# A Recent Shift in the Monsoon Centers Associated with the Tropospheric Biennial Oscillation

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(Manuscript received 24 May 2017, in final form 2 October 2017)

ABSTRACT

The tropospheric biennial oscillation (TBO) is conventionally considered to involve transitions between the Indian and Australian summer monsoons and the interactions between these two monsoons and the underlying Indo-Pacific Oceans. Here it is shown that, since the early 1990s, the TBO has evolved to mainly involve the transitions between the western North Pacific (WNP) and Australian monsoons. In this framework, the WNP monsoon replaces the Indian monsoon as the active Northern Hemisphere TBO monsoon center during recent decades. This change is found to be caused by stronger Pacific–Atlantic coupling and an increased influence of the tropical Atlantic Ocean on the Indian and WNP monsoons. The increased Atlantic Ocean influence damps the Pacific Ocean influence on the Indian summer monsoon (leading to a decrease in its variability) but amplifies the Pacific Ocean influence on the WNP summer monsoon (leading to an increase in its variability). These results suggest that the Pacific–Atlantic interactions have become more important to the TBO dynamics during recent decades.

## 1. Introduction

The tropical Indo-Pacific sector encompasses two of the most active monsoons in our climate system: the Indian monsoon located in the northern part of the sector and the Australian monsoon in the southern part. The tropospheric biennial oscillation (TBO) is a major variation of the Indian–Australian monsoon system, with years of strong summer rainfall more likely to be followed by years of weak rainfall and vice versa (Meehl 1987, 1994, 1997; Meehl and Arblaster 2002a). Previous studies have already revealed that the biennial variations in the Indian summer monsoon and the Australian summer monsoon tend to be related to each other (e.g., Matsumoto 1992; Yu et al. 2003; Hung et al. 2004). This relation is manifested as an in-phase transition from the Indian summer monsoon to the Australian summer monsoon (i.e., a strong Indian monsoon in boreal summer is followed by a strong Australian summer monsoon in the following austral summer and vice versa) and an out-of-phase transition from the Australian summer monsoon back to the Indian summer monsoon in the next year (i.e., a strong Australian summer monsoon is followed by a weak Indian summer monsoon and vice versa). These in-phase and out-of-phase transitions between the Indian and Australian monsoons are two key features of the TBO (Yu et al. 2003) and are referred to as the biennial monsoon transitions in this study.

Much effort has been expended trying to understand TBO dynamics and the associated monsoon transitions, resulting in substantial advances in our understanding of this important climate phenomenon. It is generally agreed that the interactions between the monsoons and the tropical Indian and Pacific Oceans play a central role in the TBO dynamics (e.g., Nicholls 1978, 1979, 1984; Meehl 1987, 1993; Clarke et al. 1998; Chang and Li 2000;

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DOI: 10.1175/JCLI-D-17-0349.1

Meehl et al. 2003; Li et al. 2006; Zheng et al. 2008). This is based on the observational finding that the biennial variations in monsoon rainfall are associated with significant variations at similar time scales in the sea surface temperatures (SSTs) of the tropical Indian and Pacific Oceans (Rasmusson and Carpenter 1983; Meehl 1987; Kiladis and van Loon 1988; Ropelewski et al. 1992; Lau and Yang 1996; Meehl and Arblaster 2002b; Yu et al. 2003). The associated SST anomalies are characterized by an El Niño–Southern Oscillation (ENSO)-type pattern in the Pacific Ocean and a basin-scale warming or cooling in the Indian Ocean, or an east–west dipole along the equatorial Indian Ocean that is known as the Indian Ocean dipole (IOD) or Indian Ocean zonal mode (Webster et al. 1999; Saji et al. 1999).

Meehl (1993) proposed that the in-phase monsoon transition is produced by local monsoon–ocean interactions in the Indian and western Pacific Oceans associated with the southeastward migration of convection during the annual cycle. The Indian summer monsoon winds force SST anomalies around Australian (via ocean upwelling/downwelling and mixing) that persist into the following boreal winter and affect the strength of the Australian summer monsoon. For the out-of-phase monsoon transition, the Australian summer monsoon wind forces oceanic waves in the western Pacific Ocean that propagate into the eastern Pacific to influence SST anomalies there. These SST anomalies later influence the strength of the Indian summer monsoon through the large-scale east–west atmospheric circulation. In this theory, it is the interaction between the monsoons and the eastern Pacific Ocean that provides the phase-reversal mechanism for the TBO. Chang and Li (2000) also related the monsoon–ocean interactions to the TBO but did not emphasize the interaction between the Australian monsoon and the eastern Pacific. Instead, they suggested that a strong Australian summer monsoon enhances the Walker circulation over the Indian Ocean and produces strong westerly anomalies in the central Indian Ocean. These wind anomalies help to cool Indian Ocean SSTs through processes such as wind and latent and sensible heat fluxes. The cold SST anomalies persist into the following boreal summer to produce a weak Indian summer monsoon by reducing the moisture available. Regardless of the differences, these theories emphasize the interactions between Indian–Australian summer monsoons and the Indo-Pacific Ocean to explain how the TBO and the monsoon transitions can be produced.

However, in addition to the Indian monsoon, there are other subcomponents in the Asian monsoon system including the East Asian monsoon and the western North Pacific (WNP) monsoon (e.g., Wang et al. 2001). Lee et al. (2014) examined the interdecadal changes in global monsoon variability and noticed that the interannual variability of the WNP monsoon has increased and become a dominant component of the Asian summer monsoon variability after 1993. They found the WNP summer monsoon variability after 1993 tends to be related to central-Pacific (CP)-type ENSO events (Yu and Kao 2007; Kao and Yu 2009), which have also occurred more frequently since the early 1990s (Yu et al. 2012, 2015). These recent changes in ENSO and the monsoons may have modified the characteristics of the TBO in recent decades, particularly the associated biennial monsoon transitions. Furthermore, recent studies on the early-1990s change in ENSO types have suggested that this change may be related to the increased influences of the Atlantic Ocean on the Pacific climate (Yu et al. 2015; Lyu et al. 2017; Wang et al. 2017), which is associated with a change of the Atlantic multidecadal oscillation (AMO; Schlesinger and Ramankutty 1994; Kerr 2000) from a negative to a positive phase around that time. There is also increasing evidence to support a significant influence of the tropical Atlantic SST on the variability of the Indian summer monsoon (Kucharski et al. 2007, 2008; Cash et al. 2013) and Pacific climate variability (Rodríguez-Fonseca et al. 2009; Keenlyside et al. 2013; Hong et al. 2014; Yu et al. 2015; Li et al. 2016). Therefore, it is possible that the Atlantic Ocean may have become important to the TBO dynamics. It is necessary to know how these Atlantic influences may have modified the biennial monsoon transitions of the TBO in recent decades.

In this study, we perform statistical analyses using observations and reanalysis products to examine the decadal changes in the TBO and its biennial monsoon transitions since 1948. This paper is organized as follows. The data used and the analysis procedures are described in section 2. The changes in the biennial monsoon transitions in the TBO since the early 1990s are presented in section 3. Section 4 illustrates the influences of tropical Atlantic SST anomalies on the recent change in the biennial monsoon transitions. Conclusions and discussion are given in the final section (section 5).

## 2. Data and methods

The SST data used here are the monthly extended reconstructed SST (ERSST) analyses (Smith et al. 2008). The atmospheric fields are from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) monthly reanalysis (Kalnay et al. 1996) that begin in 1948. The monthly rainfall data used is NOAA's Precipitation Reconstruction (PREC) (Chen et al. 2002), which also begins in 1948 and was obtained from http://www.esrl.noaa.gov/psd/.

The following dynamic monsoon indices are used in the analysis. 1) The WNP monsoon index is defined, following Wang and Fan (1999), as the difference in 850-hPa zonal winds between a southern region (5°–15°N, 100°–130°E) and a northern region (20°–30°N, 110°–140°E). 2) The Indian summer monsoon index is defined as the difference in the 850-hPa zonal winds between a southern region (5°–15°N, 40°–80°E) and a northern region (20°–30°N, 70°–90°E) (Wang et al. 2001). 3) The Australian monsoon index is defined as the 850-hPa zonal wind averaged over the area (5°–15°S, 110°–130°E), following Kajikawa et al. (2010). These dynamic monsoon indices based on 850-hPa winds have been shown to be consistent with monsoon rainfall variability (Wang et al. 2001; Kajikawa et al. 2010). Following Kwon et al. (2005), an East Asian (EA) summer rainfall index is defined as the June–August precipitation anomaly averaged over the area 30°–50°N, 115°–150°E.

The Niño-3.4 index is used to represent ENSO intensity and is defined as the SST anomalies averaged over (5°S–5°N, 170°–120°W). The Indian Ocean Dipole Mode Index (DMI) is defined as the difference between SST anomalies in the western (10°S–10°N, 50°–70°E) and eastern (10°S–0°, 90°–110°E) equatorial Indian Ocean (Saji et al. 1999). Following Kucharski et al. (2008), a tropical South Atlantic SST index is defined as the SST anomalies averaged over 20°S–0°, 30°W–10°E. Following Hong et al. (2014), a tropical North Atlantic SST index is defined as the SST anomalies averaged over 0°–20°N, 80°W–25°E.

Anomalies are calculated by removing long-term trend first and then the mean seasonal cycle for the period 1971–2000. The interannual variability was obtained by applying a 7-yr high-pass filter to the anomalies. The intensity of the interannual variability during a period is measured by the standard deviation of the interannual time series during that period.

We classified each year during the analysis period to be a strong, weak, or normal monsoon year based on whether the anomalies in the summer monsoon index during that year are, respectively, above, below, or close to a threshold value. Following Wu and Kirtman (2004), the 0.43 standard deviation was used as the threshold value to ensure that the three categories (i.e., strong, weak, and normal monsoon years) have nearly equal numbers of years. We have also repeated the classification using three other threshold values (i.e., 0.40, 0.45, and 0.50 standard deviations) and found similar results. Thus, the results reported in this study are not particularly sensitive to the threshold value used in the classification. Considering that the biennial monsoon transitions comprise both in-phase and out-of-phase transitions between monsoons, we then selected the in-phase and out-of-phase monsoon transition years for composite analysis. An "in-phase" transition case is identified if a strong (weak) summer monsoon in one hemisphere (such as the Indian and WNP monsoon in the Northern Hemisphere or the Australian monsoon in the Southern Hemisphere) is followed by a strong (weak) summer monsoon in the other hemisphere. Similarly, an "out-of-phase" transition case is identified if a strong (weak) summer monsoon in one hemisphere is followed by a weak (strong) summer monsoon in the other hemisphere. For the sake of comparison in the composite analysis, if a transition involves a strong Australian summer monsoon, we refer to it as a "positive in-phase" or "positive out-of-phase" transition case. If the transition involves weak Australian summer monsoon, that transition case is referred to as a "negative in-phase" or "negative out-of-phase" transition case. The composites for the in-phase transition were then constructed as the means of the positive in-phase transition cases minus the means of the negative in-phase transition cases. Similarly, the composites for the out-of-phase transitions were constructed as the means of the positive out-of-phase transition cases minus the means of the negative out-of-phase transition cases. These composites are designed to reveal the atmospheric and oceanic conditions involved in the transition from a strong or weak Indian/WNP summer monsoon to a strong Australian summer monsoon or from a strong Australian summer monsoon back to a strong or weak Indian/WNP summer monsoon.

The analysis methods used in this study include correlation and composite analyses. We determine the statistical significance levels based on the two-tailed $P$ values using a Student's $t$ test.

## 3. Decadal changes in the biennial monsoon transitions of the TBO

To elucidate possible decadal changes in the biennial monsoon transitions, we first performed a 21-yr sliding correlation analysis between the Indian monsoon index during boreal summer [June–August (JJA)] and the Australian monsoon index during the following austral summer [December–February (DJF)] for the period 1948–2016 (Fig. 1a). As expected, the correlation coefficients are positive throughout the analysis period, which indicates that in-phase transitions from a strong (weak) Indian summer monsoon to a strong (weak) Australian summer monsoon dominate the time series. However, these in-phase transitions are statistically

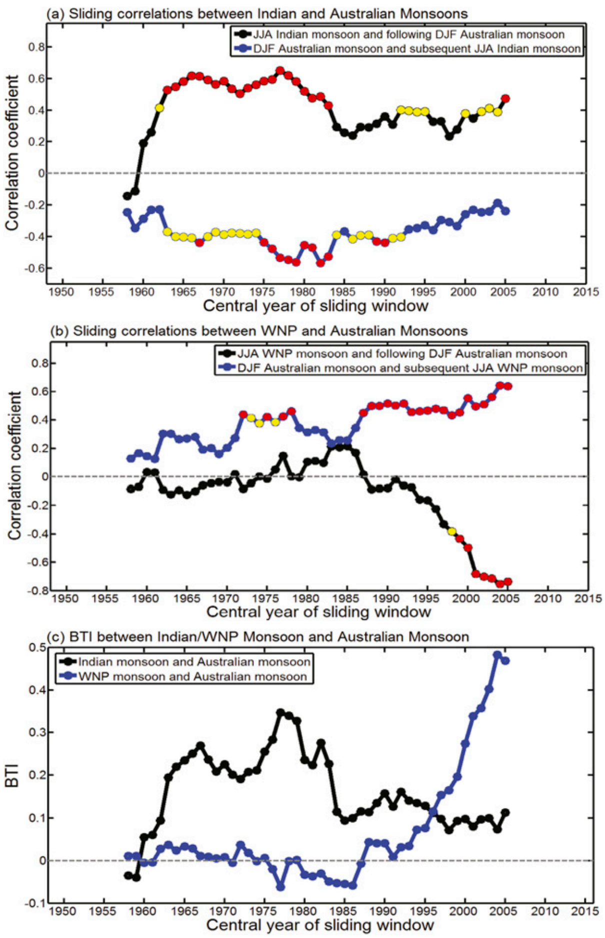


FIG. 1. (a) The 21-yr sliding correlation coefficients (e.g., the correlation coefficient in 2000 represents the period 1990–2010) between the JJA Indian monsoon index and the DJF Australian monsoon index for the period 1948–2016. (b) As in (a), but for the JJA WNP monsoon index and the DJF Australian monsoon index. The red (yellow) dots in (a) and (b) represent correlations that are significant at the 95% (90%) confidence level. (c) The BTI between the JJA Indian/WNP monsoon and the DJF Australian monsoon.

significant at the 95% confidence level only during the early 1960s to the early 1980s, having weakened remarkably afterward. A similar sliding analysis between the DJF Australian monsoon index and the subsequent JJA Indian monsoon index shows negative correlation coefficients throughout the analysis period (Fig. 1a), which indicates the expected out-of-phase transition from the Australian summer monsoon to the following Indian summer monsoon. The negative correlations are statistically significant at the 90% or 95% confidence levels during the early 1960s to the early 1990s, becoming insignificant after the early 1990s. This analysis indicates that the biennial monsoon transitions between the Indian and Australian monsoons were strong during the 1960s to 1980s but have weakened in recent two decades.

We then performed the same 21-yr sliding correlation analyses between the boreal summer WNP monsoon and the austral summer Australian monsoon to examine the decadal changes in their relationships. The analysis results (Fig. 1b) show two main differences from the relationships found between Indian and Australian summer monsoons. First, the transition from the WNP summer monsoon to the following Australian summer monsoon is out of phase (i.e., negative values in their correlation coefficients) whereas the transition from the Australian summer monsoon to the subsequent WNP summer monsoon is in phase (i.e., positive correlation coefficients). These monsoon transitions are exactly the opposite of the transitions between the Indian and Australian summer monsoons. This is related to the fact that the interannual variations of the WNP summer monsoon tend to be out of phase with the variations of Indian summer monsoon, which has been documented by Gu et al. (2010). They attributed this phenomenon to the different relationships between these two monsoons and El Niño. During the summer of a developing El Niño, the anomalously warm eastern Pacific tends to induce cyclonic wind shear over the WNP that strengthens the WNP monsoon but suppresses convection over the Indian monsoon region. During the subsequent summer, as El Niño decays, a weak WNP monsoon tends to occur due to the persistence of a local anomalous anticyclone, whereas a strong Indian monsoon tends to occur due to the ENSO-induced basinwide Indian Ocean warming. The other main feature to note in Fig. 1b is that the correlations between the WNP and Australian summer monsoons become stronger after the early 1990s, which is the period when the correlations between Indian and Australian summer monsoons weaken (see Fig. 1a). These sliding correlation analyses suggest that the WNP monsoon has replaced the Indian monsoon in the monsoon biennial transitions associated with the TBO after the early 1990s.

To quantitatively assess the strength of the biennial monsoon transitions, we define a biennial transition index (BTI) as follows:

$$\mathrm{BTI} = (-1) \times \mathrm{Cor1} \times \mathrm{Cor2}. \tag{1}$$

Here, Cor1 is the 21-yr sliding correlation coefficient between the JJA Indian or WNP monsoon index and the subsequent DJF Australian monsoon index, while Cor2 is the sliding correlation coefficient between the DJF Australian monsoon index and the subsequent JJA Indian or WNP monsoon index. The factor −1 is included in Eq. (1) to reflect the fact that the biennial monsoon transitions are the combination of an in-phase transition (i.e., a positive correlation coefficient) and an

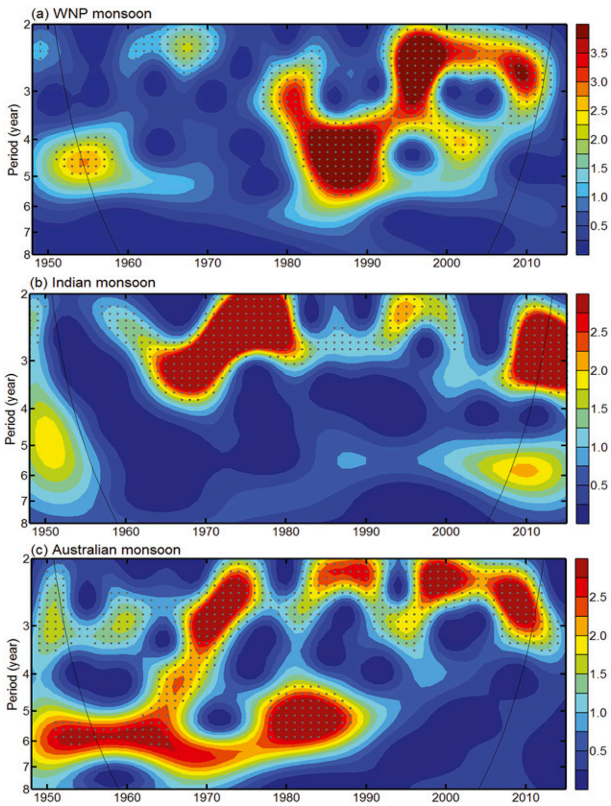


FIG. 2. Wavelet power spectrum of the summer monsoon indices for (a) WNP monsoon, (b) Indian monsoon, and (c) Australian monsoon. The regions exceeding the 95% confidence level against red noise are dotted.

out-of-phase transition (i.e., a negative correlation coefficient). Thus, the larger the BTI value, the stronger the biennial transitions between the two monsoons. Figure 1c shows the BTI values calculated from the sliding correlations between the Indian/WNP summer monsoons and the Australian summer monsoon. The most obvious feature in the figure is that the biennial transitions were stronger between the Indian and Australian summer monsoons before the early 1990s but stronger between the WNP and Australian summer monsoons afterward. We also use a regime shift detection method developed by Rodionov (2004) to confirm that the shift really occurs during the early 1990s. This detection method uses a regime shift index (RSI) to objectively determine the time when a time series undergoes a regime shift. Previous studies have also found an early-1990s shift in the WNP monsoon (e.g., Kwon et al. 2005; Lee et al. 2014) and the WNP subtropical high (e.g., Sui et al. 2007; Paek et al. 2016).

Since the EA monsoon is another important subcomponent of the Asian monsoon system, we also examined the biennial relationship between the JJA EA monsoon index and the Australian monsoon index. We find that the correlation coefficient is weak and mostly insignificant throughout the analysis period (not shown). The BTI values calculated for the EA–Australian monsoon transitions were much smaller than those calculated for the Indian–Australian and WNP–Australian monsoon transitions (not shown). These results suggest that there were no significant biennial monsoon transitions between the EA and Australian monsoons during the analysis period. This may be due to the fact that the EA monsoon is located at higher latitudes than the other subcomponents of the Asian monsoon system and receives more extratropical influences (such as those associated with midlatitude jet stream variations) than the other monsoon components.

According to these results, the TBO has evolved from mainly involving biennial monsoon transitions between the Indian and Australian monsoons to mainly involving biennial monsoon transitions between the WNP and Australian monsoons since the early 1990s. Since the early 1990s, the relations between the biennial tendencies in the Indian monsoon and the Australian monsoon have weakened remarkably; on the contrary, the relations between biennial tendencies in the WNP monsoon and the Australian monsoon have been greatly enhanced. To further demonstrate this change, we show in Fig. 2a wavelet analysis of the indices of the WNP, Indian, and Australian summer monsoons. The wavelet power spectrum in Fig. 2a shows that the quasi-biennial (QB) band (e.g., 2–3 yr) of the WNP monsoon variability significantly increased after the early 1990s, which is consistent with the observed fact that the leading periodicities of the summer western Pacific subtropical high also shifted from the low-frequency (LF) band (e.g., 3–7 yr) to the QB band in

TABLE 1. Comparison of the correlations between the Indian/WNP and Australian monsoon indices during P1 and P2. Correlations that are significant at the 95% confidence level are shown in boldface.

| | P1 (1967–87) | P2 (1994–2014) |
|---|---|---|
| JJA Indian monsoon and following DJF Australian monsoon | **0.65** | 0.39 |
| DJF Australian monsoon and subsequent JJA Indian monsoon | **−0.54** | −0.19 |
| JJA WNP monsoon and following DJF Australian monsoon | 0.15 | **−0.75** |
| DJF Australian monsoon and subsequent JJA WNP monsoon | 0.38 | **0.64** |

TABLE 2. Selected cases of monsoon transitions during P1 (I-to-A and A-to-I transitions) and P2 (W-to-A and A-to-W transitions). The years are shown are the years of the JJA Indian monsoon or the JJA WNP monsoon involved in the transitions.

| | Positive | Negative |
|---|---|---|
| I-to-A (P1) | 1967, 1970, 1973, 1975, 1980 | 1969, 1972, 1974, 1987 |
| A-to-I (P1) | 1974, 1981, 1987 | 1970, 1973, 1975, 1978, 1983 |
| W-to-A (P2) | 1996, 1998, 2003, 2007, 2008, 2010 | 1997, 2004, 2006, 2009, 2012 |
| A-to-W (P2) | 1997, 2001, 2004, 2009 | 1998, 2003, 2007, 2010 |

the early 1990s (Sui et al. 2007; Paek et al. 2016). However, the QB band of the Indian monsoon variability weakened substantially after the 1980s (Fig. 2b). For the Australian monsoon, significant power in the QB band is found during nearly the entire analysis period (Fig. 2c). No significant changes before and after the early 1990s are found for the QB band of the Australian monsoon, although the LF band of the Australian monsoon was found to be suppressed greatly after the early 1990s.

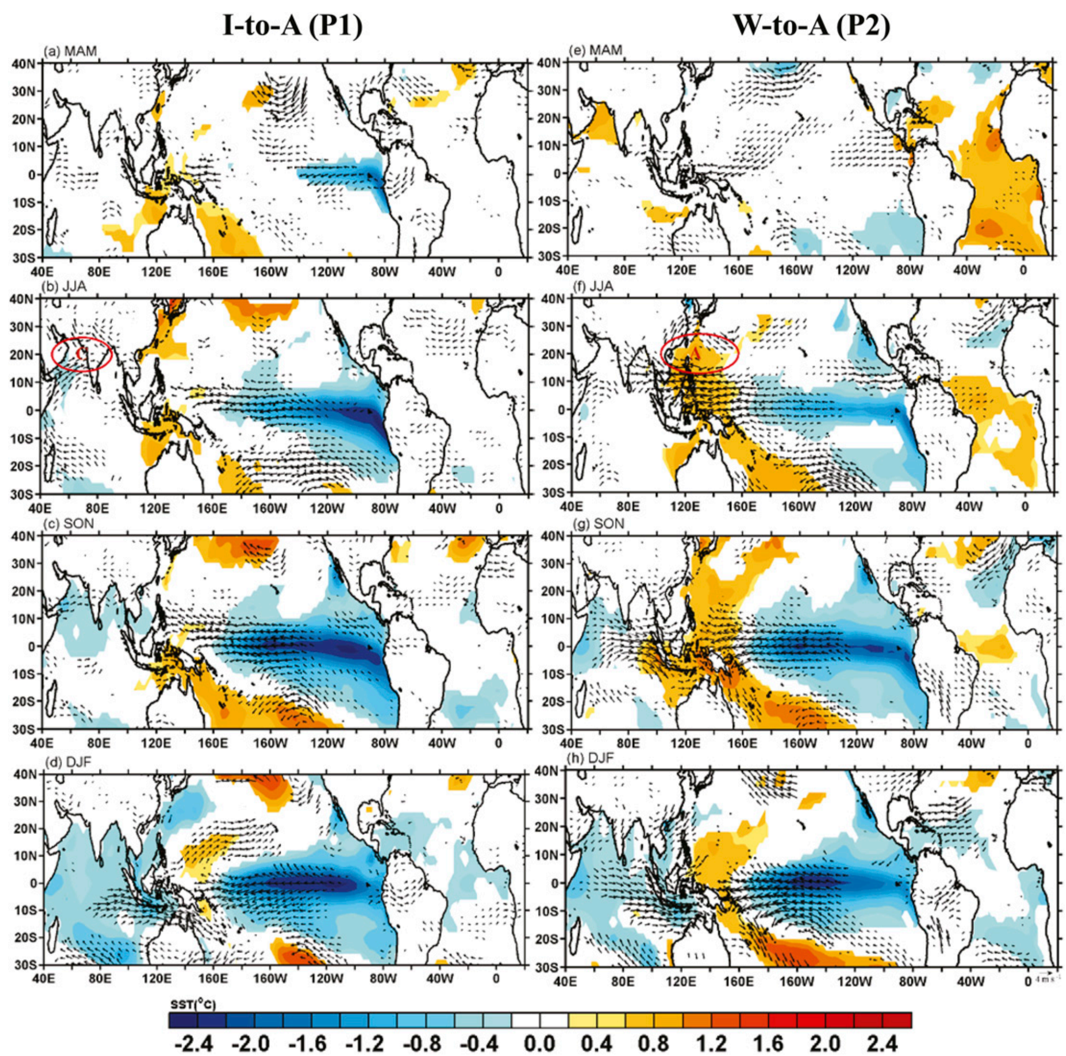


FIG. 3. Differences in SSTs (shading) and 850-hPa winds (vectors) between the so-called positive and negative monsoon composites from MAM to the following DJF for (a)–(d) the I-to-A transitions during P1 and (e)–(h) the W-to-A transitions during P2. Only the values at the 90% confidence level or higher are shown. The red circles in (b) and (f) represent the summertime low-level circulation anomalies over the Indian/WNP monsoon region, with A (C) denoting anticyclone (cyclone).

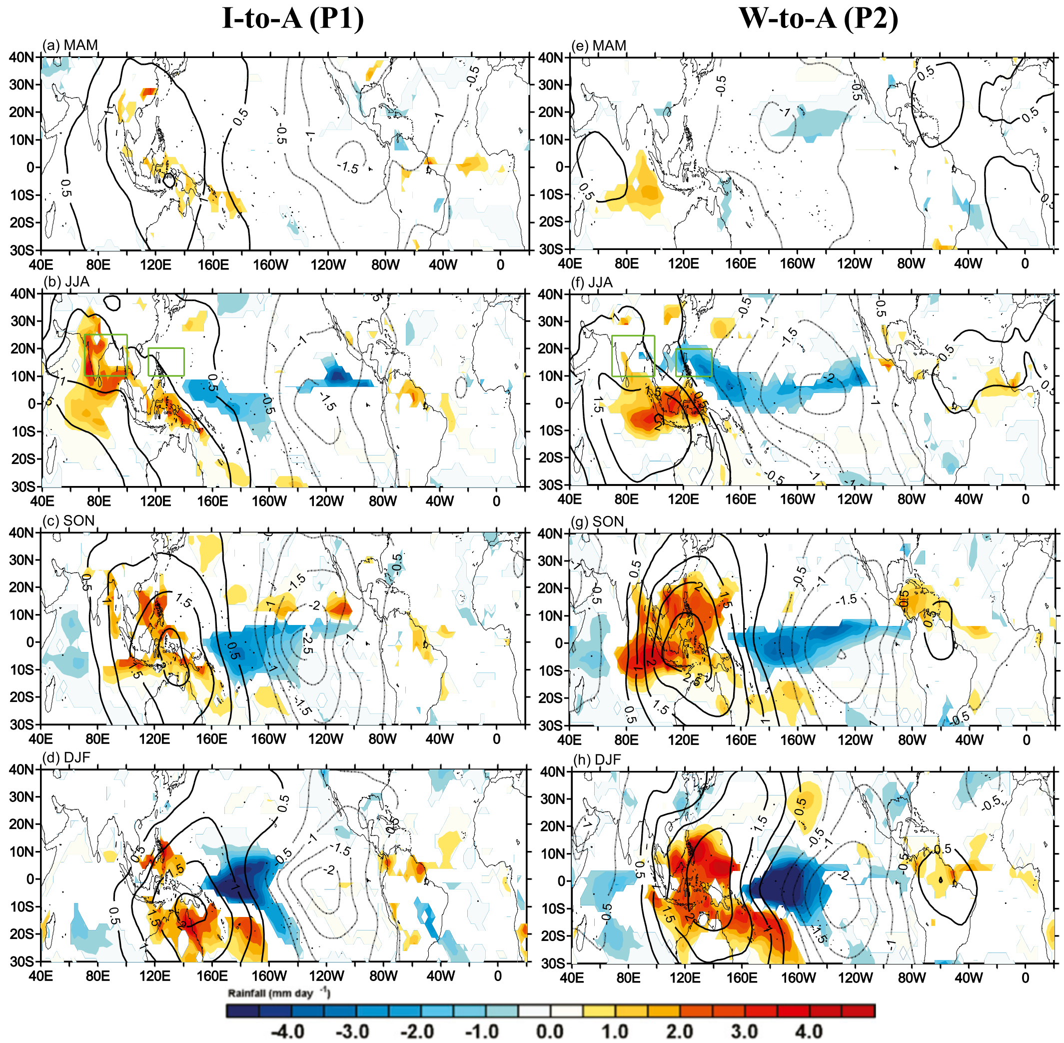


FIG. 4. As in Fig. 3, but for the rainfall (shading) and 850-hPa velocity potential (contours, $10^6\,m^2\,s^{-1}$, solid for positive and dashed for negative values). The green boxes in (b) and (f) encompass the core regions of the Indian and WNP monsoon according to Wang et al. (2001).

## 4. Influences of Atlantic SSTs on the recent TBO changes

Why are different biennial monsoon transitions observed in the TBO before and after the early 1990s? The biennial monsoon transitions before and after the early 1990s are likely to be accompanied by different SST anomalies, whose interactions with the monsoons should be considered to explain the TBO changes (e.g., Meehl 1987, 1993; Clarke et al. 1998; Chang and Li 2000; Meehl et al. 2003; Li et al. 2006).

To identify the possible differences in SST anomalies involved in the biennial monsoon transitions, we composited SST anomalies for the biennial transitions between the Indian and Australian summer monsoons for the period before the early 1990s (1967–87, referred to as P1 hereafter) and between the WNP and Australian summer monsoons in a period after the early 1990s (1994–2014, referred to as P2 hereafter). These two periods correspond, respectively, to the 21-yr periods in Fig. 1c that produce the maximum BTI values for the Indian–Australian monsoon transitions (centered around 1977) and the WNP–Australian monsoon transitions (centered around 2004). During P1, the BTI value for the Indian–Australian monsoon transitions is 0.347 but is only −0.062 for the WNP–Australian monsoon transitions. In contrast, the BTI value during P2 is only 0.073 for the Indian–Australian monsoon transitions but 0.483 for the WNP–Australian monsoon transitions. The correlation coefficients (Table 1) confirm

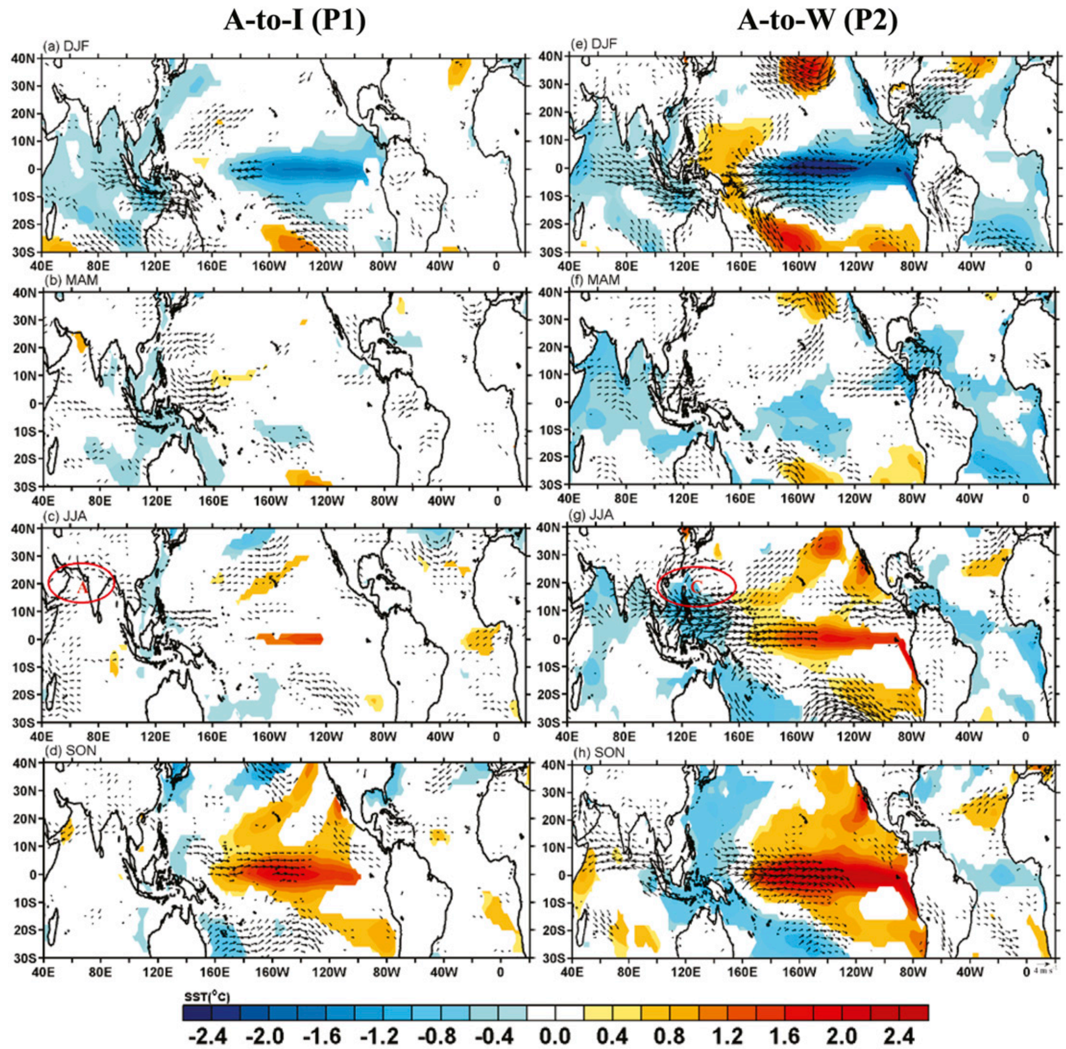


FIG. 5. Differences in SSTs (shading) and 850-hPa winds (vectors) between the positive and negative monsoon composites from DJF to the following SON for the (a)–(d) A-to-I transitions during P1 and (e)–(h) A-to-W transitions during P2. Only the values at the 90% confidence level or higher are shown. The red circles in (c) and (g) represent the summertime low-level circulation anomalies over the Indian/WNP monsoon region, with A (C) denoting anticyclone (cyclone).

that the in-phase and out-of-phase transitions are both strong and significant between the Indian and Australian monsoons during P1 and between the WNP and Australian monsoons during P2. During P1, 9 cases were selected for the in-phase transition from the Indian to Australian summer monsoons (i.e., the I-to-A transition) and 8 cases were selected for the out-of-phase transition from the Australian to Indian summer monsoons (i.e., the A-to-I transition). During P2, 11 cases were selected for the out-of-phase transition from the WNP to Australian summer monsoons (i.e., the W-to-A transition) and 8 cases were selected for the in-phase transition from the Australian to WNP summer monsoons (i.e., the A-to-W transition). The cases selected for the composite are listed in Table 2.

As mentioned previously, the biennial monsoon transitions during P1 mainly consist of an in-phase I-to-A transition and an out-of-phase A-to-I transition. During the I-to-A transitions (Figs. 3a–d), significant SST anomalies in the composite are located mostly in the Pacific and Indian Oceans but not in the Atlantic Ocean. The anomalies are characterized by a developing La Niña event, with the cold anomalies emerging in the central-to-eastern Pacific during boreal spring, spreading westward during boreal summer and autumn, and peaking in the central Pacific during austral summer. In

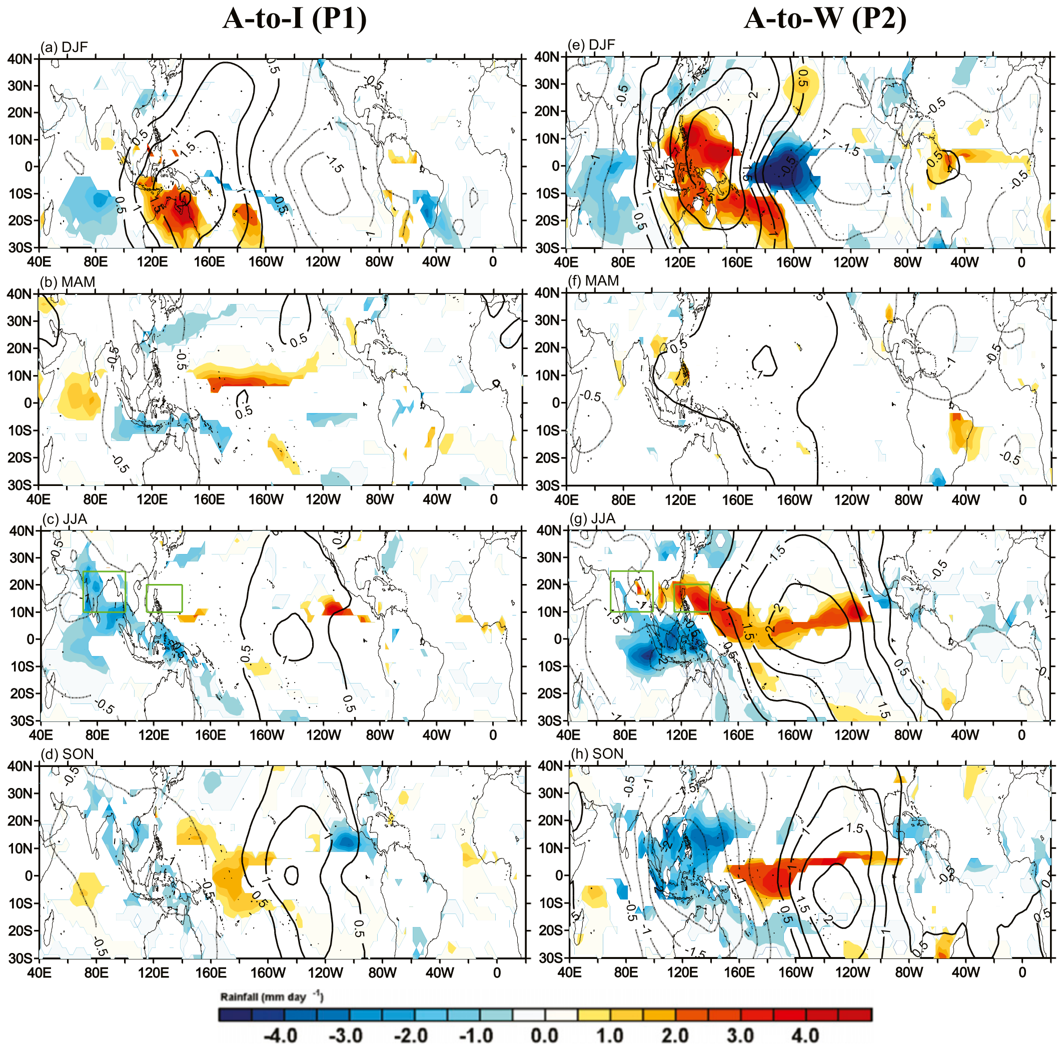


FIG. 6. As in Fig. 5, but for the rainfall (shading) and 850-hPa velocity potential (contours, $10^6\ m^2\ s^{-1}$, solid for positive and dashed for negative values). The green boxes in (b) and (f) encompass the core regions of the Indian and WNP monsoon according to Wang et al. (2001).

the Indian Ocean, the composite SST anomalies resemble the typical Indian Ocean response to a La Niña event: a negative-phase IOD appears during boreal summer and autumn [September–November (SON)] that later evolves into a basinwide cooling in boreal winter (i.e., austral summer). These composite anomalies are consistent with those found in previous studies suggesting that the TBO is accompanied by El Niño–like (La Niña–like) anomalies in the Pacific and the IOD and basinwide warming (cooling) in the Indian Ocean (Loschnigg et al. 2003; Meehl et al. 2003). A developing La Niña is known to be capable of strengthening the Walker circulation to intensify the Indian monsoon during JJA and the Australian monsoon in DJF (e.g., Chang and Li 2000; Yu et al. 2003; Gu et al. 2010). Therefore, the in-phase I-to-A transition can be maintained by the developing La Niña. The cold anomalies in the Indian Ocean may also contribute to the strong Australian summer monsoon (e.g., Taschetto et al. 2011). The composites in atmospheric fields reveal a strong summer Indian monsoon during this transition phase that is characterized by an anomalous low-level cyclonic circulation (Fig. 3b) and enhanced rainfall (Fig. 4b) around the Indian peninsula together with anomalous descending motion (represented by negative values of anomalous 850-hPa velocity potential) over the eastern Pacific and anomalous ascending over the Indian monsoon region (Fig. 4b).

As for the out-of-phase A-to-I transitions during P1 (Figs. 5a–d), the composite SST anomalies are most prominent in the Pacific Ocean. The anomalies are characterized by a phase reversal from a decaying La Niña in DJF to an El Niño in JJA that continues to develop thereafter. The decaying La Niña in DJF supports a strong Australian summer monsoon, whereas the developing El Niño in JJA results in a weak Indian summer monsoon. Therefore, the out-of-phase A-to-I transition can be explained reasonably by this SST evolution pattern. Cold SST anomalies in the Indian Ocean are large only during DJF when they may strengthen the Australian summer monsoon through a Gill-type atmospheric response and by inducing anomalous ascending motions over Australia (Taschetto et al. 2011; Cai and van Rensch 2013). The composited atmospheric fields are also characterized by a weakened summer Indian monsoon, with an anomalous low-level anticyclonic circulation (Fig. 5c), suppressed rainfall (Fig. 6c), and anomalous descending motion over the Indian monsoon region (Fig. 6c). Therefore, the biennial monsoon transitions between the Indian and Australian summer monsoons during P1 can be established by the biennial component of the ENSO with some contributions from Indian Ocean SSTs. SST anomalies in the Atlantic Ocean are not involved in these Indian–Australian monsoon transitions.

As for the biennial monsoon transition during P2, it consists of an out-of-phase WNP to Australian (W-to-A) transition and an in-phase Australian to WNP (A-to-W) transition. The SST anomalies composited for the W-to-A transition (Figs. 3e–h) are characterized by a developing La Niña in the Pacific, an IOD and a basin-wide cooling in the Indian Ocean, and a warming in the tropical Atlantic. The SST anomalies in the Pacific and Indian sectors are mostly similar to the anomalies composited for the I-to-A transition during P1 but are very different in the Atlantic sector. The 850-hPa velocity potential anomalies (Figs. 4e–h) indicate that there are significant differences between the ascending and descending anomalies associated with the Walker circulations during P2 and P1. Anomalous ascent occurred over the tropical Atlantic during the summer of P2 (Fig. 4f) that was not observed during P1 (Fig. 4b). The ascending anomalies above the Indian region during P1 are displaced southeastward during P2 (Fig. 4f) and as a result the rainfall anomalies in the Indian monsoon region became weaker (Fig. 4f). In the Pacific, the descending anomalies extend farther westward during P2 (Fig. 4f), inducing low-level anticyclonic circulation anomalies (Fig. 3f) that suppress the WNP monsoon rainfall (Fig. 4f).

Previous studies (Kucharski et al. 2007, 2008; Wang et al. 2009; Cash et al. 2013) have shown that warm tropical Atlantic SST anomalies can excite atmospheric

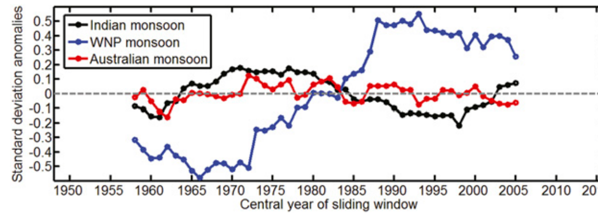


FIG. 7. Time series of anomalies in the intensity of the interannual variations in the three monsoon indices. The intensity of the interannual variations is defined as the 21-yr running standard deviation of the interannual monsoon indices. The anomalies are then calculated by subtracting the long-term mean intensity of the interannual variations during the entire analysis period.

wave responses over the Indian Ocean that can weaken the Indian summer monsoon. This weakening effect from the tropical Atlantic can cancel out the strengthening effect produced by the developing La Niña. This cancelling has been verified in numerical modeling experiments by Kucharski et al. (2007). This cancelling should also reduce the interannual variability of Indian monsoon during P2, which is confirmed by a running variance analysis (see Fig. 7). Thus, the increased influence of the tropical Atlantic SST anomalies seems to be the reason why similar Indo-Pacific SST anomalies can support a strong I-to-A transition during P1 but not during P2. While they cancel out the La Niña influences on the Indian summer monsoon variability, the tropical Atlantic SST anomalies can at the same time reinforce the La Niña influence on the WNP summer monsoon variability. Hong et al. (2014) have shown that warm tropical Atlantic SST anomalies can intensify the WNP subtropical high and weaken the WNP summer monsoon via an anomalous zonally overturning circulation, which ascends over the tropical Atlantic and descends over the equatorial central Pacific. This anomalous descending motion can then excite a low-level anticyclonic anomaly to the west and therefore weaken the WNP monsoon. Also, it is known that negative SST anomalies in the tropical central Pacific associated with the La Niña can weaken the WNP monsoon during boreal summer by inducing anticyclonic circulation anomalies through a Gill-type response (Gill 1980; Gu et al. 2010). Thus, the tropical Atlantic SSTs and Pacific La Niña reinforce each other to produce large negative anomalies in the WNP summer monsoon. Because of this reinforcing effect, the intensity of the interannual variability in the WNP monsoon was observed to increase markedly (Fig. 7) during the P2 period. During DJF, Figs. 3g and 3h show that the tropical Atlantic SST anomalies decay while the Pacific La Niña continues to grow and support a strong Australian summer monsoon. Therefore, the Pacific La Niña and tropical Atlantic SST

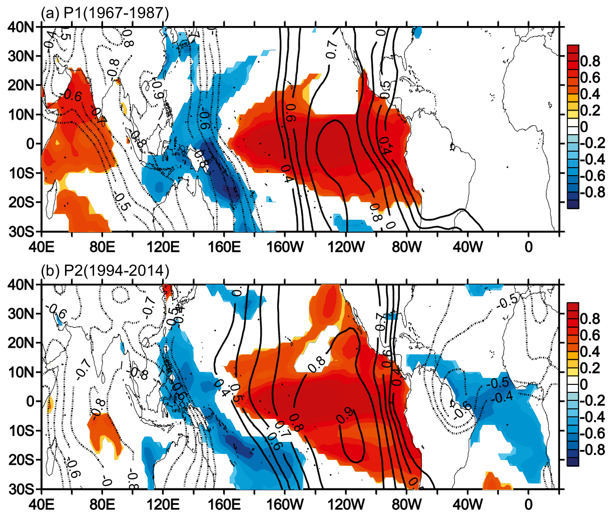


FIG. 8. The correlations of the JJA Niño-3.4 index with JJA SST (shading) and 850-hPa velocity potential (contours, solid for positive and dashed for negative values) at each grid point (a) during P1 and (b) during P2. Only the values at the 90% confidence level or higher are shown.

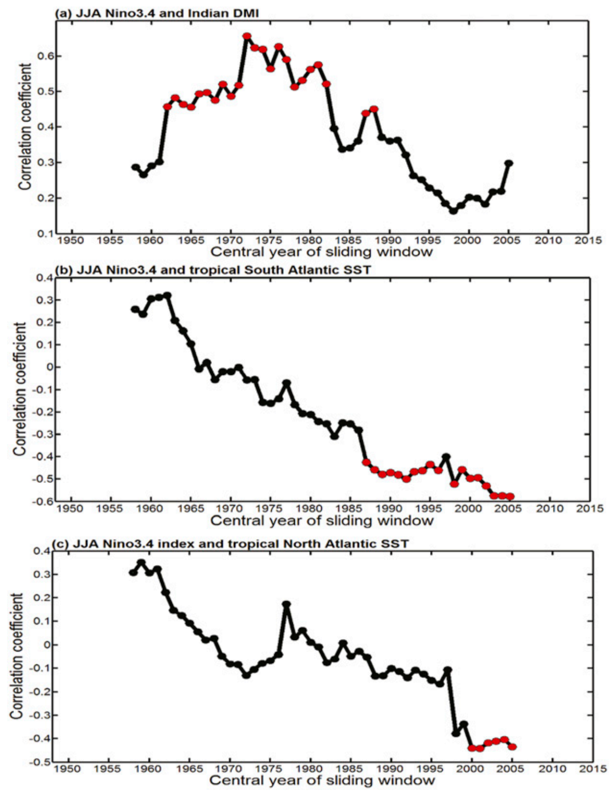


FIG. 9. (a) The 21-yr sliding correlation coefficients (e.g., the correlation coefficient in 2000 represents the period 1990–2010) between the JJA Niño-3.4 and Indian DMI during 1948–2016. (b) As in (a), but for the JJA Niño-3.4 and tropical South Atlantic SSTs. According to Kucharski et al. (2008), the warm tropical South Atlantic SSTs could weaken the summer Indian monsoon. (c) As in (a), but for the JJA Niño-3.4 and tropical North Atlantic SSTs. According to Hong et al. (2014), the warm tropical North Atlantic SSTs could enhance the WNP subtropical high and weaken the summer WNP monsoon. The red dots represent correlations that are significant at the 95% confidence level.

anomalies together support a transition from a weak WNP summer monsoon in JJA to a strong Australian summer monsoon in DJF (i.e., an out-of-phase W-to-A transition) during P2.

For the in-phase A-to-W transition during P2, the composite SST anomalies (Figs. 5e–h) evolve from a decaying La Niña in DJF to a developing El Niño in JJA, which is similar to the composite SST anomalies observed for the out-of-phase A-to-I transition (see Figs. 5a–d). However, the A-to-W transition during P2 is associated with significant SST anomalies in the tropical Atlantic during the transition seasons (i.e., from DJF to the subsequent JJA). The tropical Atlantic cold SST anomalies again act to cancel out the effect of the Pacific El Niño on the Indian monsoon (Kucharski et al. 2007) but reinforce the El Niño influence on the WNP monsoon (Hong et al. 2014). The 850-hPa circulation (Fig. 5g) and rainfall anomalies (Fig. 6g) also confirm that the most active summer monsoon center during P2 is the WNP monsoon rather than the Indian monsoon. Notable anomalous descent occurred over the tropical Atlantic during the JJA of P2 (Fig. 6g), in association with cold SST anomalies in the tropical Atlantic (Fig. 5g). These anomalies were not observed during P1 (Fig. 6c). The region of anomalous descent over the Indian Ocean moved off the Indian peninsula during P2 (Fig. 6g). In the Pacific, anomalous ascent expanded westward greatly during P2 (Fig. 6g). These changes in ascending/descending anomalies lead to a shift of the monsoon anomaly center from the Indian summer monsoon to the WNP monsoon. These analyses indicate that Pacific and Atlantic SST anomalies together support a transition from a strong Australian summer monsoon to a strong WNP summer monsoons (i.e., an in-phase A-to-W transition) during P2.

One important feature to note from the SST composite analyses is that the Pacific Ocean SST anomalies during the TBO monsoon transitions tend to be accompanied by Indian Ocean SST anomalies of the same sign during P1 but with Atlantic Ocean SST anomalies of the opposite sign during P2. To further confirm this impression, we performed a correlation analysis of the tropical SST anomalies with the JJA Niño-3.4 index. As shown in Fig. 8a, the significant correlation coefficients are characterized by an El Niño in the Pacific and an IOD in the Indian Ocean during P1. El Niño during this period has little correlation with Atlantic SST anomalies. In contrast, the significant correlation coefficients

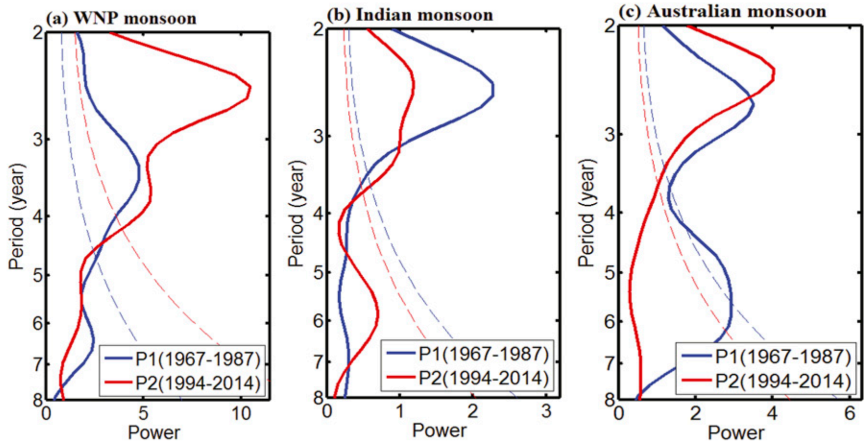


FIG. 10. Comparison of power spectra for the monsoon indices before and after the early 1990s: distribution of power spectrum for (a) WNP monsoon, (b) Indian monsoon, and (c) Australian monsoon. The dashed line denotes the 95% confidence interval against red noise.

during P2 show that the El Niño in the Pacific is accompanied by cold tropical Atlantic SST anomalies (Fig. 8b). The correlation with the Indian Ocean SST is very small. A sliding correlation analysis also indicates a weakened correlation between the Niño-3.4 index and the Indian DMI index after the early 1990s (Fig. 9a) but enhanced correlations between the Niño-3.4 index and North and South Atlantic SST indices (Figs. 9b,c). These results suggest that there was a stronger Pacific–Indian Ocean coherence or coupling during P1 and a stronger Pacific–Atlantic coherence or coupling during P2. Associated with the changes in the interbasin SST correlations, differences were observed in the ascending and descending branches of the Walker circulations between these two periods (Fig. 8). Anomalous ascent and descent were confined within the Indo-Pacific regions during P1 (Fig. 8a) but extended into the Atlantic during P2 (Fig. 8b). During the latter period, anomalous descent developed over the tropical Atlantic where cold anomalies occurred. At the same time, the descent anomalies over the Indian Ocean moved southwestward, leading to a decrease in their influence on the Indian summer monsoon, whereas the ascent anomalies in the Pacific expanded westward, leading to an increase in their influence on the WNP summer monsoon. Therefore, the shift from the strong Pacific–Indian Ocean coupling to the strong Pacific–Atlantic coupling may support the shift of biennial monsoon transitions from the Indian–Australian monsoon transitions to the WNP–Australian monsoon transitions. The stronger Pacific–Atlantic coupling during recent decades is the primary reason why we observe a shift in the monsoon centers in the biennial monsoon transitions associated with the TBO.

Consistent with the shift in biennial monsoon transitions from the Indian–Australian monsoons during P1 to the WNP–Australian monsoons during P2, stronger biennial variability is observed in the power spectrum of the Indian monsoon during P1 but in the power spectrum of the WNP monsoon during P2 (Figs. 10a,b). For the Australian monsoon, relatively strong biennial variability is found during both P1 and P2 (Fig. 10c).

## 5. Summary and discussion

In this study, we performed statistical analyses using observations and reanalysis products to show that two important changes to the TBO have occurred since the early 1990s: 1) The biennial monsoon transitions associated with the TBO have shifted from involving the Indian–Australian summer monsoons to involving the WNP–Australian summer monsoons, and 2) tropical Atlantic SST anomalies have become an important part of the monsoon–ocean interactions associated with the TBO. Figure 11 illustrates how the different SST anomaly patterns before and after the early 1990s can induce the shift of the summertime monsoon centers. During the period before the early 1990s, strong Pacific–Indian SST coupling or coherence confined most of the anomalous ascent or descent to the Indo-Pacific region (Fig. 11a), with one of the anomaly centers located right over the Indian monsoon region. During the period after the early 1990s, strong Pacific–Atlantic SST coupling/coherence displaced the locations of the anomalous ascent/descent centers and shifted one center to the WNP monsoon region (Fig. 11b). Therefore, the changes in the monsoon centers associated with the TBO are

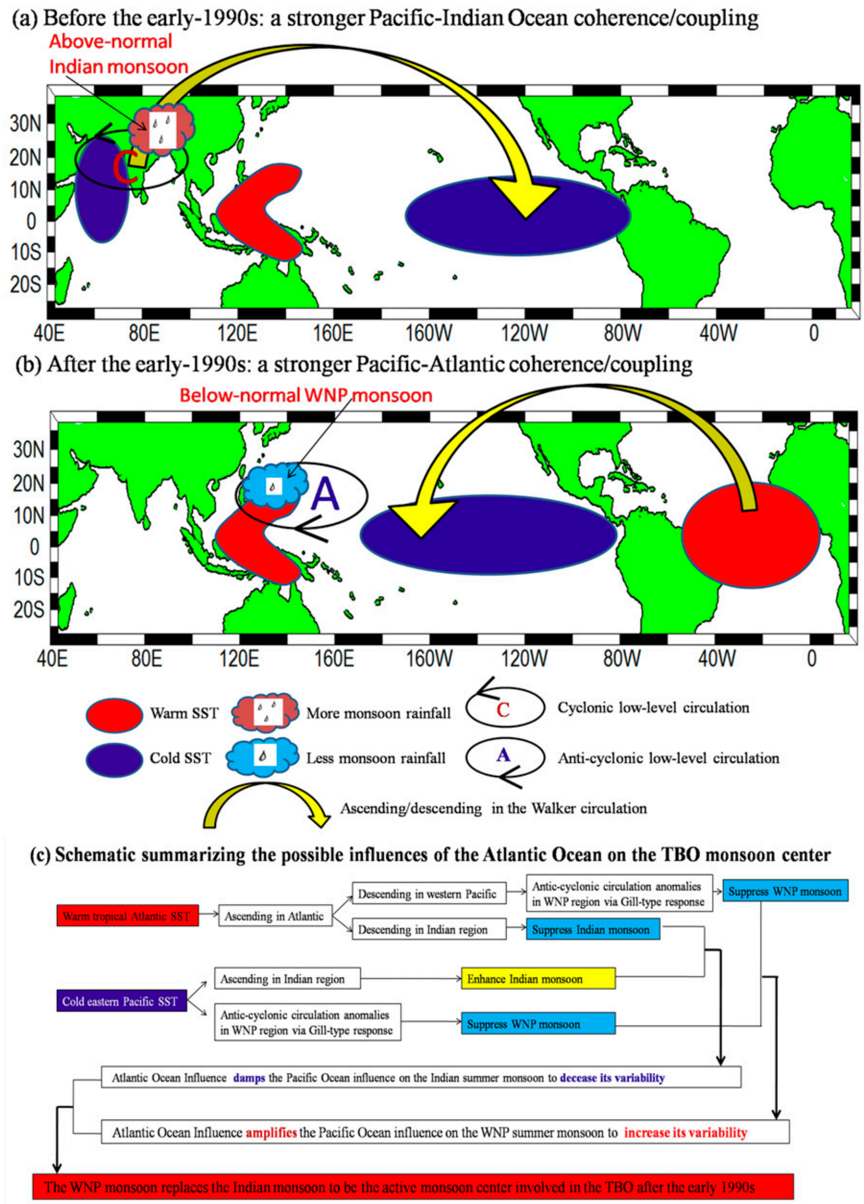


FIG. 11. A schematic diagram showing the physical processes to induce the recent shift in the monsoon centers of the TBO. (a) The Pacific–Indian Ocean coherence/coupling before the early 1990s supports the Indian summer monsoon to be the monsoon center involved in the TBO. (b) The Pacific–Atlantic Ocean coherence/coupling after the early 1990s supports the WNP summer monsoon to be the monsoon center involved in the TBO. (c) The schematic summarizing the possible influences of the Atlantic Ocean on the shift of the summertime monsoon centers involved in the TBO.

related to an increased influence of tropical Atlantic SST anomalies on the Indian and WNP summer monsoons. The increased Atlantic SST influence acts to weaken the ENSO influence on the Indian summer monsoon leading to a decrease in its variability but enhances the ENSO influence on the WNP summer monsoon to increase its variability (Fig. 11c). As a result, during the last two decades the WNP monsoon has replaced the Indian monsoon to become a major component in the biennial monsoon transitions associated with the TBO. These results highlight the important roles of Pacific–Atlantic interactions in the TBO dynamics in recent decades.

Much effort has been expended attempting to understand the TBO dynamics, most of which involve interactions between the Indian–Australian summer monsoons and SST anomalies in the Indo-Pacific sector (e.g., Meehl 1987, 1993;Chang and Li 2000; Loschnigg et al. 2003; Yu et al. 2003; Meehl et al. 2003). Our results suggest that the Pacific–Atlantic Ocean sector may have become involved in the TBO dynamics after the early 1990s, in contrast to the pre-1990s TBO dynamics that mainly involved the Pacific–Indian Ocean sector (e.g., Chang and Li 2000; Meehl et al. 2003; Yu et al. 2003; Loschnigg et al. 2003). This shift in the TBO dynamics may be part of the early-1990s climate shift (Qian et al. 2014; Yu et al. 2015; Paek et al. 2016; Lyu et al. 2017; Wang et al. 2017) that has been linked to a phase change of the AMO and global warming trends. The main message of this study is that the TBO and its underlying dynamics may have to be studied separately for the periods before and after the early 1990s.

The change in the biennial monsoon transitions associated with the TBO during the recent two decades may affect the strategies used for seasonal climate predictions in the Asian–Australian monsoonal regions. For example, tropical Atlantic SSTs may need to be considered and incorporated more than previously when predicting the Asian–Australian monsoon. Also, the characteristics of the Australian summer monsoon may become more useful for predicting the WNP summer monsoon during the following boreal summer. Since the intensity of the EA summer monsoon generally has a negative correlation with that of the WNP summer monsoon (e.g., Kwon et al. 2005), a better prediction of the WNP summer can also benefit predictions of the EA summer monsoon that is closely associated with the Chinese mei-yu, the Korean changma, and the Japanese baiu.

Our results suggest an increase in the importance of Pacific–Atlantic interactions to the TBO dynamics in recent decades, at around the time when the AMO changed from a negative to a positive phase (in the early 1990s). The phase of the AMO may be playing a role in the increased influence of the Atlantic during some decades but not others. Different phases of the AMO may produce different interbasin SST gradients between the Pacific and Atlantic and different impacts on the strengths of Walker circulations above the basins (Wang 2006; Chikamoto et al. 2015; Zhang and Karnauskas 2017), which may separately or together affect the role of the Atlantic Ocean in the TBO dynamics. Further studies are needed to better understand these interbasin interaction processes.

*Acknowledgments.* We thank the three anonymous reviewers for their comments and suggestions that have greatly improved the manuscript. This work was supported by the National Science Foundation's Climate and Large Scale Dynamics Program under Grant AGS-1505145. L.W.'s work was also supported by the National Natural Science Foundation of China (Grant 41776031), the Guangdong Natural Science Foundation (Grant 2015A030313796), and the National Program on Global Change and Air–Sea Interaction (GASI-IPOVAI-04). L.W. thanks the China Scholarship Council for supporting his visit to the University of California, Irvine.

## REFERENCES

Cai, W., and P. van Rensch, 2013: Austral summer teleconnections of Indo-Pacific variability: Their nonlinearity and impacts on Australian climate. *J. Climate*, **26**, 2796–2810, https://doi.org/10.1175/JCLI-D-12-00458.1.

Cash, B. A., X. Rodo, J. Ballester, M. J. Bouma, A. Baeza, R. Dhiman, and M. Pascual, 2013: Malaria epidemics and the influence of the tropical South Atlantic on the Indian monsoon. *Nat. Climate Change*, **3**, 502–507, https://doi.org/10.1038/nclimate1834.

Chang, C.-P., and T. Li, 2000: A theory for the tropical tropospheric biennial oscillation. *J. Atmos. Sci.*, **57**, 2209–2224, https://doi.org/10.1175/1520-0469(2000)057<2209:ATFTTT>2.0.CO;2.

Chen, M., P. Xie, J. E. Janowiak, and P. A. Arkin, 2002: Global land precipitation: A 50-yr monthly analysis based on gauge observations. *J. Hydrometeor.*, **3**, 249–266, https://doi.org/10.1175/1525-7541(2002)003<0249:GLPAYM>2.0.CO;2.

Chikamoto, Y., and Coauthors, 2015: Skilful multi-year predictions of tropical trans-basin climate variability. *Nat. Commun.*, **6**, 6869, https://doi.org/10.1038/ncomms7869.

Clarke, A. J., X. Liu, and S. Van Gorder, 1998: Dynamics of the biennial oscillation in the equatorial Indian and far western Pacific Oceans. *J. Climate*, **11**, 987–1001, https://doi.org/10.1175/1520-0442(1998)011<0987:DOTBOI>2.0.CO;2.

Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, https://doi.org/10.1002/qj.49710644905.

Gu, D., T. Li, Z. Ji, and B. Zheng, 2010: On the phase relations between the western North Pacific, Indian, and Australian monsoons. *J. Climate*, **23**, 5572–5589, https://doi.org/10.1175/2010JCLI2761.1.

Hong, C.-C., T.-C. Chang, and H.-H. Hsu, 2014: Enhanced relationship between the tropical Atlantic SST and the summertime western North Pacific subtropical high after the early 1980s. *J. Geophys. Res. Atmos.*, **119**, 3715–3722, https://doi.org/10.1002/2013JD021394.

Hung, C.-W., X. D. Liu, and M. Yanai, 2004: Symmetry and asymmetry of the Asian and Australian summer monsoons. *J. Climate*, **17**, 2413–2426, https://doi.org/10.1175/1520-0442(2004)017<2413:SAAOTA>2.0.CO;2.

Kajikawa, Y., B. Wang, and J. Yang, 2010: A multi-time scale Australian monsoon index. *Int. J. Climatol.*, **30**, 1114–1120, https://doi.org/10.1002/joc.1955.

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kao, H.-Y., and J.-Y. Yu, 2009: Contrasting eastern-Pacific and central-Pacific types of ENSO. *J. Climate*, **22**, 615–632, https://doi.org/10.1175/2008JCLI2309.1.

Keenlyside, N. S., H. Ding, and M. Latif, 2013: Potential of equatorial Atlantic variability to enhance El Niño prediction.

*Geophys. Res. Lett.*, **40**, 2278–2283, https://doi.org/10.1002/grl.50362.

Kerr, R. A., 2000: A North Atlantic climate pacemaker for the centuries. *Science*, **288**, 1984–1986, https://doi.org/10.1126/science.288.5473.1984.

Kiladis, G. N., and H. van Loon, 1988: The Southern Oscillation. Part VII: Meteorological anomalies over the Indian and Pacific sectors associated with the extremes of the oscillation. *Mon. Wea. Rev.*, **116**, 120–136, https://doi.org/10.1175/1520-0493(1988)116<0120:TSOPVM>2.0.CO;2.

Kucharski, F., A. Bracco, J. H. Yoo, and F. Molteni, 2007: Low-frequency variability of the Indian monsoon–ENSO relationship and the tropical Atlantic: The ''weakening'' of the 1980s and 1990s. *J. Climate*, **20**, 4255–4266, https://doi.org/10.1175/JCLI4254.1.

——, ——, ——, and ——, 2008: Atlantic forced component of the Indian monsoon interannual variability. *Geophys. Res. Lett.*, **35**, L04706, https://doi.org/10.1029/2007GL033037.

Kwon, M., J.-G. Jhun, B. Wang, S.-I. An, and J.-S. Kug, 2005: Decadal change in relationship between East Asian and WNP summer monsoons. *Geophys. Res. Lett.*, **32**, L16709, https://doi.org/10.1029/2005GL023026.

Lau, K.-M., and S. Yang, 1996: The Asian monsoon and predictability of the tropical ocean–atmosphere system. *Quart. J. Roy. Meteor. Soc.*, **122**, 945–957, https://doi.org/10.1002/qj.49712253208.

Lee, E.-J., K.-J. Ha, and J.-G. Jhun, 2014: Interdecadal changes in interannual variability of the global monsoon precipitation and interrelationships among its subcomponents. *Climate Dyn.*, **42**, 2585–2601, https://doi.org/10.1007/s00382-013-1762-4.

Li, T., P. Liu, X. Fu, B. Wang, and G. A. Meehl, 2006: Spatiotemporal structures and mechanisms of the tropospheric biennial oscillation in the Indo-Pacific warm ocean regions. *J. Climate*, **19**, 3070–3087, https://doi.org/10.1175/JCLI3736.1.

Li, X., S.-P. Xie, S. T. Gille, and C. Yoo, 2016: Atlantic-induced pan-tropical climate change over the past three decades. *Nat. Climate Change*, **6**, 275–279, https://doi.org/10.1038/nclimate2840.

Loschnigg, J., G. A. Meehl, P. J. Webster, J. M. Arblaster, and G. P. Compo, 2003: The Asian monsoon, the tropospheric biennial oscillation, and the Indian Ocean zonal mode in the NCAR CSM. *J. Climate*, **16**, 1617–1642, https://doi.org/10.1175/1520-0442(2003)016<1617:TAMTTB>2.0.CO;2.

Lyu, K., J.-Y. Yu, and H. Paek, 2017: The influences of the Atlantic multidecadal oscillation on the mean strength of the North Pacific subtropical high during boreal winter. *J. Climate*, **30**, 411–426, https://doi.org/10.1175/JCLI-D-16-0525.1.

Matsumoto, J., 1992: The seasonal changes in Asian and Australian monsoon regions. *J. Meteor. Soc. Japan*, **70**, 257–273, https://doi.org/10.2151/jmsj1965.70.1B_257.

Meehl, G. A., 1987: The annual cycle and interannual variability in the tropical Pacific and Indian Ocean regions. *Mon. Wea. Rev.*, **115**, 27–50, https://doi.org/10.1175/1520-0493(1987)115<0027:TACAIV>2.0.CO;2.

——, 1993: A coupled air–sea biennial mechanism in the tropical Indian and Pacific regions: Role of the ocean. *J. Climate*, **6**, 31–41, https://doi.org/10.1175/1520-0442(1993)006<0031:ACASBM>2.0.CO;2.

——, 1994: Coupled land–ocean–atmosphere processes and South Asian monsoon variability. *Science*, **266**, 263–267, https://doi.org/10.1126/science.266.5183.263.

——, 1997: The South Asian monsoon and tropospheric biennial oscillation. *J. Climate*, **10**, 1921–1943, https://doi.org/10.1175/1520-0442(1997)010<1921:TSAMAT>2.0.CO;2.

——, and J. M. Arblaster, 2002a: The tropospheric biennial oscillation and Asian–Australian monsoon rainfall. *J. Climate*, **15**, 722–744, https://doi.org/10.1175/1520-0442(2002)015<0722:TTBOAA>2.0.CO;2.

——, and ——, 2002b: Indian monsoon GCM sensitivity experiments testing tropospheric biennial oscillation transition conditions. *J. Climate*, **15**, 923–944, https://doi.org/10.1175/1520-0442(2002)015<0923:IMGSET>2.0.CO;2.

——, ——, and J. Loschnigg, 2003: Coupled ocean–atmosphere dynamical processes in tropical Indian and Pacific Oceans and the TBO. *J. Climate*, **16**, 2138–2158, https://doi.org/10.1175/2767.1.

Nicholls, N., 1978: Air–sea interaction and the quasi-biennial oscillation. *Mon. Wea. Rev.*, **106**, 1505–1508, https://doi.org/10.1175/1520-0493(1978)106<1505:ASIATQ>2.0.CO;2.

——, 1979: A simple air–sea interaction model. *Quart. J. Roy. Meteor. Soc.*, **105**, 93–105, https://doi.org/10.1002/qj.49710544307.

——, 1984: The Southern Oscillation and Indonesian sea surface temperature. *Mon. Wea. Rev.*, **112**, 424–432, https://doi.org/10.1175/1520-0493(1984)112<0424:TSOAIS>2.0.CO;2.

Paek, H., J.-Y. Yu, F. Zheng, and M.-M. Lu, 2017: Impacts of ENSO diversity on the western Pacific and North Pacific subtropical highs during boreal summer. *Climate Dyn.*, https://doi.org/10.1007/s00382-016-3288-z, in press.

Qian, C., J.-Y. Yu, and G. Chen, 2014: Decadal summer drought frequency in China: The increasing influence of the Atlantic multi-decadal oscillation. *Environ. Res. Lett.*, **9**, 124004, https://doi.org/10.1088/1748-9326/9/12/124004.

Rasmusson, E. M., and T. H. Carpenter, 1983: The relationship between eastern equatorial Pacific sea surface temperature and rainfall over India and Sri Lanka. *Mon. Wea. Rev.*, **111**, 517–528, https://doi.org/10.1175/1520-0493(1983)111<0517:TRBEEP>2.0.CO;2.

Rodionov, S. N., 2004: A sequential algorithm for testing climate regime shifts. *Geophys. Res. Lett.*, **31**, L09204, https://doi.org/10.1029/2004GL019448.

Rodríguez-Fonseca, B., I. Polo, J. García-Serrano, T. Losada, E. Mohino, C. R. Mechoso, and F. Kucharski, 2009: Are Atlantic Niños enhancing Pacific ENSO events in recent decades? *Geophys. Res. Lett.*, **36**, L20705, https://doi.org/10.1029/2009GL040048.

Ropelewski, C. F., M. S. Halpert, and X. Wang, 1992: Observed tropospheric biennial variability and its relationship to the Southern Oscillation. *J. Climate*, **5**, 594–614, https://doi.org/10.1175/1520-0442(1992)005<0594:OTBVAI>2.0.CO;2.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. *Nature*, **401**, 360–363.

Schlesinger, M. E., and N. Ramankutty, 1994: An oscillation in the global climate system of period 65–70 years. *Nature*, **367**, 723–726, https://doi.org/10.1038/367723a0.

Smith, T. M., R. W. Reynolds, T. C. Peterson, and J. Lawrimore, 2008: Improvements to NOAA's historical merged land–ocean surface temperature analysis (1880–2006). *J. Climate*, **21**, 2283–2296, https://doi.org/10.1175/2007JCLI2100.1.

Sui, C.-H., P.-H. Chung, and T. Li, 2007: Interannual and interdecadal variability of the summertime western North Pacific subtropical high. *Geophys. Res. Lett.*, **34**, L11701, https://doi.org/10.1029/2006GL029204.

Taschetto, A. S., A. Sen Gupta, H. H. Hendon, C. C. Ummenhofer, and M. H. England, 2011: The contribution of Indian Ocean sea surface temperature anomalies on Australian summer rainfall during El Niño events. *J. Climate*, **24**, 3734–3747, https://doi.org/10.1175/2011JCLI3885.1; Corrigendum, **26**, 9727–9728, https://doi.org/10.1175/JCLI-D-13-00513.1.

Wang, B., and Z. Fan, 1999: Choice of South Asian summer monsoon indices. *Bull. Amer. Meteor. Soc.*, **80**, 629–638, https://doi.org/10.1175/1520-0477(1999)080<0629:COSASM>2.0.CO;2.

——, R. Wu, and K.-M. Lau, 2001: Interannual variability of Asian summer monsoon: Contrast between the Indian and western North Pacific–East Asian monsoons. *J. Climate*, **14**, 4073–4090, https://doi.org/10.1175/1520-0442(2001)014<4073:IVOTAS>2.0.CO;2.

Wang, C., 2006: An overlooked feature of tropical climate: Inter-Pacific-Atlantic variability. *Geophys. Res. Lett.*, **33**, L12702, https://doi.org/10.1029/2006GL026324.

——, F. Kucharski, R. Barimalala, and A. Bracco, 2009: Teleconnections of the tropical Atlantic to the tropical Indian and Pacific Oceans: A review of recent findings. *Meteor. Z.*, **18**, 445–454, https://doi.org/10.1127/0941-2948/2009/0394.

Wang, L., J.-Y. Yu, and H. Paek, 2017: Enhanced biennial variability in the Pacific due to Atlantic capacitor effect. *Nat. Commun.*, **8**, 14887, https://doi.org/10.1038/ncomms14887.

Webster, P. J., A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled oceanic–atmospheric dynamics in the Indian Ocean during 1997–98. *Nature*, **401**, 356–360, https://doi.org/10.1038/43848.

Wu, R., and B. P. Kirtman, 2004: The tropospheric biennial oscillation of the monsoon–ENSO system in an interactive ensemble coupled GCM. *J. Climate*, **17**, 1623–1640, https://doi.org/10.1175/1520-0442(2004)017<1623:TTBOOT>2.0.CO;2.

Yu, J.-Y., and H.-Y. Kao, 2007: Decadal changes of ENSO persistence barrier in SST and ocean heat content indices: 1958–2001. *J. Geophys. Res.*, **112**, D13106, https://doi.org/10.1029/2006JD007715.

——, S.-P. Weng, and J. D. Farrara, 2003: Ocean roles in TBO transition of the Indian–Australian monsoon system. *J. Climate*, **16**, 3072–3080, https://doi.org/10.1175/1520-0442(2003)016<3072:ORITTT>2.0.CO;2.

——, M.-M. Lu, and S. T. Kim, 2012: A change in the relationship between tropical central Pacific SST variability and the extratropical atmosphere around 1990. *Environ. Res. Lett.*, **7**, 034025, https://doi.org/10.1088/1748-9326/7/3/034025.

——, P.-K. Kao, H. Paek, H.-H. Hsu, C.-W. Hung, M.-M. Lu, and S.-I. An, 2015: Linking emergence of the central Pacific El Niño to the Atlantic multidecadal oscillation. *J. Climate*, **28**, 651–662, https://doi.org/10.1175/JCLI-D-14-00347.1.

Zhang, L., and K. B. Karnauskas, 2017: The role of tropical interbasin SST gradients in forcing Walker circulation trends. *J. Climate*, **30**, 499–508, https://doi.org/10.1175/JCLI-D-16-0349.1.

Zheng, B., D. J. Gu, A. L. Lin, and C. H. Li, 2008: Spatial patterns of tropospheric biennial oscillation and its numerical simulation. *Adv. Atmos. Sci.*, **25**, 815–823, https://doi.org/10.1007/s00376-008-0815-9.