Indo-Pacific Climate Interactions in the Absence of an Indonesian Throughflow

JULES B. KAJTAR, AGUS SANTOSO, AND MATTHEW H. ENGLAND

ARC Centre of Excellence for Climate System Science, and Climate Change Research Centre, University of New South Wales, Sydney, New South Wales, Australia

WENJU CAI

CSIRO Marine and Atmospheric Research, Aspendale, Victoria, Australia

(Manuscript received 5 February 2014, in final form 9 September 2014)

ABSTRACT

The Pacific and Indian Oceans are connected by an oceanic passage called the Indonesian Throughflow (ITF). In this setting, modes of climate variability over the two oceanic basins interact. El Niño–Southern Oscillation (ENSO) events generate sea surface temperature anomalies (SSTAs) over the Indian Ocean that, in turn, influence ENSO evolution. This raises the question as to whether Indo-Pacific feedback interactions would still occur in a climate system without an Indonesian Throughflow. This issue is investigated here for the first time using a coupled climate model with a blocked Indonesian gateway and a series of partially decoupled experiments in which air–sea interactions over each ocean basin are in turn suppressed. Closing the Indonesian Throughflow significantly alters the mean climate state over the Pacific and Indian Oceans. The Pacific Ocean retains an ENSO-like variability, but it is shifted eastward. In contrast, the Indian Ocean dipole and the Indian Ocean basinwide mode both collapse into a single dominant and drastically transformed mode. While the relationship between ENSO and the altered Indian Ocean mode is weaker than that when the ITF is open, the decoupled experiments reveal a damping effect exerted between the two modes. Despite the weaker Indian Ocean SSTAs and the increased distance between these and the core of ENSO SSTAs, the interbasin interactions remain. This suggests that the atmospheric bridge is a robust element of the Indo-Pacific climate system, linking the Indian and Pacific Oceans even in the absence of an Indonesian Throughflow.

1. Introduction

The El Niño–Southern Oscillation (ENSO) is the dominant global climate mode on interannual time scales, exerting profound impacts upon the environment and economies worldwide. Originating in the equatorial Pacific Ocean, ENSO impacts air–sea processes over remote oceans (Klein et al. 1999; Lau and Nath 2003; Liu and Alexander 2007; Du et al. 2009), generating anomalous sea surface temperature anomalies (SSTAs). These remote SSTAs can, in turn, feed back onto ENSO variability in the Pacific. A particularly strong feedback is exerted by the Indian Ocean [see Santoso et al. (2012) and references therein], to which the Pacific Ocean is directly connected via the Indonesian Throughflow.

During their growth phase, El Niño and La Niña events occasionally induce positive and negative Indian Ocean dipole (IOD) events, respectively, which peak in boreal autumn (Saji et al. 1999). The IOD is a coupled mode of variability, involving a seesaw pattern in SSTAs in the western and eastern tropical Indian Ocean. A positive phase of the IOD corresponds with anomalously cool SST in its eastern pole and anomalously warm SST in the western pole (vice versa for the negative phase). As El Niño and La Niña events peak in boreal winter, they tend to induce basinwide warming and cooling, respectively, over the tropical Indian Ocean, which is commonly referred to as the Indian Ocean basinwide mode (IOBM; Klein et al. 1999; Lau and Nath 2003; Du et al. 2009). These Indian Ocean modes of variability in turn feed back onto ENSO processes in the Pacific, influencing its period, amplitude, and thus its
predictability (e.g., Kug and Kang 2006; Luo et al. 2010; Izumo et al. 2010; Santoso et al. 2012).

It has long been believed that the primary way in which Indian Ocean climate modes feed back onto ENSO is via their influence on the atmospheric Walker circulation (Lau and Nath 2000, 2003; Alexander et al. 2002; Wu and Kirtman 2004; Annamalai et al. 2005; Behera et al. 2006; Potemra and Schneider 2007a). This can be inferred, for instance, by suppressing air–sea interactions over the Indian Ocean in climate models (e.g., Wu and Kirtman 2004; Behera et al. 2006; Dommeng et al. 2006; Santoso et al. 2012). Warm IOBM has been found to drive easterly wind anomalies over the western Pacific that act to dampen El Niño events (Santoso et al. 2012) and accelerate the transition to a La Niña event (Kug and Kang 2006; Kug et al. 2006). These easterly wind anomalies induce eastward propagating upwelling Kelvin waves along the equatorial Pacific that eventually terminate the El Niño event (Wang et al. 1999). The IOD has also been thought to influence ENSO dynamically processes in the Pacific through the atmospheric bridge, via an alteration of the Walker circulation (Izumo et al. 2010, 2014). During a negative IOD for instance, which tends to co-occur with a developing La Niña, the anomalously warm eastern Indian Ocean induces easterly wind anomalies that deepen the thermocline over the western Pacific warm pool. The eastward propagating upwelling Kelvin waves reinforce the shallow thermocline in the eastern Pacific, thus enhancing development of the ensuing La Niña. The abrupt termination of these wind anomalies at the end of the negative IOD event in November allows the anomalous warming to spread toward the eastern Pacific, preconditioning for an El Niño in the following year (Izumo et al. 2010).

The atmospheric bridge is thus an important element of the Indo-Pacific climate system. However, its role in coupling the two basins is likely to be complicated by the presence of the Indonesian Throughflow (ITF). Yuan et al. (2011, 2013) argued that the ocean channel mechanism is more important than the atmospheric bridge in the coupling between ENSO and the IOD at longer time lags. With both models and observations, they showed that IOD events can generate upwelling anomalies in the eastern tropical Indian Ocean, inducing Kelvin waves that propagate through the Indonesian seas to the equatorial Pacific Ocean.

The idea of a dominant ocean channel mechanism seems reasonable given that the ITF transports a significant volume of water and heat from the Pacific to Indian Ocean. On average, the flow rate is 10–15 Sv (1 Sv = 10$^6$ m$^3$ s$^{-1}$; Potemra 1999; Gordon 2005; Wijffels et al. 2008), and the heat transport is on the order of 0.5–1.0 PW (Vranes et al. 2002; England and Huang 2005). In addition, the ITF exhibits significant interannual variability (e.g., Meyers 1996; England and Huang 2005; van Sebille et al. 2014) controlled by pressure differences between the western Pacific and eastern Indian Ocean, which are in turn linked to ENSO and Indian Ocean variability. The ITF can also be directly influenced by changes in oceanic circulation induced by both ENSO and the IOD (e.g., England and Huang 2005; Potemra and Schneider 2007b; Sprintall et al. 2009; Sprintall and Révelard 2014).

Blocking the Indonesian Throughflow in coupled climate models significantly alters the global climate and ENSO (Schneider 1998; Wajsowicz and Schneider 2001; Song et al. 2007; Santoso et al. 2011). For instance, blocking the ITF results in weaker trade winds, flatter equatorial thermocline, and weaker upwelling across the Pacific Ocean, leading to an alteration of ENSO characteristics. This demonstrates that the ITF is an important component of the Indo-Pacific climate system. Thus, it is illuminating to assess the significance of the atmospheric bridge mechanism in light of the prominent presence of the ITF. An understanding of the role of the two available pathways may help to improve ENSO, IOD, and IOBM forecasting, and it can shed light on Indo-Pacific feedback interactions throughout Earth’s history, over which the ITF has varied substantially (e.g., Cane and Molnar 2001; Kuhnt et al. 2004).

One way to evaluate the role of the atmospheric bridge is to utilize a climate model with a closed ITF and conduct partially decoupled experiments in which the air–sea interactions are suppressed over each oceanic basin independently. Each of the two experimental elements—that is, a closed ITF and suppressed air–sea interactions—has been analyzed separately in previous studies (Santoso et al. 2011, 2012). Here we combine these elements for the first time. This experimental design enables us to assess the extent to which Indo-Pacific coupled interactions occur when the ITF is blocked. In this case, the interbasin feedback interactions, if any, will necessarily occur through the atmospheric bridge.

We proceed by outlining the numerical model and experimental design in section 2. In section 3, the change to the mean climate with ITF closed is presented, along with the transformations of the dominant Pacific and Indian Ocean climate modes. In section 4, the importance of the atmospheric bridge is assessed with the aid of partially decoupled experiments. A discussion and summary follows in section 5.

2. The climate simulations

The simulations were conducted using version 1.2 of the Commonwealth Scientific and Industrial Research
Mean climate and modes in the closed ITF control experiment

Blocking the ITF results in significant changes to the mean climate. Figure 1 shows the annual mean climate in CTRL_open and CTRL_clsd and the differences between the two experiments, featuring SST, surface wind stress, ocean surface currents, rainfall, and sea level pressure. Changes to the Pacific Ocean shown by Santoso et al.
(2011) with the earlier version of the CSIRO Mk3L model are reproduced here in Fig. 1 and briefly described for completeness. Closing the ITF dramatically changes the ocean circulation, with notable strengthening of the East Australian Current and weakening of the Agulhas Current. The equatorial Pacific thermocline slope declines, resulting in a warmer eastern Pacific that leads to higher and lower sea level pressure (SLP) in the western and eastern Pacific, respectively. This drives westerly wind anomalies and weakens the westward equatorial surface currents and upwelling. As a result, rainfall increases in the eastern Pacific and decreases in the western Pacific.

The changes to the mean climate over the Indian Ocean in the model, not discussed by Santoso et al. (2011), include cooling of the eastern Indian Ocean waters resulting from the absence of heat transported from the western Pacific via the ITF. The concurrent increase in SLP drives stronger southeasterly winds that promote equatorial upwelling and lifts the thermocline depth, thereby further cooling the eastern Indian Ocean. The most significant change to precipitation is a large decrease over the southeastern Indian Ocean, which is consistent with the cooler SST and higher SLP in that region.

The intense cooling in the eastern Indian Ocean and lesser warming in the eastern Pacific render a slowdown of the Walker circulation across the two basins. Zonal atmospheric wind speeds over the Pacific Ocean are typically reduced by half. These changes to the mean climate and Walker circulation are illustrated as a schematic in Fig. 2 and are qualitatively consistent with previous studies that used coupled climate models to study the issue (Wajsowicz and Schneider 2001; Song et al. 2007). The other elements of Fig. 2 are discussed in section 4.

Closing the ITF significantly alters the modes of climate variability as a result of changes to the mean climate upon which they evolve (Song et al. 2007; Santoso et al. 2011). Figures 3 and 4 show the spatial patterns of
the dominant empirical orthogonal function (EOF) modes for SST. The EOF analyses were performed on the full 200-yr CTRL\textsubscript{open} and CTRL\textsubscript{clsd} sets for each ocean separately and bounded by 20\degree S–20\degree N. In the Indian Ocean, the IOBM (Fig. 3a) and the IOD (Fig. 3b) are the leading modes of climate variability, explaining 22\% and 19\% of the total variance, respectively. However, when the ITF is closed, the Indian Ocean essentially exhibits only a single mode, with the first EOF mode (EOF-1) explaining 31\% of the total variance, (EOF-2 and EOF-3 correspond to only 9\% and 8\%, respectively). The spatial pattern exhibits a broad warming (or cooling) signature that extends westward from the eastern Indian Ocean.

The pattern and temporal characteristics, as shown in section 4, are unlike that of either the IOD or the IOBM in CTRL\textsubscript{open}. The mode, whose SSTA pattern is of uniform polarity, bears closer resemblance to an El Niño signature in the Indian Ocean given the equatorial region in CTRL\textsubscript{clsd} is marked by strong upwelling, with trade wind and oceanic current patterns similar to those in the Pacific Ocean. For simplicity, we will refer to this mode in the CTRL\textsubscript{clsd} experiment as the Indian Ocean mode and abbreviate it to IOM\textsubscript{clsd} to emphasize its occurrence is unique to the closed ITF experiments.

Consistent with observations, ENSO in CTRL\textsubscript{open} (Fig. 4a) is the leading mode, explaining 41\% of the total variability. ENSO-like variability persists in CTRL\textsubscript{clsd} (Fig. 4b) despite having its characteristics altered, in agreement with that reported by Santoso et al. (2011) in the earlier version of the model. The core of the ENSO SSTAs is confined farther to the east in CTRL\textsubscript{clsd}. The overall variability is reduced, primarily through the collapse of the decadal component, as the magnitude of the interannual component is largely retained (see Fig. 6f of Santoso et al. 2011), without involving any apparent change in seasonality. These alterations to ENSO are a result of the changes to the Walker circulation, which drives weaker easterly wind stresses over the equatorial Pacific (see schematic in Fig. 2).

4. Effect of suppressed air–sea interactions

To examine the response of the modes in CTRL\textsubscript{clsd} to suppressed air–sea interactions, we constructed representative SST indices for each oceanic basin that best capture the modes of variability. An Indian Ocean central index (IOC\textsubscript{I}) was constructed by averaging SST over the region of strongest variability (5\degree S–5\degree N, 50\degree–100\degree E) for the Indian Ocean mode (IOM\textsubscript{clsd}), as indicated in Fig. 3c. Note that the overall results do not change when the averaging box is shifted slightly to the south. To account for the eastward shift in the core region of the ENSO SSTAs in CTRL\textsubscript{clsd}, the Niño-3 index (5\degree S–5\degree N, 150\degree–90\degree W) was adopted for this analysis, as indicated in Fig. 4b, which is captured better by the Niño-3.4 index in CTRL\textsubscript{open}. The monthly standard deviation, power spectral densities, and autocorrelations of these two
indices are shown in Fig. 5, for CTRL$_{clsd}$, DCPI$_{PO}$, and DCPI$_{IO}$. A notable feature of IOM$_{clsd}$ is that the variability peaks during May–July (Fig. 5a), in contrast to the IOD (which peaks during August–November) and the IOBM (peaking during January–May). This seasonal phase locking in CTRL$_{clsd}$ is consistent with the peak of the southeasterly winds and equatorial upwelling in austral winter (not shown).

Suppressing air–sea interactions in either oceanic basin results in amplification of the overall variability of both Niño-3 and the IOCI. This occurs without any change to the seasonality (Figs. 5a,b). The modes operate on notably different time scales: interdecadal for IOM$_{clsd}$ (Fig. 5c) and interannual for ENSO (Fig. 5d).

The tendency for an increase in the decorrelation time scale (Figs. 5e,f), more prominently for the IOCI, corroborates a shift in the modes toward longer periodicity.

The partially decoupled experiments show that the removal of the SST mode from one basin strengthens the other, relative to the CTRL$_{clsd}$ simulations. Therefore, we conclude that damping occurs between the Pacific and Indian Ocean SST modes. Such interactions necessarily occur through the atmospheric bridge, since the ITF is blocked. Unlike the situation in CTRL$_{open}$, in which IOD and IOBM are strongly correlated with ENSO (Santoso et al. 2012), IOM$_{clsd}$ tends to occur more independently from ENSO, as evidenced by a weak positive correlation between Niño-3 and the IOCI, with a maximum correlation coefficient of approximately 0.2, occurring at zero lag (not shown). Although it is statistically significant at the 95% confidence level, the weak correlation also implies that the cool phase of IOM$_{clsd}$ can co-occur with an El Niño and the warm phase with a La Niña. The tendency for slightly more frequent occurrences of paired warm IOM$_{clsd}$ phase with El Niño and cool IOM$_{clsd}$ phase with La Niña allows the damping to occur as explained below.

The atmospheric bridge underpins the coupling between the Indian and Pacific Oceans in CTRL$_{clsd}$.
Figure 6a shows that suppressing air–sea coupling over the Indian Ocean results in strengthened equatorial zonal wind stress ($\tau^z$) variability over the eastern Pacific. This is the signature of the enhanced Niño-3 variability seen in Fig. 5b. The significant weakening of $\tau^z$ variability over the Indian Ocean (between 50° and 100°E) during March–June is due to the absence of $\text{IOM}_{\text{clsd}}$ in DCPLIO $\text{clsd}$ (Fig. 6a). The weakened variability extends across to the western Pacific (150°E–160°W) over the latter half of the year. Thus we infer, and reinforce later, that the weakened $\tau^z$ variability over the western Pacific represents weaker $\tau^z$ anomalies in that region, which leads to enhanced $\tau^z$ anomalies over the eastern Pacific and thus permits stronger ENSO events in the absence of $\text{IOM}_{\text{clsd}}$. The damping effect of ENSO on $\text{IOM}_{\text{clsd}}$ is apparent by the strengthening of $\tau^z$ variability over the Indian Ocean in DCPLPO $\text{clsd}$ (Fig. 6b). The enhanced $\tau^z$ variability manifests over the western Pacific (between 100° and 150°E) during July–October because of the removal of ENSO, and extends over the Indian Ocean (50°–100°E) during November–March.

The composite SSTA evolution of the warm and cool phases of IOM $\text{clsd}$, shown in Figs. 7a and 7b, respectively, illustrates the weak correlation between the modes. In Fig. 7 and later figures, Jul(0) corresponds to July (calendar months abbreviated) in the year of the warm or cool event, and −1 or 1 in parentheses denotes the year before or after the event. In addition to the weak correlation between warm or cool events in the Indian and Pacific Oceans, there appears to be a degree of nonlinearity. Specifically, the composites show that while the warm IOM $\text{clsd}$ phase coincides with some anomalous warming in the Pacific [Fig. 7a; east of 160°W during Jul(−1)–Jul(0)], the cool phase does not as frequently
co-occur with a La Niña (hence the weaker cool SST signature in the eastern Pacific in Fig. 7b).

The influence of IOMclsd becomes apparent when the Pacific Ocean is decoupled. Figure 7c shows easterly \( \tau^x \) anomalies over the western Pacific between Jul(0) and Oct(0) following the warm IOMclsd phase and, conversely, westerly \( \tau^x \) anomalies following the cool phase (Fig. 7d). Since this is in CTRLclsd, it is difficult to infer the origin of these anomalies, and they may, in fact, be induced by either or both the IOMclsd and ENSO modes. By decoupling the Pacific Ocean, Figs. 7e,f show that the origin of the \( \tau^x \) anomalies over the western Pacific is due in large part to the Indian Ocean SSTAs, since ENSO is absent. However, the westward shift of the anomalies in DCPIclsd indicates a degree of coupling to ENSO in CTRLclsd. The \( \tau^x \) anomalies correspond with a response in SLP that is anomalously low over the Indian Ocean and anomalously high over the Pacific for the warm IOMclsd phase, with opposite anomalies for the cool phase. This influence of IOMclsd, which is strongest during the latter half of the year (around the mature phase of ENSO), is consistent with a Kelvin wave response to zonally uniform diabatic heating over the Indian Ocean (Annamalai et al. 2005). It is further evidenced by spatially uniform rainfall changes, which can be seen by comparing Figs. 8a, b with Figs. 8c,d.

Composites of SSTAs associated with El Niño and La Niña events in Figs. 9a,b reaffirm the weak and asymmetric pairing with the warm and cool phases of IOMclsd respectively. A warming signature can be seen over the Indian Ocean during Jan(1)–Apr(1) in Fig. 9a, but a corresponding cooling signature is absent in Fig. 9b. The Indian Ocean induced easterly wind anomalies over the western Pacific enhance the wind components that are directly related to El Niño evolution (and the opposite for La Niña), which in turn exert a damping effect on ENSO variability (Santoso et al. 2012). Figures 9g,h show the differences in magnitudes of \( \tau^x \) anomalies between DCPIclsd (Figs. 9e,f) and CTRLclsd (Figs. 9c,d) for El Niño and La Niña events. For El Niño (Fig. 9g), the weakened \( \tau^x \) anomalies in the western Pacific [between 150°E and 160°W during Jul(0)–Jan(1)] due to the removal of IOMclsd are apparent, but for La Niña (Fig. 9h) it is less so. Conversely, the enhancement of the \( \tau^x \) anomalies in the eastern Pacific (between 160° and 110°W) are more pronounced for La Niña. The weakening of the \( \tau^x \) anomalies in the western Pacific is expected to be masked to some extent by the concurrent enhancement of the \( \tau^x \) anomalies to the east, since the two are linked via ENSO amplitude. Nevertheless, the ENSO magnitude for both phases is consistently enhanced because of the removal of IOMclsd induced wind stress anomalies in the western Pacific.

The damping effect of ENSO on IOMclsd is apparent by the strengthening of the \( \tau^x \) anomalies over the Indian Ocean sector in DCPIclsd (Figs. 7c,d,f) relative to that in CTRLclsd (Figs. 7c,d). The strengthening occurs over the entire 24-month span that is shown, further illustrating the shift in IOMclsd to longer periods in DCPIclsd. El Niño events induce easterly \( \tau^x \) anomalies in CTRLclsd over the western Pacific (Fig. 9c). In the absence of IOMclsd, the easterly \( \tau^x \) anomalies are stronger across the Indian Ocean basin during Jan(1)–Jul(1) (Fig. 9g). These induced \( \tau^x \) anomalies over the Indian Ocean are in response to the high SLP anomalies associated with anomalous cooling seen in Fig. 9a between 150°E and 140°W commencing in Jan (1), which appears to propagate eastward resulting from the more dominant ENSO thermocline feedback in CTRLclsd than in CTRLopen (Santoso et al. 2011). The easterly \( \tau^x \) anomalies induced by El Niño are favorable for upwelling and latent-heat-driven cooling in the Indian Ocean, and hence they exert a damping effect on the co-occurring warm phase of IOMclsd. La Niña and cool IOMclsd events interact similarly but with \( \tau^y \), SST, and SLP anomalies of the opposite signs to the scenario described for El Niño and warm IOMclsd events. Thus, when the Pacific Ocean is decoupled, the wind stress variability associated with IOMclsd is enhanced, including that over the western Pacific (Fig. 5b). This allows IOMclsd to grow stronger and persist longer.

The interactive feedback between the warm IOMclsd phase and El Niño in CTRLclsd is summarized in the
schematic of Fig. 2. In CTRLclsd, divergent $\tau^x$ anomalies manifest over the Pacific Ocean during the growth phase of an El Niño (large gray arrows). When the Indian Ocean is decoupled, the easterly component (western Pacific) is weakened, while the westerly component (eastern Pacific) is enhanced (dark green arrows) in association with the stronger El Niño. Thus the effective influence of the warm phase of IOMclsd on El Niño in CTRLclsd is through the strengthened easterly $\tau^x$ anomalies over the western Pacific (depicted by the difference between the $\tau^x$ anomaly in CTRLclsd and DCPLIOclsd; bright green arrow). Over the Indian Ocean, a westerly $\tau^x$ anomaly emerges during the warm IOMclsd phase (large gray arrow). When the Pacific Ocean is decoupled, the $\tau^x$ anomaly is enhanced (dark purple arrow) in association with the stronger IOMclsd. In this case, the effective influence of the developing El Niño on the warm IOMclsd phase is marked by the weakened westerly $\tau^x$ anomaly over the Indian Ocean (bright purple arrow denoting the difference between the $\tau^x$ anomaly in CTRLclsd and DCPLIOclsd).

5. Discussion and conclusions

Using a suite of coupled and partially decoupled climate model experiments, this study examined the importance of the atmospheric bridge on feedback interactions between the Pacific and Indian Oceans. To isolate the atmospheric bridge, any possible influence of the Indonesian Throughflow (ITF) was negated by introducing a land bridge across the Maritime Continent. First, it was shown that closing the ITF resulted in significant changes to modes of variability linked to changes in the mean climate. Over the Pacific, the core of the ENSO SSTAs shift eastward into the Niño-3 region, and the IOBM and IOD collapse into a single dominant Indian Ocean mode (IOMclsd).

With the ITF closed, further experiments were conducted with air–sea interactions suppressed, firstly over the Indian Ocean, and then separately over the Pacific. Decoupling in this way eliminates any possible influence of modes of variability in that ocean basin on the other, since SSTAs in that basin are prohibited from perturbing
the atmosphere. It was shown that when one ocean is decoupled, SST variability over the other ocean basin is enhanced. Thus, it was inferred that the modes of variability in opposite basins act to dampen one another. This negative feedback occurs despite the fact that the occurrences of Indian Ocean SSTAs appear to be independent of the ENSO mode, and vice versa, unlike in the case when the ITF is open (Santoso et al. 2012). Nevertheless, the simulations produce a slightly stronger tendency for a warm Indian Ocean to co-occur with an El Niño. The damping effect is shown to occur through this combination. Specifically, the warm Indian Ocean SSTAs induce easterly winds over the western Pacific that exert a damping effect on the ensuing El Niño. Removing IOMclsd weakens these wind anomalies, and thus strengthens the ENSO mode. El Niño is also shown to induce easterly wind anomalies over the Indian Ocean that has a cooling effect through upwelling and evaporation. Such conditions are not favorable for the generation of the warm IOMclsd phase. This mechanism is similar but with anomalies of the opposite sign, for La Niña and the cool IOMclsd phase. The removal of the mode in each basin through the partial decoupling technique thus strengthens the mode in the other basin. With the ITF blocked, Indo-Pacific interactions can only occur via the atmospheric bridge.

The damping influence of the Indian Ocean on ENSO with the ITF closed is analogous to the results with ITF open. Santoso et al. (2012) found that variability of ENSO is enhanced by a similar magnitude when the Indian Ocean is decoupled. With the ITF open, they revealed that the IOBM influences equatorial zonal wind stress $\tau_x$ over the Pacific, which acts to dampen ENSO. Specifically, the warm phase of the IOBM, which generally follows the peak of an El Niño, induces easterly $\tau_x$ anomalies over the western Pacific Ocean. This weakens the westerly $\tau_x$ anomalies that are conducive for the Bjerknes coupled air–sea feedback, so this consequently results in a weaker El Niño phase. The converse applies for the cool IOBM phase and La Niña. Thus, suppressing air–sea interactions in the Indian Ocean with the ITF open weakens $\tau_x$ variability over the western Pacific during January–April because of the removal of the IOBM, which generally coincides with the decay phase of ENSO events. We have found that a similar mechanism exists when the ITF is closed, but with an altered seasonality. The variability of the dominant climate mode in the Indian Ocean peaks during May–July with a much longer persistence, so it tends to influence ENSO during the latter half of the year, corresponding to its growth phase.

The results of Santoso et al. (2012) were consistent with Dommenget et al. (2006), who also found that Indian Ocean variability acts to dampen ENSO. Both sets of authors also agreed that the ENSO period becomes longer with the Indian Ocean decoupled. Some earlier studies had concluded that coupling ocean modes tends to increase the variability of ENSO (e.g., Barsugli and Battisti 1998; Yu et al. 2002; Wu and Kirtman 2004), but many of these were based on a single experiment with a shorter run time (on the order of 50 yr) or used an overly simplified GCM.
In the present study, it is somewhat surprising that the atmospheric bridge mechanism remains strong despite the fact that, when the ITF is closed, the core region of the ENSO SSTAs is shifted eastward, thereby increasing the spatial separation from the Indian Ocean. The maintenance of the interbasin interactions is also reflected in the influence of the Indian Ocean mode, which is significant in CTRLclsd. This is consistent with the uniform polarity pattern of the tropical SSTA of IOMclsd. Such structure is more conducive for a stronger Kelvin wave response than if it were of a dipole pattern, which would otherwise generate an interference of Kelvin waves of opposite signs (Annamalai et al. 2010). We also note that the potential influence of SSTAs over the Indonesian seas on the coupling between the two basins (Annamalai et al. 2010) has not been explicitly assessed here. The present study demonstrates nonetheless that the atmospheric bridge is a robust element of the Indo-Pacific climate that would allow complex climate feedback interactions to occur even in the absence of the oceanic channel.

While the ITF has never been completely blocked in the real system [see Santoso et al. (2011) and references therein] and the behavior of the atmospheric bridge may be a function of changes in the mean climate, our results point to the possibility that modes of variability in the Indian and Pacific Oceans have been in constant interaction throughout Earth’s history. This could have important implications for our understanding of Indo-Pacific climate variability in the context of past and future climates.

Acknowledgments. This study was supported by the Australian Research Council (ARC) Centre of Excellence for Climate System Science. The model simulations were conducted on the NCI National Facility in Canberra, which is supported by the Australian Commonwealth government. W. Cai is supported by the Australian Climate Change Science Programme. We thank the three anonymous reviewers for their
comments and suggestions, which helped to greatly improve the manuscript.

REFERENCES


