To further examine the relationship of the IPCO with the EASM and the rainfall, the composite differences in circulations between the positive and negative IPCO phases are shown as Figure 4. In the positive IPCO phase, anticyclonic anomalies appear over the NIO and the western Pacific (east of Japan), while cyclonic anomalies appear over the WNP and northeastern Asia near Sakhalin Island at 850 hPa (Figure 4(a)). These anomalies form a teleconnection from the Indian Ocean across the East Asia to the high latitudes which coincides with the Indo-Asia-Pacific (IAP) pattern (Li and Hu, 2011). Meanwhile, strong divergence over the middle and lower reaches of the Yangtze River valley, Korea and Japan is observed, which may result in less precipitation in these regions during the Meiyu-Changma-Baiu season. At 500 hPa (Figure 4(b)), the IAP pattern is still clearly identified. The Ural and Okhotsk blocking highs are weaker than normal, and the western Pacific subtropical high (WPSH) weakens and shrinks. These anomalies have been manifested to favor a strong EASM (Wu *et al*., 2009, 2012). The zonal-vertical anomaly circulation over the Indo-western Pacific associated with the positive phase of the IPCO (Figure 4(c)) is anticlockwise, i.e. the downward anomalies over the NIO, upward over the WNP, westerly at low levels, and easterly at high levels. This circulation integrates the two poles of the IPCO, connecting the out-of-phase fluctuation between convection over the two poles. Anomalies are opposite in the negative IPCO phases.

4. Summary and discussion

In this article, we determine the characteristics of the dominant interannual variation modes of the JJA convection over the Indo-western Pacific region. The MPCO (out-of-phase convection anomalies over the maritime continent and equatorial Pacific) and IPCO (out-of-phase convection anomalies over the NIO and the WNP) are identified using statistical methods. The latter is highlighted. The IPCO displayed a quasi-biennial variability during the period 1979–2010, and is closely related to the EASM and rainfall in the middle reach of the Yangtze River valley. Years with a positive (negative) IPCOI phase are associated with a stronger (weaker) EASM and less (more) rainfall in the middle reach of the Yangtze River valley due to the combined action of the weaker (stronger)-than-normal blocking highs over the Ural mountain and Okhotsk Sea at middle to high latitudes, and the weakening and shrinking (strengthening and extending) WPSH at low latitudes.

Different from the previous studies (Lau and Chan, 1986; Zhu and Wang, 1993), we here highlight the interannual variability of the out-of-phase fluctuation between summer convection over the tropical Indian Ocean and WNP and its potential influences on the EASM as well. By considering the significant relationship between the East Asian summer rainfall and the quasi-biennial IPCO, this study provides a possible new clue to predict East Asian summer rainfall on interannual time scale. Preceding studies (Lau and Chan, 1988; Chen and Houze, 1997) have manifested that the relationship between tropical convection variability and the ENSO is complex. The possible linkage between the IPCO and ENSO is an important topic in our future study.

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