Decadal Variability of Asian–Australian Monsoon–ENSO–TBO Relationships

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ABSTRACT

A set of dynamically coupled ocean–atmosphere mechanisms has previously been proposed for the Asia–Pacific tropics to produce a dominant biennial component of interannual variability [the tropospheric biennial oscillation (TBO)]. Namely, a strong Asian–Australian monsoon is often associated with negative SST anomalies in the equatorial eastern Pacific and a negative Indian Ocean dipole in northern fall between the strong Indian monsoon and strong Australian monsoon, and tends to be followed by a weak monsoon and positive SST anomalies in the Pacific the following year and so on. These connections are communicated through the large-scale east–west (Walker) circulation that involves the full depth of the troposphere. However, the Asia–Pacific climate system is characterized by intermittent decadal fluctuations whereby the TBO during some time periods is more pronounced than others. Observations and models are analyzed to identify processes that make the system less biennial at certain times due to one or some combination of the following:

1) increased latitudinal extent of Pacific trade winds and wider cold tongue;
2) warmer tropical Pacific compared to tropical Indian Ocean that weakens trade winds and reduces coupling strength;
3) eastward shift of the Walker circulation;
4) reduced interannual variability of Pacific and/or Indian Ocean SSTs.

Decadal time-scale SST variability associated with the interdecadal Pacific oscillation (IPO) has been shown to alter the TBO over the Indo-Pacific region by contributing changes in either some or all of the four factors listed above. Analysis of a multicentury control run of the Community Climate System Model, version 4 (CCSM4), shows that this decadal modulation of interannual variability is transferred via the Walker circulation to the Asian–Australian monsoon region, thus affecting the TBO and monsoon–Pacific connections. Understanding these processes is important to be able to evaluate decadal predictions and longer-term climate change in the Asia–Pacific region.

1. Introduction

For nearly a century, research has been done to attempt to quantify the nature and origins of large-scale connections between interannual variability of the Asian–Australian monsoon system, South Asian land processes, and Indian and Pacific Ocean sea surface temperatures (SSTs). Sir Gilbert Walker (1923, 1924) is generally credited with being the first to note an interannual opposition of sign of sea level pressure anomalies between the Indian and Pacific sectors in his attempts to discover predictive indicators of monsoon strength. Walker and Bliss (1930) coined the term “Southern Oscillation” to describe this large scale seesaw of mass that connected climate in the Indian and Pacific regions. Bjerknes (1969) named the large-scale east–west circulation linking the eastern and western

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During El Niño (documented, e.g., by Rasmusson and Carpenter 1982), an associated decrease of Indian monsoon rainfall was quantified in papers by, for example, Pant and Parthasarathy (1981) and Rasmusson and Carpenter (1983). Other studies also showed a decrease of Australian monsoon rainfall (usually defined as area-averaged southern summer rainfall over a region encompassing parts of Indonesia, Papua New Guinea, and northern Australia) during El Niño (Tanaka 1981; Allan 1983). Influences from the Indian monsoon on tropical Pacific SSTs were addressed by Yasunari (1990).

Relatively early on there were studies that noted a biennial tendency of many phenomena in the Indo-Pacific climate system (Troup 1965; Trenberth 1975; Nicholls 1978, 1979, 1984; Mooley and Parthasarathy 1983). The role of air–sea coupling in producing such a biennial tendency was suggested by Brier (1978). Meehl (1987) proposed a mechanism to account for the biennial tendency in the Indo-Pacific region, termed the tropospheric biennial oscillation (TBO; Meehl 1997). Invoking large-scale dynamically coupled interactions, the mechanism was based on the premise of air–sea coupling being strong one season per year in the Indo-Pacific region during the passage of the “convective maximum” (a mass of convection and precipitation that transitioned from the Indian monsoon in northern summer southeastward to the Australian monsoon in southern summer and out into the tropical Pacific in northern spring in the climatological seasonal cycle). Through local heat balance arguments, Meehl (1987) showed how this strong once-per-year air–sea coupling could transition positive SST anomalies to negative SST anomalies, which would then persist until the following year when there would then be anomalies in precipitation opposite in sign to the year before. The persistence of these anomalies from one year to the next was maintained through upper-ocean heat content anomalies (Meehl 1993) and the large-scale east–west circulation in the atmosphere (Meehl 1987). Subsequently, there was the added concept of convective heating anomalies in the Indo-Pacific region helping to sustain atmospheric circulation anomalies over South Asia to preserve the memory of land surface temperature anomalies that contributed to anomalous meridional temperature gradients and consequent biennial monsoon variability (Meehl 1994a,b). The fundamental nature of the dynamically coupled processes involved with the TBO has been additionally documented in a number of global coupled climate model simulations (e.g., Meehl 1997; Ogasawara et al. 1999; Loschnigg et al. 2003). Given the large-scale east–west circulations noted above, the position of the diabatic heating in the western Pacific drives thermal contrasts associated with these Indo-Pacific connections (Webster et al. 1998) and, has been shown to involve circulation in the upper troposphere as indicated by the strength of the upper-level South Asian high (Kucharski et al. 2011).

Regional patterns of SST anomalies in the TBO involve El Niño and La Niña in the Pacific, and SST anomaly patterns in the Indian Ocean during the northern fall season following the South Asian monsoon (see Meehl 1987, his Fig. 11) that subsequently became known as the Indian Ocean dipole (IOD; e.g., Webster et al. 1999; Saji et al. 1999). An important component of the Indo-Pacific climate system that characterizes the TBO is a transition of SST anomalies in the equatorial eastern Pacific Ocean from positive to negative (or vice versa) in northern spring, so the seasons leading up to those transitions are crucial to the TBO (e.g., Meehl and Arblaster 2002a,b). Thus, a “negative IOD” in northern fall (negative SST anomalies in the western tropical Indian Ocean, and positive SST anomalies in the eastern tropical Indian Ocean), with negative SST anomalies in the equatorial eastern Pacific (Fig. 1), can transition to basin-wide negative SST anomalies across the Indian Ocean in northern winter (Fig. 2), with positive SST anomalies in the eastern equatorial Pacific in the following northern spring and summer in the TBO (Meehl et al. 2003).

Recently, Izumo et al. (2010) made use of these transition processes in the TBO to document El Niño forecast skill by monitoring the state of the IOD in northern fall. Since the system is characteristically biennial (i.e., the TBO), as noted above there would tend to be a transition of the sign of SST anomalies in the equatorial eastern Pacific from negative to positive in the next few seasons after a negative IOD, thus giving rise to El Niño forecast skill from the TBO.

The nature of these transition processes going from northern fall to the following northern spring/summer invoke dynamic air–sea coupling in the far western equatorial Pacific. Thus, convective heating anomalies in the Indian Ocean associated with the IOD (e.g., Annamalai et al. 2005), or the Pacific (Wu and Kirtman 2004), or a combination of southeastern Indian Ocean and western Pacific (Li et al. 2006) affect the southeastern Indian Ocean and western North Pacific anticyclones to produce wind stress anomalies in the equatorial western
Pacific. These force ocean Kelvin waves that then contribute to changing the depth of the thermocline in the eastern equatorial Pacific and consequently the sign of SST anomalies there (Lau and Wu 2001; Turner et al. 2007).

Two questions that logically arise at this point are: given these seemingly self-sustaining dynamically coupled processes that tend to make the Indo-Pacific climate system biennial, thus making it possible to identify the TBO, why is the system not perfectly biennial, and what can disrupt the connections between the Asian–Australian monsoon and tropical Pacific SSTs? There are actually at least several factors that could produce these disruptions. The first is an external forcing that has a longer time scale but produces SST anomaly patterns in the tropical Pacific that are similar to those of the TBO in Figs. 1 and 2. For example, there is evidence that occasional volcanic eruptions could produce such SST anomaly patterns (e.g., Mann et al. 2005). Peaks in the 11-yr solar cycle have been shown to produce weak La Niña–like SST anomalies in the equatorial Pacific (van Loon et al. 2007; van Loon and Meehl 2008; Meehl et al. 2008, 2009a) followed a couple of years later by low-amplitude El Niño–like SST anomalies (White et al. 1998; Meehl and Arblaster 2009; Meehl et al. 2009a). Second, there is evidence that decadal time-scale climate variability in the Atlantic sector could affect climate in the Indo-Pacific region (e.g., Kucharski et al. 2007). Third, decadal time-scale variability associated with the interdecadal Pacific oscillation (IPO; Power et al. 1999) produces low-amplitude positive and negative SST anomalies in the tropical Pacific with connections to Asian–Australian climate (Power et al. 1999; Arblaster et al. 2002; Meehl and Hu 2006) that could potentially disrupt or modify biennial variability associated with the TBO in the Indo-Pacific region. It is this last possibility that we will explore in the present paper.

Such decadal time-scale variability when the tropical Pacific is in an anomalously warm state (positive IPO) tends to weaken the connections between interannual SST variability in the eastern equatorial Pacific and Australian rainfall variability (Power et al. 1999). This

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**FIG. 1.** Positive minus negative TBO composites for the September–November season following a strong Indian monsoon, for equatorial upper-ocean temperature differences (°C), SST (°C), surface wind stress (scaling arrow = 0.03 N m⁻²), precipitation (mm day⁻¹), surface winds (scaling arrow = 1.0 m s⁻¹), and schematic representation of large-scale Walker circulation anomalies [i.e., western Walker cell (WWC) and eastern Walker cell (EWC)]. Contour intervals are 0.5 mm day⁻¹ for precipitation (values between −1 and +1 mm day⁻¹ are not plotted), 0.25°C for SST, and 0.5°C for upper-ocean temperature (after Meehl et al. 2003).

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has to do with changes in the magnitude of interannual variability in the tropical Pacific as well as an eastward shift of the Walker circulation (Arblaster et al. 2002).

As noted above, there is evidence that points to decadal time-scale processes in the tropical Pacific, including literature focused on the mid-1970s climate shift in the observations (e.g., Wang and An 2001) and comparable changes to base states in climate models (e.g., Neale et al. 2008), that could produce less biennial variability and consequently a weakened TBO. However, as alluded to previously, there is evidence that coupled processes in the Indian Ocean are playing an active role in modulating TBO variability as well. For example, eliminating Indian Ocean SST variability tends to drastically reduce the TBO in the entire Indo-Pacific region (Chung and Nigam 1999; Kim and Lau 2001; Yu et al. 2009). Yu et al. (2009) substituted observed climatological SSTs in the Indian Ocean in the Community Climate System Model, version 3 (CCSM3). This had the effect of producing a cooler Indian Ocean surface (the standard model’s Indian Ocean SSTs were warmer than observed) that then contributed to weakening the strength of the trade winds right across the equatorial Indian and Pacific Oceans, reducing the air–sea coupling strength (Zebiak and Cane 1987; Neelin et al. 1994), warming the equatorial Pacific (not unlike the mid-1970s shift in observations) and producing drastically reduced biennial variability in Niño-3.4 SSTs and the Asian–Australia monsoon. Other possibilities for these changes in that experiment could include the lack of interannual variability in the Indian Ocean or the removal of dynamically coupled air–sea interaction in the Indian Ocean.

In another modeling study, Kug and Kang (2006) contrasted conditions with active Indian Ocean SST variability (with their strong connection to equatorial Pacific SSTs) to conditions where the Pacific SSTs had little connection to the Indian Ocean. In the former, there was a stronger TBO compared to the latter, pointing to the importance of coupled processes in the Indian Ocean and an active TBO. The reason for this, as documented by Kug and Kang (2006) and Kug et al. (2006), is that in the cases with the active Indian Ocean SST connections to the Pacific SSTs, there was a positive IOD in northern fall followed by Indian Ocean basin-wide warming in northern winter. This seasonal SST evolution in the Indian Ocean was associated with a strengthened southeastern Indian Ocean anticyclone and western North Pacific anticyclone with easterly wind...
stress anomalies in the equatorial western Pacific. These easterly wind stress anomalies produced upwelling Kelvin waves that crossed the equatorial Pacific to shallow the thermocline in the east and contributed to a transition from positive to negative SST anomalies there in northern spring and summer. In the cases without this connection to Indian Ocean SST variability, the easterly anomalies in the far western equatorial Pacific did not form in northern fall and winter, and there was no SST transition in the eastern equatorial Pacific in northern spring and summer. Without that strong seasonal transition, the same anomalous SST conditions in the equatorial eastern Pacific tended to persist to the following year, thus reducing the TBO.

Even though equatorial Pacific SST and associated convective heating anomalies can influence the western Pacific anticyclone in similar ways (Wu and Kirtman 2004), the effects of interannual variability of Indian Ocean SSTs on convective heating anomalies, the western Pacific, and southeastern Indian Ocean anticyclones, and resulting zonal wind stress anomalies in the far western equatorial Pacific are crucial for an active TBO (see also Annamalai et al. 2005; Lau and Wu 2001; Li et al. 2006; Turner et al. 2007; Nagura and Konda 2007; Izumo et al. 2010).

Thus, there is evidence that dynamically coupled processes in both the Pacific and Indian Oceans can act to produce either a stronger or weaker TBO, and decadal variability in either basin, either connected or not, could produce decadal time-scale variability in TBO amplitude. However, in this paper we will focus mainly on the effects of decadal SST variability in the tropical Pacific and the associated influences on Asian–Australian monsoon and Pacific SST connections.

Section 2 includes a description of observed datasets and models used in the present study. Section 3 examines the example of the mid-1970s climate shift as a manifestation of a change in sign of the IPO from negative to positive and the associated effects on connections between interannual SST variability in the tropical Pacific with the monsoon regimes of Australia–Asia and the TBO. In section 4, two global coupled climate model simulations with different base-state climates and TBO amplitudes will be analyzed for clues as to how their different base states could be analogous to the multidecadal changes in the IPO that can make the system either more or less biennial, producing either a stronger or weaker TBO. Section 5 contains the conclusions.

2. Model and observed data descriptions

The standard CCSM3 (e.g., Collins et al. 2006) with a very regular TBO will be compared to the new Community Climate System Model, version 4 (CCSM4; Gent et al. 2011). The CCSM3 has a T85 atmospheric model with 26 levels in the vertical, and is coupled to a 1° ocean. Characteristics of the Asia–Australia monsoon simulations in CCSM3 are described by Meehl et al. (2006). As in other model versions, the monsoon rainfall in the tropical Indian Ocean tends to be greater than observed and shifted westward, with less than observed rainfall in the Bay of Bengal. Similar biases exist in CCSM4 (Meehl et al. 2011, manuscript submitted to J. Climate). However, the correlations between Indian monsoon area-averaged rainfall and tropical Pacific SSTs are comparable to the observations that can be as strong as about −0.7 depending on the time period (see discussion later with regard to Fig. 6) and are similar for CCSM3 and CCSM4 with concomitant multidecadal and multicentury variations of this connection.

The CCSM4 includes a finite-volume 1° version of the Community Atmosphere Model, version 4 (CAM4), with improved components of ocean, land, and sea ice compared to CCSM3 (Gent et al. 2011). Grid points in the atmosphere are spaced roughly every 1° latitude and longitude, and there are 26 levels in the vertical. The ocean is a version of the Parallel Ocean Program (POP) with a nominal latitude–longitude resolution of 1° (down to 1/2° in the equatorial tropics) and 60 levels in the vertical. No flux adjustments are used in either CCSM3 or CCSM4.

As noted earlier, the CCSM3 has an overly biennial ENSO signal and thus a very strong and regular TBO (Deser et al. 2006). In CCSM4, modifications have been made to the convection scheme that then produces a much less biennial ENSO and thus a weaker TBO (Neale et al. 2008; Deser et al. 2011, manuscript submitted to J. Climate). Earlier developmental versions of CCSM4 showed that these changes also produced a less biennial Indian monsoon as well (not shown).

Observed SSTs and ocean data are from the Simple Ocean Data Assimilation (SODA) ocean reanalysis product that also includes surface wind stress forcing (Carton et al. 2000). All-India rainfall are monsoon-averaged data (June–September) described by Partheyraty and Mooley (1978). Additional observed SSTs are used from the Hadley Centre Sea Ice and SST model (HadISST; Rayner et al. 2003) and the National Oceanic and Atmospheric Administration (NOAA) Extended Reconstructed SST (ERSST v3; Smith et al. 2008) datasets, and the 200-hPa velocity potential results are computed from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) reanalysis data (Uppala et al. 2005).
3. The mid-1970s climate shift in the Pacific and the effects on the TBO and Asia–Pacific climate

The mid-1970s saw a significant shift of climate in the Indo-Pacific region, with the tropical Pacific SSTs transitioning from anomalously cool to warm, accompanied by many other changes in the Pacific region (e.g., Trenberth and Hurrell 1994; Mantua et al. 1997; Meehl et al. 2009). Not only was there a warming of SSTs in the tropical eastern Pacific, but the trade winds decreased in strength on average in the equatorial Pacific. This was associated with changes to interannual variability in tropical Pacific SSTs, with the post-1970s shift being less biennial (i.e., a weaker TBO) than the pre-1970s shift period (e.g., Wang and An 2001). This is reconfirmed by global wavelet spectra [see Wittenberg (2009) for a description of global wavelet spectra] of observed Niño-3.4 SSTs in Fig. 3b, where the amplitude of the wavelet spectrum power in the 2-yr band is about half the amplitude in the post-1970s shift period compared to the pre-1970s shift era, with greater power near 5 yr in the latter. Meanwhile, a new aspect of this shift is that the Indian monsoon also became less biennial after the shift, with the amplitude of monsoon rainfall power in the 2-yr band postshift also about half that in the preshift time frame (Fig. 3a).

To illustrate in a different way the changes of base state noted by Wang and An (2001) and others, Fig. 4a shows SST differences, postshift (average for 1976–2004) minus preshift (1958–75). The largest SST differences of over 0.5°C are seen in the eastern equatorial Pacific, with considerably less warming and even some small amplitude cooling in the western equatorial Pacific. However, as noted above, processes in the Indian Ocean can affect TBO amplitude, and the Indian Ocean shows relatively larger warming than the western equatorial Pacific after the 1970s shift, with positive anomalies of several tenths of a degree on this decadal time scale. The corresponding surface wind stress anomalies associated with these changes in SST are superimposed on the SST differences in Fig. 4. Focusing on the dynamically important equatorial region, westerly anomalies along the equator extend right across from

![Fig. 3. Global wavelet spectra for (a) all-India rainfall for pre-1970s shift (1958–75 average, dashed line) post-1970s shift (1976–2004 average, dotted line), and long term (1871–2006, solid line). (b) As in (a), but for Niño-3.4 SSTs.](image)

![Fig. 4. Change in base-state SSTs (colors) and surface wind stress (arrows), post-1970s shift (1976–2004 average) minus pre-1970s shift (1958–75 average). Scaling arrow at lower right is 0.045 N m⁻². Zonally averaged SST differences are at right and are calculated across the domain 154°E–120°W.](image)
the Indian Ocean to the Pacific, with largest wind stress anomalies in the eastern equatorial Pacific approaching 0.04 N m\(^{-2}\) or about 10% of the mean values. As noted above, such a weakening of the trade winds along the equator is often interpreted as a reduction of air–sea coupling strength (Zebiak and Cane 1987; Neelin et al. 1994) that is associated with longer time period interannual variability. This is consistent with the weakening of the TBO in the postshift period noted in Fig. 3.

As could be expected with a shift from negative to positive IPO (preshift to postshift; Meehl et al. 2009), the domain of the Walker circulation should also shift eastward somewhat (Arblaster et al. 2002). Figure 5 shows the annually averaged velocity potential averages at 200 hPa for preshift (Fig. 5a) and postshift (Fig. 5b). Annual means are shown to illustrate the decadal base-state changes. The differences for postshift minus preshift (Fig. 5c) show largest positive anomalies between the date line and 120°W, with the greatest negative anomalies east of 120°W. This indicates a weakening as well as a suggestion of an eastward shift (see further discussion in Arblaster et al. 2002).

Power et al. (1999) and Arblaster et al. (2002) noted that a positive phase of the IPO with the associated eastward shift of the Walker circulation (e.g., post-1970s) has a reduced correlation on the interannual time scale between Niño-3.4 and Australian monsoon rainfall. This is because an eastward shift of the rising branch of the Walker circulation over the western Pacific moves the center of upward motion away from the Australian–Asian region, and acts to isolate that area from its connection to the eastern tropical Pacific. This is

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**Fig. 5.** Velocity potential at 200 hPa for (a) pre-1970s shift (1958–75 average, contour interval = 2.0 × 10^6 m^2 s\(^{-1}\)), (b) post-1970s shift (1976–2004 average; contour interval = 2.0 × 10^6 m^2 s\(^{-1}\)), and (c) differences (postshift minus preshift, contour interval = 0.4 × 10^6 m^2 s\(^{-1}\)).
Earlier work by Capotondi et al. (2006) and Neale et al. (2008) indicated that less biennial variability (and thus weaker TBO) in equatorial Pacific SSTs is associated with a widening of the meridional domain of trade wind anomalies that are connected with eastern equatorial Pacific SST anomalies, as well as an eastward shift of the center of action for the connections between wind stress and SST anomalies. Regression of the bandpass-filtered (1–5 yr) zonal component of wind stress on to Niño-3.4 SSTs for the preshift and postshift periods (Figs. 7a,b) shows that when Niño-3.4 SSTs are anomalously warm, there are westerly wind stress anomalies (positive regression values) between about 130°W and 150°E roughly centered on the equator, with negative values east of around 130°W in the far eastern equatorial Pacific, and west of about 150°E across the Indian Ocean. These wind stress anomalies are consistent with the depiction of circulation anomalies in Fig. 1b (with signs reversed for negative SST anomalies in the eastern

![Fig. 6. Observed correlations between all-India season-averaged rainfall (JJAS) and Niño-3.4 SSTs, pre-1970s shift (1958–75, hatched bar), and post-1970s shift (1976–2004, stippled bar).](image1)

![Fig. 7. Regression of bandpass-filtered (1–5 yr) zonal component of surface wind stress onto Niño-3.4 SSTs for (a) pre-1970s shift (1958–75), (b) post-1970s shift (1976–2004), and (c) differences between postshift and preshift periods. Zonal averages are shown at right and are calculated across the domain 154°E–120°W.](image2)
equatorial Pacific) associated with the eastern and western Walker cells that link the Indo-Pacific climate system through large-scale east–west circulations in the atmosphere. For the postshift minus preshift regression differences in Fig. 7c, positive differences near 10°N and 10°S indicate a widening of the meridional domain of the trade wind anomalies in the latter period consistent with wind-forced ocean Rossby waves forming farther poleward. With their slower propagation speeds there is less of a tendency to produce transitions in SSTs on the biennial time scale (Capotondi et al. 2006). Weak positive differences from about 160°E to 160°W in Fig. 7c and negative differences to the west and east suggest an eastward shift of the coupled wind stress–SST regime in the equatorial Pacific, which would also contribute to a reduced biennial signal (Capotondi et al. 2006). However, the expanded meridional extent of the trade wind teleconnections is the stronger signal.

Thus, the mid-1970s climate shift was characterized by the IPO transitioning from a negative to a positive state with the characteristic warming of eastern tropical Pacific and Indian Ocean SSTs. This was associated with, on average, a reduced TBO in both tropical Pacific SSTs and Asia–Australia monsoon rainfall, weaker trade winds and reduced air–sea coupling strength, an eastward shift of the Walker circulation, and a widening and eastward shift of the coupling between surface wind stress and Niño-3.4 SST anomalies.

4. Climate model results for changes in base state

To examine coupled climate model realizations of conditions associated with decadal TBO variability identified above in order to gain insight into the relevant processes, comparison of 500-yr preindustrial control simulations of the CCSM3 and the CCSM4 shows that CCSM4 undergoes greater warming in the equatorial central and eastern Pacific compared to the western equatorial Indian Ocean (Fig. 8). Though there are some differences in the comparison to the mid-1970s shift in Fig. 4 in the off-equatorial regions, there are also slightly cooler conditions in the dynamically important far eastern Indian and western equatorial Pacific Oceans with a consequent weakening of the trade winds along the equator (westerly anomaly wind stress values in Fig. 8). As a consequence the TBO is weaker in CCSM4 than in CCSM3 with greater power at periods greater

![Fig. 8. Base-state differences for SST (°C) and surface wind stress (N m⁻²), 500-yr averages, CCSM4 minus CCSM3. Scaling arrow shown at lower right.](image-url)
than 2 yr in Indian monsoon rainfall (Fig. 9a) and in Niño-3.4 SSTs in the Pacific (Fig. 9b). The latter compares better to observations (Gent et al. 2011; Deser et al. 2011, manuscript submitted to J. Climate) where, for much shorter time series, there is more power near 4 yr than 2 yr in CCSM4 (Fig. 9b) and observations (Fig. 3b). For Indian monsoon rainfall in CCSM4 in Fig. 9a, the power is spread from 2 to 4 yr more uniformly as in the observations (Fig. 3a) as compared with CCSM3 where there is proportionately more power near 2 yr than near 4 yr (Fig. 9c). However, the periodicity in CCSM4 still has too little spread compared to observations.

Associated with the decreased TBO in CCSM4, the domain of the zonal wind stress anomaly connections to Niño-3.4 SSTs in the Pacific (Fig. 9b) and in Niño-3.4 SSTs in the Pacific (Fig. 9b) and observations (Fig. 3b). For Indian monsoon rainfall in CCSM4 in Fig. 9a, the power is spread from 2 to 4 yr more uniformly as in the observations (Fig. 3a) as compared with CCSM3 where there is proportionately more power near 2 yr than near 4 yr (Fig. 9c). However, the periodicity in CCSM4 still has too little spread compared to observations.

5. Climate model results for decadal IPO variability affecting TBO amplitude

There remains the question as to whether IPO-type shifts of base state in the models can produce comparable changes to the amplitude of the TBO as in the observed mid-1970s shift. Figure 12a shows the correlation between the 2–3-yr band of global wavelet spectra from Niño-3.4 and the same quantity for the Indian monsoon for the 500-yr control simulations of CCSM3 and CCSM4. CCSM3 with its large regular TBO shows a significant correlation of \( r = 0.47 \) (significant at the 5% level after taking into account autocorrelation in the unsmoothed time series following Leith (1973); this test is applied to subsequent correlation calculations in this paper). This quantity in CCSM4, with its low-amplitude TBO, drops to a statistically insignificant value of \( r = 0.11 \). This indicates that the CCSM3, with the base state more like the pre-1970s

![Fig. 10. Regression of bandpass-filtered (1–5 yr) zonal component of surface wind stress onto Niño-3.4 SSTs for (a) 500-yr CCSM3 control run, (b) 500-yr CCSM4 control run, and (c) differences for CCSM4 minus CCSM3. Zonal averages are shown at right and are calculated across the domain 154°E–120°W.](image)
period in observations, has a stronger monsoon–Pacific connection when the TBO is stronger. If the correlation between Niño-3.4 and the Indian monsoon is first computed, and then correlated with the Indian monsoon TBO in Fig. 12b (the 2–3-yr band of global wavelet spectra of Indian monsoon precipitation), the CCSM3 with its active TBO has a significant correlation of $0.52$ (significant at the 5% level), while this value for CCSM4 drops to $0.16$, again indicative of the weaker TBO in CCSM4 with its base state more like the post-1970s period (positive IPO) compared to CCSM3. A similar calculation using the biennially filtered Niño-3.4 SSTs shows a similar sign, but somewhat smaller correlations that are not statistically significant.

A separate but not unrelated mechanism from the TBO for modifying the magnitude of the monsoon–Pacific connection is the amplitude of Niño-3.4 interannual variability. Larger-amplitude SST variability in the eastern equatorial Pacific can force the east–west atmospheric connections more aggressively and thus produce a stronger monsoon–Pacific connection (e.g., Arblaster et al. 2002; Cai et al. 2009). This appears to be a contributing factor for CCSM4 compared to CCSM3 in Fig. 12c. First computing the correlation of annual mean values of Niño-3.4 with Indian monsoon rainfall, and then calculating the correlations of those values with the 10-yr sliding window standard deviation of Niño-3.4 SSTs, shows that during periods of higher-amplitude Niño-3.4 variability in CCSM4 (larger values of Niño-3.4 standard deviations on the decadal time scale), there is a stronger monsoon–Pacific SST connection (see Fig. 12c; cf. larger-amplitude correlations for CCSM4 of $0.38$ and for CCSM3 of $-0.18$).

Finally, there is the question of whether internally generated IPO variability in the control runs can modulate the amplitude of the TBO in the models. To
document the spatial structure of the IPO in the models compared to observations, the first EOF of low-pass-filtered (13 yr) detrended SSTs from the 500-yr control runs of CCSM3 and CCSM4 are shown in Figs. 13a,b, compared to what is typically interpreted as the IPO pattern from observations [second EOF of low-pass-filtered SSTs, Fig. 13c (Meehl and Hu 2006)]. In the observations, the increasing GHGs produce a warming trend that is the first EOF. Then the internally generated decadal variability, the IPO, is the second EOF. In the long control runs from the models, there are no changes in external forcing so that the internally generated IPO is the first EOF (see discussion in Meehl et al. 2009b).

The time scale of the IPO in models and observations is in the multidecadal range (spectral peaks of 10–30 yr, not shown), and all have comparable patterns. However, the CCSM3 has more of a zonally uniform sign of SST anomalies across the Indian and Pacific in the IPO, while the CCSM4 has opposite-sign anomalies in the eastern Indian and western Pacific Oceans. The observations in Fig. 13c are somewhere in between the two model simulations. However, these differences in simulated IPO patterns produce a better connection of IPO variability with the TBO in CCSM4 compared to CCSM3 in Fig. 14a. There is a stronger negative correlation of the 2–3-yr (or TBO) band of Niño-3.4 SSTs with the phase of the IPO in CCSM4 (−0.47, significant at the 5% level) compared to CCSM3 (−0.28). Additionally, the CCSM4 shows a larger negative correlation of standard deviation of Niño-3.4 (using a 10-yr sliding window standard deviation of Niño-3.4 SSTs) with the IPO (−0.55, significant at the 5% level) relative to CCSM3 (−0.22). This shows that when the IPO is in its negative phase in the models, there is higher-amplitude interannual variability of Niño-3.4 in both models (as also shown in Arblaster et al. 2002), but this connection is stronger in CCSM4.

To test this apparently stronger connection between the IPO, Niño-3.4 SSTs, and the Indian monsoon in the CCSM4, correlations between June–September (JJAS) Niño-3.4 SSTs and all-India rainfall (area average from 5°–40°N, 60°–110°E) computed from CCSM4 are computed for each IPO segment greater than 8 yr long with a normalized index of over 0.5 (Fig. 15a). Figure 15b shows that when the CCSM4 IPO is negative, there is a stronger connection between JJAS Niño-3.4 SSTs and the Indian monsoon in the model (correlation of −0.41, significant at the 5% level) compared to when the IPO is positive (−0.29).

As noted in Fig. 12c, the amplitude of Niño-3.4 SST variability can be a factor in the Indo-Pacific teleconnections. In a positive IPO phase, El Niño events occur superimposed on a warmer tropical Pacific SST base state that can provide somewhat larger absolute amplitude warm events and show a strong connection to droughts in the South Asian monsoon region. The converse can also occur for negative IPO, La Niña events, and certain monsoon floods (Krishnamurthy and Goswami 2000).

6. Conclusions

Nearly a century of research on connections between the Indian monsoon and the tropical Pacific has provided a framework within which to understand dynamically coupled air–sea processes and large-scale tropical circulations that produce variability on different time scales,
from the TBO to ENSO (likely larger-amplitude events in the TBO), the Indian Ocean dipole (IOD; the SST pattern in northern fall in the TBO), and decadal variability in the Pacific (the IPO) that could modulate Indo-Pacific connections and the TBO.

TBO mechanisms suggest that the Indo-Pacific region should be perfectly biennial, but it is periodically less biennial due to decadal variability associated with some combination of the following:

1) increased latitudinal extent of Pacific trade winds and wider cold tongue;
2) warmer tropical Pacific compared to tropical Indian Ocean that weakens trade winds and reduces air–sea coupling strength;
3) eastward shift of the Walker circulation; and
4) reduced interannual variability of Pacific and/or Indian Ocean SST.

Relevant to the present results, previous studies have shown that convective heating anomalies associated with Indian Ocean SST anomalies that transition from the IOD in northern fall to Indian Ocean basin-wide SST anomalies in northern winter in the TBO can produce changes in equatorial western Pacific wind stress and associated ocean Kelvin waves. These act to transition SST anomalies in the eastern equatorial Pacific to conditions opposite to that the year before, thus contributing to a stronger TBO.

The mid-1970s climate shift (negative to positive IPO with warmer SSTs in the tropical Pacific) is an example from observations of how the system went from more...
biennial and thus higher-amplitude TBO (preshift) to less biennial and consequently lower-amplitude TBO (postshift). Global coupled climate model simulations (cf. CCSM3 and CCSM4) illustrate how the mid-1970s shift-type changes in base state can make the Indo-Pacific monsoon–SST system less biennial with lower-amplitude TBO due to the combinations of the four factors listed above.

Analysis of the 500-yr control run from CCSM4 shows how internally generated variability associated with the IPO can modulate the TBO and Indo-Pacific connections in some ways comparable to the mid-1970s shift in observations.

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