Coupling Modes of Climatological Intraseasonal Oscillation in the East Asian Summer Monsoon

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ABSTRACT

The present study used harmonic and multivariate empirical orthogonal function (MV-EOF) analyses to identify the existence of climatological intraseasonal oscillation (CISO) in the diabatic heating, precipitation, and circulation of the East Asian summer monsoon (EASM). The strongest CISO signals are found in the north of the western North Pacific, possibly because of the horizontal gradient of diabatic heating induced by the seasonal land–sea thermal contrast. Further, the phase relationship between the diabatic heating components maintains the EASM CISO. The first two coupling modes of EASM CISO in the circulation are robust during May through August, with a period of 40–80 days, and exhibit phase locking to the stepwise establishment of the EASM, which reveals the coaction of the Mongolian cyclone (MC) around Lake Baikal at 850 hPa, the western North Pacific subtropical high (WNPSH) at 500 hPa, and the South Asian high (SAH) over the Tibetan Plateau (TP) at 200 hPa. The first mode shows that the jointly enhanced MC, WNPSH, and SAH correspond to a tripole rainfall anomaly with strong mei-yu and baiu fronts over East Asia. The second mode, however, indicates the eastward and northwestward propagation of MC and WNPSH, respectively, with suppressed SAH, as well as a dipole rainfall anomaly over East Asia. Both the observations and numerical simulation verify the importance of daily diabatic heating and SST in maintaining the CISO modes over the WNP, where the condensation heating related to atmospheric forcing determines the local intraseasonal air–sea interaction.

1. Introduction

The East Asian summer monsoon (EASM) includes the tropical South China Sea (SCS) and the western North Pacific (WNP), along with the subtropical summer monsoon north of 20°N over the East Asia–Pacific region (Lau and Li 1984; Zhu et al. 1986; Tao and Chen 1987; Wang and LinHo 2002; He et al. 2007; Zhu et al. 2011; Wang et al. 2012; Chi et al. 2015). The rain belts of the EASM with stronger low-level southwesterly winds span from the subtropical WNP to the higher latitudes of eastern Russia and become the most northern component in the global monsoon system (Wang et al. 2012). A number of studies have suggested that the WNP subtropical high (WNPSH) in the mid-to-lower troposphere (Figs. 1a,b) and the upper-level South Asian high (SAH) over the southern Tibetan Plateau (Fig. 1c) are the major circulations of the EASM to regulate its rainfall. During the boreal summer, the enhanced WNPSH transports more warm moisture by strengthening the southwesterly winds, thus resulting in heavier summer rainfall in the mid-to-lower reaches of the Yangtze River in China (i.e., mei-yu fronts), in conjunction with the stronger ascent induced by the strengthened SAH (Tao and Zhu 1964; Ren et al. 2007; Lu et al. 2014). Stronger mei-yu fronts are also closely associated with enhanced rainstorm frequency and intensity due to the cold and warm airmass interaction between the tropical and mid-to-high latitudes (Tao 1980; Lau and Li 1984; Chen et al. 2001; Zhu et al. 2003; Lu et al. 2014), which is associated with the Mongolian cyclone (MC) and cold vortex over northern China (Li and Wang 2003; Miao et al. 2006a,b; He et al. 2007).

In climatology, the EASM circulation and rainfall begin in early April, peak in August, and prevail over the East Asia–Pacific region until finally withdrawing in October (Zhu et al. 2005; Zhao et al. 2007; He et al. 2007;
The EASM rain belts exhibit a stepwise and subsequent northward jump with the seasonal migration of the EASM (Fig. 1d), thus resulting in the peak heavy rainfall in southern China, the Yangtze River, and northern China in early April, mid-June, and late July, respectively (He et al. 2008; Tian and Yasunari 1998; Zhu et al. 2011). The timing of each enhanced rainfall fluctuation indicates the subtropical EASM, SCS, and tropical WNP monsoon onset, and the rapid northward shift of the rainband marking the start of the summer rainy season in southern China, mei-yu front, and northern China, respectively (e.g., Lau et al. 1988; Zhu et al. 2011).

The seasonal progression of the EASM relative to its annual cycle exhibits subseasonal variation, which manifests as seasonal fluctuations of the EASM rainbands during May–August (Fig. 1d). This subseasonal variation of the EASM strongly modulates the waxing and waning of summer monsoon rainfall (e.g., Lau and Li 1984; Chen and Jin 1984; Lau et al. 1988; Wang and Rui 1990; Wang and Xu 1997; He et al. 2015; Chi et al. 2015). Wang and Xu (1997), by using climatological pentad mean outgoing longwave radiation (OLR) and the ECMWF winds have revealed that the subseasonal variation in the Asian summer monsoon is directly associated with the activity of the climatological intraseasonal oscillation (CISO). They focused on the time series of OLR at 15°N, 140°E and have confirmed the statistical significance of the CISO signal relative to the annual cycle in the WNP domain. Their results suggest that the summer monsoon CISO results from phase locking between the transient ISO and the annual cycle, which presents a dynamically coherent structure between enhanced convection and lower-level convergent (upper-level divergent) cyclonic (anticyclonic) circulation. The extreme phases of the CISO characterize monsoon singularities (i.e., monsoon events occurring on a fixed pentad with usual regularity; Wang and Xu 1997), thus implying the potential ability of CISO to predict the subseasonal variation of the EASM.

Although the CISO of the EASM has been observed since the 1990s, its existence and detailed structure remain uncertain. Because the general circulation, particularly in the tropics, is driven by atmospheric diabatic heating (Yanai et al. 1976; Hantel and Baader 1978), which receives feedback from the circulation through atmospheric instabilities (Lau and Peng 1987; Emanuel...
et al. 1994; Liu et al. 2004; Hazra and Krishnamurthy 2015), it has been argued that the annual and diurnal variations of radiative forcing with shortwave radiative heating and longwave radiative cooling produce the most significant climatological modes of the atmospheric circulation at the same time scale. However, it remains unclear whether the shortwave radiative heating and the longwave cooling intrinsically have an intraseasonal component. If so, can the CISO be treated as the atmospheric response to the intraseasonal variation of radiative heating and cooling? Moreover, because the EASM-related circulation exhibits both internal vertical and meridional coupling, with evident seasonal land–sea thermal contrast over the EASM, it is interesting to further investigate the coupling modes of the EASM CISO, their effects on the subseasonal variations of summer rainfall, and the possible impact from local intraseasonal air–sea interaction.

The present study will first discuss the existence of CISO in the EASM by investigating the relationship among atmosphere diabatic heating, precipitation, and large-scale monsoon circulation over the East Asia–Pacific region. Subsequently, the coupling modes of EASM CISO and their possible impacts on the subseasonal rainfall activity will be further described. In addition, we will explore the possible regulation of air–sea interaction by the EASM CISO. The paper is organized as follows. The data and methods used are described in section 2. The existence of the CISO EASM modes is presented in section 3. Section 4 demonstrates the coupling modes of the EASM CISO in circulation, as revealed by the multivariate empirical orthogonal function (MV-EOF) analysis, and their impacts on the activity of subseasonal rainfall. The role of subseasonal air–sea interaction over the WNP in maintaining the EASM CISO is explored in section 5. Finally, a summary and discussion are provided in section 6.

### 2. Data and methods

The data utilized in this study include the atmospheric components for the period 1981 to 2010 from the National Centers for Environmental Prediction–U.S. Department of Energy (NCEP–DOE) AMIP-II reanalysis (R-2; Kanamitsu et al. 2002), along with the pentad-averaged precipitation from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997), with a horizontal resolution of 2.5° × 2.5°. The daily NOAA high-resolution SST data (Reynolds et al. 2007) were provided by the NOAA/OAR/ESRL Physical Sciences Division (PSD), Boulder, Colorado, from their website (http://www.esrl.noaa.gov/psd/).

The vertically integrated atmospheric heat source $Q_1$ (Hantel and Baader 1978; Yanai et al. 1992; Yanai and Tomita 1998; Liu et al. 2004) was calculated using the R-2 data as follows:

$$Q_1 = SH + CO + RC,$$

where $SH$, $CO$, and $RC$ represent the surface sensible heating, condensation heating released by convection, and net radiative heating of the air column, respectively. Here, $SH$ was extracted from the NCEP–DOE reanalysis, and $CO$ and $RC$ were calculated using

\[ CO = Pr \times Lw \times \rho \]  
\[ RC = R_w - R_0 = (S_w^i - S_w^o) - (S_0^i - S_0^o) - (F_0^i - F_0^o) - F_c. \]

In Eq. (2), $Pr$, $Lw$, and $\rho$ are the precipitation amount, condensation heat coefficient, and water density, respectively. In Eq. (3), $R_w$ and $R_0$ are the net radiation values at the top of atmosphere (TOA) and the land surface, respectively. The variables $S$ and $F$ denote the shortwave (SW) and longwave (LW) radiation fluxes, their subscripts $\infty$ and 0 denote the TOA and the ground surface, and the superscripts $\downarrow$ and $\uparrow$ represent downward and upward transport, respectively.

The climatological data was defined as the 30-yr arithmetic average value from 1981 to 2010, and the CISO was defined as the 20–90-day harmonic components relative to the annual cycle. Following previous studies (Wang and Xu 1997; Wheeler and Hendon 2004), we applied harmonic analysis to extract the CISO component for each grid as follows:

\[ A(t) = \bar{A} + SVC(t) + ISVC(t) + R(t), \]
\[ SVC(t) = \sum_{m=1}^{3} A_m \exp \left( i \frac{2\pi m}{T_0} t \right), \]  
\[ ISVC(t) = \sum_{m=4}^{18} A_m \exp \left( i \frac{2\pi m}{T_0} t \right), \]

where $A(t)$ and $\bar{A}$ indicate the climatological daily mean and annual average for each element, respectively. The seasonal variation component $SVC(t)$ was defined as the sum of the first three harmonics ($m = 1, \ldots, 3$) and corresponds to the annual and semiannual components. The term $ISVC(t)$ represents the sum of harmonics 4 through 18 and has a period of 20–90 days. The residual in the harmonic analysis $R(t)$ was for the transient components with periods less than 20 days.

Here, we focused on the CISO in the Asia–Pacific region of 0°–60°N, 60°–160°E during boreal summer (April–October). To identify the dominant modes of
vertical and meridional circulation interaction in the EASM, we used MV-EOF analysis (Wang 1992; Wang et al. 2008; He et al. 2015) on the wind fields at various levels (i.e., 850, 500, and 200 hPa). In addition, non-dimensional eigenvectors are used to standardize the combined meteorological fields.

3. Existence of CISO in EASM

a. CISO in diabatic heating

The annual modes of $Q_1$ indicated by the first harmonic component can be considered to be the atmospheric response to the annual solar radiation forcing. However, the phase peak of the annual cycle (AC) of $Q_1$ is different from region to region, even at the same latitude, and it is very distinct between the land and ocean, owing to their different thermal capacities (Fig. 2a). For instance, to the east of 105°E in the East Asia–Pacific region, $Q_1$ AC over the continent reaches its peak in April and May at approximately 25° and 40°N in southern and northern China, respectively. The peak maximum of the $Q_1$ AC to the south of 30°N over Asia generally appears before July, but over the Indian and Pacific Oceans, it occurs after July in August, with a 1–2-month lag. The positive amplitudes of the $Q_1$ AC are centered to the south of 20°N, with three branches over the Indian subcontinent, where it adjoins oceanic regions, Indo-China, and the eastern Philippine Sea (Fig. 2a). During May–August, the spatial pattern of the climatological $Q_1$ is similar to that of the $Q_1$ AC (Figs. 2a,b). Specifically, to the west of 105°E, the maximum standard deviation (STD) of CISO, located over the tropical Indian Ocean, is accompanied by the climatological $Q_1$ center, whereas in the western Pacific, the CISO STD is centered over the SCS and the WNP regions, without the maximum of the climatological $Q_1$. 

FIG. 2. (a) The first harmonic-indicated distribution of the climatological annual cycle mode of diabatic heating, represented by its phase (vector) and amplitude (shading; W m$^{-2}$); (b) distribution of averaged $Q_1$ during May–August (contour; W m$^{-2}$) and the standard deviation (shading; W m$^{-2}$) of CISO; (c) seasonal variation of the $Q_1$ AC between 115° (shading) and 140°E (contour); and (d) the difference of the AC (shading) and CISO (contour) in terms of $Q_1$ between 115° and 140°E (115°E minus 140°E). The phase of the cycle in (a) is shown as a 12-month clock, with a northward arrow indicating maximum heating in July. The arrows rotate clockwise with eastward, southward, and westward arrows indicating October, January, and April, respectively, and the arrow length defines the amplitude of the cycle. The contour interval of the $Q_1$ AC is 50 W m$^{-2}$, and the positive (solid) and negative (dashed) green lines of the CISO interval in (d) is 30 W m$^{-2}$, where the amplitude below 30 W m$^{-2}$ is omitted.
Moreover, the amplitude of CISO depends on the horizontal $Q_1$ gradient during the seasonal match of the EASM. Figure 2c shows that the seasonal evolution of the $Q_1$ AC between 115°E and 140°E is asymmetric in the meridional direction. The maximum of the $Q_1$ AC at 115°E is located to the north of 20°N during March–August, but it appears over the tropical ocean at 140°E to the south of 20°N in May–December (Fig. 2c). Thus, the negative meridional gradient of the $Q_1$ AC ($\partial Q_1/\partial y < 0$) becomes evident in May–August over East Asia and the ocean to its east. Meanwhile, the negative zonal gradient of the $Q_1$ AC ($\partial Q_1/\partial x > 0$) exists over the East Asia–Pacific region north of 20°N from March to August, whereas the negative counterparts appear over the tropical western Pacific latitudes during all seasons (Fig. 2d). Therefore, in May–August, a negative vorticity source is formed to the north and east of the $Q_1$ AC maximum, giving rise to the anticyclonic vorticity in the air column over East Asia and the surrounding ocean to the north of 20°N (Liu et al. 2001). Subsequently, the air column contracts to enhance the local ascending and the $Q_1$ AC, as required by the geostrophic relationship and the atmospheric continuous equation. Figures 3c,d also indicate that the CISO in $Q_1$ is active in May–August over East Asia, with enhanced ascending and $Q_1$ AC, where the CISO STD is centered (Fig. 2b). This result suggests that the existence of CISO depends partly on the distinct seasonal evolution of the $Q_1$ AC between East Asia and its surrounding oceans due to the land–sea thermal contrast and that the CISO is closely associated with the vertical motion over East Asia.

In addition to the land–sea thermal contrast, the CISO in $Q_1$ of the EASM is maintained by the phase relationship between the $Q_1$ components. We applied a direct method to estimate each component separately (Newell et al. 1969, 1975). Our results suggest that the subseasonal variation of $Q_1$ primarily derives from the AC-removed residuals of CO heating during May–October. The large STDs of $Q_1$ and CO are equivalent over the WNP (20°–40°N, 115°–140°E) from May to August (figures not shown), but the contributions of SH, SW, and LW are less than 10% of the total. Similarly, we decomposed each component of $Q_1$ into AC, CISO, and the residual terms by applying Eqs. (4)–(6). As shown in Fig. 3a, the AC of SW heating and LW cooling are evident with their peaks occurring in mid-June and early August, respectively. The maximum of SH occurs in the boreal winter, whereas the CO heating reaches its maximum in August. There is a lead–lag relationship between the AC of the $Q_1$ components. The AC-removed variation of CO exhibits a larger fluctuation than do those of SW and LW, particularly during May–October (Fig. 3b). The subseasonal variation of SW and LW follows that of SH and CO, which is quite different from the case in AC mode (Fig. 3a). Specifically, the SH varies initially, followed by the CO variation;
subsequently, changes in SW and LW occur over the EASM region at a subseasonal time scale (Fig. 3c). Such a lead–lag relationship is better organized in the correlations of the CISO (Fig. 3d). In physics, the increase in SH is able to enhance the atmospheric convective instability, and the monsoon convection releases CO heating, thus resulting in increased cloud, which in turn modulates the SW and LW in situ. As a result, the subseasonal variation of SW and LW lags behind the CO heating, following the SH oscillation. Wavelet analysis of AC-removed $Q_1$ and its components shows a significant 40–80-day oscillation during May–October (figures not shown). This periodicity is prominent in the SW, LW, and CO, but it is weak in the SH. The 40–80-day oscillation is consistent with the dominant periodicity of the CISO components, particularly the CO heating; this finding implies that the significance of CISO in $Q_1$ can be mainly ascribed to the CO heating over the WNP and EASM region during the boreal summer. As a result, the evident CISO in $Q_1$ is maintained by the lead–lag phase relationship between the $Q_1$ components, in which the feedback between SH and CO variations is dominant. In addition, the variations of SW heating and LW cooling are possible responses to the CO heating, instead of the origin of CISO in $Q_1$.

b. CISO of EASM rainfall and circulation

The CISO over the EASM region not only exists in $Q_1$ but also is embedded in the monsoon rainfall and circulation. The summer rainfall related to the mei-yu front in the mid-to-low reaches of the Yangtze River indicates the remarkable activity of the EASM, where the rainfall at 30°N, 115°E exhibits an outstanding subseasonal fluctuation relative to its AC mode (Fig. 4a). The time series of the AC-removed residual exhibits an intra-seasonal oscillation that coincides with its CISO component during May–August; its positive and negative phases indicate the activity and cessation of EASM rainfall, respectively (Fig. 4b). A similar feature is also presented in the subseasonal variation of rainfall at

![Fig. 4.](http://journals.ametsoc.org/jcli/article-pdf/29/17/6363/4071590/jcli-d-15-0794_1.pdf)
40°N, though with a smaller amplitude (Fig. 4c). To verify the existence of CISO in the EASM circulation, we adopted an EASM index (EASMI), which is defined on the basis of land–sea thermal contrasts by the composite of the normalized standard difference of sea level pressure (SLP; SLP at 160°E minus SLP at 110°E) and the vertical zonal wind shear ($U_{850hPa} - U_{200hPa}$) in the region of 0°–10°N, 100°–130°E (Zhu et al. 2005). The positive and negative phases of the AC in EASMI indicate the seasonal alternation between the summer and winter monsoon and the reversal of the land–sea thermal contrast over the East Asia–Pacific region. The seasonal variation of EASMI presents a stepwise enhancement with subseasonal fluctuation relative to its AC component (Fig. 4d). The EASMI without AC exhibits an intraseasonal oscillation with a significant 40–80–day period during May–August, which is consistent with its CISO component (Figs. 4e,f). Consequently, the CISO does exist in the EASMI and manifests as the residual between the daily climatology and its AC, with a significant 40–80–day period from May to August. As a result, the CISO of the EASM is closely associated with the subseasonal variation of CO heating ascribed to the zonal and meridional land–sea thermal contrast in the boreal summer monsoon season, and it is maintained via a lead–lag relationship between the $Q_1$ components.

4. Coupling modes of EASM CISO and its impact on rainfall

a. Coupling CISO modes in EASM

The significant ISO in the diabatic heating, rainfall, and EASMI implies the possible existence of CISO modes in the vertical circulation over the East Asia–Pacific region. As shown in Fig. 1, the EASM rainfall is closely associated with the vertical coupling of circulation at different levels (i.e., the coupling modes). Because the CISO is significant in EASM rainfall, it is speculated that the coupling modes should also play an important role in the EASM CISO. To reveal the coupling modes of CISO in EASM circulation, we applied the MV-EOF analysis to the standard normalized components of CISO in the wind fields at 850, 500, and 200 hPa from April to October. Figure 5 shows the first two leading modes of the MV-EOFs, which account for 17.1% and 14.6% of the total variance, respectively. These two modes are statistically distinguished by the criterion of North et al. (1982). The first MV-EOF mode shows a stronger MC around Lake Baikal at 850 hPa, an enhanced WNPSH over the northwestern Pacific at 500 hPa, and a strengthened SAH over the Tibetan Plateau (TP) at 200 hPa (Figs. 5a, 5c, and 5e, respectively). This pattern represents a cooperatively enhanced MC, WNPSH, and SAH. The MC and WNPSH are highly pronounced, with well-organized barotropic structures. The second MV-EOF mode, however, is featured by an enhanced MC at 850 hPa to the east of Lake Baikal and a northwestward shift of WNPSH at 850 and 500 hPa, showing an eastward shift of MC over northeastern China in the vertical levels and an inland northwestward propagation of WNPSH over the WNP (Figs. 5b and 5d, respectively). In contrast, the weakened SAH manifests as a southward cyclone over the TP at 200 hPa (Fig. 5f). Thus, the first mode of the EASM’s CISO is characterized by the simultaneous enhancement of the MC, WNPSH, and SAH, whereas the second mode shows an eastward and northwestward shift of the MC and WNPSH accompanied by the weakening SAH. Therefore, the first two modes compose the coupling mode of CISO in the EASM circulation at the intraseasonal time scale in boreal summer, which features the interaction among MC, WNPSH, and SAH.

Figure 6a presents the phase relationship between the first two principal components (PCs) of CISO in the EASM. PC1 and PC2 exhibit an ISO with an alternative appearance in terms of the minimum and maximum phases. In the boreal summer, the maximum PCs occur in June, when the mei-yu reaches its peak phase, whereas the minimum PCs occur in late July, corresponding to the rainfall season over northeastern China. Wavelet analysis suggests that PC1 and PC2 exhibit a significant 40–80–day period in boreal summer (figures not shown), where the lag correlation between PC1 and PC2 indicates that PC1 leads PC2 by approximately 20 days (Fig. 6b). The phase difference between PC1 and PC2 suggests that the CISO state can be depicted in a two-dimensional phase space by using the two PCs (Fig. 6c). This pattern shows that the sequential days move clockwise around the origin and that the CISO becomes active with a large amplitude from June to August, thus indicating CISO’s critical role in modulating the EASM’s intraseasonal variability.

Furthermore, three circulation indices are defined according to the vorticity of the 850-hPa MC (40°–50°N, 100°–120°E), 500-hPa WNPSH (25°–37.5°N, 120°–140°E), and 200-hPa SAH (25°–37.5°N, 75°–100°E) to examine their association with the coupling mode of EASM CISO. Figure 7 shows the normalized time series of the seasonal cycle, annual cycle, and CISO of the MC index (MCI), WNPSH index (WNPSHI), and SAH index (SAHI), respectively. In daily climatological and AC modes, the MCI first becomes positive in April, followed by the SAHI in early May and the WNPSHI in June, during the seasonal transition from boreal winter...
to summer (Figs. 7a,b). These three indices subsequently reach their peak in July and weaken and become negative in September, November, and December, respectively. Therefore, the evolution of the critical circulations in the EASM exhibits a stepwise seasonal development, which is consistent with the subsequent onset of the subtropical, SCS, and WNP summer monsoons of the EASM (Zhu et al. 2011). Moreover, the withdrawal of the EASM is related to the phase change in the annual cycle of these three circulation indices. The
CISOs of the MCI, WNPSHI, and SAHI are evident and present large amplitudes from April to August that alternately reach their peak during the boreal summer (Fig. 7c). This feature persists until late September. Moreover, there are phase lags between the circulation indices during the EASM CSIO cycle (Figs. 7d,e). The maximum of the SAHI leads PC1 by 2 days, whereas the counterparts of the MCI and WNPSHI lag PC1 by 0 and 15 days, respectively (Fig. 7d). Meanwhile, the MCI leads PC2 by 10 days, whereas the SAHI exhibits a significant relationship with PC2 (Fig. 7e). Thus, the intraseasonal variation of the MCI, WNPSHI, and SAHI indicates that the MC, WNPSH, and SAH are the three critical circulations in the dynamical wheel of the EASM and that their various phases determine the coupling modes of the EASM’s CISO.

b. Impacts on rainfall

The EASM rainfall is prominent between 100°E and 120°E over East Asia (Fig. 1d), where the CISO of rainfall is remarkable in the latitudinal belt of 20°–35°N (Figs. 4b,c). The influences of the coupling modes in the EASM CISO on rainfall are identified by the winds and rainfall regressed against the first two PCs (Fig. 8) and by their correlation with the three circulation indices (Fig. 9). For the positive PC1, stronger southwesterly winds are located over southern China in the lower troposphere, as are the strengthened WNPSH and an anticyclone over the Okhotsk Sea in far-eastern Russia (Fig. 8a). The PC1-regressed circulation in the upper troposphere presents a 200-hPa anticyclone centered over the TP (Fig. 8c). Meanwhile, the PC1-regressed rainfall is enhanced along the Yangtze River and is suppressed over the northeastern China and WNP regions (Fig. 8a), consistently with the synthesis between WNPSHI- and SAHI-associated precipitation (Figs. 9c,d). The PC1-regressed rainfall mode is similar to that associated with the Pacific–Japan (P-J) or East Asia–Pacific (EAP) pattern (Nitta 1987; Huang and Li 1988), thus showing the robust coupling between the WNPSH and SAH in the EASM’s CISO.

In contrast, the PC2-regressed winds and rainfall anomalies exhibit a dipole mode that is characterized by the weakening of the MC and WNPSH with enhanced and suppressed rainfall to the north and the south of the Yangtze River, respectively (Fig. 8b). However, the SAH in the upper troposphere is weakened in the PC2-regressed wind anomalies at 200 hPa (Fig. 8d). The
rainfall and upper-level circulation regressed against PC2 is opposite to the counterpart regressed against the SAHI (Fig. 9d), indicating the decoupling of the SAH to the MC and WNPSH in the mid-to-lower troposphere. The circulation regressed against the PC2 shows the comprehensive effect of MC and WNPSH on the CISO in terms of EASM rainfall. Thus, the CISO of tripole precipitation over the EASM region with its center over the mid-to-lower reaches of the Yangtze River can be explained by the joint impacts of the MC, WNPSH, and SAH at intraseasonal time scale. In particular, the CISO of SAH is closely related to a tripole mode of rainfall over the WNP and the area to its north, implicating the importance of SAH in the first coupling mode of EASM CISO. In addition, the second mode of EASM CISO featured by the coupling between the MC and WNPSH results in a dipole mode of rainfall anomalies (Figs. 9b,c). We further conducted EOF analysis on the CISO rainfall during April–October to verify whether the PC-regressed rainfall would resemble the primary characteristics of monsoon precipitation. Notably, the PC1- and PC2-regressed rainfall anomalies resemble the first two leading EOF modes for rainfall (shading in Figs. 8c,d), and the correlation coefficients between the first two PCs of circulation and their rainfall counterparts are +0.67 and +0.66, respectively. The PC1 and PC2 of rainfall anomalies also exhibit a 40–80-day period, and PC2 lags behind PC1 by 15 days, with a maximum correlation of +0.6. Therefore, the first two leading EOF modes of CISO of EASM rainfall suggest snapshots during the northeastward shift.
of the EASM’s rain belts, which are directly regulated by the EASM’s CISO, as indicated by the coupling of the MC, WNPSH, and SAH.

5. Linkages to the air–sea interaction

a. SST–$Q_1$ modes

Because of the tropical ocean’s high heat capacity, the sea surface temperature (SST) exhibits a slow change during its seasonal cycle; therefore, the oscillation in the atmosphere over a single month can be considered to be a response to the SST forcing in the tropical ocean (Hoskins 2013). The gradual variation in SST may couple with $Q_1$ and modulate the EASM’s CISO. Here, the MV-EOF approach is again applied to the CISO components of the SST and $Q_1$ over the region 0°–40°N, 40°–160°E in boreal summer. The results represent the feedback between atmospheric circulation and SST over the EASM region.

Figure 10 shows the first two leading modes of CISO in the SST and $Q_1$, which account for 37.4% and 28.3% of the total variance, respectively. In the first MV-EOF mode of the SST and $Q_1$, anomalous diabatic heating occurs over continental Asia, with its center settling over southern India, the southern edge of the TP, and Japan (Fig. 10a). Conversely, the diabatic cooling anomalies mainly occur over the Arabian Sea and the WNP, with three cooling centers situated over western Sumatra, the SCS, and the open ocean near southern Japan. Further, tropical SST anomalies exhibit a tripole mode, manifesting as cooling to the south of 10°N and around the Japan Sea and warming over the Bay of Bengal (BOB), SCS, and WNP (Fig. 10c). However, the tropical SST anomalies display a zonal dipole pattern in the northern Indian Ocean (0°–20°N, 45°–90°E). The configuration between the SST and $Q_1$ is quite different over the WNP versus the Arabian Sea. Because tropical diabatic heating is primarily contributed by condensation heating, the diabatic heating (cooling) over the tropical ocean suggests local enhanced (suppressed) convection. Over the WNP, the warm (cold) SST is overlaid by diabatic cooling (warming), thus suggesting that the local convection is suppressed (encouraged) to warm (cool) the SST by
increasing (decreasing) the shortwave radiation and decreasing (increasing) the surface latent heat flux. However, diabatic cooling is superimposed on the cold SST over the Arabian Sea. This pattern demonstrates that the suppressed convection related to diabatic cooling should result from the local cold SST in the Arabian Sea, which decreases the supply of energy to the atmosphere and the convective instability via the air–sea interaction. As a result, the anomalous tropical convection is the reason for rather than the result of the local SST anomalies over the WNP, whereas the SST anomalies determine the convection over the Arabian Sea.

In the second SST–$Q_1$ mode (Figs. 10b,d), the anomalous diabatic cooling and heating centered over land straddle the Yangtze River over East Asia (Fig. 10b), whereas over the ocean, the diabatic cooling is centered over the Arabian Sea and the SCS. In addition, warm SST anomalies are distributed from the Arabian Sea to the SCS crossing the BOB (Fig. 10d). A distinct zonal gradient of SST anomalies is then generated between the SCS and the Philippines. The SST anomalies exhibit a meridional alternative pattern, but their phase differs from that of the first mode in the northwestern Pacific Ocean. Anomalous diabatic cooling is collocated with the warm SST over the Arabian Sea and the SCS, thus implying a local SST response to atmospheric forcing.

The interaction between the EASM CISO and the surrounding SST anomalies can be indicated by their leading modes. Figures 10e,f show the first two PCs of CISO with respect to the EASM and SST–$Q_1$ anomalies. The correlation coefficients of the first two PCs are +0.64 and +0.71, respectively, with statistical confidence at the 95% level. Notably, the EASM CISO leads to changes in SST for several days during June–September, implicating the EASM CISO modulation on the surrounding SST anomalies, which, in turn, affect the underlying condition of the EASM. During the positive PCI phase, the coupled mode of the EASM generally exhibits an enhanced WNPSH with suppressed rainfall over the WNP, leading to local clear skies and thus greater shortwave radiation to warm the SST. To the south and north of the WNPSH, however, abundant rainfall and heavy cloud cover may reduce...
shortwave radiation reaching the sea surface. Subsequently, the warming and cooling SST in the WNP may not only enhance the meridional SST gradient (Fig. 10c) but also modulate the summer rainfall over the Yangtze River. We noted that the first two PCs of EASM CISO always lead the counterparts of CISO in the SST–Q1 mode, implying atmospheric forcing on the SST at the intraseasonal time scale. Although the PCs of the EASM CISO lead the SST–Q1 CISO by 10 days, the relationship between WNPSH and SST–Q1 CISO is distinct in the two leading modes (Figs. 11a,b). The SAHI and MCI lead the PC1 of SST–Q1 CISO by 15 and 5 days, whereas the WNPSH lags the PC1 by 5 days (Fig. 11a). The SAHI, MCI, and WNPSH lead the PC2 of SST–Q1 CISO by 30, 20, and 5 days, respectively (Fig. 11b). This result indicates that the feedback between the WNPSH associated with the EASM CISO and the air–sea coupling system over East Asia may support the life cycle of the EASM CISO in boreal summer. The SAH and MC may cause the WNPSH to regulate the local air–sea interaction over the WNP. Specifically, the enhanced SAH and MC may enhance the midlatitude westerly winds, which strengthen the WNPSH to shift northward. Subsequently, the sea surface would receive more shortwave radiation to warm the SST because of the lower cloud cover and weaker convection related to the enhanced WNPSH. In addition, the air–sea coupling system over East Asia may affect the strength of the EASM CISO by changing Q1 over the WNP. If Q1 is suppressed with warmer SSTs over the WNP, a positive vorticity source (∂Q1/∂y > 0) would emerge to the north of the WNP, where the cyclonic vorticity of the air column in situ is increased, thereby weakening the ascending and EASM convection. Finally, the EASM CISO would be dampened, leading to enhanced convection and colder SSTs over the WNP.

b. Model simulation

An atmospheric general circulation model (AGCM) of ECHAM5.4 (Roeckner et al. 2003) is used to further
examine the role of SST forcing in maintaining the EASM CISO. We used the version with a spectral T63 horizontal resolution and 31 vertical levels (T63L31); the vertical domain of the model covered the surface level to 10 hPa. The daily SSTs used to drive this model were the same as those in the data analysis. The 30-yr (1982–2011) averaged circulation, forced by the daily SSTs, was defined as the climatology of the model simulation.

Figure 12 shows the model-simulated climatological mean winds and rainfall during May–August and the seasonal variation of rainfall at 115°E. In contrast, the model is able to accurately simulate the observed wind patterns, the tropical low-level westerly jet, the WNPSH at 850 hPa, and the tropical rainbands. The WNPSH is also clearly observed at 500 hPa in the simulation results (Fig. 12b). The SAH, characterized by a large-scale anticyclone centered on the TP, is present in the upper troposphere (Fig. 12c). The simulated seasonal variation of subtropical rainfall along 115°E exhibits a northward shift during May–August, which is much weaker than

![Figure 12](https://example.com/fig12.png)

**Fig. 12.** As in Fig. 1, but for the model simulations.
the observations and is limited to south of 30°N (Figs. 1d and 12d). However, a subtropical rainband originating from 25°N is still present, and it rapidly shifts northward after mid-May and reaches its maximum in August. Relative to the AC component, the seasonal variation of subtropical and tropical rainfall exhibits a clear fluctuation during this period.

The EASM CISO was also simulated. The simulated seasonal variation of rainfall at 30°N, 115°E presents two peaks occurring in late June and August, respectively (Fig. 13a), corresponding to the double AC peaks, which are distinct compared with the single peak in the observations. In addition, the observed maximum rainfall is approximately 12 mm day$^{-1}$, but it is only 4 mm day$^{-1}$ in the simulation (i.e., only one-third that of the observations). The simulated maximum rainfall occurs in August, which is later than in the observations (late June). Therefore, this model fails to simulate the observed seasonal variation of rainfall in the lower reaches of the Yangtze River. However, the enhanced rainfall CISO is captured by the model (Fig. 13b), indicating the enhancement of local rainfall at the intraseasonal time scale during the mei-yu season. Although the time series of the AC-removed rainfall at 40°N, 115°E exhibits a CISO during May–August (Fig. 13c), its amplitude and evolution are quite different from the observations (Fig. 4c). In contrast, the seasonal variation and the AC of EASMI are well reproduced by the model (Fig. 13d), indicating a strong ability in simulating the seasonal transition between winter and summer. Moreover, the time series of EAMI without AC shows a marked CISO (Fig. 13e), but its 40–80-day period is insignificant in the wavelet analysis.

The coupling modes of CISO in EASM can also be partly captured by the AMIP run. Figure 14 shows the first two model-simulated MV-EOF modes of the EASM's CISO. In simulation, the first mode of circulation at different levels (Figs. 14a,c,e) is similar to that of the observations, which is characterized by the jointly enhanced MC, WNPSH, and SAH (Figs. 5a, 5c, and 5e, respectively). In the second mode, however, the simulated 500-hPa WNPSH is weaker and the 200-hPa SAH is stronger than the observations (Figs. 14d,f), and both are located farther to the north (Figs. 5d,f). Therefore, the model has a spatial bias and fails to simulate the second mode of the observed EASM CISO.

Figure 15 shows the first two PCs and the time series of the MCI, WNPSHI, and SAHI in the observations and the model simulation, respectively. The high correlation coefficient between the simulated and observed PC1 suggests that the SST forcing can accurately reproduce

**Fig. 13.** As in Fig. 4, but for the model simulations. (f) Wavelet analysis (Torrence and Compo 1998) of the residual (shading), where the enclosed contours indicate regions of 95% confidence for a red-noise process, and the thick blue dashed line denotes the cone of influence, outside of which the edge effects become important.
the first leading mode of the EASM CISO. However, the AGCM cannot accurately represent the observed PC2. For each component of the EASM CISO, the MCI could be well simulated, whereas the AGCM cannot capture the observed evolution of the WNPSHI and SAHI. Because the SST in the observed second mode of SST–$Q_1$ acts as a response of SST to the atmospheric circulation, the AGCM cannot accurately represent the observed air–sea interaction, resulting in the failure in simulating PC2 of the EASM CISO.

6. Summary and discussion

a. Summary

The climatological intraseasonal oscillation (CISO) of the East Asian summer monsoon (EASM) is characterized
by the vertical and meridional interactions of monsoon circulation, with a stepwise northward shift of front-related rain belts during boreal summer, particularly from May to August. To reveal the vertical structure and the internal modes of the EASM CISO, as well as their interaction with the surrounding SSTs, we conducted harmonic and multivariate empirical orthogonal function (MV-EOF) analyses on the climatological daily winds, rainfall, diabatic heating, and SST for the period 1981–2010.

The EASM CISO exists not only in diabatic heating, primarily contributed by condensation heating, but also in rainfall and circulation. The existence of the EASM’s CISO is determined by both the different timing of diabatic heating $Q_1$ over East Asia due to the land–sea thermal contrast and the distinct phase relationship between the $Q_1$ components. During the EASM season, the $Q_1$ phase associated with condensation heating is centered over the WNP and East Asia. Thus, the horizontal gradient of $Q_1$ produces a negative vorticity source to the north and east of the $Q_1$ centers, favoring the local background ascending that triggers the CISO. Moreover, the EASM CISO initiates with the change in surface sensible heating, which modulates the atmospheric convective instability, followed by the variation of monsoon convection, releasing condensation heating, as well as the regulation of radiative heating.

The MV-EOF results reveal that the first two modes of the EASM CISO are mainly characterized by the coupling of the Mongolian cyclone (MC) around Lake Baikal at 850 hPa, the WNP subtropical high (WNPSH) at 500 hPa, and the South Asian high (SAH) over the Tibetan Plateau (TP) at 200 hPa. The first leading mode shows a jointly enhanced MC, WNPSH, and SAH accompanied by a tripole rainfall anomaly of the strong mei-yu and baiu fronts around the lower reaches of the Yangtze River and southern Japan, whereas the rainfall was suppressed over northeastern Asia and the WNP. The second leading mode, which indicates the eastward and northwestward propagation of the enhanced MC and WNPSH with the weakened SAH, is associated with a dipole of the rainfall anomaly, with abundant and deficient rainfall over the northeastern and southeastern Asia–Pacific regions, respectively. The first two CISO
modes of daily SST and atmospheric $Q_1$ are closely associated with the counterparts of the EASM CISO, except for a phase difference. Warming SST caused by the suppressed convection would enhance the WNPSH and cause the triple rain belts over East Asia through the coupling modes of the EASM CISO.

In this study, we also applied ECHAM5.4, with a version at T63L31, to perform the AGCM simulations forced by daily SST from 1981 to 2010. The results show that the simulated first CISO leading mode is consistent with the observed mode, suggesting the critical role of seasonal variations of daily SST in the EASM CISO over the Asia–Pacific region. However, the AGCM runs fail to realistically resemble the second CISO leading mode. The MC CISO is well reproduced by the AGCM, but the simulated CISO of the WNPSH and SAH is poor compared with the observations owing to the lack of feedback of atmosphere forcing.

b. Discussion

Since the MJO is the most prominent ISO in the tropics, one might instinctively consider it as the main source of the EASM CISO. Note that the MJO events are distinct in the individual year, and the effects of MJO on the climate-mean state exist possibly due to their phase lock to the annual cycle. To examine the phase locking of MJO to the annual cycle, we have applied the statistical method of a box-and-whisker plot (Tukey 1977) to the daily real-time multivariate MJO indices (RMM1 and RMM2), MJO intensity, and phase during April–October for the period 1981–2010 (Wheeler and Hendon 2004). There are three MJO-like events observed during May–August, but its period is less than 40 days. In addition, less than 40% of OLR CISO over the southern WNP could be linked to the MJO events, implicating the limited contribution of MJO to the EASM CISO modes.

The possible mechanism of EASM CISO could be attributed to the SST–$Q_1$ CISO over the WNP, which exhibits a remarkable northward propagation of anomalous SST and $Q_1$, as shown in Figs. 10a and 10b, respectively. Previous studies have ascribed the intraseasonal northward shifting of SST–rainfall pattern over the WNP to the local air–sea interaction (Lau and Nath 2009). A similar process is observed in the SST–$Q_1$ CISO mode over the WNP (figures not shown). Since the basic flow over the WNP is featured by the strong southeasterly over the southern WNP and weak northwesterly over the northern WNP during boreal summer (May–October), the more-than-normal rainfall over the cold SST anomalies in the WNP could induce an abnormal near-surface cyclone in situ. It could enhance (weaken) the evaporation latent heating over the northern (southern) WNP via the wind–evaporation–SST feedback (Xie and Philander 1994), which supports the northward propagation of SST CISO mode. In the meantime, the zonal land–sea thermal contrast over East Asia is modified to alter the EASM rainfall and circulation as required by the thermal wind relationship (He et al. 2008; Zhu et al. 2011; He and Liu 2016). The robust cycle of EASM CISO starts with the changes in WNPSH and MC in the lower troposphere, followed by the upper-level SAH variation when the rainfall anomalies over East Asia are enhanced. Finally, the EASM CISO and the SST–$Q_1$ CISO over the WNP are coupled together to move northward.

Wang and Xu (1997) have speculated that the regulation of seasonal stepwise planetary-scale circulation related to the atmospheric heating may be responsible for the phase-locking behavior of the CISO to an annual cycle. This appears to be the result of complex nonlinear interactions among atmosphere, ocean, and land. In our results, we found that the CISO is most evident during May–August, instead of the whole boreal summer season, and that it exhibits phase locking to the EASM establishment and the subsequent summer monsoon onset in the Asia–Pacific region. During this period, the annual cycle maximum of diabatic heating over the land and sea exhibits distinct phase lags (Fig. 2a), which may be responsible for the generation of the CISO signal in the summer monsoon region. Therefore, to better understand the mechanism of CISO, further research should focus on the impacts of local air–sea interactions on the CISO over the Asia–Pacific region. In addition, the CISO can be observed over other summer monsoon regions in boreal summer (figure not shown), but its strength is weaker than that of the EASM CISO. The features of CISO over other monsoon regions and the mechanism for their formation also require further investigation.

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