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The critical role of Indian summer monsoon on the remote forcing between Indian and Northwest Pacific during El Niño decaying year

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Recent studies have found a connection between Indian Ocean Basin Warming and the anomalous Northwest Pacific Anticyclone (ANPWA) during El Niño decaying year. This study focuses on the necessary condition for this connection by using observation and numerical simulation. The seasonal transition of the Indian Ocean sea surface wind is critical to the climatic effect of Indian Ocean Basin Warming. When the South Asian Summer Monsoon reaches its peak, the background wind becomes desirable for basin warming, which then affects the climate in the Northwest Pacific. Via the Kelvin waves and Ekman divergence, the wind anomalies exist in the lower atmosphere east of the Indian Ocean warm Sea Surface Temperature (SST) anomalies, and intensify and sustain the ANWPA throughout the El Niño decaying summer. This impact plays an important role in the inter-annual variability of the East Asian Summer Monsoon.

El Niño decaying year, Indian Ocean Basin Warming, South Asian Summer Monsoon, Northwest Pacific anticyclonic anomaly, East Asian Summer Monsoon

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During an El Niño cycle, there is a teleconnection between SST anomalies in the tropical Pacific and winds in East Asia. In El Niño developing years, the East Asian Winter Monsoons are weaker, whereas in the decaying years the summer monsoons are stronger with abnormal precipitation [1–7]. Wang et al. [5] pointed out that the anomaly Northwest Pacific Anticyclone (ANWPA) is a key system to bridging Equatorial Pacific El Niño events and East Asian monsoons. This anomalous lower-tropospheric anticyclone develops rapidly in late fall of the year when El Niño reaches the extreme phases and persists until the following spring or early summer. They proposed that a positive feedback between Northwest Pacific (NWP) Sea Surface Temperature (SST) anomalies and local lower-tropospheric vorticity anomalies plays a critical role in this persistence. In response to an El Niño event, cyclonic Rossby waves (cyclonic anomaly) propagate westward. East of this cyclonic anomaly, increasing winds intensify the evaporation and result in sea surface cooling. The negative SST anomaly in turn reinforces Rossby waves and causes an anticyclonic anomaly and further cooling of SST, creating a positive SST-evaporation feedback. However, the prevailing Northeast trade winds change in early summer as the South Asia Summer Monsoon breaks out. Therefore, this SST-

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evaporation feedback becomes invalid later and transforms into a negative feedback [8–10].

The Indian Ocean can act as a capacitor to preserve the east-central Pacific SST anomaly signal, leading to a uniform basin warming mode over tropical Indian Ocean Basin (IOB) [6, 11–18]. In the summer of El Niño decaying years, SST anomalies almost dissipate over the central and east Pacific but persist in the Indian Ocean [11] and South China Sea [6, 10]. Kawamura et al. [17] found that the meridional gradient of SST anomalies in the Indian Ocean could induce an asymmetric atmosphere circulation pattern in spring, with northeast (northwest) wind anomalies to the north (south) of equator. Such pattern was attributed to the wind-evaporation-SST (WES-V) feedback [19] triggered by background easterlies wind in winter and spring. The first EOF mode of Indian Ocean precipitation from March to May is coherent with such asymmetric mode, which is usually found during El Niño decaying years [21]. Du et al. [20] further showed that the asymmetric wind pattern could persist to the next summer owing to the warming of SWIO (Southwest Indian Ocean). Throughout winter and spring, the stronger northeast wind induces more surface evaporation and generates a weak warming in the north Indian Ocean. However, as the southeast monsoons start, the northeast wind anomalies become a warming mechanism, leading to evident SST warming over the north Indian Ocean and South China Sea in summer [22]. According to Xie et al. [6], because of the warming of the Indian Ocean, the tropospheric temperature increases through moist-adiabatic adjustment in deep convection, which generates a baroclinic Kelvin wave propagating into the Pacific. In NWP, the equatorial Kelvin wave induces easterly wind anomaly, and the resultant divergence to the north of the easterly area suppresses convection and triggers the lower-tropospheric anomalous anticyclone. They further compared the contribution of remote Indian Ocean forcing and anomalous local SST anomaly to anomalous anticyclonic using Atmospheric General Circulation Model (AGCM). The result implies that the Indian Ocean warming can induce anomalous easterly wind as well as local negative SST anomaly. Such wind anomaly results in suppressed convection and lowertropospheric anomalous by WES-V mechanism.

As Yang et al. [14] stated, the duration of IOB warming can persist from late fall in El Niño developing years to the following summer. Does the Kelvin wave-induced Ekman divergence mechanism [6] work all the time? Hu et al. [10] showed that the outbreak of Asian summer monsoon will destroy the WES feedback. Known as a basin with significant tropical monsoon, the Indian Ocean has distinct atmosphere circulation in spring and summer. What's the role of the Indian Ocean seasonal transition in the Kelvin wave-induce Ekman divergence mechanism?

1 Data and methods

1.1 Data

This study is based on the 'capacitance effect' of the tropical Indian Ocean, which is notable mainly after 1979 [16]. So the data we used include: 1) the sea surface temperature (Reynolds SST), which is the global grid data, with the spatial resolution being 1°×1°, the temporal resolution being month, and the time span being from January in 1982 to December in 2009 (http://www.nhc.noaa.gov/aboutsst.shtml). 2) the global monthly average climatic reanalysis data (NCEP/DOE Reanalysis 2, 2.5°×2.5°) [15], which are from National Centers for Environment Prediction (NCEP), including wind speed, sea level pressure, surface latent flux, surface sensible flux, convection precipitation rate, and 200 hPa geopotential height. NCEP-DOE R2 (hereafter referred to as NCEP-2) uses the up to date analysis/forecast system to do data assimilation, which is the continuation of the NCEP R1 (hereafter referred to as NCEP-1) reanalysis data plan. They have corrected the known artificial error, being regarded as the improving edition of NCEP-1 (http://nomads.ncdc.noaa.gov). Therefore, taking into account the accuracy of the NCEP-2 data and the 1979 Indian Ocean warming in the equatorial Pacific El Niño effect over the western North Pacific anticyclone more significant features, NCEP-2 data are used in this research.

1.2 Methods

In this paper, we try to reveal the evolution of Indian Ocean SST anomalies and the consequent climatic effects in an El Niño cycle (the developing year and the decaying year). The climatological mean is removed for all fields. To exclude the influence of long-term climate change, linear trends are removed from the monthly anomalies by the least squares fit at each grid point for all fields. The first EOF modes of Pacific and Indian Ocean sea surface temperature anomalies (SSTA) are consistent with El Niño and IOB warming, respectively. Therefore, the principal component time series associated with the leading EOF modes of SSTA variability over the tropical Pacific Ocean (120°E–60°W, 26°S–26°N) and the tropical Indian Ocean (30°E–120°E, 30°S–26°N) are defined as the Nino and IOB warming indices, respectively. We set the threshold of El Niño (IOB Warming) episodes as the Nino (IOB) index exceeds its 0.75 standard deviation for consecutive 5 months. The chosen years are described in Table 1

Table 1 The cases of El Niño and IOB

El Niño events	IOB warming events
1982. 09–1983. 08	1982. 10–1983. 03
1986. 12–1987. 12	1987. 05–1988. 10
1991. 11–1992. 05	1990. 12–1991. 06
1997. 05–1998. 05	1997. 09–1998. 10
2006. 08–2007. 01	

From the results above, the El Niño events can be divided into two categories. One is El Niño events with IOB warming and the other is those without. They will be referred to as (EI) and (E) events, respectively. The years of (EI) events are 1982–1983, 1987–1988, and 1997–1998. The years of (E) events are 1991–1992 and 2006–2007. Using ERSST.v3b and NCEP-1 data in 1950–2009, Hu et al. [10] also proved the existence of (EI) and (E) events.

Three EI events have the same evolution process—beginning from March, maturing in winter, and completely fading in the following summer, whereas IOB warming usually starts in May and lasts more than a year to next winter (Figure 1). Because all the (EI) events and IOB warming events exhibit a tight phase locking to the annual cycle, it is meaningful to make a composite analysis. Hereafter, the three (EI) episodes and two (E) episodes will be termed as the composite (EI) and (E) events, respectively.

1.3 Model and experiments

We also use FOAM1.5 (Fast Ocean-Atmosphere Model 1.5) to study the influence of IOB warming on NWP. FOAM1.5 is a coupled atmosphere-ocean model developed by Department of Mathematics and Computer Science of Argonne National Laboratory and Space Science Technology Center of University of Wisconsin-Madison [22, 23]. It is of fast calculation speed and suitable to study long-term natural variability in the climate system. FOAM1.5 uses the combination of a low-resolution (R15) atmosphere model with 18 vertical levels in sigma coordinates and a highly efficient medium-resolution ocean model with 24 vertical levels, and no flux corrections are used. It compares well with higher resolution models in the quality of the simulated climate.

The experiment setting is inspired by Du et al. [13], who showed that, when there is a basin-wide warming in the Indian Ocean during an El Niño cycle, a baroclinic Kelvin wave is induced north of the Equator and that the north Indian Ocean plays a principal role in heating the atmosphere [6, 21, 24–26]. However, their research did not address i) whether the heating effect of IOB warming on the atmosphere remains during the annual cycle and ii) how IOB warming generates an eastward propagating baroclinic Kelvin wave over NWP. To answer these two questions, we adopt a 200-yr FOAM control run (CTRL) and carry out three sets of sensitivity experiments. Each set of sensitivity experiment includes 50-ensemble members, which are initialized on a particular date from different years of CTRL, with positive SST anomalies added over the North Indian Ocean (40.7812°–108.281°E, 0.70312°–30.2344°N). We refer to them as the IOB_W3 (initial SST anomaly is added on March 1, at the time before the South Asian Monsoons break out), the IOB_W5 (on May 1 when the South Asian Monsoon begins to break out) and IOB_W8 (on August 1 when South Asian Monsoon breaks out completely), respectively. As an initial condition, temperature anomalies are prescribed as follows in the top 50 m (in the mixed layer of the model): the anomaly field boundary to 0°C, from the border towards the middle temperature increases as a sine

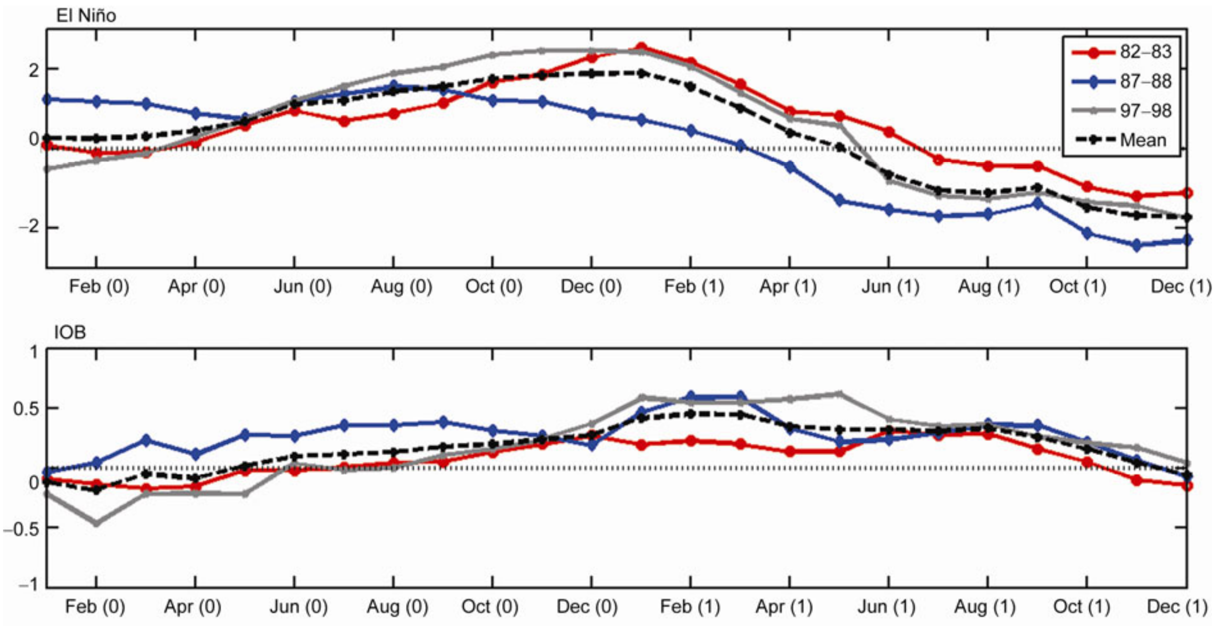

Figure 1 Phase lock of El Niño and IOB SSTA of first class event in nino3.4 (5°N–5°S, 170°W–120°W, upper picture) and the Indian Ocean (30°E–120°E, 30°S–26°N，the picture below). Dash line shows the composite of three cases. (0) means months in El Niño developing year, (1) means months in El Niño decaying year; the same below.

function, up to 4°C. The model is integrated for 1 yr. The result of each experiment is taken from the average of 50-member ensemble runs with different initial fields. The ensemble mean is analyzed to isolate the coupled response to an initial warming over the North Indian Ocean.

2 The link between the Indian Ocean Warming and ANWPA

2.1 Observation

As mentioned above, IOB event is not necessarily concurrent with El Niño. According to Figure 2(b), anticyclonic vorticity exists in both (EI) and (E) events before El Niño dissipates. However, the anticyclonic vorticity in the (EI) event persists even after El Niño begins to decay, which is not the case in the (E) event. Since August, the anticyclonic anomaly in the (E) event dissipates rather quickly with the decay of El Niño and gradually turns to a cyclonic anomaly. In contrast, the anticyclonic anomaly in the (EI) event intensifies in May and August. Particularly in August, the anticyclonic anomaly reaches its peak. This suggests that IOB warming assists in maintaining the ANWPA. Notably, during the decaying year of El Niño, the intensity of IOB warming stays nearly constant while the ANWPA strengthens abruptly.

The seasonal mean of SSTA and surface wind anomalies in (EI) and (E) events during El Niño cycle are shown in Figure 3 (both variables have passed the *t* test). In the (EI) event, the anomalous anticyclone, centered on 10^{o}N and 140°E, persists through summer and enhances in August, whereas for (E) events, it is weak and completely disappears in JJA (Figure 3(f)). During the winter of an El Niño developing year, positive SSTA over the Pacific of (EI) event is stronger than that of (E). In the following summer, the former turns into negative SSTA, while the latter persists positively in a lower value. In the Indian Ocean, the SSTA of (EI) event keeps positive from DJF to JJA, while changes little in (E). Furthermore, the SST warming is not spatially homogeneous. In an El Niño decaying year, IOB warming is located mainly in the south Indian Ocean and tropical Indian Ocean in spring, but migrates to the north Indian Ocean in summer.

During the (EI) event, the positive SSTA over north Indian Ocean always wanes. However, the ANWPA enhances at the same time. Moreover, only in summer is there a persistent easterly wind anomaly over the east tropical Indian Ocean and west tropical Pacific, which can be attributed to the Kelvin wave-induced Ekman divergence [6]. According to the analysis above, we find that IOB warming persists along the El Niño decaying year. Now the focus is why the Kelvin wave emanates in summer, the very moment when IOB warming begins to wane?

2.2 Model results

To address this issue, we have carried out three experiments by adding the same SSTA to the north Indian Ocean in three different months (Section 2).

In Figure 4, we show the monthly evolution of IOB_W3. The positive SST anomaly added in March over the Indian Ocean persists for a considerably long time and dissipates in August. Particularly in the first three months, the warm signal decays slowly. However, a strong positive SST anomaly dominates the north Indian Ocean, either from March to May when wind anomalies over Indian Ocean are much stronger than those over NWP, or in June and July when the

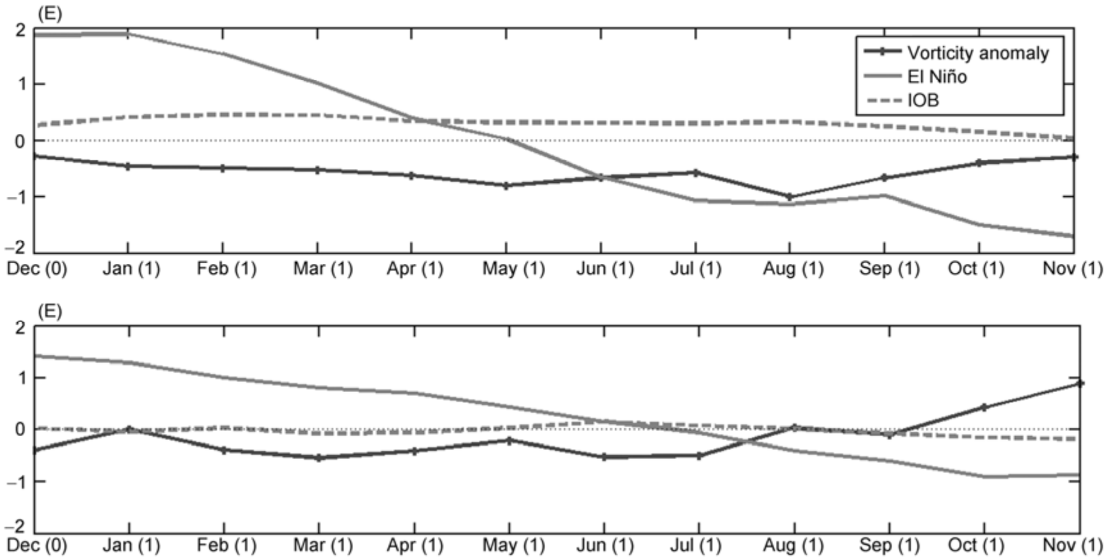

Figure 2 Evolution of monthly mean Niño-3.4 SST anomalies (°C), Indian Ocean SST anomalies (°C) and NWP (125°–147.5°E, 10°–25°N) vorticity of (EI) and (E) from Dec to the following Nov.

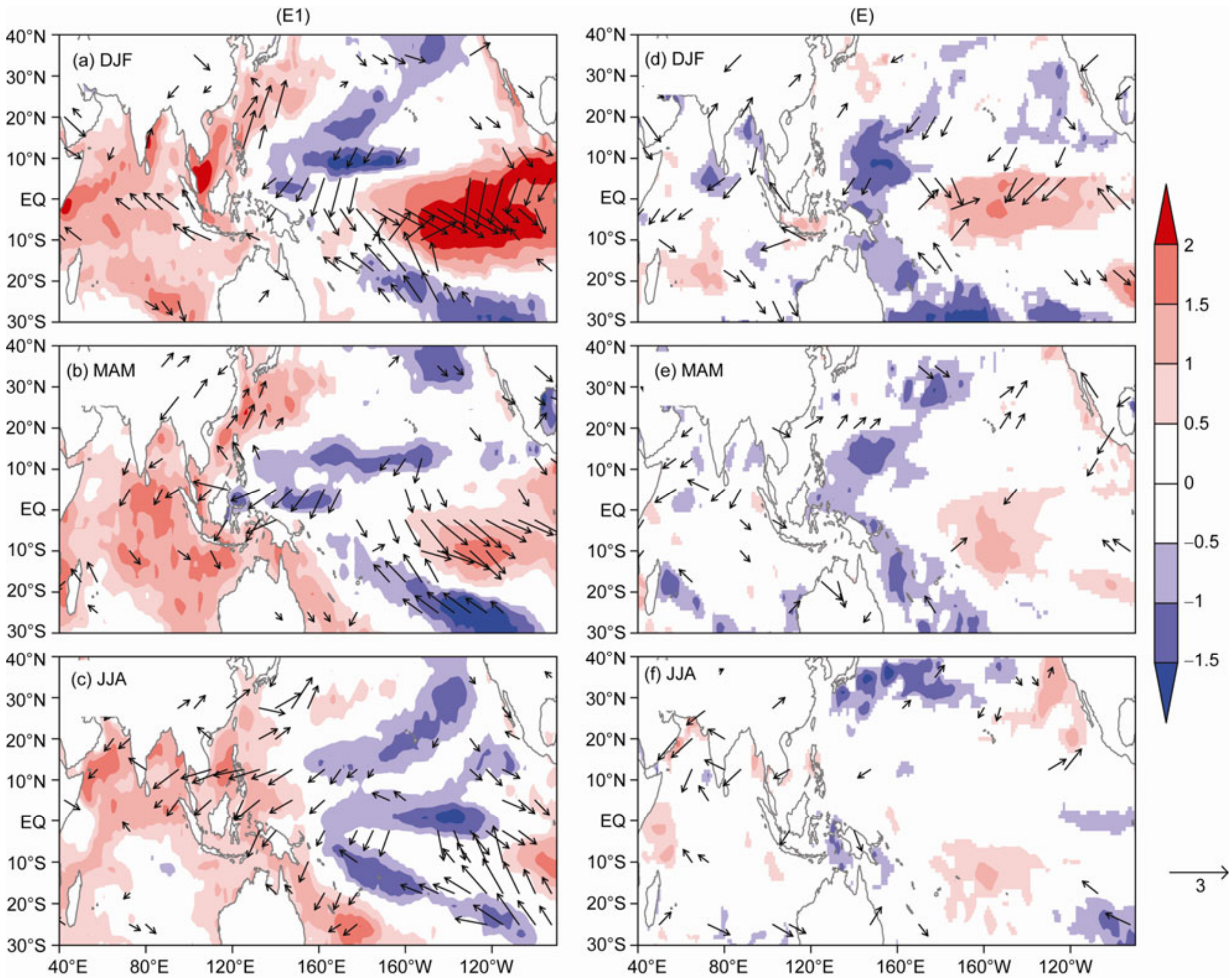

Figure 3 Spatial distribution of SST anomalies (shaded) and wind anomalies (vector) over the Indian-west Pacific in (EI) and (E) events in winter, spring and summer, respectively (SSTA and wind anomalies have passed *t* test).

warm signal begins to wane quickly. It is until August that a weak easterly anomaly starts over east of the Indian Ocean and NWP, accompanied with a near-normal SST anomaly in the Indian Ocean. An anticyclonic anomaly is formed in NWP. The model result is consistent with observation. According to Izumo et al. [25], the main cause of north Indian Ocean warming is the continuous atmosphere-ocean interaction of the Indian Ocean as El Niño dissipates. Du et al. [26] further showed that the teleconnection of ENSO results in anticyclonic anomalies over the Indian Ocean, leading to downwelling Rossby wave and positive SSTA via feedback of thermocline. This positive SSTA propagates westward slowly in spring and early summer over the southwest Indian Ocean, which generates an asymmetric wind pattern, with northeast wind anomalies in the north Indian Ocean. In winter and the following spring, the wind anomalies and mean wind are in the same direction, slowing down the warming in the north Indian Ocean. The contrary case happens when the southwest monsoon breaks out. As a result, the positive SSTA is strengthened [21, 22].

To better understand the climatic effect in different warming time, we plot the wind and SST response of our model experiments over the north Indian Ocean for IOB_W3, IOB_W5, IOB_W8 (Figure 5). In IOB_W3, wind responses are weak in the NWP, with few easterlies anomalies. Positive SSTA is located north of the Equator, where warm SSTA is added. In IOB_W5, wind responses are also weak in the NWP, but anomalous easterlies are observed off Indonesia, near the Equator. Besides, positive SSTA extends as far as 10°S. In IOB_W8, significant easterlies anomalies dominate in the NWP and Indonesia, with anticyclone anomalies centering at 20°N and 130°E. Compared with IOB_W3 and IOB_W5, area and intensity of the anomalous warming are smaller in IOB_W8. There is even an area of negative SSTA between 10°S and the Equator. Therefore, we conclude that the same SSTA has distinct effects over NWP if added in different seasons, which also has an influence in the SST response over the Indian Ocean.

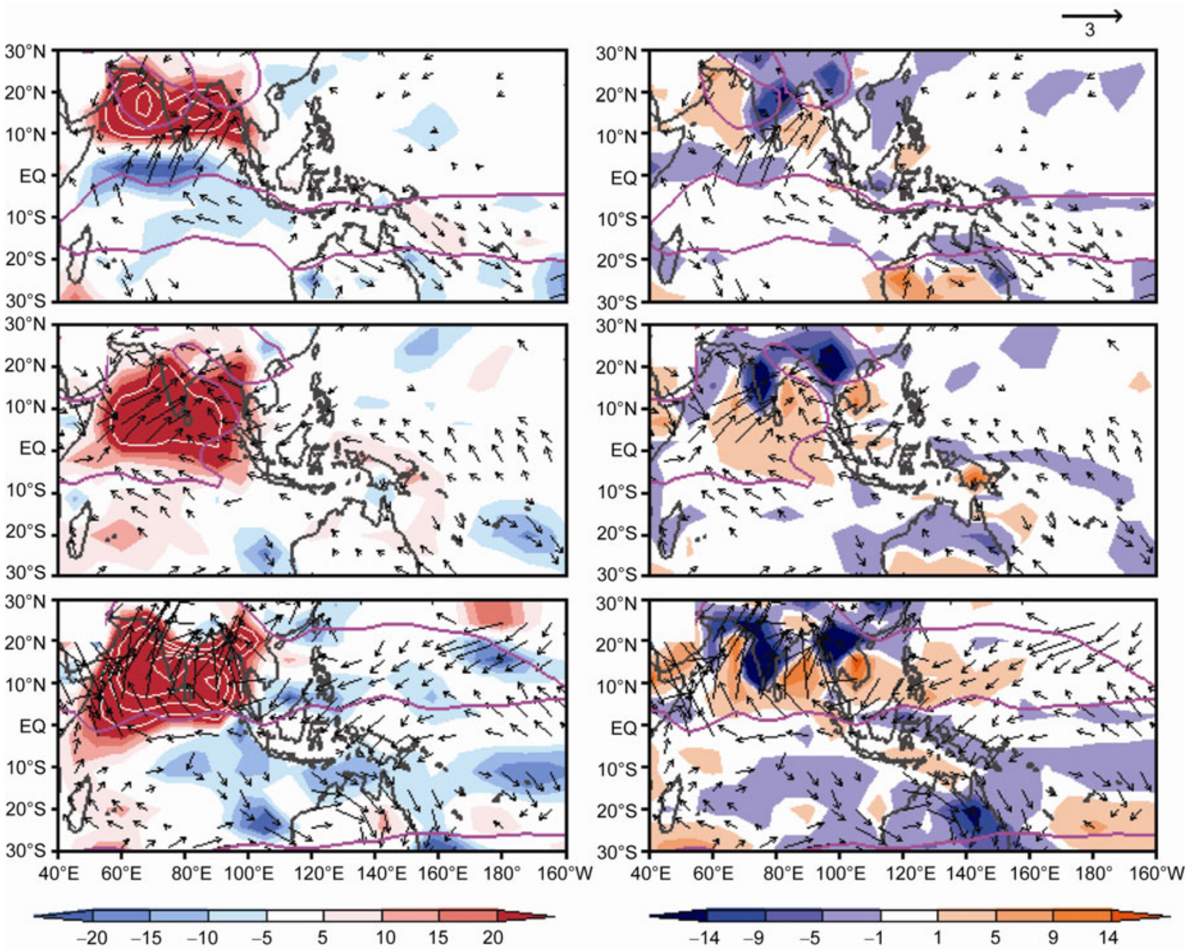

Figure 4 Evolution of SST anomalies (shaded) and surface wind anomalies (vector) in IOB_W3 by FOAM (Dash rectangle is the area where an SST anomaly is added. Grey lines are the zero line of U component in CTRL experiment. All the variables have past *t* test).

As a result, the atmosphere over the Indian Ocean does not respond to a warm signal in it immediately, since other factors may also control the heating effect.

Now the question is what these factors are and how they work. Xie et al. [6] proposed that the easterly anomaly in the NWP is forced by IOB warming via an eastward Kelvin wave. According to Matsuno [27], the prerequisite condition for emanating a Kelvin wave is that the heating source is strong enough. Thus, heating budget of the upper atmosphere over the Indian Ocean is the key. For simplicity, we demonstrate our point with latent heat flux and sensible heat flux anomalies. Figure 6 presents the spatial distribution of latent heat flux anomalies and sensible heat flux anomalies in the very month when SST anomalies are added in IOB_W3, IOB_W5, and IOB_W8, respectively. In IOB_W3, the latent heat flux anomaly is confined within the north Indian Ocean, exactly to the north of 10°N. In IOB_W5, anomalies extend to almost north Indian Ocean, as bounded by the equator. So the total latent heat flux is larger than that of IOB_W3, regardless of the weaker centering intensity. Among these three experiments, the latent heat flux anomaly of IOB_W8 is the largest in both intensity and extent. Compared with the latent heat flux anomaly, the sensible heat flux anomaly is smaller, generally ranging from 1 to 4 W m^{-2}. Besides, its spatial distribution is well correlated with SST anomaly. We conclude that the sensible heat flux anomaly is mostly affected by SST, whose contribution is limited by the wind anomalies or heating the atmosphere. Then how can it produce such distinct difference among three experiments in latent heat flux anomaly?

Indian Ocean Basin is known for its monsoon climate. The local atmospheric circulation changes completely in spring-summer transition, which may produce distinct responses to the same warming signal. In March, before the South Asian Summer Monsoon breaks out, the prevailing wind over the north Indian Ocean is weak northeasterly (Figure 7). At this time, the sea surface evaporation is weak and the latent heat flux is limited. Therefore, the warm signal persists during spring. In May, the South Asian Monsoon breaks out. Southeasterly trade wind crosses the Equator and turns into southwesterly trade wind. With this

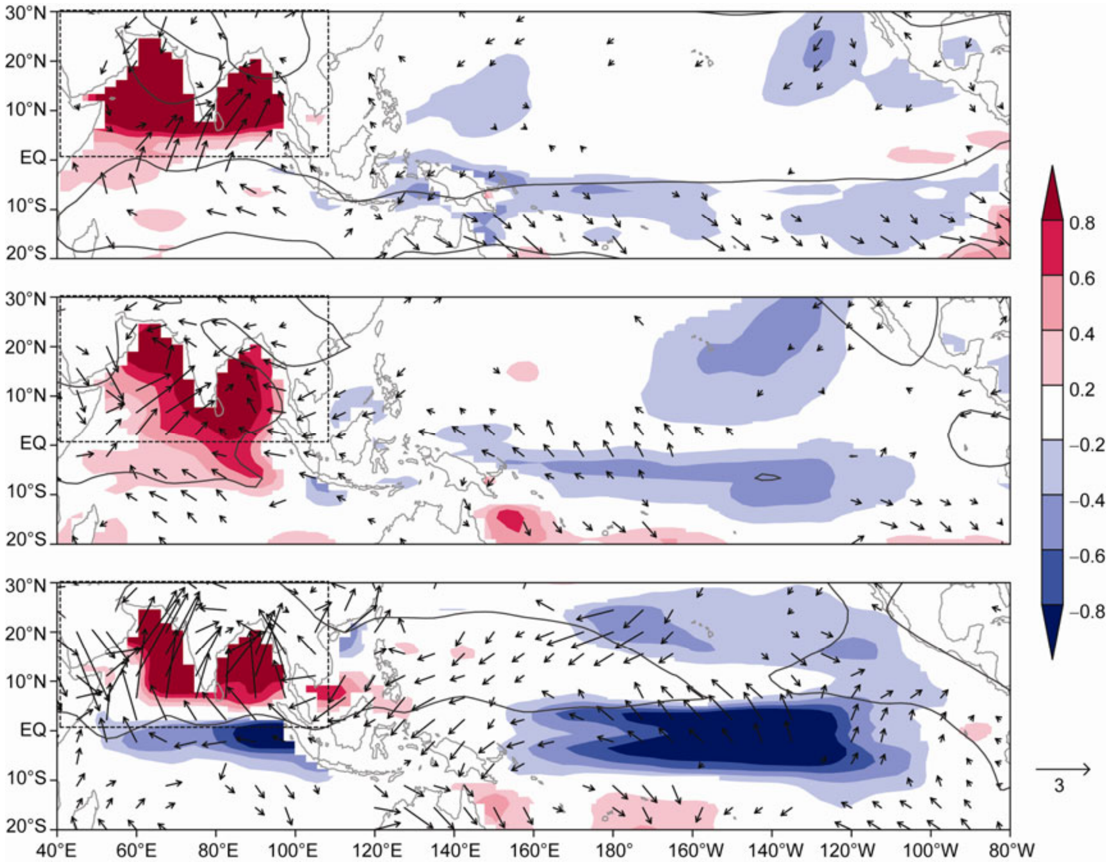

Figure 5 SST and wind responses by FOAM after a positive SSTA is added on Mar 1st, May 1st and Aug 1st, respectively.

strong wind, evaporation is enhanced as well as latent heat flux. It is not until August that the South Asian Monsoon comes to its peak. And the air-ocean heat exchange becomes most active, which finally forms a strong heating source over the Indian Ocean for emanating a Kelvin wave.

Precipitation response of three experiments confirms this point. In IOB_W8, southwest monsoon prevails over the north Indian Ocean, which is much stronger than that in IOB_W3 and IOB_W5. Convective precipitation anomaly is also substantially larger than the other two. Compared with easterlies anomaly in Figure 6(c), we conclude that the main cause for the east wind anomalies and Kelvin wave is the convective precipitation that releases a large quantity of latent heat to the atmosphere. Forced by the positive SSTA in the north Indian Ocean, anticyclonic anomalies are strengthened via Ekman divergence mechanism. Before the outbreak of the South Asian Monsoon, the background wind is weak (in IOB_W5) or northeastward (in IOB_W3), which are unfavorable for heating release. Thus, convective precipitation anomalies, latent heat flux anomalies, and east wind anomalies in IOB_W8 are the most significant among three experiments. Huang et al. [16] confirmed this phenomenon in observation.

In conclusion, the seasonal transition of prevailing wind in the Indian Ocean plays a crucial role in the teleconnection between IOB and ANPWA.

3 Discussion and conclusions

The remote forcing between the equatorial Pacific and East Asia affects both the strength and scope of the East Asian Summer Monsoon. The ANWPA is the key in the remote response and plays an important role in the forecast of the East Asian Summer Monsoon. Wang et al. [5] and Xie et al. [6] showed that both the local cold sea surface temperature (SST) anomalies in the Northwest Pacific and the warm SST anomalies in the North Indian Ocean may contribute to the development and persistence of the ANWPA during the El Niño decaying year. However, it is not sufficiently explained for the observed abrupt increase of the ANWPA during El Niño decaying summer.

Based on previous studies, we discussed the responding time of NWP to IOB warming. Considering the seasonal transition of background wind over the Indian Ocean, we showed the outbreak of the South Asian Summer Monsoon as the key bridge between the North Indian Ocean SST anomalies and the strength of the ANWPA. Combined with

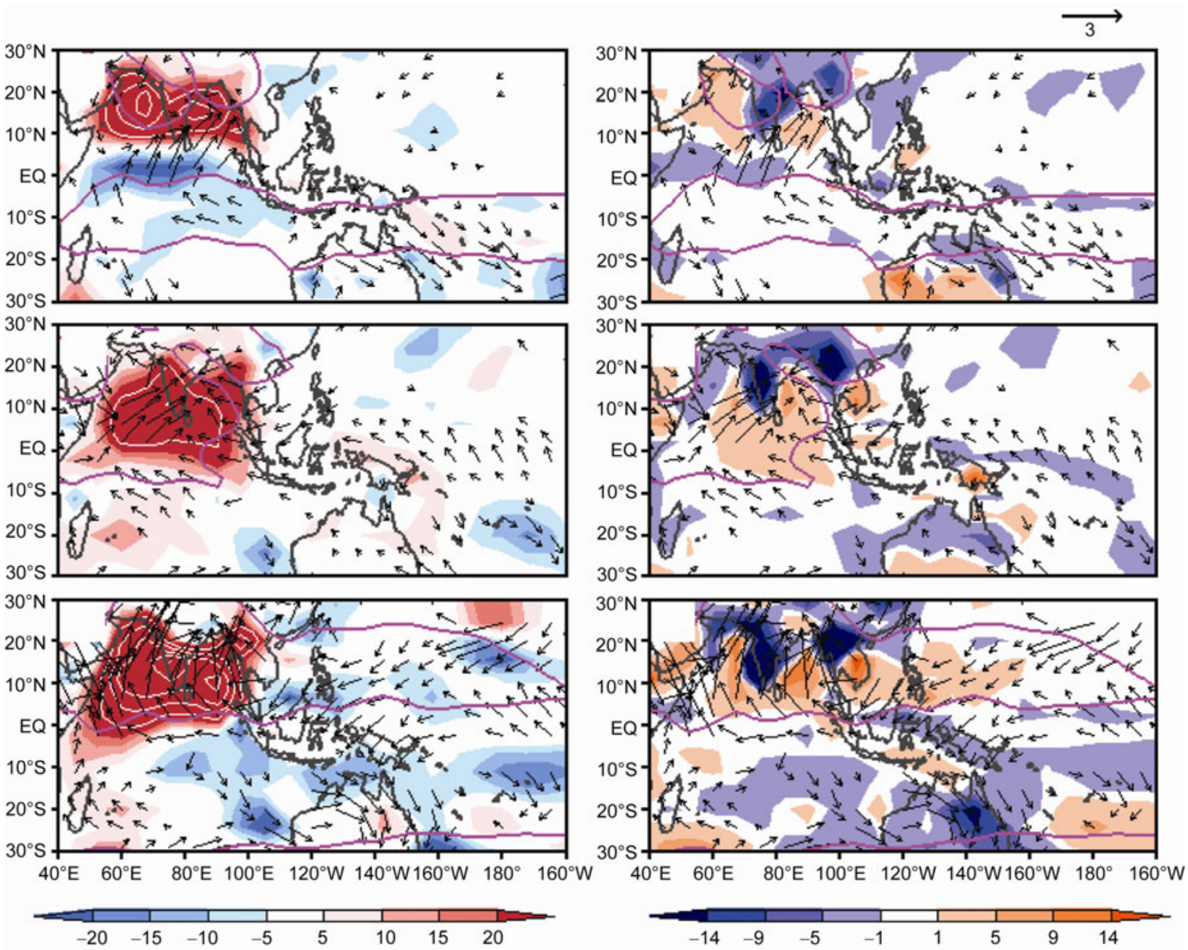

Figure 6 Spatial distribution of latent heat flux anomaly (shaded, left), sensible heat flux anomaly (shaded, right) and wind responses (vectors) in the very month when SSTA are added. White contours begin with 30 W/m^2, with an interval of 20 W/m^2. Purple lines are the zeros lines of U component in CTRL experiment.

observation, we further demonstrated that the ANWPA is enhanced by the remote North Indian Ocean warming effect during El Niño decaying summer. Initialized with the same SST anomaly in the North Indian Ocean using FOAM, the responses of the surface wind anomalies, the latent heat flux anomalies, the sensible heat flux anomalies and the convective precipitation anomalies in the whole Indo-West Pacific were discussed at three different South Asia Summer Monsoon period (the period before South Asia Summer Monsoon onset, the period during the South Asia Summer Monsoon outbreak, and the period after the outbreak). The anomalous SST introduced before or after the South Asia Summer Monsoon outbreak persists longer than that introduced during the monsoon outbreak, with more latent heat release in the case of Summer Monsoon onset by local convection. This positive latent heat flux anomaly leads to the Matsuno-Gill pattern over the Indo-West Pacific. After that, the Kelvin Waves cause anomalous over the NWP, which increase the ANWPA by the local positive feedback. Although both the outbreak time and the strength of the South Asian Summer Monsoon are not independent of internual scales, both are greatly related with the SST anomalies in the North Indian Ocean. So the South Asian Summer Monsoon has large changes while there is an IOB warming anomaly in the Indian Ocean. As shown in Figure 7, as the North Indian Ocean warming, the precipitation and the latent heat release anomalies appear over the south side of the Indian Peninsula and induce the Southwest wind anomalies exist over the North Indian Ocean. Particularly in IOB_W8 test, the South Asian Summer Monsoon is largely enhanced, with the larger Southwest wind anomalies and precipitation anomalies in the North India Ocean and South India peninsula.

In addition, there are always easterly wind anomalies in three experiments over the Northwest Pacific during the period of South Asia Summer Monsoon outbreak even with the weak warm SST anomalies in the North Indian Ocean. It seems that the onset of the South Asia Summer Monsoon is an important factor in the remote influence of the Indian Ocean Warming SST anomalies on the ANPWA. This article focuses on exploring the Indian Ocean warming effect in

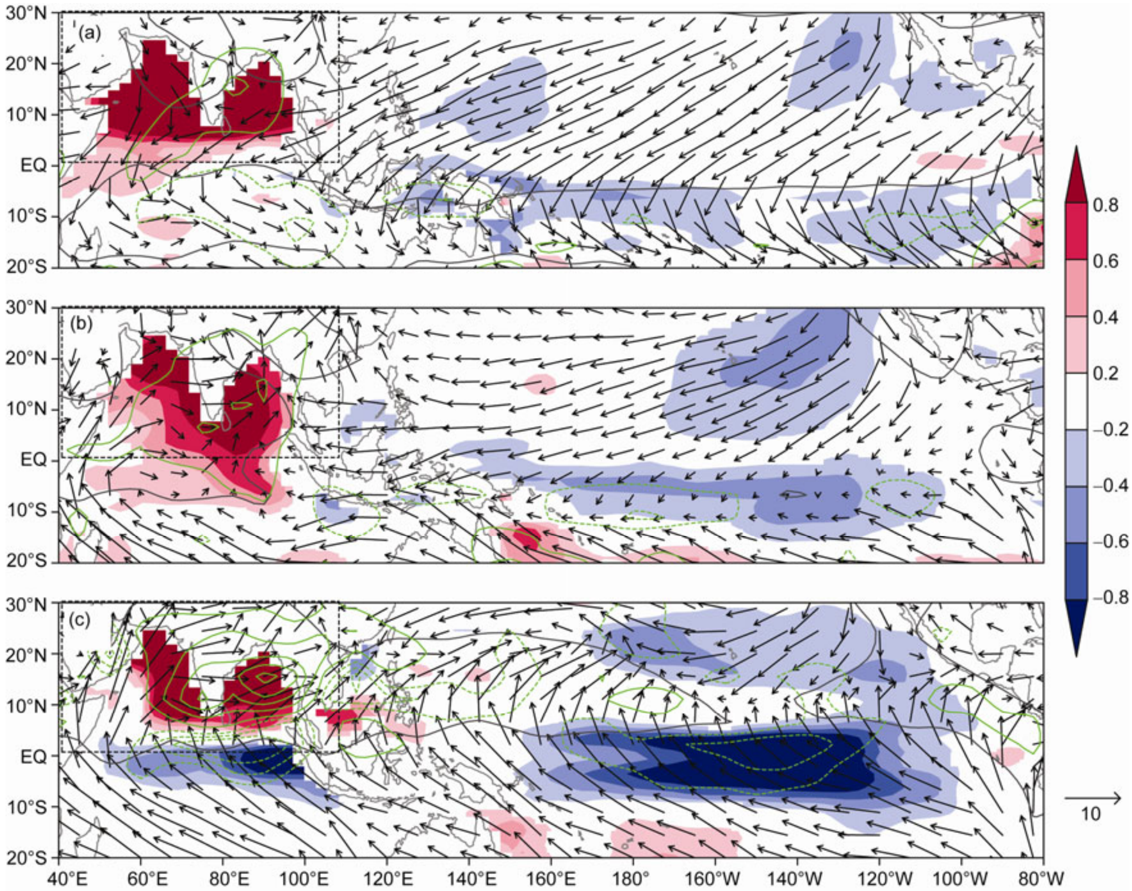

Figure 7 Spatial distribution of precipitation (contours), wind (vector) and SST (shaded) responses in IOB_W3, IOB_W5 and IOB_W8, respectively. Contours begin at 1 mm/day, with an interval of 3 mm/day.

spring and summer, whereas other factors, such as the positive feedback between the negative NWP SST anomalies in spring and the ANWPA, are not included. We will discuss the latter in the future.

We appreciate website http://nomads.ncdc.noaa.gov/data.php#cfsr for providing global monthly mean reanalysis data NCEP-2. We thank Professor Wu Lixin for the guidance of CGCM used in this article. We also thank the valuable suggestions and comments given by the two anonymous reviewers. This work was supported by the National Key Program for Developing Basic Science (Grant Nos. 2010CB428504 & 2012CB956002), National Natural Science Foundation of China (Grant Nos. 40906005, 41105059, 41275069, GYHY201106017 & GYHY201206038), and the National Key Technologies R&D Program of China (Grant No. 2009BAC51B01).

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