A dipole pattern in the Indian and Pacific oceans and its relationship with the East Asian summer monsoon

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Abstract
This study demonstrates a robust relationship between the Indo-Pacific warm pool (IPWP) and North Pacific Ocean dipole (IPOD) and the East Asian summer monsoon (EASM) using observational datasets and sensitivity tests from the Community Atmosphere Model version 3.1 of the National Center for Atmospheric Research. The IPOD, which is a significant pattern of boreal summer SSTA in the Indian and Pacific oceans characterized by positive (negative) sea-surface temperature anomalies (SSTA) in the North Pacific and negative (positive) SSTA in the IPWP, appears around May, intensifies in the following months, and weakens in September. In summers with a positive IPOD phase, the western Pacific subtropical high (WPSH) weakens and shrinks with the axis of the WPSH ridge moving northwards, which favours an intensified EASM and a decrease in summer rainfall in the Yangtze River valley, and vice versa.

Keywords: Indo-Pacific warm pool and North Pacific Ocean dipole, East Asian summer monsoon, western Pacific subtropical high

1. Introduction

The East Asian summer monsoon (EASM) has a significant impact on China, Japan, Korea, and the surrounding seas (Tao and Chen 1987). The significant interannual and interdecadal variability of the EASM can cause floods and droughts, which have often caused large economic losses and casualties in the affected areas (Huang et al 2004, 2007).

Previous studies have suggested that the EASM is regulated by sea-surface temperatures (SST) in both the Pacific and Indian oceans (Huang and Wu 1989, Wu and Liu 1995, Wu et al 1995). The effects of the Indian Ocean on the EASM have been extensively discussed (Guan and Yamagata 2003, Li et al 2008, Wu et al 2009, Xie et al 2009, Ding et al 2010, Li and Hu 2011). The Indo-Pacific warm pool (IPWP) contains some of the warmest ocean water in the world, and it is an important source of energy for driving atmospheric circulation and an important sink of atmospheric fresh water (Wang and Mehta 2008). The IPWP moves northwards during the boreal summer and southwards during the boreal winter, following the annual march of the Sun (Kim et al 2012). The variations of the IPWP have a significant impact on climatic conditions in the tropics (Ma and Li 2008, Feng and Li 2013) and also at higher latitudes, including the South Asian monsoon (D’Arrigo et al 2006, Annamalai et al 2013), the EASM (Huang and Sun 1992, Feng et al 2011), and the Northeast Asian summer monsoon (Shin et al 2011).

Although the atmospheric signal over East Asia associated with mid-latitude SST anomalies (SSTA) is less robust...
than that associated with tropical SSTA, the feedback processes associated with mid-latitude air-sea coupling also have an important effect on atmospheric anomalies over East Asia (Lau 1997, Liu and Wu 2004, Wu et al 2009). SSTA in the North Pacific (NP) also influence the East Asian climate by affecting summer circulation and precipitation in this region (Wang and Wang 2002, Liu et al 2006, Frankignoul and Semmchael 2007, Tan et al 2009), and the dominant mode in the NP–Pacific decadal oscillation (PDO) can modulate the impact of El Niño/Southern Oscillation (ENSO) events on summer climate variability in China (Zhu and Yang 2003).

The IPWP and the NP are both important areas of SSTA affecting the EASM, but only a few studies have investigated the joint impacts of the IPWP and the NP on the EASM. Li and Hu (2011) proposed a cross-basin dipole pattern that correlates remarkably well with the variations of the EASM. This pattern, characterized by positive (negative) SSTA in the NP and negative (positive) SSTA in the IPWP, is termed the IPWP and North Pacific Ocean dipole (IPOD). The IPOD index (IPODI) is defined by the difference in normalized SSTA between the NP (120°–160°E, 36°–44°N) and the IPWP (80°–130°E, 4°–24°N).

This paper focuses on characterizing the relationship of the IPOD with the EASM. The remainder of the paper is organized as follows. The datasets and methodology are described in section 2, the results are presented in section 3, and a summary is provided in section 4.

2. Data and methodology

The EASM index (EASMI) used here, which has a good skill in measuring the intensity of the EASM (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. Such the DNS index in the nth month of the nth year is given by

\[ \delta_{m,n} = \frac{\| \mathbf{V}_m - \mathbf{V}_{m,n} \|}{\| (\mathbf{V}_m + \mathbf{V}_{m,n}) / 2 \|}, \]

(1)

where \( \mathbf{V} \) are the January climatology wind vector and the mean of January and July climatological wind vectors, respectively, and \( \mathbf{V}_{m,n} \) is the monthly wind vector for the year \( n \) and the month \( m \). Then, we can define the EASMI as the JJA DNS index averaged over the East Asia (EA; 110°–140°E, 10°–40°N)

\[ \text{EASMI} = \{ \delta \}_{1,4}, \]

(2)

More details on the DNS index and the EASMI are described by Li and Zeng (2002, 2003), Li et al (2010). The EASMI can be downloaded from the website: http://ljp.lasg.ac.cn/dct/page/65577.

For SST data, we used the NOAA Extended reconstructed SST V3b on a 2° × 2° grid (Smith et al 2008) for the period 1948–2012. Atmospheric reanalysis data were obtained from the National Center for the Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis 1 for the period 1948–2012 (2.5° × 2.5° grid) including geopotential height, zonal wind and meridional wind (Kalnay et al 1996). The NOAA’s Precipitation Reconstruction dataset on a 2.5° × 2.5° grid for the period 1948–2012 (Chen et al 2002), and University of Delaware Precipitation on a 0.5° × 0.5° grid for the period 1948–2010 (obtained from www.esrf.noaa.gov/psd/) were also used in the study.

The analysis is based on the boreal summer (June to August, JJA) mean data. The two key regions of interest in this work are the NP (120°–160°E, 36°–44°N) and the IPWP (80°–130°E, 4°–24°N). The empirical orthogonal function/principal component analysis (EOF/PCA) (e.g. Zhang et al 2010) and rotated EOF/PCA (e.g. Barlow et al 2001) are often applied to extract the modes of SST variability. In this paper, the unrotated EOF method was employed in the domain (30°E–60°W, 40°S–60°N) to determine the dominant modes of JJA SST variability in the Pacific and Indian Oceans. The statistical techniques used in this study were correlation, partial correlation, composite analysis, and linear regression. The nine-year lowpass and highpass components of a variable were obtained by a Gaussian type of filter (Ma and Li 2008). The significance of the correlation between two lowpass time series was assessed using the effective number of degrees of freedom (Bretherton et al 1999, Li et al 2013). The positive and negative IPOD years are determined as the years when the IPODI is greater or less than one standard deviation. According to this criterion, the positive phase of the IPOD occurred in 1948, 1950, 1951, 1952, 1955, 1961, 1967, 1972, 1994, and 2000, whereas negative phase occurred in 1983, 1986, 1987, 1988, 1993, 1995, 1996, 1998, 2003, and 2007.

To verify the results of climatic diagnosis based on reanalysis data and observations, the NCAR Community Atmosphere Model (CAM) version 3.1 (Collins et al 2006) was used to investigate the possible effect of the IPOD events on the climate over East Asia. The model configuration used for this study has a 42-wave triangular spectral truncation (T42, about 2.8° in latitude and longitude) and 26 levels in the vertical. The SSTA in the NP and IPWP referred to as the positive (negative) IPOD events were obtained from the composite difference in observed SST (ERSST) between the positive (negative) IPOD years and the climatology. For the resolution of ERSST is 2.0° in latitude and longitude, the SSTA were interpolated onto the T42 grid. The control experiment (E1) used the monthly average SST to force the CAM 3.1 model, experiment E2 (E3) used the SSTA of the positive (negative) IPOD events, and experiment E4 (E5) imposed the SSTA of the positive IPOD events in the NP (IPWP). The five tests were all run for 28 years, with the first 3 years excluded for the model spin-up, meaning that the remaining 25 years were used in the analysis.
3. Results

3.1. The IPOD and the EASM

Figure 1(a) shows the correlations between the EASMI and JJA SST, and there are two significant areas of correlation located in the IPWP and the NP. The interannual variability of the EASM is not only negatively correlated with SST in the IPWP, but it is also positively correlated with SST in the NP (extending from the Yellow Sea to the ocean east of Japan). There appears to be no significant correlation between the EASMI and JJA SST in the eastern equatorial Pacific. The correlations using the HadISST and Kaplan SST datasets also exhibit a similar structure (figure not shown). Consequently, the ERSST data were used in the analysis that follows. Figure 1(a) shows that the cross-basin dipole pattern of SST variability in the Indian and Pacific oceans is significantly correlated with variations of the EASM. This pattern is termed the IPWP and North Pacific Ocean dipole (IPOD) (Li and Hu 2011).

Figure 1(b) shows the correlations between the IPOD and JJA SST. It can be seen that the positive IPOD phase shows warmer than average SST in the NP and cooler than average SST in the IPWP, whereas the negative IPOD phase shows the opposite anomalies. Figures 1(c) and (d) display the time series of the EASMI and the IPOD. As shown in table 1, the correlation coefficient between the raw time series of the EASMI and the IPOD is 0.63, and the correlation coefficient between the interannual (decadal) components of the EASMI and the IPOD is 0.57 (0.80). They all exceed the 99% confidence level based on the Student’s t-test, showing that the IPOD and the EASM are significantly correlated on both the interannual and decadal timescales. When the IPOD is in its positive phase, the EASM is enhanced, and vice versa. It can also be seen that there are significant and abrupt interdecadal changes in both time series around 1970; that is, the IPOD switches from a positive phase, to a negative phase, whereas the EASM switches from an enhanced phase to a weakened phase. This indicates that the IPOD is closely related to the EASM, and studying the IPOD and its underlying mechanisms may therefore help to advance our understanding of the EASM.

Table 1. Correlation coefficients between the IPOD and EASMI, PC1 and PC2 of JJA SSTA in the Indian and Pacific oceans for 1948–2012.

| IPOD (L, H) | EASMI (L, H) | PC1 (L, H) | PC2 (L, H) |
|------------|--------------|------------|------------|
| 0.63***    | 0.63***      | 0.54****   | -0.31***   |
| 0.80***    | (0.70*,      | (0.51***   | -0.30**)   |
| 0.57***    |               |            |            |

The letters L and H in brackets represent the nine-year lowpass and highpass time series, respectively, which may represent the decadal and interannual components of a variable. The bold with ***, ** and * indicate significant at the 99%, 95% and 90% confidence levels using the Student’s t-test, respectively. The significance of the correlation between two lowpass time series was accessed using the effective number of degrees of freedom (Bretherton et al 1999, Li et al 2013).
understanding of the interannual and decadal variability of the EASM.

Figure 2 shows the composite differences in SST and near-surface (sigma-995 level) winds between the positive IPOD years and the negative IPOD years. It can be seen that southwesterly winds first appear over the Indochina Peninsula in May, while warm SSTA occur in the NP near Japan; this stage may be regarded as the pre-onset phase of the IPOD. Cool SSTA in the IPWP occur in June, marking the onset of the active phase of the IPOD, while southwesterly winds blow from the IPWP to the NP, covering most of Eastern China and indicating the enhanced EASM. These patterns intensify further and peak in July–August. However, by September–October, the strength and areal extent of the SSTA become greatly reduced, and this stage may be regarded as the decaying phase of the dipole event. These significant anomalies appear around May, intensify in the following months, and weaken in September; consequently, seasonal phase locking is also an important characteristic of the IPOD.

Figure 3 shows the spatial patterns of the first two EOFs of JJA SSTA in the Indian and Pacific oceans and their corresponding PC time series. The EOF1 mode has an IPOD-like spatial pattern, and the spatial anomaly correlation coefficient (ACC) between the spatial patterns of the IPODI regressed to SST and EOF1 is 0.70, significant at the 99% confidence level. The correlation coefficient between IPODI and PC1 is 0.54, also significant at the 99% confidence level. The correlation coefficient between the IPODI and PC1 after detrending is 0.34, which is still significant at the 99% confidence level. Further analysis shows that the correlation between the IPODI and PC1 are both significant on the decadal and interannual timescales (table 1). These findings indicate that the IPOD is significantly related to the EOF1 mode.

The EOF2 mode in figure 3 is a pan-Pacific ENSO-like pattern that includes a weak Indian Ocean component, which is similar to the second rotated EOF mode of the annual mean SST (Schubert et al. 2009). PC2 is a mixture of interannual and decadal variability over the Pacific and Indian Oceans.
The correlation coefficient between the Niño 3 index and PC2 is 0.67, which is significant at the 99% confidence level, showing that ENSO is related to EOF2. Although the correlation coefficient between the IPODI and PC2 is −0.31, it is much smaller than that between the IPODI and PC1, and the correlation is only significant on the interannual timescale (table 1). The correlation between the IPODI and PC2 indicates the possible linkage between the IPOD and ENSO, which will be investigated in another article.

3.2. The atmospheric circulations over East Asia associated with the IPOD

The IPOD may be a coupled ocean-atmosphere phenomenon. Figure 4(a) shows the composite differences in circulations between the positive IPOD years and the climatology. At 500 hPa, there are large negative geopotential height anomalies over the Indo-China Peninsula, but positive geopotential height anomalies over Japan and the region near Japan. At 850 hPa, notable cyclonic circulation anomalies appear over the South China Sea, whereas anticyclonic circulation anomalies appear over Japan and the region near Japan. In the lower troposphere, the positive Pacific–Japan (P–J) pattern is characterized by the cyclonic anomalies in the South China Sea and the anticyclonic anomalies in the southeast of Japan (Nitta 1987, Kosaka and Nakamura 2006). The circulation anomalies associated with the positive IPODI phase are in agreement with the positive P–J pattern, which favours a strong EASM (Wu et al. 2009, Li et al. 2013), and vice versa.

Although local atmospheric forcing has a significant impact on the SSTA in the NP (Wu et al. 2006), the mid-latitude ocean may also influence the atmospheric circulation (Kushnir et al. 2002). In addition, the atmospheric response over the North Pacific is nonlocal (Liu and Wu 2004), so the SSTA could remotely affect atmospheric circulation over subtropical East Asia. Therefore, the CAM 3.1 model can still be used to investigate the remote forcing of the IPOD on the atmosphere anomalies over subtropical East Asia. Figure 4(b) shows the differences in the JJA 500 hPa geopotential heights between experiments E2 and E1, and figure 4(d) shows the differences in the JJA 850 hPa horizontal wind fields between experiments E2 and E1. It is found that the positive IPOD events tend to enhance convection over the South China Sea but suppress convection over Japan and the region near Japan, which is consistent with the results based on observational datasets.

To examine whether both poles of the dipole pattern are important in forcing the atmospheric response, experiments...
E4 and E5 were performed. As shown in figures 4(e) and (f), the differences in the JJA geopotential heights and horizontal wind fields between experiments E4/E5 and E1 is similar to that between E2 and E1. This similarity in the spatial distribution of the atmospheric response indicates that the two poles of the IPOD are both important in forcing the atmospheric response. It also can be found that the amplitude of the anomalies in E4 and E5 are both smaller than the anomalies in E2, indicating that the joint impact of two poles is greater than the individual effect of a single pole. Furthermore, it is evident that the SSTA in the IPWP (NP) play a greater part in the atmospheric response over the South China Sea (Okhotsk Sea), whereas the SSTA in the NP and IPWP play almost equally important parts in the atmospheric response over Japan and the region near Japan.

3.3. The IPOD, western Pacific subtropical high (WPSH) and summer precipitation over East Asia

On the basis of the above analysis, it appears that the IPOD is related to the WPSH. Figures 5(a) and (b) show the composite of the WPSH and the axis of the WPSH ridge for the positive IPOD cases, the negative IPOD cases, and the climatological mean. The axis of the WPSH ridge is defined by the line connecting the centers at each level (Li et al 2010). In summers with a positive IPOD phase, the WPSH weakens and shrinks with the axis moving northwards; in summers with a negative IPOD phase, the WPSH strengthens and extends with the axis moving southwards. The influence of the IPOD on the shift of the WPSH can be explained by the slope equation of the high ridge-axis (Li and Zhu 2010, Li et al 2010). For a high-pressure system, the slope \( k \) of the high
ridge-axis meets the following relationship:

\[ k \equiv \left( \frac{\partial y}{\partial z} \right) = -\frac{1}{T} \left( \frac{\partial T}{\partial y} \right)_p \left( \frac{\partial^2 z}{\partial y^2} \right)_p. \]  

(3)

According to equation (3), for the high ridge-axis, \( \left( \frac{\partial^2 z}{\partial y^2} \right)_p < 0 \), the axis tilts to the warm side. Therefore, the axis of the WPSH ridge tends to tilt northwards under the positive IPOD conditions, whereas the axis tends to tilt southwards under negative IPOD conditions (figure 5(b)).

Figures 5(c) and (d) show the correlations between the IPODI and precipitation. Significant positive correlations are observed in the western Pacific, and significant negative correlations are found in the middle and lower reaches of the Yangtze River, Korea, and Japan. These anomalies can also be reproduced by general climate model simulations (figure 5(e)). The SSTA in the IPWP and the NP both play important roles in affecting JJA precipitation. The precipitation anomalies over the NP in the correlation map between the IPODI and observed summer precipitation using PREC datasets (1948–2012) are less significant than those shown in figure 5(c), pointing to the importance of mid-latitude air-sea interaction on precipitation. The IPOD, which involves the tropical and mid-latitude SST signals, is more closely associated with the monsoonal precipitation anomalies over East Asia than the IPWP.

To further examine the relationship between the IPOD and precipitation, the partial correlations between the IPODI and precipitation with the WPSH index (WPSHI) removed are shown in figure 5(f). The WPSHI used here is the domain-
averaged geopotential height over the region (130°–180°E, 12.5°–22.5°N) at 500 hPa, which represents the southwards shift of the WPSH. The partial correlations suggest that the correlations decrease in significance after removing the WPSHI signal. This implies that the IPOD influences JJA rainfall in the middle and lower reaches of the Yangtze River valley via the bridge of the WPSH. The precipitation anomalies over East Asia are the result of large-scale motion related to the WPSH rather than of the thermal state of the atmosphere (Cha et al. 2011).

Moreover, the cloud-cover anomalies (figure not shown) act as a positive feedback for the SSTA. In summers with a positive IPOD phase, with the axis of the WPSH ridge moving northwards, negative anomalies in cloud cover appear over the northern flank of the WPSH (near the NP with warm SSTA), whereas positive anomalies in cloud cover appear over the southern flank of the WPSH (near the IPWP with cool SSTA). Negative (positive) anomalies in cloud cover increase (decrease) the downwards-directed shortwave radiation over the NP (IPWP) and make it warmer (cooler), thus introducing a positive feedback mechanism. This positive feedback mechanism is also effective for negative IPOD events.

4. Summary

In this paper, we have demonstrated a cross-basin dipole pattern of SSTA–IPOD, which appears to be a significant pattern of boreal summer SST in the Indian and Pacific oceans. The IPOD is characterized by positive (negative) SSTA in the NP, and negative (positive) SSTA in the IPWP. The IPOD has strong seasonal phase locking: it appears around May, intensifies in the following months, and then weakens in September. The IPOD exhibits a considerable correlation with variations of the EASM. In summers, a positive (negative) IPOD phase shows the weakened and shrunk (strengthened and enlarged) WPSH with the axis of the WPSH ridge moving northwards (southwards), which favours an intensified (weakened) EASM and a decrease (increase) in summer rainfall in the Yangtze River valley.

In contrast to previous studies (Huang and Sun 1992, Tan et al. 2009), we highlight here the joint impact of the IPWP and the NP on variations of the EASM. The correlation between the IPODI and the EASMI is more robust than the correlation between the IPWP or NP index and the EASMI. The tropical SST and mid-latitude SST signals are both involved in the IPOD, thus the IPOD is more closely associated with the monsoonal precipitation than the IPWP, which may provide a new perspective to understand the EASM and rainfall over East Asia.

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