Role of North Indian Ocean Air–Sea Interaction in Summer Monsoon Intraseasonal Oscillation

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ABSTRACT

Air–sea coupling processes over the north Indian Ocean associated with the Indian summer monsoon intraseasonal oscillation (MISO) are investigated. Observations show that MISO convection anomalies affect underlying sea surface temperature (SST) through changes in surface shortwave radiation and surface latent heat flux. In turn, SST anomalies may also affect the MISO precipitation tendency (dP/dt). In particular, warm (cold) SST anomalies can contribute to increasing (decreasing) precipitation rate through enhanced (suppressed) surface convergence associated with boundary layer pressure gradients. These air–sea interaction processes are manifest in a quadrature relation between MISO precipitation and SST anomalies. A local air–sea coupling model (LACM) is formulated based on these observed physical processes. The period of the LACM is proportional to the square root of seasonal mixed layer depth H , assuming other physical parameters remain unchanged. Hence, LACM predicts a relatively short (long) MISO period over the north Indian Ocean during the May–June monsoon developing (July–August monsoon mature) phase when H is shallow (deep). This result is consistent with observed MISO characteristics. A 30-day-period oscillating external forcing is also added to the LACM, representing intraseasonal oscillations propagating from the equatorial Indian Ocean to the north Indian Ocean. It is found that resonance will occur when H is close to 25 m, which significantly enhances the MISO amplitude. This process may contribute to the higher MISO amplitude during the monsoon developing phase compared to the mature phase, which is associated with the seasonal cycle of H .

1. Introduction

The Indian summer monsoon is associated with abundant rainfall and a pronounced large-scale circulation pattern, and therefore its variation and prediction have enormous implications for the agriculture and economy of South Asia (Lau and Waliser 2012). The substantial heating source associated with the abundant monsoon rainfall can also affect climate worldwide (Ting 1994; Ding and Wang 2005). In boreal summer, the monsoon exhibits pronounced intraseasonal variability, manifest as northward-propagating alternating wet and dry anomalies

originating from the equatorial Indian Ocean, with periods of 30–60 days (Sikka and Gadgil 1980; Hartmann and Michelsen 1989; Annamalai and Slingo 2001). Understanding the monsoon intraseasonal oscillation (MISO) is of vital importance to extended-range weather prediction. Hence, the MISO has been extensively studied (Webster 1983; Annamalai and Slingo 2001; Kemball-Cook and Wang 2001; Fu et al. 2003; Jiang et al. 2004; Krishnamurthy and Shukla 2007; Joseph and Sabin 2008; Roxy et al. 2013; Li et al. 2016).

Various mechanisms have been proposed for the northward propagation of the MISO. Webster (1983) suggested that the interaction between circulation/rainfall and land surface sensible heat flux favors northward propagation. Bollasina and Ming (2013) found that land surface processes associated with interactions between soil moisture, precipitation, and circulation are important for the northward propagation of MISOs. The important

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role of land surface hydrology in MISO has also been noted in [Rajendran et al. \(2002\)](#). On the other hand, MISO propagation has been attributed to moist equatorial Rossby waves emanating from the eastward-propagating convection anomalies associated with the Madden–Julian oscillation (MJO) over the equatorial Indian Ocean ([Wang and Xie 1996](#); [Xie and Wang 1996](#); [Kemball-Cook and Wang 2001](#)). [Jiang et al. \(2004\)](#) pointed out that the generation of vorticity anomalies to the north of equatorial convection produced by vertical motion anomalies interacting with vertical wind shear could also contribute to the northward MISO propagation. [DeMott et al. \(2013\)](#) indeed found that the barotropic vorticity effect and moisture advection are the primary mechanisms that favors the northward propagation of MISOs. Furthermore, [Wu et al. \(2006\)](#) suggest that the off-equatorial convection anomalies are closely associated with the east–west dipole-like heating over the equatorial Indian Ocean associated with precipitation anomalies. [Adames et al. \(2016\)](#) showed that the west–east moisture gradient over India and Arabian Sea as well as the southwesterly monsoon circulation are important for the MISO meridional propagation. [Abhik et al. \(2013\)](#) suggested that the vertical eddy heat transport and the low-level condensation heating may contribute to triggering deep convections to the north of the existing convection associated with MISOs.

In addition to atmospheric internal dynamics and land–atmosphere interaction, air–sea interaction may also play an important role in the MISO amplitude and its propagation. [Krishnamurti et al. \(1988\)](#) found evident 30–50-day oscillations of sea surface temperature (SST) over the Bay of Bengal (BoB), which were attributed to surface heat flux anomalies. Coherent relationships between intraseasonal variabilities of SST, precipitation, and associated winds were later confirmed by field observations ([Bhat et al. 2001](#); [Webster et al. 2002](#)). Through analysis of satellite datasets, [Sengupta et al. \(2001\)](#) and [Sengupta and Ravichandran \(2001\)](#) also found prominent intraseasonal variability of SST over the Indian Ocean, which was further attributed to variations of the surface latent heat flux and the solar radiation. Despite the importance of surface heat fluxes, recent studies showed that intraseasonal SST variability over the Indian Ocean is also influenced by complicated oceanic dynamical processes (e.g., [Li et al. 2017a](#)). While MISO precipitation anomalies can have a prominent impact on SST, it has also been suggested that the SST anomalies could have feedbacks onto MISO rainfall ([Vecchi and Harrison 2002](#); [Maloney and Sobel 2004](#); [Joseph and Sabin 2008](#); [Roxy and Tanimoto 2007](#); [Sharmila et al. 2013](#); [Li et al. 2016](#)). [Vecchi and Harrison](#)

(2002) proposed a mechanism for intraseasonal oscillations over the BoB, which relates precipitation and wind anomalies to variations of SST gradient between northern and southern BoB. [Joseph and Sabin \(2008\)](#) and [Shankar et al. \(2007\)](#) suggested that the interaction between SST, precipitation, and low-level jet stream over the BoB may play an important role in MISOs. [Kemball-Cook and Wang \(2001\)](#) found that the enhanced surface evaporation, which may help precondition the atmosphere for the MISO convection, occurs ~10 days prior to positive surface convergence and precipitation anomalies over the north Indian Ocean, suggesting an important role of the air–sea interaction in MISO dynamics. By conducting numerical experiments, [Fu et al. \(2003\)](#) showed that the simulation of the MISO, particularly its northward propagation, is more realistic in a coupled model than in an atmosphere-only model.

Despite the recent progress, how SST affects precipitation at the intraseasonal time scale remains not well understood. [Fu et al. \(2003\)](#) showed that intraseasonal variability of SST is in quadrature with that of precipitation in coupled climate model simulations, as found in observations ([Li et al. 2016](#)). In contrast, in an atmosphere-only model SST variability is generally in phase with precipitation, which may be associated with the underestimated intraseasonal variability of SST ([Klingaman et al. 2008](#)). However, [Sharmila et al. \(2013\)](#) further showed that an air–sea coupled model outperforms an atmosphere-only model in terms of MISO simulation, even if the latter was forced by daily SST with realistic intraseasonal amplitude. Several studies showed that MISO precipitation anomalies lag SST anomalies underneath because positive SST anomalies associated with MISO destabilize the lower troposphere, which favors generating convective activities (e.g., [Roxy and Tanimoto 2007, 2012](#)). The time lags between SST and precipitation anomalies, however, vary in different summer monsoon regions ([Wu et al. 2008](#); [Roxy et al. 2013](#); [Xi et al. 2015](#)). For example, precipitation lags SST anomalies by ~5 days in the East Arabian Sea (EAS) but by ~12 days in the BoB, and this difference was attributed to the stronger background surface wind convergence in the EAS ([Roxy et al. 2013](#)). As we will show below, the time lag between MISO precipitation and SST anomalies also varies substantially from event to event even for the same region. Indeed, how the local SST anomaly affects MISO precipitation is complicated and involves complex dynamic and thermodynamic processes ([Sobel 2007](#)) and thus requires further investigation.

In this study, we first explore MISO air–sea interaction processes over the north Indian Ocean through analyzing observational data. Based on the observed

relationships, we then formulate a theoretical model that depicts essential physics involved in MISO air–sea coupling and helps to improve our understanding of MISO mechanisms. The rest of this paper is organized as follows. [Section 2](#) provides a detailed description of observational datasets employed in this study. [Section 3](#) analyzes the observed precipitation–SST relation associated with the MISO. [Section 4](#) develops a theoretical air–sea interaction model and applies this model to understanding MISO dynamics. Finally, [section 5](#) provides a summary and discussion.

2. Data and methods

a. Observational data

In this study, we analyze daily precipitation from Tropical Rainfall Measuring Mission (TRMM) Multisatellite Precipitation Analysis (TMPA) level 3B42 V7 product for the 1998–2015 period ([Huffman et al. 2007](#)), and daily SST from the Optimum Interpolation Sea Surface Temperature (OISST) with the Advanced Very High Resolution Radiometer (AVHRR) data from 1998–2015 ([Banzon et al. 2016](#); [Reynolds et al. 2007](#)). The spatial resolution of precipitation and SST datasets is $0.25^\circ \times 0.25^\circ$. Note that an OISST dataset that includes both the AVHRR and the Advanced Microwave Scanning Radiometer on the Earth Observing System (AMSR-E) is also available, but it only covers the period of 2002–11 and thus was not used in our analysis based on the consideration for data consistency. The OISST includes bias adjustment to buoy observations ([Reynolds et al. 2007](#)), and hence it represents to a large extent the bulk temperature within the top layer of the ocean surface. For this reason, SST intraseasonal variability may be underestimated compared to the skin temperature.

For the analysis of MISO precipitation impacts on SST variability, we use the surface net shortwave and long-wave radiation from the Clouds and the Earth’s Radiant Energy System (CERES) ($1^\circ \times 1^\circ$; 2000–15) ([Wielicki et al. 1996](#); [Loeb et al. 2001](#)) dataset and surface latent and sensible heat fluxes from the Objectively Analyzed Air–Sea Fluxes (OAFlux) ($1^\circ \times 1^\circ$; 1998–2015) ([Yu and Weller 2007](#)) dataset. The Cross-Calibrated Multi-Platform (CCMP) gridded surface winds dataset version 2 ([Atlas et al. 2011](#)) with $0.25^\circ \times 0.25^\circ$ spatial resolution for 1998–2015 was analyzed to examine the atmospheric circulation anomaly pattern associated with the MISO. In addition, mixed layer depth was calculated using $1^\circ \times 1^\circ$ monthly ocean temperature and salinity data for 2001–14 from the “Grid Point Value of the Monthly Objective Analysis” (MOAA GPV) dataset using the Argo data ([Hosoda et al. 2008](#)). To examine the relative roles of the

deep troposphere and boundary layer in causing MISO surface convergence anomalies (see [section 3b](#)), daily surface wind and pressure, SST, 850-hPa wind, and geopotential height from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) ([Dee et al. 2011](#)) during 1998–2015 are also analyzed.

b. Mixed layer model

As we shall see in [section 3](#), MISO precipitation anomalies are in phase with surface wind convergence anomalies, while causes for the latter are complex. The anomalous latent heating associated with MISO precipitation can drive anomalous large-scale low-level convergence ([Gill 1980](#)). On the other hand, anomalous SST can also cause surface convergence anomalies ([Lindzen and Nigam 1987](#)). To investigate the relative role of these two processes in causing surface convergence anomalies associated with MISO, we employ the mixed layer model (MLM) of [Back and Bretherton \(2009\)](#), which is a steady-state momentum equation for the atmospheric boundary layer, in which dominant terms are the Coriolis force, the pressure gradient force, the downward momentum mixing and the friction:

$$f\mathbf{k} \times \mathbf{U} + \rho_0^{-1} \nabla P_s = \frac{\omega_e(\mathbf{U}_T - \mathbf{U})}{h} - \frac{\omega_d(\mathbf{U})}{h},$$

where f is the Coriolis parameter, ρ_0 is the air density, P_s is surface pressure, $\mathbf{U}$ is horizontal wind in the boundary layer, and $\mathbf{U}_T$ is free troposphere wind (at 850 hPa). Also, ω_e and ω_d are two physical parameters that are set as constants, and h is the depth of the boundary layer. Following [Back and Bretherton \(2009\)](#), we set $\omega_e/h = 2 \times 10^{-5} \text{ s}^{-1}$ and $\omega_d/h = 1.5 \times 10^{-5} \text{ s}^{-1}$. Other realistic values have also been tested, and our results are not sensitive to their choices (figure not shown).

In MLM, the surface pressure P_s is further separated into the free troposphere contribution (at 850 hPa) and the boundary layer contribution. Therefore, solutions for the surface wind $\mathbf{U}$ in MLM are associated with free troposphere processes (wind and pressure) and surface pressure associated with boundary layer processes (i.e., SST anomalies). [Back and Bretherton \(2009\)](#) found that the seasonal and annual mean surface convergence are predominantly caused by SST gradients in the tropical region, which can be viewed as a cause rather than consequence of convection. Here, we employ the same model to quantify the relative role of the free troposphere and boundary layer in causing MISO surface wind convergence anomalies (see [section 3b](#)). In particular, we focus on intraseasonal variability during the boreal summer season, while the MLM was originally

developed to understand causes for the annual mean surface convergence in the tropics (Back and Bretherton 2009). As a result, the relative roles of intraseasonal SST anomalies (i.e., boundary layer) and free troposphere processes may be different from those of Back and Bretherton (2009). Interested readers are referred to Back and Bretherton (2009) for more details of this model.

c. Methods

To obtain intraseasonal variations of various atmospheric and oceanic variables associated with the MISO, we applied a 20–90-day Butterworth filter to anomalies of variables for the entire 1998–2015 period analyzed in this study, with the climatological seasonal cycle removed. Using a Lanczos bandpass filter yields very similar results (figure not shown). The seasonal cycle was obtained by calculating the monthly climatology of each variable over the 1998–2015 period. The monthly climatology is similar to the seasonal cycle obtained by the first four harmonic fits (annual + semiannual + 120-day + 90-day cycles). We then conducted composite analysis of objectively selected MISO events (see section 3a).

3. MISO air–sea interaction

a. Precipitation–SST relation

First, MISO events are selected for the EAS (65°–75°E, 10°–20°N) and the BoB (85°–95°E, 10°–20°N), separately. This is because these two regions have the largest intraseasonal variability of precipitation during boreal summer, as found in previous studies (e.g., Li et al. 2016) and shown in our analysis (see Fig. S1 in the online supplemental material). Larger boxed areas with southern and northern boundaries extended to 7.5° and 22.5°N have also been tested, and the MISO events are not sensitive to these choices (figure not shown). MISO precipitation anomalies over these two regions are well correlated with 20–90-day filtered precipitation anomalies over central India ($r \sim 0.5$), with EAS (BoB) precipitation leading rainfall over India by 5 (0) days (Fig. S2). The selected region for the central India has been used to define active-break spells (Rajeevan et al. 2010; Narapusetty et al. 2016). Similar correlations have been calculated using daily precipitation anomalies without filtering, and the coefficients remain statistically significant at the 90% confidence level, despite the much reduced ($\sim 50\%$) correlation values compared to Fig. S2. Consequently, the selected regions are important for Indian summer rainfall forecasts. We first calculate the MISO precipitation indices for the two regions, which are the time series of 20–90-day filtered,

domain-averaged precipitation over the EAS and BoB (Fig. 1). Then we select May–October MISO events when the indices are greater than one standard deviation. Note that the MISO events defined for the EAS (64 events) and the BoB (71 events) do not always coincide with each other, suggesting a certain degree of independence between the two regions.

Then, composite analysis for the selected MISO events are performed (Fig. 2). The period of the composited MISO events is approximately 30 days on average for both regions. MISO precipitation reaches a minimum on day -15 and increases to a peak on day 0. Meanwhile, SST anomalies are positive and reach a peak on day -6 (-9) in the EAS (BoB). Of particular interest is that the composited precipitation tendency (dP/dt) is in phase with positive SST anomalies. After day 0, the SST anomaly becomes negative along with dP/dt . MISO precipitation anomalies are also associated with a response in SST. The precipitation anomaly exhibits an out-of-phase relation with the SST tendency ($dSST/dt$) at intraseasonal time scales (Fig. 2); positive (negative) precipitation anomalies induce negative (positive) SST tendencies. This process is primarily through changes in surface heat fluxes as shown in previous studies, which will also be discussed in detail in section 3b.

MISO precipitation and SST anomalies exhibit a quadrature relation due to air–sea coupling processes (Fig. 2), as found in previous studies (Fu et al. 2003; Li et al. 2016). The largest correlation between MISO SST and precipitation anomalies appears when SST leads precipitation by 7 (11) days for the EAS (BoB) region (Fig. 3). The shorter time lag of MISO precipitation response to SST anomaly over EAS compared to BoB qualitatively agrees with Roxy et al. (2013), although the difference between the two regions is larger (5 vs 12 days) in their study. However, SST anomalies and precipitation tendency (dP/dt) exhibit an in-phase relation with almost no time lag (Fig. 2), with the correlation between SST and dP/dt being maximum near day 0 (Figs. 3b and 3d). The analysis of the precipitation–SST relation for individual MISO events (Figs. S3 and S4) shows that the top 25 percentiles of MISO events sorted by the correlation between SST and dP/dt exhibit a correlation for both EAS and BoB of $r \geq 0.68$. For these well-correlated events, the time lag between the peaks of intraseasonal SST and precipitation varies from 3 days to 14 days. The time lag in the fourth quartile (less-correlated events for SST and dP/dt) also varies substantially from event to event.

These results suggest that intraseasonal SST anomalies may play an important role in affecting precipitation tendency of the MISO. Note that the composited SST anomalies associated with MISO are relatively small

FIG. 1. The 20–90-day bandpass filtered daily precipitation anomalies (mm day^{-1}) averaged over the (a) EAS (65° – 75°E , 10° – 20°N ; Fig. S1) and (b) BoB (85° – 95°E , 10° – 20°N ; Fig. S1). Boxes are marked in Fig. 4 and Fig. S3, respectively. Red asterisks denote peaks of MISO events, which are defined as the date when peak positive precipitation anomalies are greater than one standard deviation (dashed blue lines) during May–October.

($\sim 0.2\text{ K}$; Fig. 2) because each event is different and thus their average is weak, as pointed out by previous studies (e.g., Schiller and Godfrey 2003). For some individual MISO events, the SST anomaly amplitude can be much larger and reach $\sim 1\text{ K}$ (Figs. S3 and S4). Note also that the AVHRR-only OISST product has been adjusted to buoy data, which largely represent the bulk SST variability, whose amplitudes can be significantly weaker than those of skin temperature (see section 2). Besides, given that the mean state SST over the north Indian Ocean is high ($>28^{\circ}\text{C}$) and exceeds the threshold for tropical deep convection during the summer monsoon season (Fig. S5; note that SST exceeds 28°C during early and late monsoon seasons; it is lower during July and August but still higher than 27.5°C), a small SST anomaly may have a prominent contribution to precipitation anomalies (Gadgil et al. 1984; Graham and Barnett 1987; Waliser et al. 1999). Physically, positive (negative) SST anomalies can contribute to the positive (negative) moisture tendency through enhancing (weakening) moisture convergence (Lindzen and Nigam 1987). The moistening (drying) atmosphere is then associated with increasing (decreasing) precipitation rate. Surface convergence

anomalies indeed agree with precipitation anomalies (Fig. 2). The analysis shown in section 3b below will support the role of SST anomalies in causing the surface convergence. Note that the precipitation–SST relation breaks down for some MISO events (Figs. S3 and S4), which may be due to the interference of synoptic-scale weather noise on precipitation and/or oceanic internal instabilities that affect SST.

b. MISO air–sea coupling

To better illustrate MISO air–sea coupled processes, we show lead–lag composited spatial patterns of 20–90-day filtered precipitation and SST anomalies for BoB MISOs at 5-day intervals relative to lag 0 (Figs. 4a,c). On day -15 , pronounced negative precipitation anomalies appear over the BoB region, along with positive SST anomalies. Meanwhile, the equatorial Indian Ocean receives anomalous positive rainfall. Subsequently, the equatorial positive precipitation anomalies propagate northward and dominate the southern BoB region on day -10 . During this time, BoB warm SST anomalies strengthen and reach maximum magnitude. As the MISO precipitation anomalies continue propagating northward from day -5 to

FIG. 2. (a) Composites of 20–90-day filtered, domain-averaged precipitation (solid pink; cm day^{-1}) and SST (solid black; K) over the EAS for the 64 MISO events identified in Fig. 1, labeled on the left axis. The right axis shows precipitation tendency (dashed pink; mm day^{-2}), SST tendency (dashed black, 0.01 K day^{-1}), and surface divergence anomalies (solid green; 10^{-6} s^{-1} ; negative means convergence) associated with MISO. Day 0 is the date when MISO precipitation peaks. (b) As in (a), but for the composites of 71 MISO events over the BoB.

day 0, positive precipitation anomalies become stronger, while positive SST anomalies gradually weaken to almost zero over the BoB. Dry anomalies develop in the equatorial Indian Ocean at this time. From day 0 to day 15, a similar evolution of SST and precipitation anomalies occurs but with opposite sign. The quadrature relation between MISO precipitation and SST anomalies over the BoB region is quite clear.

We next examine spatial patterns of precipitation and SST anomaly tendencies (dP/dt and $dSST/dt$) associated with MISO (Figs. 4b,d). Precipitation anomalies exhibit a good in-phase relation with SST tendency, suggesting that MISO convection-related processes affect the intraseasonal variability of SST. SST anomalies may in turn affect precipitation tendency through their impact on the moisture field, and therefore contribute to the intraseasonal variability of precipitation. Our results suggest that MISO convection and SST may interact through affecting each other's tendency.

MISO precipitation anomalies affect SST primarily through changing surface heat fluxes (Figs. 5a,b). Positive (negative) precipitation anomalies are associated with more (less) cloud cover, which reduces (increases) surface downward shortwave radiation and cools (warms) the ocean surface. In addition, positive (negative) rainfall anomalies are also associated with strengthening (weakening) of surface latent heat flux, which also contributes to BoB SST cooling (warming). Surface latent heat flux anomalies are primarily caused by surface wind speed anomalies associated with MISO

(Fig. 5c); positive (negative) rainfall anomalies are associated with low-level westerly (easterly) anomalies, as has been noted in previous studies to be associated with a frictionally modified Gill-type response to the MJO heating (Wang and Li 1994; Maloney and Hartmann 1998). As a result, the southwesterly summer monsoon circulation is strengthened (weakened) in the wet (dry) phase of MISO, increasing (decreasing) the surface wind speed and subsequently increasing (decreasing) the surface latent heat flux. Changes in the surface longwave radiation and the sensible heat flux associated with MISO are relatively small compared to the surface shortwave radiation and the latent heat flux, and thus not shown. Results for EAS MISO composites also support these relationships (Figs. S6 and S7).

MISO precipitation anomalies are associated with surface wind convergence anomalies (Fig. 5d), which as discussed above may have contributions from MISO SST anomalies. However, in addition to the possible role of SST anomalies in causing the surface convergence, the substantial latent heating associated with the deep convection anomaly itself may also drive convergence anomalies (Gill 1980). There has been an extensive debate on the relative roles of these two mechanisms in causing precipitation anomalies in tropics (Wang and Li 1993; Battisti et al. 1999; Sobel and Neelin 2006).

To separate the role of deep tropospheric processes (pressure gradient and downward momentum mixing) and boundary layer pressure gradients (caused by SST gradients) in causing the tropical surface convergence,

FIG. 3. Lead-lag correlation between 20–90-day filtered (a) MISO SST and precipitation anomalies and (b) SST anomalies and dP/dt over the EAS region. (c),(d) As in (a) and (b), but for the BoB region. Negative (positive) days denote SST anomalies leading (lagging) precipitation anomalies. Gray lines denote correlations for each year (May–October) for 1998–2015, and black lines denote the averaged results. Maximum correlation (denoted by cyan lines) occurs on day -7 in (a), day $+2$ in (b), day -11 in (c), and day -2 in (d).

we employ the MLM of Back and Bretherton (2009) (see section 2b). Hereafter, we refer to the former process as MLM-850 hPa, as variables at 850 hPa (above the planetary boundary layer) are used as inputs, and the latter process as MLM-PBL. Although there are some limitations for the MLM, such as the simple form of friction, the choice of 850 hPa as the bottom of free troposphere, and the empirically determined constant parameters, this model provided qualitative estimates of the relative roles of the free troposphere process versus the boundary layer process in causing the surface convergence.

Associated with peak precipitation anomalies in the composite of MISOs over EAS (Fig. S6), there are prominent surface convergence anomalies over the EAS region, accompanied by anomalous divergence over the equatorial region (Fig. 6a). Note that Fig. 6a is based on the MISO composite over the EAS region, which should be compared with the observational results of other fields in Figs. S6 and S7. The composites based on the BoB MISOs are shown in Figs. 4 and 5 (see also Fig. S8).

The MLM simulates the pattern of convergence anomalies in the tropical Indian Ocean reasonably well, although it underestimates divergence anomalies in the equatorial Indian Ocean (Fig. 6b). Additionally, the surface convergence anomalies in MLM lead those in observations by ~ 2 – 3 days, suggesting a bias in MLM. Surface convergence anomalies in MLM can be further separated into contributions from the deep troposphere and the boundary layer processes, which are shown to play comparable roles in causing the total convergence associated with the MISO (Figs. 6c,d,e). For instance, the boundary layer (i.e., SST anomalies) effect contributes $\sim 46\%$ to the total MISO convergence in the EAS when the effect of SST anomalies peaks, and the free troposphere effect contributes 64% during its peak (Fig. 6). Note that at different stages of the MISO, contributions of these two processes vary. The spatial correlation between MLM and MLM-850 hPa (MLM-PBL) over the tropical Indian Ocean is 0.71 (0.83). In MLM-PBL, surface convergence anomalies slightly lead those in MLM-850 hPa, which are associated with

FIG. 4. Composites of 20–90-day filtered (a) MISO precipitation anomalies (mm day^{-1}), (b) SST tendency (K day^{-1}), (c) SST anomalies (K), and (d) precipitation tendency (mm day^{-2}). Shown are results from day -15 to day 15 at 5-day intervals. MISO events are defined based on BoB precipitation index in Fig. 1b. Shown are results that are statistically significant at the 90% confidence level.

FIG. 5. As in Fig. 4, but for anomalies of (a) surface shortwave radiation (W m^{-2}), (b) surface latent heat flux (W m^{-2}), (c) surface wind (vectors; m s^{-1}) and surface wind speed (shading; m s^{-1}), and (d) surface divergence (10^{-6} s^{-1} ; negative means convergence). Positive means downward in maps of surface heat flux anomalies (SST warming). Shown are results that are statistically significant at the 90% confidence level.

FIG. 6. (a) Composite of 20–90-day filtered surface divergence anomalies associated with MISO on day 0 (10^{-6} s^{-1} ; negative means convergence). MISO events are defined based on the EAS precipitation index in Fig. 1a. (b) As in (a), but for simulated surface divergence anomalies in the mixed layer model (Back and Bretherton 2009). MISO composites of 20–90-day filtered anomalies of 850-hPa wind, geopotential height, and surface pressure are used as inputs. Values of parameters are the same as Back and Bretherton (2009). (c),(d) As in (b), but for the results from MLM-850 hPa and MLM-PBL, in which the surface divergence anomalies are caused by deep tropospheric processes and planetary boundary layer process, respectively. The number in parentheses in (c) is the spatial correlation between surface divergence in MLM-850 hPa and MLM over the region $30^{\circ}\text{--}120^{\circ}\text{E}$, $20^{\circ}\text{S--}30^{\circ}\text{N}$. In (d), the number is for the correlation between MLM-PBL and MLM. Contours in (d) are for MISO SST anomalies (0.05-K interval). (e) Composited surface divergence anomalies averaged over the EAS region ($65^{\circ}\text{--}75^{\circ}\text{E}$, $10^{\circ}\text{--}20^{\circ}\text{N}$) from day -15 to day 15 . The solid black line represents the observational result, dashed black the MLM result, blue MLM-850 hPa, and red MLM-PBL. Surface wind and pressure, 850-hPa wind, geopotential height, and SST are from ERA-Interim for the period 1998–2015.

preceding warm SST anomalies relative to the peak of precipitation anomalies (Figs. 2 and 6). These results suggest that MISO SST anomalies play an important role in driving surface convergence anomalies, and the associated moistening (drying) may contribute to the increasing (decreasing) precipitation rate. Note that

positive SST anomalies and their associated surface convergence anomalies are located in the northern part of the EAS region on day 0 (Fig. 6). This result is consistent with existing studies in that the northward propagation of SST anomalies of MISO leads precipitation (Fig. 4).

In the BoB MISO composite, MLM-850 hPa plays a more dominant role compared to MLM-PBL, while the latter is still not negligible (Fig. S8). For BoB, the maximum effects of SST anomalies and free troposphere processes contribute 46% and 63% to the total surface convergence anomalies during their peak, respectively (Fig. S8). Hence, the relative roles of free troposphere process and SST anomalies in causing the anomalous surface convergence and subsequent precipitation associated with MISO may vary in different regions. On the other hand, the composited BoB convergence is notably overestimated in MLM (Fig. S8), indicating an overestimated effect of the free troposphere process as represented by MLM-850 hPa. In observations, convergence anomalies are in phase with positive precipitation anomalies with no apparent lead (Fig. 2b), suggesting the prominent role of the free troposphere process in causing surface convergence anomalies (Fig. 6; see also Fig. S8).

In summary, observations suggest that strong air–sea coupled processes over the north Indian Ocean play an important role in the MISO, which leads to a quadrature relation between MISO precipitation and SST anomalies. The impact of MISO precipitation anomalies on the SST tendency through changing surface heat fluxes is consistent with previous studies (Krishnamurti et al. 1988; Bhat et al. 2001; Sengupta and Ravichandran 2001; Sengupta et al. 2001; Webster et al. 2002; Vialard et al. 2012; Sharmila et al. 2013). Conversely, MISO precipitation anomaly can be a response to SST anomaly, but with a time lag (Roxby et al. 2013; Wu et al. 2008). We demonstrated above that this time lag may be explained by the fact that the precipitation tendency is strongly affected by SST anomalies through their impact on the surface convergence. A theoretical model that encompasses these air–sea coupled feedbacks using observational constraints will be developed next.

4. Local air–sea coupling model

a. Theoretical framework

Based on the observed air–sea interaction processes over the north Indian Ocean associated with MISO, we develop a theoretical framework and discuss its implications for observed MISO characteristics. We begin with an ocean mixed layer heat budget that governs the SST variability. Previous studies found that MISO SST anomalies are largely determined by changes in surface heat fluxes (Sengupta et al. 2001; Sengupta and Ravichandran 2001; Vialard et al. 2012), and therefore we focus only on surface heat flux forcing and neglect oceanic advection and entrainment terms. Given these assumptions the approximate equation for the ocean heat budget is

$$\frac{d\delta T}{dt} = \frac{\delta Q_{sw} - \delta Q_{lh}}{\rho_o C_p H}, \quad (1)$$

where δT represents MISO SST anomalies, δQ_{sw} and δQ_{lh} stand for surface shortwave radiation and latent heat flux anomalies associated with the MISO; ρ_o denotes the density of mixed layer seawater, C_p is the specific heat capacity, and H is the seasonal mixed layer depth (MLD). Note that the intraseasonal H variation over the north Indian Ocean associated with the MISO is neglected here because it is small compared to the seasonal-to-interannual variations of H (Li et al. 2017a,b). Surface longwave radiation and sensible heat flux anomalies are neglected, as they are much smaller compared to shortwave radiation and latent heat flux anomalies (Fig. 7).

As shown by the observational analysis above, both δQ_{sw} and δQ_{lh} are closely associated with MISO precipitation anomalies. The former relation is dominated by changes in the cloud cover due to the MISO convection, which is mathematically represented by the following equation:

$$\delta Q_{sw} = -\gamma_1 \delta P, \quad (2)$$

where γ_1 is a positive physical parameter that represents the linear relation between the precipitation and δQ_{sw} . To relate δQ_{lh} to the surface wind speed, we use the bulk formula for Q_{lh} :

$$\begin{aligned} Q_{lh} &= V L_v \rho_a C_E (q_s - q_a) \\ &= V L_v \rho_a C_E q_s (T) (1 - RH \times e^{\lambda' S}). \end{aligned} \quad (3)$$

In the above, V denotes the surface wind speed, L_v the latent heat of evaporation, ρ_a the surface air density, C_E the heat exchange coefficient of Q_{lh} , RH the relative humidity of surface air, and q_s and q_a the specific humidity of sea surface and surface air. The Clausius–Clapeyron equation is used to relate q_a to q_s with a temperature difference $S = T_a - T_s$, and $\lambda' = L_v / R T_s T_a$, in which R is the ideal gas constant for water vapor and $\lambda' \sim 0.06 \text{ K}^{-1}$ for the tropical region. A more detailed derivation of Eq. (3) can be found in previous studies (Richter and Xie 2008; Xie et al. 2010; Zhang and Li 2014). Assuming δQ_{lh} is primarily associated with changes in V and $q_s(T)$, if one relates changes in the surface wind speed to precipitation anomalies (Wang and Li 1994) (i.e., $\delta V = \gamma_2 \delta P$) we obtain

$$\delta Q_{lh} = \gamma_2 \delta P \frac{\overline{Q_{lh}}}{\overline{V}} + \lambda' \delta T \overline{Q_{lh}}. \quad (4)$$

In Eq. (4), the overbar represents mean state variables. We treat $\overline{Q_{lh}}/\overline{V}$ as a constant estimated by observed values, which is evaluated to be ~ 15 during the Indian summer monsoon season (Fig. S9).

FIG. 7. (a) Composites of 20–90-day filtered, domain-averaged precipitation (pink; cm day^{-1}), SST (solid black; K), and surface wind speed anomalies (black dashed; 10 m s^{-1}) for EAS MISOs on the left axis, together with composites of surface heat flux anomalies (positive means downward), including surface latent (solid green) and sensible heat fluxes (dashed green) and shortwave (solid red) and longwave radiation (dashed red) anomalies (W m^{-2}) on the right axis. (b) As in (a), but for BoB MISOs.

With Eqs. (1), (2), and (4), it can then be shown that

$$\frac{d\delta T}{dt} = \frac{-\gamma_1 - \gamma_2}{\rho_0 C_p H} \delta P - \frac{\lambda' \overline{Q_{lh}}}{\rho_0 C_p H} \delta T = -\alpha \delta P - \lambda \delta T, \quad (5)$$

where $\gamma_2 = \gamma'_2 \overline{Q_{lh}} / \overline{V}$, $\lambda = \lambda' \overline{Q_{lh}} / \rho_0 C_p H$, and $\alpha = (\gamma_1 + \gamma_2) / \rho_0 C_p H$. Hence, MISO SST anomalies are determined by precipitation-related surface heat flux anomalies and the Newtonian damping process; the Newtonian damping term comes from δQ_{lh} associated with SST anomalies [second term in Eq. (4)], which is a cooling effect given a positive SST anomaly (i.e., a greater surface latent heat flux).

MISO precipitation anomalies can be in turn affected by SST anomalies through changes in the surface convergence (i.e., an SST anomaly produces a surface convergence anomaly that induces a dP/dt anomaly). This process can be empirically written as

$$\frac{d\delta P}{dt} = \beta \delta T, \quad (6)$$

where β relates MISO SST anomalies to precipitation tendency anomalies and can be estimated based on observations. Here we assume that the MISO precipitation is predominantly affected by SST anomalies, while we neglect other processes that may also cause rainfall anomalies at intraseasonal time scales. This is a caveat that one needs to bear in mind.

Finally, combining Eqs. (5) and (6) yields

$$\frac{d^2 \delta P}{dt^2} + \lambda \frac{d\delta P}{dt} + \alpha \beta \delta P = 0. \quad (7)$$

Equation (7) is a *damped oscillating system*, which depicts physical processes associated with the local air–sea interaction over the north Indian Ocean due to the MISO. This is a one-dimensional *local air–sea coupling model* (LACM). The frequency of LACM ω is determined by the following equation:

$$\omega = \sqrt{\alpha \beta - \lambda^2 / 4} \approx \sqrt{\alpha \beta}. \quad (8)$$

In Eq. (8), the damping term is small and thus can be neglected when considering the frequency of the system, which is determined by multiple physical parameters associated with air–sea coupling processes as described below. Note that this frequency/period is essentially the intrinsic frequency/period that the system prefers because of the local air–sea coupling. Consequently, MISOs propagating from the equatorial Indian Ocean into the north Indian Ocean are modified by the local air–sea coupling processes.

b. MISO in LACM

In this section, we estimate key parameters α (proportional to $\gamma_1 + \gamma_2$) and β that determine the frequency ω of the LACM based on composited MISO events in observations (Fig. 7). It is worth noting that α is inversely proportional to the mean mixed layer depth H , and ω is related to the square root of H . Consequently,

the seasonal cycle of H can substantially affect the MISO amplitude and frequency in LACM. The role of the MLD in MISO SST anomalies has been noted in previous studies. It has been shown that a shallower MLD prompts a larger SST response given the same surface heat flux forcing associated with the MISO precipitation (Fu et al. 2003; Waliser et al. 2004; Vialard et al. 2012; Li et al. 2016, 2017a). It has also been shown that the mean monsoon flow and moisture gradient play important roles in MISO (Jiang et al. 2004; Wu et al. 2006; DeMott et al. 2013; Adames et al. 2016). Although these factors are not considered in LACM, the seasonal variation of H itself can be associated with the large-scale seasonal variation of the circulation and the freshwater flux (Pookkandy et al. 2016). Consequently, the LACM results also reflect the impact of the seasonal variation of atmospheric conditions to some extent. Similarly, although the ocean dynamical effect is not explicitly considered in LACM, its role in modulating MISO SST anomalies via affecting the seasonal variations of MLD is included, as shown in Eq. (1).

Positive (negative) rainfall anomalies are associated with large (small) cloud cover that reduces (increases) the surface incoming shortwave radiation, as shown by observations (Fig. 7). Indeed, we find a good linear relation between the 20–90-day filtered precipitation anomalies and associated δQ_{sw} (Fig. 8). A linear regression analysis yields $\gamma_1 = 3.74$ (4.1) for the EAS (BoB) region. Similarly, we obtain that $\gamma_2 = 0.152$ (0.208) based on the relation between surface wind speed and precipitation anomalies. The parameter β , which is associated with the response of the MISO precipitation tendency to SST anomalies, is equal to 4.31 (5.73). Note that the value β in Fig. 8 represents the rainfall tendency *per day* for a 1° change of the SST anomaly, and this value needs to be converted so that it corresponds to the rainfall tendency in units of *per second* (divided by $60 \times 60 \times 24$). These physical parameters are estimated based on the composited MISO event (Fig. 7), which provides useful information on MISO statistics. For each individual MISO event, however, the parameters can exhibit quite different characteristics (Figs. S3 and S4), which leads to different relations between physical variables and thus different values of LACM parameters. These discrepancies might be attributed to complicated causes for the MISO precipitation (e.g., atmospheric internal variability, high-frequency weather noise, and oceanic processes) that are not considered in the LACM.

The MISO period in LACM, which is the inverse of frequency ($\omega \sim 2\pi/T$), is determined by physical parameters γ_1 (rainfall–surface shortwave radiation), γ_2

(convection–surface wind speed), β (SST– dP/dt), and the seasonal MLD H . Here we show the sensitivity of the LACM period to variations of these parameters (Fig. 9). Consistent with Eq. (8), the LACM period decreases with γ_1 , γ_2 , and β , and increases with H . For instance, with observed values of γ_1 and γ_2 , the LACM period increases from 27 (38) days to 57 (80) days when β decreases from 9 to 2, if one sets $H = 25$ (50) m (Fig. 9a). Also note that the LACM period is more sensitive to variations in β and H than those in γ_1 and γ_2 (Fig. 9).

With empirically determined values of parameters ($\gamma_1 = 4$, $\gamma_2 = 0.2$, $\beta = 5$), we obtain an oscillation period of ~ 36 days when $H = 25$ m and 51 days when $H = 50$ m, and the LACM produces correct phase relations between precipitation, SST, and surface heat flux anomalies as expected (Fig. 10): the quadrature relation between precipitation and SST anomalies, and the cooling (warming) effect due to δQ_{sw} and δQ_{lh} associated with positive (negative) rainfall anomalies, are all consistent with observations (cf. Figs. 7 and 10). The selected H values are based on the observations of 25 (50) m over the EAS region during the developing (mature) phase of the Indian summer monsoon (Fig. 11). In this paper, we define May and June as the developing phase, July and August as the mature phase, and September and October as the decaying phase. Over the BoB region, the seasonal cycle of H is similar, but with a smaller amplitude (Fig. 11). Consistent with the LACM results, previous studies have indeed identified two peaks in the MISO period. One is ~ 30 days and the other ~ 50 days (Krishnamurthy and Shukla 2008; Krishnamurthy and Achuthavarier 2012). Given that the only physics in LACM is the local air–sea coupling and its period essentially represents the intrinsic period that the local air–sea coupled system prefers, the qualitative agreement between LACM results and observations suggest an important modification effect of the local air–sea coupling on MISOs over the north Indian Ocean.

c. Resonance in LACM

As mentioned above, the current version of the LACM is a damped oscillating system, with no instability, as shown by Eq. (7). Therefore, external forcing is required to trigger/maintain oscillations in LACM. As found by previous studies, MISOs over the north Indian Ocean originate from convection anomalies over the equatorial Indian Ocean that propagate northward during boreal summer (Sikka and Gadgil 1980; Hartmann and Michelsen 1989; Annamalai and Slingo 2001). The northward propagation of MISOs is clear over both EAS and BoB regions (Fig. 4; see also Figs. S6 and S10). The MISO precipitation variability is notably greater over the north Indian Ocean compared to its

FIG. 8. (top) Scatterplots of composited MISO precipitation anomalies against surface shortwave radiation anomalies, based on which the parameter γ_1 is evaluated using the linear regression method. (middle) The relation between precipitation and surface wind speed anomalies and the value of parameter γ_2' . (bottom) The relation between SST anomalies and the precipitation tendency and the value of parameter β . Results are shown for (a) EAS MISOs and (b) BoB MISOs. Considering the small lags between various variables (Fig. 7), the scatterplot and the regression coefficient between MISO precipitation and surface shortwave radiation is evaluated when precipitation leads by 1 (1) day over EAS (BoB). The surface wind speed lags precipitation by 6 (2) days over EAS (BoB) in the middle panels, and the precipitation tendency leads (lags) SST by 2 (1) days over EAS (BoB) in the bottom panels.

initiation region, the equatorial Indian Ocean (Fig. S10). There also seems to be a northward jump of MISO signals between 5° and 10°N , right before the MISO amplification. Hence, MISOs over the north Indian Ocean may be caused by a combined effect of the local air–sea interaction and the “external forcing” from the equatorial Indian Ocean. In the north Indian Ocean where the summertime SST is high, the local air–sea interaction is active and may play an important role in MISO, while intraseasonal oscillations propagating into this region from the equatorial Indian Ocean may act

as a trigger. To simulate this external equatorial forcing effect, we add an oscillating external forcing with period ~ 30 days to Eq. (7), and obtain

$$\frac{d^2\delta P}{dt^2} + \lambda \frac{d\delta P}{dt} + \alpha\beta\delta P = F \cos(\omega_{\text{ext}}t), \quad (9)$$

where $\omega_{\text{ext}} = 2\pi/30$ days, and F denotes the amplitude of the external forcing. The period of the external forcing is determined based on the composited intraseasonal oscillations over the initiation region (i.e., the equatorial

FIG. 9. (a) LACM period (day) predicted by the local air–sea coupling model with varying parameter β and mean-state mixed layer depth H , and fixed parameters $\gamma_1 = 4$ and $\gamma_2 = 0.2$. Black stars represent period values with $\beta = 5$ and $H = 25$ (50) m, which are determined based on observations. (b) LACM period with varying γ_1 and γ_2 , and $\beta = 5$ and $H = 25$ m.

Indian Ocean). This will be discussed in more detail in section 4d.

In Eq. (9), the MISO frequency is not solely determined by the internal, local air–sea coupling processes anymore, but rather is also affected by the external

forcing. More importantly, a resonance will occur when the frequency of the external forcing is close to the internal frequency, which can lead to MISO amplification. Hereafter, we refer to the internal frequency of LACM determined by the local air–sea interaction

FIG. 10. Simulated MISO precipitation (solid pink; cm day^{-1}) and SST anomalies (solid black; K) in LACM on the left axis. On the right axis, surface heat flux anomalies (positive means downward; W m^{-2}), including surface shortwave radiation (solid red) and latent heat flux (solid green). Blue is the sum of the two surface heat fluxes. Parameters are chosen as $\gamma_1 = 4$, $\gamma_2 = 0.2$, and $\beta = 5$. Results are shown for (a) $H = 25$ m and (b) $H = 50$ m. Note that the lengths of the x axes are different between (a) and (b), due to different periods of simulated MISOs given different mixed layer depths H . The model was initialized with $\delta P = -5 \text{ mm day}^{-1}$ and a small positive perturbation in δSST in both experiments.

FIG. 11. (a) Seasonal cycles of domain averaged standard deviation of 20–90-day filtered precipitation anomalies (mm day^{-1}) over the EAS (red), the BoB (blue), and the initiation region (black; $65^{\circ}\text{--}95^{\circ}\text{E}$, $2.5^{\circ}\text{S--}2.5^{\circ}\text{N}$; Fig. S1). (b) As in (a) but for the seasonal cycle of the climatological monthly mixed layer depth (m).

[Eq. (8)] as ω_{int} , to distinguish from the frequency of the external forcing ω_{ext} . As mentioned earlier, ω_{int} is determined by the seasonal MLD, assuming other parameters are constants. Below, we choose different values of H to evaluate changes in the MISO amplitude in LACM (Fig. 12). The amplitude of simulated oscillations is much larger when $H = 25$ m (internal period ~ 36 days). In contrast, either $H = 10$ m (internal period ~ 23 days) or $H = 50$ m (internal period ~ 51 days) yields much weaker oscillations. These results are consistent with previous finding in Sobel and Gildor (2003) and Maloney and Sobel (2004), who found a non-monotonic response of intraseasonal precipitation variability to increase in the MLD, with the optimal $H \sim 20$ m corresponding to the strongest intraseasonal precipitation variability.

These LACM results qualitatively agree with the observed seasonal cycle of MISO amplitude and MLD (Fig. 11). The MLD in the north Indian Ocean is shallower during the monsoon developing phase when it is close to 25 m. As predicted by the LACM, the MISO amplitude should be larger (smaller) during the

developing (mature) phase. Indeed, the standard deviation of 20–90-day filtered precipitation peaks in May and June and decreases in July and August. Note also that the seasonal variation of the MISO amplitude over the EAS region is much greater than that over BoB during the summer monsoon season, which is consistent with the much greater variation in the seasonal MLD in the former region.

d. MISO characteristics during different monsoon phases

The climatological mean state over the north Indian Ocean exhibits evident seasonal variations, which leads to distinct MISO characteristics during different phases of the Indian summer monsoon (Figs. 11 and 13). Analysis of the composited MISO over the EAS and BoB regions shows that the MISO period during the monsoon developing phase is shorter than that during the mature phase (Figs. 13a,c), which qualitatively agrees with LACM and the seasonal variation of MLD, which is shallow (deep) during the developing (mature) phase. Given that individual MISO events may be different (Figs. S3 and S4), percentiles of the MISO period for all the MISO events are also shown (Fig. 14). It is found that values of the mean, median, and 75th percentile of the MISO period are indeed smaller during the developing phase than the mature phase. Over the EAS (BoB), the mean MISO period is around 31 (32) days during the developing phase, and around 34 (35) days during the mature phase, and this difference is statistically significant at the 80% (70%) confidence level. It is noted that the difference of the MISO period during different monsoon phases is smaller than that predicted by Eq. (8) with different H (Figs. 9 and 10). This discrepancy may be due to the influence of the external forcing [Eq. (9)]; when the oscillating external forcing is added to the LACM, the model period is determined by both the internal period and the period of the external forcing. Consequently, the MLD variation can only induce a small difference in the model period. During the decaying phase, the MISO period becomes short over the BoB due to the shallow MLD (Figs. 13c and 14c). Over the EAS, the decrease of the MISO period from the mature phase to the decaying phase is small, although the 75th percentile of the MISO period during the decaying phase is clearly smaller than that during the mature phase.

In contrast with the north Indian Ocean, the period of intraseasonal oscillations over the equatorial Indian Ocean (i.e., the initiation region where MISOs originate) exhibits very different seasonal variation (Figs. 14b,d). For MISOs over the BoB region, the seasonal variation of associated intraseasonal oscillations over the initiation

FIG. 12. Simulated MISOs in LACM with an oscillating external forcing (period = 30 days). Precipitation (pink; cm day^{-1}) and SST anomalies (solid black; K) in LACM are shown on the left axis. The right axis shows surface heat flux anomalies (positive means downward; W m^{-2}), including surface shortwave radiation (solid red) and latent heat flux (solid green). Blue is the sum of the two surface heat fluxes. Parameters are chosen as $\gamma_1 = 4$, $\gamma_2' = 0.2$, and $\beta = 5$. Shown are results for (a) $H = 10$ m, (b) $H = 25$ m, and (c) $H = 50$ m. Values of the internal period and the period of the external forcing are shown in top of each label. Note the distinct MISO amplification in (b), which is associated with resonance in LACM. The model was initialized with zero values in both δP and δSST , and a same small positive value for F in all experiments.

region is opposite to that over the BoB: the period is shorter during the mature phase and longer during the developing and the decaying phases. For MISOs over the EAS, the initiation region exhibits small seasonal variation, with the longest period occurring during the

decaying phase. These results suggest that the seasonal variation of the MISO period is not caused by changes of intraseasonal variabilities over the initiation region. The period of intraseasonal oscillations over the equatorial initiation region is around 30 days on average, which

MISO composite during different Indian summer monsoon phases

FIG. 13. Observed MISO composites during different Indian summer monsoon phases over (a) EAS, (b) the initiation region 65° – 95° E, 2.5° S– 2.5° N (based on EAS MISOs), (c) BoB, and (d) the initiation region (based on BoB MISOs), during (left) the developing phase (May and June), (center) the mature phase (July and August), and (right) the decaying phase (September and October). In each panel, the amplitude of the composited MISO is shown in the top-left corner and the MISO period in the top-right corner. The MISO period is defined as the time difference between the closest minima prior to and after the peak precipitation anomaly on day 0. MISO amplitude is defined as the difference between the peak precipitation anomaly on day 0 and the smallest value in the closest minima prior to and after the peak (dashed gray lines). For the initiation region, peak precipitation anomaly is chosen as the closest peak prior to day 0. Solid gray lines show precipitation anomalies for each individual MISO event.

FIG. 14. Percentiles of observed MISO period during different monsoon phases over the (a) EAS and (b) initiation region 65° – 95° E, 2.5° S– 2.5° N (based on EAS MISOs). Gray bars denote the 25th–75th percentiles. Vertical lines denote the 10th–90th percentiles. Hollow circles and crosses denote the median and mean values, respectively. Dashed cyan lines connect the mean values (crosses) during different monsoon phases. (c), (d) As in (a) and (b), but for BoB MISOs. The MISO period is defined as the time difference between the closest minima prior to and after the peak precipitation anomaly on day 0. For the initiation region, peak precipitation anomaly is chosen as the closest maximum prior to day 0.

corresponds to the period of the external forcing that we add to LACM (section 3c).

MISO amplitude also exhibits evident seasonal variation (Figs. 13 and 15). Over both EAS and BoB, the amplitude of the composited MISO is higher during the developing phase compared to the mature phase (Figs. 13a,c), and such seasonal variation of the MISO amplitude is more evident over EAS compared to BoB. Further, it is found that the 25th percentile of the MISO amplitude over EAS during the developing phase is close to the 75th percentile of that during the mature phase (Fig. 15a). Difference of the mean MISO amplitude over EAS between the developing and the mature phases is statistically significant at the 95% confidence level. Such seasonal variation of MISO amplitude is also consistent with LACM results (section 3c). The internal

period associated with the local air–sea coupling is close to that of the external forcing during the developing phase, which may lead to MISO amplification (resonance effect); the deep MLD during the mature phase leads to a longer internal period, and hence weaker MISOs. Over the initiation region, the amplitude of intraseasonal oscillations is strongest during the monsoon decaying phase (Figs. 13 and 15), which is different from MISOs over the north Indian Ocean.

It is also found that the MISO amplitude is small during the decaying phase (Figs. 15a,c), which is inconsistent with LACM results given a shallow MLD. Such discrepancy could be due to the longer period of intraseasonal oscillations over the initiation region during the decaying phase than the developing and the mature phases (Figs. 13b,d and 14b,d); the spread of

FIG. 15. As in Fig. 14, but for percentiles of MISO amplitude. MISO amplitude is defined as the difference between the peak precipitation anomaly on day 0 and the smallest value in the closest minima prior to and after the peak. For the initiation region, peak precipitation anomaly is chosen as the closest maximum prior to day 0.

these periods is also larger during the decaying phase than the developing phase. These changes of the initiation region could reduce the resonance effect as suggested by LACM, contributing to the weak MISOs during the decaying phase. In addition, this model discrepancy could also be associated with effects of atmospheric internal dynamics and subsurface ocean dynamics that are missing in LACM. For instance, the barrier layer thickness (BLT) over the BoB is deeper during the monsoon decaying phase compared with the developing phase (Li et al. 2017a), which could cause weaker intraseasonal variability of SST (Li et al. 2017b).

The evident seasonal dependence of MISO statistics over the north Indian Ocean, as well as its inconsistency with the seasonal variation of intraseasonal oscillations over the initiation region, suggests that different behaviors of MISOs during different monsoon phases are not caused by the latter. Instead, the qualitative agreement between the LACM and observations suggest an important

modification effect of the local air–sea coupling over the north Indian Ocean on MISO. However, other factors such as the seasonal evolution of the monsoon flow and the moisture distribution may also play important roles in causing the seasonal variation of MISO (Jiang et al. 2004; Wu et al. 2006; DeMott et al. 2013; Adames et al. 2016). Nevertheless, the theoretical framework of the LACM put forward here accounts for an important part of the complicated physical processes underlying the MISO, and it is useful for explaining some of the observed MISO characteristics and providing insights into some processes. Since the local air–sea interaction in the north Indian Ocean is less explored in existing MISO studies, our effort complements the knowledge of MISO dynamics.

5. Summary and discussion

The Indian summer monsoon exhibits strong intraseasonal variability, manifest as northward propagating

convection anomalies originating from the equatorial Indian Ocean. Over the north Indian Ocean, the monsoon intraseasonal oscillation (MISO) precipitation produces prominent SST anomalies on the underlying ocean surface; these SST anomalies may in turn feed back onto precipitation anomalies, suggesting that the local air–sea coupling plays an important role in the MISO.

The impact of MISO precipitation anomalies on SST is primarily through changing the surface shortwave radiation and the latent heat flux. While the former is related to cloudiness anomalies, the latter is primarily caused by surface wind speed anomalies. Positive precipitation anomalies are associated with low-level westerly anomalies enhancing the background summer monsoon circulation over the north Indian Ocean, leading to greater surface latent heat flux; the opposite occurs for dry anomalies. These results are consistent with previous findings (Krishnamurti et al. 1988; Sengupta et al. 2001; Sengupta and Ravichandran 2001; Bhat et al. 2001; Webster et al. 2002; Vialard et al. 2002; Sharmila et al. 2013).

SST anomalies may affect the precipitation tendency (dP/dt) at intraseasonal time scales. Under the assumption that this occurs, positive (negative) SST anomalies induce increasing (decreasing) precipitation rates. The largest dP/dt coincides with the warmest SST anomalies. Spatial patterns of MISO SST anomalies and dP/dt also exhibit remarkable consistency. The impact of SST anomalies on dP/dt is likely associated with surface convergence anomalies. By employing the mixed layer model of Back and Bretherton (2009), which invoked the Lindzen–Nigam mechanism, we separate the effects of deep tropospheric processes and boundary layer process on the surface convergence. The results show that a large fraction of MISO surface convergence anomalies result from the anomalous SST gradients. We have also calculated the surface wind convergence induced by SST anomalies (i.e., convergence is proportional to the Laplacian of intraseasonal SST) as in Li and Carbone (2012), and the results confirm the importance of SST anomalies from the mixed layer model (not shown). The physically meaningful results obtained from the mixed layer model, and their agreement with that from the Li and Carbone approach, suggest that the mixed layer model applies to our MISO study, even though the original focus of Back and Bretherton (2009) was seasonal variability. The strong air–sea coupled processes lead to a quadrature relation between MISO SST and rainfall, consistent with the time lag of the MISO precipitation response to SST anomalies found in previous observational studies (Wu et al. 2008; Roxy et al. 2013; Xi et al. 2015). Note that the

simple models cannot fully represent the complex air–sea coupling processes associated with MISO. Further studies are needed using general circulation models to more accurately quantify the role of SST anomalies in causing MISO precipitation.

A local air–sea coupling model (LACM) is developed based on observed physical processes associated with MISO. In LACM, precipitation (SST) anomalies determine the tendency of SST (precipitation). MISO air–sea interactions represented in this manner lead to an oscillating system with a quadrature relation between the two variables, as found in observations. In the LACM, the period of the oscillation is determined by multiple physical parameters including the rainfall–surface shortwave radiation relation γ_1 , convection–surface wind speed relation γ_2 , SST– dP/dt relation β , and the mean mixed layer depth H . The first three parameters are empirically determined based on observations and are assumed to be constants with no seasonal cycle. The depth H , on the other hand, exhibits evident seasonal variations over the north Indian Ocean, which substantially affects the LACM period. A shallow (deep) MLD leads to a short (long) period of the model oscillation. It is worth noting that atmospheric internal dynamics as well as the mean-state circulation and moisture pattern, which have been shown to play important roles in MISO (Kemball-Cook and Wang 2001; Jiang et al. 2004; Wu et al. 2006; DeMott et al. 2013; Adames et al. 2016), are not considered in LACM. Therefore, LACM cannot quantify the relative importance of these processes versus the local air–sea interaction in MISO. This is one caveat of this simple model.

The LACM depicts the local air–sea coupled processes over the north Indian Ocean. With no external forcing or other instability mechanism, the period of the model oscillation should represent the intrinsic period of the air–sea coupled system. However, in observations the MISO is also affected by convection anomalies originating from the equatorial Indian Ocean. Thus, we add an oscillating external forcing with period of 30 days to represent this process. The period of LACM is then determined by both the internal period ($\sim\sqrt{H}$) and the period of the external forcing. Because of the seasonal variation of mixed layer depth over the north Indian Ocean, the internal period predicted by LACM varies substantially during different Indian summer monsoon phases. During the developing phase (May and June), H is relatively shallow, leading to an internal period of 36 days, very close to the period of external forcing from the equatorial Indian Ocean. Consequently, resonance occurs that contributes to the local amplification of MISO signals. During the mature phase (July and

August), H deepens and the internal period becomes longer (~ 51 days), while the period of the external forcing exhibits no significant increase compared to that during the developing phase. Hence, no resonance occurs, and MISO amplitude is predicted to be relatively small.

The LACM simulation is qualitatively consistent with observed MISO statistics, suggesting an important modification effect of the local air–sea coupling on MISO. In observations, MISO amplitude is indeed greater during the monsoon developing phase over both East Arabian Sea (EAS) and Bay of Bengal (BoB), with a relatively short MISO period that is close to the 30-day period of intraseasonal variability over the equatorial Indian Ocean. During the monsoon mature phase, MISO events tend to be weaker and last longer, which, based on LACM results, is associated with deeper mixed layer depth over the north Indian Ocean. During the decaying phase (September and October), H becomes smaller, and the MISO period becomes shorter. However, the MISO amplitude is relatively small during September and October, which may be associated with the relatively long period of intraseasonal oscillations over the initiation region as well as the large spread of this period, both of which may reduce the resonance effect. Neglect of subsurface and horizontal oceanic processes in our one-dimensional air–sea coupled model may also contribute to these discrepancies between observations and LACM.

It is worth noting that although the LACM qualitatively agrees with observed MISO characteristics, it does not provide quantitative estimates of physical processes associated with MISO. Consequently, the LACM cannot be used for real-time MISO prediction. Instead, it improves our understanding of the air–sea interaction processes associated with MISO, and it may also be useful for diagnosing climate model biases in terms of the MISO simulation by estimating the LACM parameters and comparing them with those from observations. A recent paper by Li et al. (2018) has utilized a simplified version of LACM to explore the causes for the MISO biases in Climate Forecast System version 2 (CFSv2) forecasts.

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