Rainy season precipitation variation in the Mekong River basin and its relationship to the Indian and East Asian summer monsoons

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Abstract

In this paper, we study the relationship between the covariability of the Indian summer monsoon (ISM) and East Asian summer monsoon (EASM) and the rainy season precipitation anomaly in the Mekong River basin (MRB) using ERA-Interim reanalysis data and precipitation from the Climate Hazards Group Infrared Precipitation station for 1981–2016. The results indicate that the interannual variability of the rainy season precipitation in the MRB is significantly modulated by the ISM and EASM covariability. Herein, the ISM mainly influences the rainy season precipitation west of the MRB, and the EASM mainly influences the rainy season precipitation over the southeastern MRB. When the Bay of Bengal (BOB) and South China Sea (SCS) are dominated by positive anomalies of diabatic heating associated with a stronger ISM and EASM, the anomalous westerly winds in the BOB and significantly anomalous easterly winds in the SCS, which carry more warm-wet air, induce a stronger than normal convergence and upward motion in the MRB. This results in much heavier rainy season precipitation in the MRB. Model runs with a linear baroclinic model confirm the observational analysis results.

1 Introduction

The Mekong River basin (MRB), one of the most important transboundary rivers in southeast Asia, starts in Tibet, flows from China's Yunnan Province to Vietnam, and finally empties into the South China Sea (Fig. 1). The entire basin is divided into two regions, the Upper MRB (UMRB), where the river is also called the Lancang Jiang, and the Lower MRB (LMRB), with a boundary at approximately 21°N (Mekong River Commission 2005). The geomorphology of the UMRB transforms from high mountains and deep valleys to low/medium mountains and wide

valleys in a northwest–southeast orientation. Compared to the UMRB, the LMRB has a flatter land surface, lower elevation, and more precipitation (Mekong River Commission 2005). The regional ecosystem, food, and many other socioeconomic activities all strongly rely on the Mekong River for sustenance (Ferguson et al. 2011). For example, the LMRB experienced the worst drought from late 2015 to mid-2016, with a relatively large area covering Vietnam, Cambodia, Laos and Thailand, which caused huge economic losses for local societies. Therefore, it is meaningful to understand the MRB rainfall variability.

The MRB experiences a typical monsoon climate (Stephen et al. 2016). A monsoon climate is a dry and wet alternation with the wet season lasting from mid-May to October (Matsumoto 1997). During the rainy season, the region around the MRB, which is influenced by the Indian summer monsoon (ISM) and East Asian summer monsoon (EASM) (Cao et al. 2012, 2016; Guo et al. 2016; Holmes et al. 2009; Tao et al. 2016), receives over 80% of its annual precipitation (Costa-Cabral et al. 2007; Kingston et al. 2011). In contrast, while there have been studies on the ISM and EASM variabilities (e.g., Chang and Wheeler 2004, 2017 and references cited therein; Wang 2006, and references cited therein), there have been relatively few studies on the key physical processes associated with the variability in the

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Fig. 1 The study area topography with climatologic surface winds averaged from June to September. The shaded area denotes the altitude (units: m)

rainy season precipitation over the MRB. Endo et al. (2009) found that the average precipitation intensity on wet days increased during the 1950s–2000s. Xue et al. (2011) suggested that the runoffs of the middle to lower reach of the MRB, resulting from monsoonal precipitation, are correlated with the ISM, and those of the lower reach are mainly correlated with the EASM. Delgado et al. (2012) studied the flood season discharge at eight stations on the Mekong River. They found that the ISM has less influence on the interannual flood regime of the LMRB than its western Pacific counterpart but shows a stronger summer precipitation signature in the northern part of the MRB. Räsänen and Kummu (2013) indicated that the MRB precipitation in Southeast Asia has experienced large interannual variations in recent decades. The precipitation decreased (increased) during El Niño (La Niña) over the period of 1981–2005. Cao et al. (2014) found that the subtropical Indian Ocean dipole-like pattern is a key external thermal forcing for the summer rainfall variations around the UMRB rather than the El Niño–Southern Oscillation. Fan and He (2015) and Wu et al. (2016) suggested that the summer precipitation in the UMRB presents a slight increase in recent decades. Tsai et al. (2015) focused on the local precipitation over Myanmar and Thailand and defined four Indochina Monsoon Indices according to wind anomalies. They found that these indices correlate better with precipitation over Myanmar and Thailand. Ge et al. (2017) obtained similar results over the Indochina peninsula. Cao et al. (2017) found that the May precipitation around the UMRB is closely related to the thermal configuration of the Bay of Bengal–Tibetan Plateau region.

Although it is well accepted that the ISM and EASM combined is the fundamental driver of the Mekong hydroclimate (Holmes et al. 2009; Xue et al. 2011; Delgado et al. 2012; Tsai et al. 2015), these studies, as reviewed above, mainly focus on the impact of the individual ISM or EASM on precipitation over part of the MRB. The quantitative

relationship between the ISM and EASM covariability, anomalous MRB precipitation during the rainy season and corresponding physical processes remains unclear. This condition motivates us to determine whether there is a significant relationship between the ISM and EASM covariability and the anomalous precipitation over the MRB during the rainy season. If so, what are the key physical processes governing these mechanisms?

The remainder of the paper is arranged as follows. The data, method and linear baroclinic model (LBM, Watanabe and Kimoto 2000) are described in Sect. 2. In Sect. 3, the relationship between the rainy season precipitation anomaly in the MRB and the ISM or EASM is investigated, revealing the possible physical processes through which the ISM and EASM covariability influences the interannual variability of the rainy season precipitation in the MRB. The model results are shown in Sect. 3 and are used to confirm the relationship and associated physical processes also revealed in Sect. 3. Finally, a summary is presented in Sect. 4.

2 Data and method

We used the empirical orthogonal function (EOF) analysis to obtain the main spatial distribution of the rainy season precipitation over the MRB and its relation to the ISM and EASM. The linear correlation analysis and composite analysis and their corresponding Student's *t* test were used to reveal the associated key physical processes. We use daily ERA-Interim reanalysis data from the European Centre for Medium-range Weather Forecasts (ECMWF) (Simmons et al. 2004; Dee et al. 2011) for the period of 1979–2016. The resolution of the ERA-Interim reanalysis data is 2° in latitude and longitude with 37 pressure levels from 1000 to 1 hPa. The precipitation was analyzed using the Climate Hazards Group Infrared Precipitation with Station data (CHIRPS) (Funk et al. 2014). The CHIRPS resolution is 0.05° in latitude and longitude. Guo et al. (2017) indicated that the CHIRPS data are suitable for studying drought in the LMRB. In this study, the rainy season was June through September. The index developed by Goswami et al. (1999) was adopted to describe the ISM variability (hereafter, the index denoted as ISMI), and the index developed by Wang and Fan (1999) was adopted to describe the EASM variability (hereafter, the index denoted as EASMI). The ISMI is defined in terms of $V_{850} - V_{200}$, where V_{850} and V_{200} are the meridional velocities at 850 hPa and 200 hPa, respectively, averaged over the region of 70°–110°E, 10°–30°N. The EASMI is defined as the difference between the westerly anomalies averaged over (5°–15°N, 100°–130°E) and the westerly anomalies

averaged over (20°–30°N, 110°–140°E). The apparent heat source (Q_1) and apparent moisture sink (Q_2) are calculated using the daily ERA-Interim reanalysis data and the same method used in Tao et al. (2016, and the corresponding references cited therein).

Because Lu and Lin (2009), Tao et al. (2016) and Cao et al. (2017) suggested that the dry version LBM (DLBM) is more suitable than the moist version in analyzing the responses of regional diabatic heating, the DLBM with a horizontal resolution of T42 and 20 sigma levels in the vertical direction was adopted in this study to test the reliability of the key physical process linking the apparent heat source and apparent moisture sink variability with the MRB precipitation anomaly during the rainy season (Watanabe and Kimoto 2000, Appendix B). The model used in this study depends on primitive equations linearized about the rainy season climatology calculated from the ERA-Interim for 1979–2016. Similar to Tao et al. (2016) and Cao et al. (2017), the diabatic heating patterns associated with the anomalous configuration of ISM and EASM will show its importance in driving the large-scale circulation influencing the rainy season precipitation in the MRB. In the DLBM, the time scales of the Rayleigh's friction and Newtonian damping are 0.5 day^{-1} for $\sigma \geq 0.9$, 1 day^{-1} for $\sigma \leq 0.03$, and 30 day^{-1} for σ between 0.9 and 0.03. The circulation response reaches the steady state after approximately day 15.

3 Results and discussion

3.1 Observational analysis

3.1.1 Correlation between the ISM, EASM and rainy season precipitation in the MRB

First, we perform the EOF analysis on the rainy season precipitation of the MRB. The EOF results indicate that the top three modes accounted for 26.2%, 12.9% and 9.3% of the rainy season precipitation variability in the MRB, respectively. Figure 2a shows that the leading mode of the rainy season precipitation in the LMRB is almost positive and alternates with positive and negative precipitation in the areas north of 21°N. Because the explained variance of the first EOF mode is double that of the second EOF mode, the first EOF mode reflects the major pattern of the rainy season precipitation variability in the MRB. The MRB precipitation index (PI) is defined as the time series corresponding to its leading EOF mode.

To explore the relationship between the summer monsoon and MRB precipitation during the rainy season, we calculated the correlation coefficients between the rainy

season PI in the MRB, ISMI and EASMI. The correlation coefficient associated with the ISMI is 0.253, which did not pass the significance test even at the 90% confidence level, and the correlation coefficient related to the EASMI is 0.294, also not passing the significance test at the 95% confidence level. These results suggest that the individual ISM or EASM may not significantly regulate the rainy season precipitation in the MRB. Holmes et al. (2009) reported that the combination of ISM and EASM rather than individually is the fundamental driver of the Mekong hydroclimate. Therefore, we construct a synthetic summer monsoon index (SSMI) simply using the normalized ISMI plus the normalized EASMI to describe the ISM and EASM covariability. Notably, the correlation coefficient between the rainy season PI and SSMI reaches 0.435, which passes the significance test above the 99% confidence level and is much higher than the previous two correlation coefficients (Fig. 2b). The results, which also agree well with previous studies, indicate that there is a close relationship between the rainy season precipitation in the MRB and the ISM and EASM covariability. When the ISM and EASM are stronger (weaker) than normal, i.e., SSMI is higher (lower) than normal, the rainy season precipitation is more (less) than normal in the LMRB and mainly presents positive precipitation in the northern UMRB and negative precipitation in the southern UMRB.

Because the first EOF mode is larger in the lower basin but smaller in the upper basin, we divided the whole MRB into the UMRB and LMRB with the boundary near 21°N and performed the EOF analysis again. The EOF results associated with the LMRB indicate that the top three modes accounted for 34.8%, 13.3% and 9.1% of the rainy season precipitation variability in the LMRB, respectively. The correlation coefficients between the time series corresponding to the first EOF mode and ISMI, EASMI or SSMI are 0.257, 0.297, and 0.441, respectively. The first two correlation coefficients fail to pass the significance test at the 95% confidence level, but the third correlation coefficient passes the significance test above the 99% confidence level (Fig. 2c, d). However, the explained variance in the first EOF mode associated with the UMRB precipitation is 28.2%. The relatively small explained variance suggests that the variability in the UMRB precipitation is more complicated than that in the LMRB precipitation. The correlation coefficients between the time series corresponding to the first EOF mode and the ISMI, EASMI and SSMI are 0.005, 0.109, and 0.091, respectively (Fig. 2e, f). The special geographical position of the northern MRB with its narrow terrain, which is located on the southeastern Tibetan Plateau, may be a main reason for the weak correlation. Because the Tibetan Plateau is a huge thermal source during the summer, the rainy season precipitation in the MRB may relate

Fig. 2 The first EOF mode of the rainy season precipitation in the MRB (a), LMRB (c), UMRB (e), and their corresponding indices over the period of 1979–2016 (b, d, f). MRB, R, ISMI, EASMI and SSMI denote the Mekong River basin, correlation coefficient, Indian summer monsoon index, East Asian summer monsoon index, and synthetic summer monsoon index, respectively

to the dynamic and thermal effects of the Tibetan Plateau (Boos and Kuang 2010; Wu et al. 2012) in addition to the ISM and EASM covariability. This issue is worth future study.

3.1.2 Rainy season precipitation anomalies in the MRB

To reveal the physical process through which the rainy season precipitation anomalies in the MRB are related to the ISM and EASM covariability, we used the SSMI time series and a criterion of ± 0.8 standard deviation (Fig. 2b) to identify 10 (8) of 36 years as positive (negative) SSMI years. The 10 positive SSMI years are 1991, 1994, 1999, 2000, 2001, 2002, 2003, 2004, 2005, 2011, and 2012. The

8 negative SSMI years are 1983, 1987, 1988, 1989, 1998, 2010, 2014, and 2015. Figure 3a shows the precipitation difference in the MRB calculated with the positive SSMI years minus the negative SSMI years selected from the SSMI time series. As their differences pass the significance test above the 95% confidence level, the rainy season precipitation in the positive SSMI years is significantly heavier than that during the negative SSMI years over the LMRB and the area between 25°N and 30°N , with maximum precipitation anomalies exceeding 200 mm. In the rest of the MRB, there is negative anomaly precipitation, which cannot pass the significance test at the 90% confidence level. The rainy season precipitation differences between the positive and negative SSMI years are consistent with the first EOF mode

Fig. 3 The rainy season precipitation difference between the positive and negative SSMI years selected from the time series SSMI (a), between the positive and negative ISMI years with normal EASMI years selected from the time series ISMI and EASMI (b), and

between the positive and negative EASMI years with normal IMI selected from the time series ISMI and EASMI (c). The green line denotes the difference passing the significance test at the 95% confidence level

to a larger degree because the spatial correlation coefficient between them reaches 0.571 and passes the significance test above the 95% confidence level after adjusting the effective freedom to 19.

To further understand which part of the MRB is more likely to be affected by the ISM or EASM, we used a criterion of $-0.8 < \text{EASMI} < 0.8$ and $|\text{ISMI}| \geq 0.8$ standard deviation (Fig. 2b) to identify 6 (7) of the 36 years as positive (negative) ISMI years. The 6 positive ISMI years are 1999, 2000, 2003, 2004, 2007, and 2011. The 7 negative ISMI years are 1981, 1982, 1984, 1987, 1989, 1992, and 2014. Figure 3b shows that the precipitation difference passing the significance test above the 95% confidence level occurs mainly over the west MRB. If we focus on the MRB, only the regions of approximately 14° – 17°N , 21°N and 27°N pass the significance test above the 95% confidence level. The spatial correlation coefficient between the first EOF mode and anomalous rainy season precipitation pattern (-0.143) cannot pass the significance test at the 95% confidence level after adjusting the effective freedom to 89. This result indicates that the ISM mainly influences the rainy season precipitation in the west MRB, which agrees with previous studies (Xue et al. 2011; Delgado et al. 2012; Tsai et al. 2015).

We used another criterion of $-0.8 < \text{ISMI} < 0.8$ and $|\text{EASMI}| \geq 0.8$ standard deviation (Fig. 2b) to identify 5 (4) of the 36 years as positive (negative) EASMI years. The 5 positive EASMI years are 1990, 1994, 2001, 2002, and 2012. The 4 negative EASMI years are 1988, 1995, 1996, and 2010. Figure 3c shows the precipitation difference

passing the significance test above the 95% confidence level, which mainly occurs over the southeastern MRB. The spatial correlation coefficient between the first EOF mode and anomalous rainy season precipitation pattern (0.201) also fails to pass the significance test at the 95% confidence level after adjusting the effective freedom to 75. This result indicates that the EASM mainly regulates the rainy season precipitation in the southeastern MRB, which is also consistent with previous studies (Xue et al. 2011; Delgado et al. 2012; Tsai et al. 2015).

3.1.3 Anomalous circulation patterns

The precipitation anomalies always result from the associated anomalous circulation. Figure 4a shows the surface horizontal wind differences between the positive and negative SSMI years. An anomalous cyclonic circulation at the Earth's surface appears in the northern Bay of Bengal (BOB), and another anomalous cyclone appears in the northern South China Sea (SCS). At the south flank of the two anomalous cyclones, significant westerly anomalies run from the southern BOB to the southern SCS, whereas significant easterly anomalies are running from the northern SCS to the southern Tibetan Plateau. The distribution of the 850 hPa horizontal wind is similar to the anomalous pattern of the horizontal wind at the Earth's surface (Fig. 4b). The negative sea level pressure (Fig. 4c), agreeing well with the anomalous horizontal winds (Fig. 4a, b), controls from the northern BOB to the SCS. As the BOB–western MRB is usually dominated

Fig. 4 The differences in the 10 m horizontal winds (unit: m s^{-1}) (a), 850 hPa horizontal wind (unit: m s^{-1}) (b), sea level pressure (unit: hPa) (c), vertical wind profile along (10°N , 90°E) to (26°N , 110°E) (unit: m s^{-1}) (d), column-integrated water vapor flux (unit: $\text{kg m}^{-1} \text{s}^{-1}$) (e), and divergence of water vapor flux (contour inter-

val = $1 \times 10^{-4} \text{ kg m}^{-2} \text{s}^{-1}$) (f) between positive and negative SSMI years during the rainy season. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

by a southwesterly monsoon and the east MRB is usually dominated by a southeasterly monsoon during the rainy season (Fig. 1), the significant anomalous pattern of the horizontal winds in the lower troposphere suggests that

the stronger southwesterly monsoon during the rainy season will be limited in the MRB during the positive SSMI years. To further illustrate the circulation anomalies, we drew a latitude-longitude-pressure section of anomalous

winds (Fig. 4d). The anomalous southwesterly winds from the BOB and anomalous southeasterly winds from the SCS converge in the MRB and force out significantly stronger ascending motion around the same area during the positive SSMI years. The anomalies of the column-integrated water vapor flux (Fig. 4e) feature nearly the same anomalous pattern as the anomalous horizontal winds in the lower troposphere (Fig. 4a, b). As a result, significant westward

anomalies of the column-integrated water vapor flux occur from the BOB to the southern MRB, and significant eastward anomalies appear in the region from north SCS to the eastern MRB. This anomalous pattern will create a sandwich-like anomalous divergence in the column-integrated water vapor flux. Herein, the anomalous water vapor convergence occurs over most of the MRB, and the anomalous divergence regions appear in the southern regions of the

Fig. 5 The same as Fig. 4 but between positive and negative ISMI years with normal EASMI during the rainy season. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

BOB and SCS and 110° – 130° E along 30° N (Fig. 4f). The coordinated configuration of the atmospheric circulation indicates that the warm-wet air from the BOB and SCS will be significantly converged and lifted around the MRB and further benefit the heavier precipitation over most of the MRB during the rainy season. Because Fig. 4 was drawn according to the anomalous SSMI years, Fig. 4 also suggests the physical meaning of the SSMI, which directly

reflects the covariability of the Indian monsoon trough and East Asian monsoon trough.

Figure 5a, b show the horizontal wind differences between the positive and negative ISMI years with normal EASMI at the Earth's surface and 850 hPa. An anomalous anticyclonic center at the Earth's surface moves to west Myanmar, and another anomalous anticyclone occurs over the SCS. The anomalous significant easterly winds

Fig. 6 The same as Fig. 4 but between positive and negative EASMI years with normal ISMI during the rainy season. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

appear from the southern SCS to southern BOB, whereas anomalous easterly winds mainly occupy the northern BOB and northern SCS–South China. The positive sea level pressure that did not pass the significance test controls from the BOB to the SCS (Fig. 5c). The latitude-longitude-pressure section of anomalous winds (Fig. 5d) shows that the stronger ascending motion mainly appears at 96° – 100° E. The anomalies of the column-integrated water vapor flux (Fig. 5e) feature nearly the same anomalous pattern as the anomalous horizontal winds in the lower troposphere (Fig. 5a, b). As a result, an anomalous water vapor convergence occurs in the west MRB, and anomalous divergence regions mainly appear in the southeastern MRB (Fig. 5f). The anomalous atmospheric circulation pattern indicates that the warm-wet air from the BOB and SCS will significantly converge and lift around the west MRB and further benefit heavier precipitation over the same region during the rainy season. Figure 6 shows the atmospheric circulation differences between the positive and negative EASMI years with a normal EASMI. The anomalous atmospheric circulation

is similar to Fig. 4 except that the anomalous easterly winds from the northern SCS to the southern Tibetan Plateau become weak to some extent, the ascending motion around the MRB fails to pass the significance test at the 95% confidence level, and the position of the anomalous water vapor convergence is farther east than that in Fig. 4f. This anomalous atmospheric circulation pattern further favors heavier precipitation over the southeastern MRB during the rainy season.

3.1.4 Apparent heat source (Q_1) and apparent moisture sink (Q_2)

Figure 7a, b display the diabatic heating difference between the positive and negative SSMI years. The positive column-integrated Q_1 mainly occurs from northeastern India to northern SCS. Hereinto, the region passing the significance test is mainly located around the north-eastern India–western Indochina peninsula, with maximum anomalies exceeding 20 W/m^2 . The region from the southern BOB to the southern SCS during the rainy

Fig. 7 The differences in the column-integrated apparent heating source (a), column-integrated apparent moisture sink (b), surface sensible heat flux (c), and surface latent heat flux (d) between the posi-

tive and negative SSMI years during the rainy season (units: W m^{-2}). The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

Fig. 8 The same as Fig. 7 but between positive and negative ISMI years with normal EASMI during the rainy season. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

season is dominated by a significantly negative apparent heating source, with maximum anomalies below -10 W/m^2 in an absolute sense. The distribution of the anomalous column-integrated Q_2 is similar to the anomalous column-integrated Q_1 pattern but with an extended area passing the significance test at $110^\circ\text{--}130^\circ\text{E}$ and 30°N . These results imply that the latent heat release anomalies may make a crucial contribution to the column-integrated Q_1 . Notably, the couplet of column-integrated Q_1 and column-integrated Q_2 anomalies in the northern BOB (rectangle A) and SCS (rectangle B) is coherent with that in the MRB precipitation (Fig. 7a). In reality, the correlation coefficients between the ISMI and column-integrated Q_1 or column-integrated Q_2 averaged in rectangle A reach 0.822 and 0.774, passing the significance test above the 99% confidence level. Additionally, the correlation coefficients between the EAMI and column-integrated Q_1 or column-integrated Q_2 averaged in rectangle B reach 0.574 and 0.692, also passing the significance test above the 99% confidence level. Figure 7c, d display the surface

latent heat flux and surface sensible heat flux differences between the positive and negative SSMI years. The surface sensible heat flux and surface latent heat flux may also have a relatively weak contribution to the column-integrated Q_1 because their values are smaller than those values associated with the column-integrated Q_1 or column-integrated Q_2 . Figure 8a, b display the diabatic heating difference between the positive and negative ISMI years with a normal EASMI. The column-integrated Q_1 and column-integrated Q_2 resemble Fig. 7a, b over the region west of 100°E to a large degree, but those values show opposite signs compared to Fig. 7a, b over the northwestern SCS. The difference in latent heat flux between Figs. 7c and 8c mainly occurs over the SCS, but the surface sensible heat flux (Fig. 8d) shares a similar pattern to that shown in Fig. 7d. These four anomalous patterns are consistent with an anomalous ISM and normal EASM. Figure 9 displays the diabatic heating difference between the positive and negative EASMI years with a normal ISMI. The anomalous column-integrated Q_1 , anomalous column-integrated

Fig. 9 The same as Fig. 7 but between positive and negative EASMI years with normal ISMI during the rainy season. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

Q_2 , anomalous latent heat flux, and anomalous surface sensible heat flux share the similar pattern to Fig. 7a–d, except the anomalous intensity of each variable over the area west of 100°E becomes weak, and the area passing the significance test largely decreases. These four anomalous patterns agree with an anomalous EASM and normal ISM.

Further, we expose the vertical structure of diabatic heating associated with SSMI. Figure 10a shows the vertical profile of the Q_1 averaged at rectangle A. The normal vertical distribution (black line) shows the maximum heating centered in 400–600 hPa, and the vertical profiles in the positive and negative SSMI years are similar to the vertical structure seen in normal years. However, the Q_1 above 800 hPa during positive SSMI years (red line) is always larger than in negative SSMI years (blue line). Figure 10c shows the vertical profile of the Q_2 in rectangle (A). The drying center in the normal years appears between 600 and 800 hPa. This pronounced feature can be observed in positive and negative SSMI years, but their values are larger in the positive SSMI years than in the

negative SSMI years for each layer. Figure 10b displays the vertical profile of the Q_1 averaged at rectangle (B). The normal vertical distribution (black line) features maximum heating located in 400–500 hPa, and the vertical profiles in the positive and negative SSMI years resemble the structure in normal years. However, the Q_1 above 925 hPa in positive SSMI years (red line) is always stronger than in negative SSMI years (blue line). Figure 10d shows the vertical profile of Q_2 in rectangle B. The drying center in normal years appears between 700 and 900 hPa. This feature is also reflected in anomalous years. Their values are higher in positive SSMI years than in negative SSMI years for each layer. These results further indicate that the anomalous latent heating release is one of the most important contributors to the apparent heating source around rectangles A and B. These vertical profiles (Fig. 10) correspond well with the column-integrated Q_1 and column-integrated Q_2 (Fig. 7a, b).

The results obtained above suggest that a positive feedback of latent heat release plays a key role over the areas around the BOB and SCS. The negative anomalous latent

Fig. 10 The vertical profiles of the apparent heating source A and apparent moisture sink (units: $^{\circ}\text{C day}^{-1}$) averaged at rectangles A (a, b) and B (c, d) during the rainy season. The rectangles A and B are defined in Fig. 7a. Black, red and blue lines denote the normal years, positive SSMI years and negative SSMI years. The circle denotes the corresponding difference between the positive SSMI years and negative SSMI years passing significance test above the 95% confidence level

heating in the southern BOB and positive anomalous latent heating in the northeastern India–Indochina peninsula (Fig. 7a, b) will force anomalous westerly winds from the southern BOB to MRB at the middle-low tropospheric levels (Fig. 4a, b). These anomalous winds are superimposed on the mean westerly winds from the BOB to MRB (Fig. 1) and make the westerly winds accelerate east of the heating center over the BOB and decelerate to the west, which further increases the evaporation over the southern BOB. Meanwhile, the negative anomalous latent heating over the southern SCS and positive anomalous latent heating over the northern SCS will force significant easterlies from the Philippines to the MRB at middle-low tropospheric levels (Fig. 4a, b). These anomalous winds are superimposed on the mean easterly winds from the Philippine Sea to the MRB (Fig. 1) and make the winds accelerate west of the heating center over the Philippine Sea and decelerate to the east, which further increases the evaporation over the SCS and decreases the evaporation over the Philippine Sea. In turn, the evaporation anomalies in the BOB and SCS will influence the latent heating anomaly, which favors greater-than-normal precipitation in the rainy season of MRB.

3.2 Model results

The DLBM was run to examine the reliability of the key physical process linking the Q_1 and Q_2 variability with the MRB precipitation anomaly during the rainy season. According to the time series of the SSMI (red line in Fig. 2b), the apparent heat sources and moisture sink in the 10 positive and 8 negative SSMI years serve as the forcing in the rectangles for the DLBM. Figure 11 shows the vertical profiles of the thermal forcing in positive (solid line) and negative (dashed line) SSMI years in the sigma coordinate system. The area is occupied by the thermal forcing (Fig. 12) resembling the observation (Fig. 7a). To obtain the atmospheric circulation anomalies responding to the thermal forcing, a time integration method was adopted. After the DLBM is integrated for 11 days, the response of the atmospheric circulation tends to approach a steady state. Therefore, the differences between positive and negative SSMI years on day 15 are analyzed as follows. Figure 12d shows the modeling $\sigma = 0.90$ horizontal wind differences between positive and negative SSMI years after superimposing the thermal forcing associated with the Q_1 on rectangle (A).

Fig. 11 The vertical profiles of the apparent heating source (units: $^{\circ}\text{C day}^{-1}$) at anomalous years averaged in rectangle A (a) and rectangle B (b) and the apparent moisture sink in rectangle A (c) and rectangle

B (d). The color solid (dashed) lines denote the vertical profiles during the positive (negative) SSMI years. The thicker black solid line denotes their climatological normal values

As the intensities of the Q_1 in the positive SSMI years are generally stronger than those in negative SSMI years in the BOB, the anomalous Q_1 thermal forcing will result in an anomalous cyclone over the BOB. The anomalous winds at the southwest flank of the anomalous cyclone are superimposed on the mean southwesterly winds and further enhance the mean southwesterly winds from the BOB to MRB. However, the circulation anomalies responding to the thermal forcing are weaker in the EASM region in comparison to the observation (Fig. 4b). Figure 12e is the same as Fig. 12d, but the thermal forcing associated with the apparent heat source is superimposed on rectangle

(B). As the intensities of the Q_1 in positive SSMI years are generally stronger than those in negative SSMI years Q_1 in the SCS, the anomalous thermal forcing will result in an anomalous cyclone over the SCS. However, the circulation anomalies responding to the thermal forcing are somewhat weak in the ISM region in comparison with the observations (Fig. 4b). Figure 12f is the same as Fig. 12d, e but with the thermal forcing associated with the superimposed on rectangles A and B. Compared with Fig. 8d, e, the anomalous circulation (Fig. 12f) best resembles the observation (Fig. 4b). The simulated $\sigma = 0.83$ horizontal wind differences (Fig. 12g, h, i) are similar to

Fig. 12 Horizontal distribution ($\sigma = 0.55$) of the apparent heat source (**a**, **d**, **g**). The contour interval is 0.1 K d^{-1} in **a**–**c**. **d**–**f** The horizontal wind anomalies (units: m s^{-1}) forced by the heating source of the positive SSMI years minus those forced by the heating source of the negative SSMI years over rectangles A and B separately, and both A

and B at $\sigma = 0.90$. Panels (**g**, **h**, **i**) are the same as panels (**d**–**f**) but at $\sigma = 0.83$. The areas shaded from light to dark denote anomalies passing the significance test at the 95% and 99% confidence levels, respectively

those at $\sigma = 0.90$ between the positive and negative SSMI years. Figure 13a shows the simulated wind differences between the positive and negative SSMI years at a latitude-longitude-sigma section after superimposing the thermal forcing associated with the Q_1 on rectangle (A). The ascending motion anomalies occupy the area west of the MRB, but easterly anomalies at the middle-low troposphere are not observed in the SCS. Figure 13b is the same as Fig. 13a but with the thermal forcing associated with the Q_1 superimposed on rectangle (B). Obviously, the easterly anomalies occur in the SCS, but the westerly anomalies are very weak in the BOB. When the thermal

forcing associated with Q_1 is superimposed on rectangles A and B (Fig. 13c), the simulated results are most similar to the observed anomalies (Fig. 4d).

Figures 14 and 15 are the same as Figs. 12 and 13 but with the thermal forcing associated with the Q_2 superimposed separately on rectangles A and B, as well as on both rectangles together. All simulated circulation anomalies along the latitude-longitude-sigma section share similar patterns related to Q_1 at $\sigma = 0.83$ and $\sigma = 0.90$. Because the thermal forcing centers associated with Q_2 generally appear between $\sigma = 0.7$ and $\sigma = 0.8$ (Fig. 11c, d), the amplitudes of the atmospheric response to the thermal

Fig. 13 a–c The vertical wind profile along (10°N, 90°E) to (26°N, 110°E) forced by the heating source of the positive SSMI years minus those forced by the heating source of the negative SSMI years over rectangles A and B separately, and both A and B, respectively. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

forcing associated with Q_2 are usually stronger than those associated with Q_1 in the lower troposphere, especially when the thermal forcing associated with Q_2 is superimposed on rectangles A and B. These modeling results suggest that the latent heat release configuration in the BOB and SCS is one of the key physical processes through which the ISM and EASM covariability impacts the rainy season precipitation in the MRB.

4 Summary

After defining the SSMI, which can be used to describe the covariability of the ISM trough and EASM trough, a significant relationship was demonstrated between the rainy season precipitation in the MRB and the covariability of the two summer monsoon systems. The correlation between the SSMI and time series corresponding to the first EOF mode of the rainy season precipitation in the MRB achieved 0.458 for the 1981 to 2016 period, which passed the significance test above the 99% confidence level. In the configuration of the ISM and EASM, the intensities of the ISM and EASM are positively correlated to the rainy season precipitation in the MRB. Herein, the individual ISM mainly modulates the rainy season precipitation west of the MRB, and the individual EASM mainly regulates the rainy season precipitation over the southeastern MRB.

The physical processes linking the ISM and EASM covariability to the rainy season precipitation anomalies in the MRB were further studied using a composite analysis. The corresponding results demonstrate that for a positive SSMI, i.e., ISM and EASM are relatively strong (weak), positive (negative) diabatic heating anomalies in the north-eastern India–northwestern Philippine Islands (southern BOB–southwestern Philippine Islands) are built up. These thermal anomalies induce the development of two anomalous cyclones in the northern BOB and northern SCS, accompanied by the appearance of westerly anomalies over the northern BOB and easterly anomalies over the northern SCS. The anomalous zonal winds transport more water vapor from the BOB and SCS in the MRB. The anomalous convergence zone associated with these anomalous zonal winds is located precisely in the MRB and leads to

Fig. 14 The same as Fig. 12 but associated with the apparent moisture sink. **a–c** The modeling results at $\sigma = 0.90$. **d–f** The modeling results at $\sigma = 0.83$. The areas shaded from light to dark denote anomalies passing the significance test at the 95% and 99% confidence levels, respectively

greater-than-normal rainy season precipitation around the same region. For a negative SSMI during the rainy season, nearly the opposite conditions occur. The counterpart conditions further lead to less-than-normal rainy season precipitation around the MRB.

The key physical process mentioned above is substantiated by the results of numerical experiments obtained from the LBM. The modeling results can also reveal the positive correlation between the rainy season precipitation in the MRB and configuration of the ISM and EASM through a positive feedback mechanism behind the key physical

processes. This mechanism is associated with the evaporation-wind feedback, which may explain the maintenance of the anomalous thermal conditions over the BOB and SCS (e.g., Neelin et al. 1987; Emanuel et al. 1994). The latent heating anomalies accelerate the westerly winds east of the heating center over the BOB, and the easterly winds west of the heating center over the SCS, further increasing the evaporation over the two regions, and in turn influencing the latent heating anomaly. Accompanied by this positive feedback, greater-than-normal rainy season precipitation in the MRB is finally observed.

Fig. 15 The same as Fig. 13 but associated with the apparent moisture sink. The areas shaded from light to dark denote the difference passing the significance test at the 95% and 99% confidence levels, respectively

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