Interactive Feedback between ENSO and the Indian Ocean

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ABSTRACT

A feedback process of the Indian Ocean SST on ENSO is investigated by using observed data and atmospheric GCM. It is suggested that warming in the Indian Ocean produces an easterly wind stress anomaly over Indonesia and the western edge of the Pacific during the mature phase of El Niño. The anomalous easterly wind in the western Pacific during El Niño helps a rapid termination of El Niño and a fast transition to La Niña by generating upwelling Kelvin waves. Thus, warming in the Indian Ocean, which is a part of the El Niño signal, operates as a negative feedback mechanism to ENSO. This Indian Ocean feedback appears to operate mostly for relatively strong El Niños and results in a La Niña one year after the mature phase of the El Niño. This 1-yr period of phase transition implies a possible role of Indian Ocean–ENSO coupling in the biennial tendency of the ENSO. Atmospheric GCM experiments show that Indian Ocean SST forcing is mostly responsible for the easterly wind anomalies in the western Pacific.

1. Introduction

Since Bjerknes (1969), considerable research effort has been devoted to gaining a better understanding of ENSO dynamics. In the accumulating of our understanding of the ENSO, various conceptual theories were suggested to explain the self-sustained oscillation of the ENSO. Among them, the delay oscillator (Schopf and Suarez 1988; Battisti and Hirst 1989) and the recharge oscillator (Jin 1997a,b) are two of the most popular theories. Both theories explain the developing mechanism of ENSO in a similar manner but suggest explicitly different phase transition mechanisms. However, both emphasize basinwide ocean adjustment processes in nonequilibrium with wind stress forcing and the role of oceanic waves in the tropical Pacific basin (An and Kang 2001).

On the other hand, several recent studies have focused on a role for the western Pacific wind in the ENSO transition (Weisberg and Wang 1997a,b; C. Wang et al. 1999; B. Wang et al. 1999, 2001). They emphasized a role for anomalous easterly (westerly) winds in the western Pacific during the mature phase of El Niño (La Niña). The wind anomalies excite equatorial Kelvin waves, which play a role in the ENSO transition. However, whether the mechanisms, suggested by previous studies, can fully explain wind anomalies in the western Pacific is being disputed.

In terms of causing wind anomalies over the western Pacific, Weisberg and Wang (1997a) and their subsequent studies emphasized a role of local cold SST anomalies induced by Ekman pumping. According to their argument, the El Niño–related anomalous convection in the equatorial central Pacific induces a pair of off-equatorial cyclones with westerly wind anomalies on the equator. These equatorial westerly winds act to deepen the thermocline and increase SST in the equatorial east-central Pacific, thereby providing a positive feedback for anomaly growth. At the same time, these off-equatorial cyclones raise the off-equatorial thermocline by Ekman pumping. Thus, a relatively shallow off-equatorial thermocline anomaly develops over the western Pacific, leading to a decrease in SST and an increase in sea level pressure (SLP). During the mature phase of El Niño, this off-equatorial high SLP initiates equatorial easterly winds over the far western Pacific.

Wang and his collaborators (B. Wang et al. 1999, 2000, 2001; Wang and Zhang 2002) have suggested that the easterly wind anomalies are associated with the anomalous Philippine Sea anticyclone. In contrast to Weisberg and Wang (1997a,b), they emphasized the role of local air–sea interaction. The anomalous anticyclone winds superposed on mean northeasterly trades.
during the boreal wintertime increase the total wind speed, thus enhancing evaporation and entrainment to the east of the anticyclone. This favors sea surface cooling in front of the anticyclone. Thus, the cold SST is located to the east of the anticyclone. The phase shift between the anticyclone and sea surface cooling implies a local feedback that can amplify and sustain the Philippine Sea anticyclonic wind. The anomalous equatorial easterly wind can be regarded as a part of the anticyclone.

Previous studies have recognized that the easterly winds are caused by the off-equatorial cold SSTs and their resultant anticyclonic flow. However, there is increasing research interest on covariability between the Indian Ocean and the western Pacific (e.g., Li et al. 2002; Watanabe and Jin 2002, 2003; Wang et al. 2003; Misra 2004; Kug et al. 2005). In particular, Watanabe and Jin (2002) showed, using their moist linear baroclinic model, that the Indian Ocean SST plays a role in developing the Philippine Sea anticyclone. In a similar manner, we will suggest that the easterly anomalies in the equatorial western Pacific are closely related to Indian Ocean warming. In addition, we will emphasize that the Indian Ocean SST plays a crucial role in the tropical Pacific ENSO variability.

So far, a number of studies have focused on the impact of ENSO on the Indian Ocean (e.g., Klein et al. 1999; Venzke et al. 2000; Baquero-Bernal et al. 2002; Xie et al. 2002; Lau and Nath 2000, 2003; Krishnamurthy and Kirtman 2003). To a large extent, they agreed that the ENSO-related Indian Ocean SST variability is mostly affected by modulation of the Walker circulation. However, little research has been devoted to the impact of the Indian Ocean SST on the ENSO. During a few recent years, some studies have suggested that the Indian Ocean variability can modulate the ENSO variability (e.g., Yu et al. 2002; Saji and Yamagata 2003; Wu and Kirtman 2004; Annamalai et al. 2005; Kug et al. 2005). In particular, Yu et al. (2002) and Wu and Kirtman (2004) examined the impacts of the Indian Ocean on the ENSO variability by performing experiments with a coupled atmosphere–ocean general circulation model (GCM), respectively. Both studies showed that a coupled GCM simulation of ENSO including both the tropical Indian and Pacific Oceans tends to be a stronger ENSO variability than that including the tropical Pacific only. However, they gave opposite results in terms of the dominant ENSO period. Yu et al. (2002) showed the dominant period of the ENSO increases from about 4 yr in the Pacific run to about 4.4 yr in the Indo-Pacific run, while Wu and Kirtman (2004) showed that the dominant ENSO period decreases from about 2.3 yr in the Indian Ocean coupled run to about 2.8 yr in the decoupled run. These controversial results should be further studied using observational data and experimental modeling.

In this study, we will show that the Indian Ocean SST anomalies affect ENSO variability using observational data. We will show that this coupling process is generated by a modulating Walker circulation and altering wind variability over the western Pacific. We will also demonstrate that the anomalous easterly wind in the western Pacific during El Niño helps a fast transition to La Niña by generating upwelling Kelvin waves.

Section 2 introduces the observed data and the model utilized. In section 3, we show a clear relationship between ENSO and the Indian Ocean SST variability using observed data, and the dynamical coupling process between ENSO and the Indian Ocean will be described. Atmospheric GCM experiments support our arguments in section 4. The summary and discussion are given in section 5.

2. Data and model

The data utilized are monthly means of sea surface temperature (SST), atmospheric circulation data, and ocean temperature over the tropical domain, 30°S–30°N. The data period is 43 yr from 1958 to 2000. Anomalies presented in this study were detrended after removing climatology for 43 yr. The SST data were obtained from the National Centers for Environmental Prediction (NCEP) and were constructed based on the EOF of observed SST (Reynolds 1988) and reconstructed after January 1981 using the optimum interpolation technique (Reynolds and Smith 1994). Atmospheric circulation data were taken from NCEP–National Center for Atmospheric Research (NCAR; Kalnay et al. 1996). They have a horizontal resolution of 2.5° × 2.5°.

The ocean temperature data utilized are the monthly means of the Simple Ocean Data Assimilation (SODA) product (Carton et al. 2000a,b). Xie et al. (2002) show that the SODA data have a good agreement with the expendable bathythermograph (XBT). It is available at 1° × 1° resolution in the midlatitudes and 0.45° × 1° latitude–longitude resolution in the Tropics, and has 20 vertical levels with 15-m resolutions near the sea surface.

An atmospheric GCM is used to reproduce the observed characteristics of atmospheric responses to given SST anomalies. The GCM is a spectral model with a triangular truncation at wavenumber 42 and has 20 vertical levels. The model was originally developed at the University of Tokyo and modified at Seoul National University (SNU). The physical processes included are
the Nakajima two-stream scheme for longwave and shortwave radiation (Nakajima et al. 1995), the Relaxed Arakawa–Shubert scheme (Moorthi and Suarez 1992), shallow convection, land surface processes, gravity wave drag, and planetary boundary layer processes (Kim et al. 1998). Kim et al. (1998) showed that the SNU GCM reasonably simulates the climatological mean patterns of tropical circulation and their anomalies during El Niño.

3. Observational analysis

As mentioned in the introduction, the equatorial wind anomalies in the western Pacific may be important for decaying the eastern Pacific SST anomalies associated with El Niño and La Niña. To examine what is of relevance to the zonal wind anomalies, a western Pacific (WP) zonal wind index is defined as a 1000-hPa zonal wind area averaged over 5°S–5°N, 120°–150°E from November to January when the typical ENSO events are in their mature phase. The WP wind index represents the strength of the zonal wind near the equatorial western Pacific boundary. In fact, the index region includes land and inland sea, where the wind cannot influence ENSO variability. It seems to be appropriate to remove the land and inland sea region when the WP wind index is calculated. However, we found that the time series with and without masking out those regions are very close, as the correlation coefficient between them is 0.97. Therefore, for simplicity we use the time series without masking out the land and inland sea regions in this study.

Figure 1 shows lag-correlation coefficients between the zonal wind index and equatorial SST averaged over 5°S–5°N during the years 1958–2001. As reported from previous studies (e.g., C. Wang et al. 1999; Wang et al. 2001), the zonal wind near the equatorial western boundary is simultaneously correlated to the eastern Pacific SST. The correlation coefficients are greater than 0.4, having a 95% confidence level. In addition, the wind index is correlated to the local western Pacific SST with a 95% confidence level as pointed out by Weisberg and Wang (1997a,b). However, Fig. 1 also shows that the wind index is well correlated to the SST in the western Indian Ocean with 1–2-month lag. The correlation coefficients are greater than 0.6, having a 99% confidence level. This means that the establishment of the equatorial zonal wind is more closely related to the Indian Ocean SST forcing rather than the eastern Pacific SST and the western Pacific SST.

It is also interesting that the WP zonal wind index is well correlated with the next year’s eastern Pacific SST with a 99% confidence level. That is, the easterly wind anomaly near the western boundary during an El Niño mature phase leads a rapid termination of the El Niño and sequentially fast phase transition to La Niña within 1 yr, judging the time scale of the typical ENSO as being 3–5 yr. One can explain the relation as follows: during an El Niño mature phase, the easterly wind anomalies generate upwelling Kelvin waves, which propagate eastward, and accelerate to break up the warm SST. Therefore, the transition from El Niño to La Niña is progressed rapidly within 1 yr. This procedure will be shown later in this section.
To show the role of the Indian Ocean SST on ENSO clearly, we defined a western Indian Ocean SST index (WISST\textsubscript{ON}),\textsuperscript{1} area averaged over 10°S–10°N, 55°–75°E, where the correlation with WP zonal wind index is relatively high during October–November. Figure 2 shows the time series of the Niño-3.4 SST and WISST\textsubscript{ON}. The Niño-3.4 SST is averaged over 5°S–5°N, 170°–120°W during November–January. The correlation coefficient between two indices is 0.62. The high correlation is not an unexpected result as several studies reported (e.g., Allan et al. 2001; Xie et al. 2002; Baquero-Bernal et al. 2002; Wang et al. 2003). That is, warming and cooling events over the western Indian Ocean are related to El Niño and La Niña events. It is worthwhile to note that not all ENSO events concur with the Indian Ocean SST events. For example, the 1965/66, 1968/69, 1976/77, 1986/87, and 1991/92 El Niño events were not accompanied by Indian Ocean warming.

Another interesting relation is found in Fig. 2. Some El Niño events concurred with the Indian Ocean warming as we mentioned. In addition, in the following year, the tropical Pacific is mostly in its cold phase. To clearly show this, we have summarized these relations in Tables 1 and 2. There are five cases where WISST\textsubscript{ON} is larger than its standard deviation as listed in Table 1.

All positive WISST events coincided with El Niño events. Following the El Niño events, the phase transition progressed rapidly, and La Niña develops in the winter of the following year. We checked that this relationship is very robust for the long historical SST (1885–2000). For this period, there were 16 warm western Indian Ocean cases, more than its one standard deviation. Among them, 13 cases led to the La Niña state in the next year (not shown).

Similarly, when the WISST\textsubscript{ON} indicates a cold event, it is mostly in the La Niña phase, in the tropical Pacific, and then the El Niño rapidly develops during the next year, as listed in Table 2. Though some exceptional cases exist, such as the years 1960 and 1974, Tables 1 and 2 show clear evidence for the close relation between ENSO and the Indian Ocean.

![Figure 2](image)

**Figure 2.** Time series of the Niño-3.4 SST during November–January (bar) and WISST\textsubscript{ON} (triangle).

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\textsuperscript{1} The subscript “ON” denotes the averaged value for October–November.

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<table>
<thead>
<tr>
<th>Positive WISST yr</th>
<th>ENSO during ND(0)J(1)</th>
<th>Next year ENSO during ND(1)J(2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1963</td>
<td>Weak El Niño</td>
<td>Weak La Niña</td>
</tr>
<tr>
<td>1972</td>
<td>El Niño</td>
<td>La Niña</td>
</tr>
<tr>
<td>1982</td>
<td>El Niño</td>
<td>Weak La Niña</td>
</tr>
<tr>
<td>1987</td>
<td>El Niño</td>
<td>La Niña</td>
</tr>
<tr>
<td>1997</td>
<td>El Niño</td>
<td>La Niña</td>
</tr>
</tbody>
</table>
There is a vigorous debate within the scientific community concerning the physical reality of the dipole mode in the Indian Ocean and its terminology (e.g., Hastenrath 2002, 2003; Yamagata et al. 2003; Lau and Nath 2004). Nevertheless, in this study we just use the terminology of “Indian Ocean dipole mode” for the sake of simplicity.

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**Table 2. Same as in Table 1, except for cold WISST events.**

<table>
<thead>
<tr>
<th>Negative WISST year</th>
<th>ENSO during ND(0) J(1)</th>
<th>Next-year ENSO during ND(1) J(2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1960</td>
<td>Normal</td>
<td>Normal</td>
</tr>
<tr>
<td>1964</td>
<td>Weak La Niña</td>
<td>El Niño</td>
</tr>
<tr>
<td>1971</td>
<td>La Niña</td>
<td>El Niño</td>
</tr>
<tr>
<td>1974</td>
<td>Weak La Niña</td>
<td>La Niña</td>
</tr>
<tr>
<td>1975</td>
<td>La Niña</td>
<td>Weak El Niño</td>
</tr>
<tr>
<td>1985</td>
<td>Normal</td>
<td>El Niño</td>
</tr>
<tr>
<td>1996</td>
<td>Normal</td>
<td>El Niño</td>
</tr>
</tbody>
</table>

Figure 3 shows lag correlations between WISST\textsubscript{ON} and the equatorial SST (5°S–5°N) during the years 1958–2000. Over the Indian Ocean, the correlation is high owing to their autocorrelation. The autocorrelations of WISST\textsubscript{ON} are significant from the previous summer to the next spring with a 95% confidence level. In the central to eastern Pacific, the correlations are also significant with a 99% confidence level. It is consistent with the results in Fig. 2 and Tables 1 and 2. Note that the correlation in the central Pacific is relatively high when tropical Pacific SST does not lag but leads WISST\textsubscript{ON}. In addition, WISST\textsubscript{ON} is negatively correlated with SST anomalies over the eastern Indian and western Pacific Oceans with a 95% confidence level. One may imagine that the negative correlation over the eastern Indian Ocean is related to the so-called Indian Ocean dipole mode\textsuperscript{2} (Saji et al. 1999). In fact, the correlation between WISST and the Indian dipole index from Saji et al. (1999) is 0.76. Most WISST events from Tables 1 and 2 were coincident with the Indian Ocean dipole events although there are some exceptional cases. For example, the 1987 warm Indian Ocean case and the 1985 cold Indian Ocean case were not coincided with the positive and negative dipole events, respectively.

In addition, WISST\textsubscript{ON} is well correlated with SST anomalies over the eastern Indian and western Pacific Oceans during the following summer as shown in Fig. 3. Once the warm (cold) SST anomalies are established, they can play a role of expediting the development of La Niña (El Niño) by generating and maintaining anomalous easterlies (westerlies) as previous studies pointed out (e.g., Weisberg and Wang 1997a; Wang et al. 2001). It is worthwhile to note that WISST is negatively correlated to the eastern Pacific SST during the following winter with a 95% confidence level. Roughly speaking, it is a 13-month lag relation. These results are shown in Tables 1 and 2.

So far, we showed that WISST\textsubscript{ON} has a close relationship with a fast ENSO transition. However, one may doubt that the relationship is just a consequence introduced by internal ENSO dynamics since WISST is well correlated with ENSO variation. To clarify this problem, we removed the partial influence of ENSO on the wind and SST anomalies. To do this, we introduce the partial correlation (Cohen and Cohen 1983). Saji and Yamagata (2003) used this method in order to show characteristics of the Indian Ocean dipole mode, which is not introduced by ENSO.

Figure 4 shows the partial correlation coefficients of the equatorial zonal wind (5°S–5°N) on Niño-3.4 SST during November–January (dashed line) and WISST\textsubscript{ON} (solid line), after accounting for the effect of the other, respectively. When the effect of WISST\textsubscript{ON} is removed, the warm Niño-3.4 SST is well correlated with westerlies in the central Pacific and the easterlies in the Indian Ocean. However, in the western Pacific, there are opposite signs between east and west of 130°E, inferring weak contribution of the Niño-3.4 SST to the western Pacific zonal wind anomaly. On the other hand, WISST\textsubscript{ON} has a strong correlation with the western Pacific wind when the effect of the Niño-3.4 SST is excluded. In addition, there is a negative correlation in the central Pacific, inferring that the Indian Ocean SST anomaly plays a role of not only enhancing the anomalous easterlies in the western Pacific but also partly reducing the ENSO-induced anomalous westerlies in the central Pacific.

Figure 5 shows the relative contributions of Niño-3.4 SST during November–January and WISST\textsubscript{ON} on the evolution of equatorial SST (5°S–5°N), representing the partial correlation coefficient. The results clearly show that the Niño-3.4 SST is related to the western Pacific SST cooling during the mature phase and to the Indian Ocean warming during the following spring (Klein et al. 1999; Xie et al. 2002). On the other hand, WISST\textsubscript{ON} is responsible for the western Pacific warming and the eastern Pacific cooling during the following year. These results support our argument that WISST\textsubscript{ON} leads to a fast transition to La Niña.

So far, we have shown a clear covariability of the tropical Pacific and Indian Oceans. These results can be interpreted as the following processes. During a summer in which El Niño is developing, the warm SST anomaly in the tropical Pacific modulates the Walker circulation by enhancing convection in the central Pa-

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cific and suppressing convection in the Maritime Continent. As a result, anomalous easterly wind stress is induced over the Indian Ocean. The wind stress and the related wind stress curl produce downwelling Rossby waves in off-equator regions and lead to SST warming in the western side of the Indian Ocean (Webster et al. 1999; Li et al. 2002; Xie et al. 2002). In addition, a weak Indian summer monsoon, which negatively correlated to the eastern Pacific SST (Shukla and Paolino 1983; Webster and Palmer 1997), tends to reduce the cross-equatorial wind over the western Indian Ocean, inferring that the SST is warming by reducing latent heat flux.

Once the warm SST anomalies are established, it can switch to a local air–sea feedback over the Indian Ocean and this then enhances the easterly wind anomaly in the equatorial Indian Ocean (Webster et al. 1999). Therefore, a warm SST anomaly further develops in the western Indian Ocean associated with ENSO. To a large extent, these processes are very similar to the developing mechanism of the Indian Ocean dipole mode (Saji et al. 1999; Webster et al. 1999), although external ENSO forcing is necessary for triggering the air–sea coupled feedback over the Indian Ocean (Allan et al. 2001; Lau and Nath 2004).

As previously mentioned, the warm SST in the west-
ern Indian Ocean induces an anomalous local Walker circulation accompanying the easterly wind stress over the Indian Ocean. Associated with the anomalous Walker circulation, the easterly wind is established over the western Pacific during the mature phase of ENSO. As a result, the Indian Ocean SST affects ENSO’s evolution by modulating the Walker circulation.

To examine the above processes, we carried out a composite analysis. For the composite, five cases were chosen in which WISST is greater than its standard deviation (0.30°C), as listed in Table 1. To compare the WISST events, we chose five other cases, which are El Niño events but not warm WISST events. The chosen events were the 1965/66, 1968/69, 1976/77, 1986/87, and 1991/92 cases. The 1977/78 and 1994/95 cases were excluded because WISST was relatively large though its value was less than the standard deviation. Hereafter we call the former the “WISST” composite and the latter is the “El Niño only” composite. The composite analysis was carried out for 2 yr, of the developing (year 0) phase and the following (year 1) phase.

Figure 6 shows the evolution of the Niño-3.4 SST and WISST index for WISST and El Niño-only composites. For the WISST case, both the SST anomalies slowly develop during the first year of the composite. It tends to be that the warming of Niño-3.4 SST leads the warming of WISST. Following the mature phase, WISST slowly decays while the Niño-3.4 SST rapidly decays and represents the La Niña state in the latter part of the second year. For the El Niño–only case, however, the WISST is nearly zero, which means that there is no relation with the Niño-3.4 SST. Two El Niños slowly decay and three El Niños develop another warm peak following the first warm peak. It is interesting that the El Niños of the WISST case are relatively strong compared to those of the El Niño–only case. It is conceivable that strong SST forcing over the tropical Pacific is required to accompany an Indian Ocean warming event.

Detailed SST patterns and their related wind anomalies are shown in Figs. 7 and 8. During the boreal summer time of the developing year, warm SST anomalies develop in the central-eastern Pacific and western Indian Ocean, in the case of the WISST composite. These warm anomalies are related to anomalous westerlies in the western-central Pacific and anomalous easterlies in the eastern Indian Ocean, respectively. Note that a cyclonic flow over the Philippine Sea tends to enhance the western Pacific westerlies.

During the boreal autumn time, the SST and wind anomalies have been further developed by the SST and surface wind feedback (e.g., Bjerknes 1969) in the Pa-
cific and Indian Oceans, respectively. Also, the westerlies in the Pacific have gradually propagated southeastward as Harrison (1987) described. This is accompanied by an eastward movement of the cyclonic flow over the Philippine Sea. In addition, an anticyclonic flow is established in the west of the cyclonic flow as Wang and Zhang (2002) mentioned. Also, an anticyclonic flow in the southern Indian Ocean (SIO) has been developing explosively (Wang et al. 2003). This anticyclonic flow is closely related to the equatorial easterlies and the WISST warming.

During the boreal wintertime, the wind system has moved eastward overall compared to previous seasons. And the wind system over the warm pool region seems to be controlled by two great anticyclonic flows. While the SIO anticyclone has weakened, the Philippine Sea anticyclone has dramatically developed. The warm SST has slightly shifted to east in the Indian Ocean, as shown in Figs. 3 and 7. Possibly related to this evolution, anomalous easterlies are established in the western Pacific. Recently, Watanabe and Jin (2002) showed, using their moist baroclinic model, that Indian Ocean warming is critical for developing the Philippine Sea anticyclone as well as the local SST forcing does. The Philippine Sea anticyclone can be directly connected to the establishment of the equatorial wind anomalies as shown in Fig. 8c. In addition, it seems that the western Pacific wind anomalies are closely related to the Indian Ocean wind anomalies including the SIO anticyclone. Annamalai et al. (2005) also showed using their atmospheric GCM experiments that the warm Indian Ocean SST anomalies generate an atmospheric Kelvin wave associated with surface easterly flow over the equatorial western-central Pacific.

After the anomalous easterlies are established, the SST anomalies in the eastern Pacific rapidly decay and fall to near zero, while the western Pacific easterlies have enhanced (Figs. 7d and 8d). And during the following summer time the SST anomalies have developed to a La Niña state, the easterlies are further developed.
in the central-western Pacific by the positive air–sea coupled feedback (Bjerknes 1969).

Unlike the WISST composite, the El Niño–only composite does not show any significant SST and surface wind anomalies over the Indian Ocean. Also, the wind anomalies in the western Pacific are relatively weak during the boreal summer and fall in the El Niño–developing year (Figs. 8a,b). During the boreal winter-
time, the wind pattern shows anomalous westerlies in the equatorial Indian Ocean and anomalous northeasterlies in the northwestern Pacific, similar to that of the WISST composite but the magnitude is weak. In addition, we found that anomalous easterlies are in the equatorial western Pacific, though they are of little significance and their magnitude is weak. There are two possible explanations for the weak magnitude. One is...
that the Indian Ocean does not work as we argued in this study. The other is that the El Niños themselves are weak compared to those of the WISST composite as shown in Figs. 6 and 7. To separate the two effects, we calculated the ratios of variables between the WISST and El Niño–only composites during November–January. The ratio for Niño-3.4 SST is about 1.52. We also calculated this ratio for the central-eastern Pacific zonal wind anomalies, averaged over 5°S–5°N, 170°E–100°W, where anomalous westerlies dominate. We found the ratio to be 1.34. The value is comparable with that of the ratio for the Niño-3.4, indicating that the wind anomalies depend on the strength of the El Niño strength. Next, we calculated this ratio for the WP zonal wind index, averaged over 5°S–5°N, 120°E–150°E. This ratio is about 7.40, clearly indicating that the Indian Ocean SST plays a very critical role in developing the wind anomalies in the western Pacific.

Figure 9 shows anomalous Walker circulations for WISST and El Niño–only composites. The zonal wind and pressure velocity are averaged over 5°S–5°N. A clear difference is found between the two composites during an El Niño–developing year. The WISST composite shows two distinctive vertical-longitude cells over the Pacific and Indian Oceans, while the El Niño–only composite shows a single cell over the Pacific basin. Note that the two cells of the WISST composite seem to move slowly eastward as El Niño develops. Therefore, the anomalous surface easterlies are confined to the Indian Ocean during the boreal summer and fall, but during winter the easterlies are extended into the western Pacific so that the Indian Ocean can affect the El Niño transition. In the case of the El Niño–only composite, a distinctive descending motion is found over the western Pacific, accompanied by a massive ascending motion associated with the warm SST in the central-eastern Pacific. Unlike the WISST composite, however, the descending motion does not accompany the ascending motion over the Indian Ocean during the boreal fall and wintertime. Therefore, the coupling between ENSO and the Indian Ocean does not operate.

Figure 10 shows a structure of ocean temperature along the equator. For the WISST composite, the subsurface temperature is overwhelmingly established in the eastern Pacific during the developing phase of an El Niño. However, a dramatic change in the ocean temperature occurs during the El Niño mature phase. The warm temperature is corrupted in the eastern Pacific, while cold temperatures develop explosively in the western Pacific. The abrupt establishment of the anomalous easterlies in the western Pacific will be responsible for the changes of the ocean temperature. During the boreal spring of the following year, a cold subsurface temperature anomaly has developed over the whole Pacific basin even though the SST represents a warm phase. This anomalous cold ocean temperature leads to a La Niña state. Note that the temperature anomaly is statistically significant, having a 99% confidence level in the central Pacific. Once the cold SST appears in the central and eastern Pacific, it is further developed by the air–sea coupled process. On the other hand, for the El Niño–only composite, the subsurface temperature anomaly is relatively weak in the western Pacific since the anomalous easterly wind is weak, as shown in Fig. 8c. Therefore, the El Niño slowly decays since its mature phase.

As shown in Fig. 10, the western Pacific wind anomalies significantly affect the structure of the ocean temperature over whole Pacific basin, though wind forcing is confined to the western Pacific during an ENSO mature phase. The western Pacific wind anomalies excite oceanic Kelvin waves, which rapidly propagate eastward, then change the oceanic vertical structure.

4. Atmospheric GCM experiments

So far, we have suggested, using the observational data, that the Indian Ocean SST warming affects the development of the anomalous easterlies in the western Pacific and so finally influences the ENSO transition. Although the statistical analysis clearly shows the relationship between the Indian Ocean and the wind anomalies, additional experimental modeling may help the understanding of our arguments in the present study. Also, experimental modeling may explicitly separate the impact of the Indian Ocean SST from that of the Pacific SST forcing.

To reveal the influence of the Indian Ocean SST on the western Pacific wind anomaly, atmospheric GCM experiments were designed, as listed in Table 3. The GCM utilized is a GCM developed at Seoul National University (SNU). All experimental conditions are identical except for SST boundary conditions. The SST boundary condition is obtained by superposing climatological SSTs on seasonally varying SST anomalies. The SST anomalies are obtained from the composites of El Niño and warm WISST events, as shown in Table 1 (1972/73, 1982/83, 1987/88, and 1997/98). The 1963/64 case was excluded due to its small amplitude. Nevertheless, the SST patterns are very similar to those in Fig. 7. In the EXPglo runs, SST anomalies are imposed in the global ocean. However, in the EXPPac and EXPfind, the SST anomalies are only used over the tropical Pacific Ocean (30°S–30°N, 120°E–80°W) and Indian Ocean (30°S–30°N, 40°–120°E) domains, respectively. These sets of the experiments are expected to
reveal the individual and combined effects of SST anomalies on the western Pacific wind variability. The model was integrated for 2 yr with the temporally varying SST anomalies in order to simulate consecutive atmospheric responses from developing stage to decaying stage of the El Niño. Using different initial conditions, 10 ensemble members were integrated. Our experimental design is very similar to that of Annamalai et al. (2005), who tried to access the relative importance of remote (Pacific) versus local (Indian Ocean) SST anomalies in determining the equatorial Indian Ocean variability.

Fig. 9. Same as in Fig. 7, except for the streamline of the zonal wind and pressure velocity anomalies averaged over 5°S–5°N.
Figure 11 shows simulated 925-hPa wind anomalies during November–January. The anomalies were obtained by difference from the EXPcntl, which used seasonally varying climatological SSTs as the boundary conditions. When the global SST anomalies are used for the boundary conditions, the GCM simulates a similar pattern to the observed pattern (see Fig. 8), but the anomalies over the Indian Ocean are relatively weak (Fig. 11a). It is conceivable that this discrepancy for weak wind simulation over the Indian Ocean is linked...
to dry bias of the model climatology over the equatorial Indian Ocean. However, the model well simulates the westerlies in the central-eastern Pacific and the easterlies in the western Pacific. Note that the model also simulates the establishment of the Philippine Sea anticyclone during the El Niño mature phase. When the SST anomalies in the only Pacific domain are considered (Fig. 11b), the anomalous wind pattern is very similar to that in Fig. 11a. The AGCM simulates the western Pacific easterlies when the SST forcing exists only in the tropical Pacific. This is consistent with the argument of Weisberg and Wang (1997a,b). However, compared to those of the EXPglo, the westerlies are stronger in the central Pacific and the easterlies are weaker in the western Pacific.

Figure 11c shows anomalous wind patterns when SST forcing exists only in the Indian Ocean. As expected, the AGCM did not simulate the anomalous westerlies in the central Pacific. However, the easterlies were well simulated over the western Pacific. Note that the amplitude is larger than that of the EXPpac. This implies the Indian Ocean warming is a more effective contributor to developing the anomalous easterlies rather than

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**Table 3. Experimental designs with the atmospheric GCM.**

<table>
<thead>
<tr>
<th>Expt</th>
<th>SST boundary condition</th>
</tr>
</thead>
<tbody>
<tr>
<td>EXPcntl</td>
<td>Climatological SST</td>
</tr>
<tr>
<td>EXPglo</td>
<td>Climatological SST + SSTA over the global ocean (120°E–80°W)</td>
</tr>
<tr>
<td>EXPpac</td>
<td>Climatological SST + SSTA over the Pacific Ocean (40°–120°E)</td>
</tr>
<tr>
<td>EXPind</td>
<td>Climatological SST + SSTA over the Indian Ocean (40°–120°E)</td>
</tr>
</tbody>
</table>

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**Fig. 11.** Simulated 925-hPa wind anomalies by atmospheric GCM: (a) EXPglo, (b) EXPpac, and (c) EXPind experiment results as listed in Table 3.
the warming in the eastern Pacific and the local cooling in the western Pacific.

To clearly show the relative contribution for the easterly anomalies, we showed the evolution of the WP zonal wind index, which was defined in the previous section. The model simulates an abrupt establishment of the easterly anomaly in November of the El Niño–developing year as shown in Fig. 12. Note that the abrupt establishment commonly occurs in both cases of the EXPpac and EXPind. It is conceivable that this feature is related to the establishment of the Philippine Sea anticyclone (Wang et al. 2000; Wang and Zhang 2002).

To compare the relative contributions of the Pacific and Indian Ocean SST, we calculated the WP zonal wind index during November–January. In the case of the EXPglo, the index is $-1.96 \text{ m s}^{-1}$. By contrast, the EXPpac and EXPind simulate $-0.84$ and $-1.28 \text{ m s}^{-1}$, respectively. The value of the EXPind is 1.5 times larger than that of the EXPpac. Although the ratio is smaller than the observational value that was presented in the previous section, this result supports our argument that the Indian Ocean SST is a crucial contributor to the easterly wind anomalies.

5. Summary and discussion

In this study, an interactive feedback between ENSO and the Indian Ocean was investigated by using observed data and atmospheric GCMs. Some El Niño events are accompanied by warming in the western Indian Ocean during the boreal fall and winter of the developing year. We have suggested that the warming in the Indian Ocean is linked to easterly wind anomalies over Indonesia and the western edge of the Pacific during the mature phase of El Niño. The anomalous easterlies induce a fast transition to La Niña by generating negative Kelvin waves. Thus, the warming in the Indian Ocean, which is a part of the El Niño signal, operates as a negative feedback mechanism to ENSO. We suggest that this Indian Ocean feedback appears to operate mostly for strong El Niños and results in a La Niña one year after the mature phase of El Niño.

In terms of the Indian Ocean feedback, the eastward extension of the anomalous Walker circulation is a crucial factor. This is linked to the establishment of the easterly wind anomalies in the western Pacific during an El Niño mature phase. Though the precise dynamics for the extension of the easterly wind are unknown, we now suggest two possible ways. First, it seems to be related to seasonal changes of the climatological basic state. There are significant seasonal changes of the surface wind in the Maritime Continent region ($100^\circ$–$150^\circ$E). Related to these changes, the Philippine Sea anticyclone is dramatically established (Wang et al. 2001; Wang and Zhang 2002), and the SIO anticyclone is suddenly decayed (Wang et al. 2003). In particular,
Wang and Zhang (2002) pointed out that the development of the Philippine Sea anticyclone depends on the seasonal basic state, which influences tropical–extratropical interaction, and monsoon–ocean interaction. Associated with the development of the Philippine Sea anticyclone and the decay of the SIO anticyclone, the anomalous Walker circulation can be expanded to the western Pacific.

The second possibility for the eastward expansion of the anomalous Walker circulation may be related to the cold anomalies in the eastern Indian Ocean. As we mentioned before, most WISST events were accompanied with the Indian Ocean dipole events except one case. This indicates that there are cold SST anomalies in the eastern Indian Ocean during these events. The warm SST in the western Indian Ocean enhances convection accompanying an anomalous low-level convergence, and the cold SST suppresses a convection accompanying an anomalous low-level divergence. The suppressed convection can play an role to confine the easterly anomalies to the Indian Ocean. However, the cold SST in the eastern Indian Ocean abruptly vanishes after November. Hence, the anomalous descending motions in the eastern Indian Ocean suddenly decrease. Therefore, the easterly wind anomalies can be extended to the western Pacific. Recently, Annamalai et al. (2005) showed from atmospheric GCM experiments that the east–west contrast in the Indian Ocean SST does not generate a significant atmospheric Kelvin wave response, and there is little effect in the tropical Pacific wind variability, while the basinwide SST anomalies can influence on the tropical Pacific wind variability. Their results somewhat support our hypothesis. Though we have suggested two possibilities here, further research should be devoted to precisely understand the eastward expansion of the anomalous Walker circulation.

The coupled feedback process, suggested in this study, may be understood using a simple analog model. Here, we have modified the analog model for the recharge oscillator (Jin 1997a). An equation for the Indian Ocean SST is added to the original recharge oscillator model (Jin 1997a) as follows:

\[
\begin{align*}
\frac{dh_w}{dt} &= -rh_w - abT_e - a_1T_I \\
\frac{dT_e}{dt} &= RT_e + \gamma h_w \\
\frac{dT_I}{dt} &= b_1T_e - c_1T_I
\end{align*}
\]  

(1)

where \(T_I\) denotes the SST anomaly in the Indian Ocean. As Jin (1997a) indicated, \(h_w\) and \(T_e\) denote the western Pacific thermocline and eastern Pacific SST anomalies, respectively. Here \(a_1\) and \(c_1\) indicate coupling strength and damping, respectively. The other coefficients \((r, \alpha, b, R, \text{and } \gamma)\) are identical to those in Jin (1997a). Though the parameter \(a_1\) has a strong seasonal dependency, we ignored it for simplicity. The Indian Ocean SST warming induces anomalous easterly wind stress, which plays a role in producing negative thermocline depth in the western Pacific. This process is additionally included in the first equation of the original recharge oscillator model. The Indian Ocean SST is affected by the tropical Pacific forcing, local air–sea interaction feedback, and local damping processes. The third equation represents these processes.

When \(a_1\) is zero, the equation is identical to Jin’s (1997a) model. We calculated the eigenmodes of Eq. (1). The value of each parameter is the same as that of Jin’s (1997a) neutral solution. As Jin (1997a) showed, the period of the dominant mode is about 3.4 yr\(^3\) when the Indian Ocean SST is decoupled from the tropical Pacific. When \(a_1 = 1\) (i.e., the ENSO is coupled to the Indian Ocean), the period of the system is significantly shortened to about 2.3 yr. It means the coupling with the Indian Ocean reduces the ENSO period. In addition, the eigenvector of the dominant mode shows that the Pacific SST leads the Indian Ocean SST by 2 months. These results are consistent with the observational analysis. The growth rate slightly increased compared to that of the noncoupling case.

This solution of the simple analog model has some implication for Indian Ocean–ENSO coupling. The tropical Pacific may have an intrinsic mode having a 3–5-yr time scale without coupling with the Indian Ocean. The classical ENSO theories such as the delayed oscillator (Schopf and Suarez 1988; Battisti and Hirst 1989), and the recharge oscillator (Jin 1997a,b), have tried to explain the self-sustained oscillation in the tropical Pacific basin. As we pointed out here, Indian Ocean SST forcing modulates the ENSO mode and so changes its periodicity and growth rate (Yu et al. 2002; Wu and Kirtman 2004). In particular, the Indian Ocean SST affects the ENSO mode by producing a wind anomaly in the western Pacific, which changes the oceanic thermocline.

Our findings, in the present study, may have useful implications for ENSO prediction. A number of El Niño prediction models were developed based on the tropical Pacific domain. Therefore, they cannot consider the western Pacific wind variability associated
with Indian Ocean SST. It means they may not correctly predict a fast phase transition of ENSO. For example, an El Niño prediction model, developed by Kang and Kug (2000), failed to predict fast-decaying and -developing El Niñas following 1982/83, 1987/88, and 1997/98 El Niños though it has quite good predictive skill (Kang and Kug 2000; Kug et al. 2001). These deficiencies will be overcome by considering the impact of the Indian Ocean SST. This kind of approach can be studied further in order to improve ENSO prediction skill.

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