The Crucial Role of Internal Variability in Modulating the Decadal Variation of the East Asian Summer Monsoon–ENSO Relationship during the Twentieth Century

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ABSTRACT

This study investigates the role of internal variability in modulating the East Asian summer monsoon (EASM)–ENSO relationship using Twentieth-Century Reanalysis (20CR) data and simulations from phase 5 of CMIP (CMIP5). Analysis of 20CR data reveals an unstable EASM–ENSO relationship during the twentieth century. During the high-correlation periods of 1892–1912 and 1979–99, an evident western Pacific anticyclone (WPAC) and dipole sea level pressure (SLP) pattern are present in the decaying El Niño summer, accompanied by Indian Ocean warming and a tropospheric temperature Matsuno–Gill pattern. However, these are weaker or absent during low-correlation periods (1914–34 and 1958–78). After removing the external forcings based on historical simulations from 15 CMIP5 models, all the above features remain almost unchanged, suggesting the crucial role of internal variability. In a 501-yr preindustrial control (piControl) simulation without external forcing variation from CCSM4, the EASM–ENSO relationship also shows significant decadal variation, with a magnitude comparable to the 20CR data. The analysis demonstrates that the EASM–ENSO relationship’s variation is modulated by the interdecadal Pacific oscillation (IPO). Compared to negative IPO phases, the warmer East China Sea in positive IPO phases weakens the western North Pacific subtropical high (WNPSH), inducing more precipitation. Thus, the Kelvin wave–induced interannual divergence suppresses more mean-state precipitation and leads to a stronger WPAC. Hence, the IPO modulates the EASM–ENSO relationship through the WNPSH, which is evident in both 20CR and the piControl simulation.

1. Introduction

The East Asian summer monsoon (EASM) is the most important climate system over East Asia. The floods, droughts, and related heat waves caused by an anomalous EASM cause huge economic losses and casualties (Zhou et al. 2013, 2014), so there are many efforts in progress to understand the interannual variability of the EASM. The El Niño–Southern Oscillation (ENSO) is regarded as one of the most important factors to influence the EASM (Chen et al. 1992; Shen and Lau 1995; Zhang et al. 1999; Chang et al. 2000; Wang et al. 2000; among many others).

In the summer following an El Niño event, an anticyclone appears over the western Pacific (WPAC; Zhang et al. 1999; Wang et al. 2003), connecting the ENSO with the EASM. The WPAC suppresses convection over the western Pacific and transports moisture to eastern China and Japan, forming a dipolar rainfall pattern: rainfall decreases in the western Pacific and increases over the mei-yu front region (28°–38°N, 105°–150°E). In an ENSO-decaying summer, the influence on the EASM is indirect and mainly mediated by tropical Indian Ocean (TIO) warming. This mediation has been demonstrated by both observational analysis (Yang...
et al. 2007; Xie et al. 2009; Wu et al. 2009) and climate models (Song and Zhou 2014a,b).

This relationship between the ENSO and EASM displays significant decadal variation. The recent strengthening after the mid-1970s has received much attention (Wu and Wang 2002; Wang et al. 2008b; Xie et al. 2010; Huang et al. 2010). The mechanisms for this strengthening are attributed to background change (Wu and Wang 2002), increased magnitude and periodicity of the ENSO, and a strengthened monsoon–ocean interaction (Wang et al. 2008a)—or the strengthening TIO response to the ENSO (Xie et al. 2010; Huang et al. 2010).

Because of limited observations, there are few studies discussing the EASM–ENSO relationship before the 1950s. Chowdary et al. (2012) investigated the ENSO–north Indian Ocean (NIO) sea surface temperature (SST) relationship from 1870 to 2007 using shipboard surface meteorological observations. They found that in both the late nineteenth and early twentieth centuries, and after the mid-1970s, the El Niño–induced NIO warming persisted into the summer and triggered significant western Pacific atmospheric anomalies. Further investigation of the EASM–ENSO relationship, and its related mechanisms, during the whole twentieth century is urgently warranted.

The possible influence of global warming on the EASM–ENSO relationship is also controversial. Chowdary et al. (2012) argued that the recent strengthened relationship is not entirely due to global warming but also reflects internal variability, since the relationship was also strong 100 years ago. The interdecadal Pacific oscillation (IPO; Power et al. 1999; Folland et al. 2002) is an important mode of internal variability, and the Pacific decadal oscillation (PDO; Mantua et al. 1997) is its North Pacific component. Through analyzing the behaviors of the EASM in an El Niño–decaying summer in different phases of the PDO, Feng et al. (2014) found that the PDO may modulate the EASM–ENSO relationship through the decay speed of El Niño during 1957–2011. Dong and Dai (2015) found that the IPO plays an important role in modulating the ENSO’s influence on the surface temperature and precipitation over many regions during 1920–2012. Meanwhile, many studies have suggested that the TIO response to the ENSO and its capacitor effect is enhanced under global warming. Zheng et al. (2011) and Hu et al. (2014) have suggested that the strengthening after the mid-1970s may be influenced by global warming. Hence, it is important to investigate the relative roles of external forcing and internal variability in the EASM–ENSO relationship during the whole twentieth century.

In this study, we aim to answer the following two questions: 1) what is the role of internal variability in the EASM–ENSO relationship during the twentieth century, and 2) what is the physical mechanism responsible for the relationship's variation? Two approaches are adopted to answer these questions. The first removes external forcing linearly in the Twentieth-Century Reanalysis (20CR) data and compares the relationship before and after the removal. The second studies the EASM–ENSO relationship in a coupled model with external forcing set at the preindustrial level. We find that internal variability plays a crucial role in the relationship’s variation throughout the twentieth century. The IPO modulates the EASM–ENSO relationship by weakening (strengthening) the western North Pacific subtropical high (WNPSH) in its positive (negative) phase.

The rest of the paper is organized as follows. In section 2, observations, model datasets, and the analysis method are described. Section 3 displays the main results, including the decadal variation of EASM–ENSO relationship in the 20CR data and the role of internal variability inferred from the 20CR data and a coupled model. Section 4 summarizes the main conclusions.

2. Observations, model datasets, and the analysis method

a. Observational datasets

We use the monthly 850-hPa winds, sea level pressure (SLP), and tropospheric temperature during 1871–2005 from the 20CR data at a horizontal resolution of 2° × 2° (Compo et al. 2011). The 20CR data comprise a 56-member ensemble and assimilate only surface pressure and SLP information. In the late nineteenth and early twentieth century, the lack of SST and SLP observations over East Asia may affect the results, so it is essential to evaluate the ensemble spread as in many previous studies (King et al. 2013; Wang et al. 2013). Hence, although the ensemble mean is used in the most analyses, the ensemble spread is also estimated by the standard deviation of the 56 ensemble members. The data have previously been extensively used to study the climatology, interannual variability, and decadal variability of the EASM (e.g., Song and Zhou 2012; Qian and Zhou 2014). We also use the monthly SST during 1871–2005 from NOAA’s Extended Reconstructed SST, version 3b (ERSST.v3b) with a horizontal resolution of 2° × 2° (Smith et al. 2008).

b. Model datasets

Historical experiments from 15 models from phase 5 of CMIP (CMIP5) are selected to represent the external forcing in the observations during 1871–2005. These models have been shown to represent well the EASM (Song and Zhou 2014b), the ENSO (Bellenger et al. 2014), and the EASM–ENSO relationship, including the decadal variation (Table 1). The model details,
including the originating institution and horizontal resolution, are described in Table 1.

To further investigate the role of internal variability on the decadal variation of the EASM–ENSO relationship, a 501-yr preindustrial control (piControl) experiment using CCSM4 (Gent et al. 2011) is analyzed. In the piControl simulation, the external forcing is fixed at the preindustrial level, so it is convenient to discuss the role of internal variability. CCSM4 is one of the best models simulating the EASM (Song and Zhou 2014b), the ENSO (Bellenger et al. 2014), and the decadal variation of the EASM–ENSO relationship (Table 1).

c. Analysis method

All the datasets are interpolated onto a 2.5° × 2.5° common grid. To focus on the interannual variability, a 2–9-yr bandpass filter is applied in the observational and model datasets with a Lanczos filter (Duchon 1979). In this study, the Pearson correlation is used in the correlation analysis except for the special statement; the Spearman rank correlation is also used for comparison. It is found that the two correlations are consistent in most places, indicating the Pearson correlation with normality assumption is reasonable in this study. For the significance test, the effective sample numbers \( N_{\text{eff}} \) for correlations of time series \( X \) and \( Y \) are evaluated by the following formula (Metz 1991; Livezey 1995):

\[
N_{\text{eff}} = RN_{\text{max}} \left[ 1,1 + 2 \sum_{\tau=1}^{\tau_{\text{max}}} \left( 1 - \tau/N \right) r_X(\tau)r_Y(\tau) \right].
\]

Here, \( R \) indicates ensemble size (\( R = 1 \) for the observation), \( N \) denotes the data length, and \( r_X(\tau) \) is an autocorrelation of time series \( X \) with a lag of \( \tau \) years. For simplicity, the maximum lag \( \tau_{\text{max}} \) is set as the maximum number less than \( N/2 \). In the running correlation analysis, we estimate the \( N_{\text{eff}} \) in each running correlation period.

To extract the IPO mode in the observational internal variability, an empirical orthogonal function (EOF) analysis is performed on the global annual mean SST after the 9-yr low-pass filter. The observational IPO index is the principal component of EOF1 (PC1), while the simulated IPO index is the projection of global annual mean SST in the model on the observational EOF1 after the 9-yr low-pass filter. Since the IPO is a coherent interdecadal variability with 10–15- and 40–60-yr periods (Deser et al. 2004; Dai 2013), it is important to take the full phases of IPO into account when considering its climate influence. Based on Feng et al. (2014), the positive (negative) IPO phases denote the years that the IPO index is greater (less) than zero in this study. An EASM index is defined by the June–August (JJA) mean difference between the 850-hPa geopotential height averaged over 22.5°–32.5°N, 110°–140°E and 5°–15°N, 90°–130°E. This index is a simple representation of the dominant mode of the EASM at the interannual time scale (Wang et al. 2008a). The boreal winter [December–February (DJF)] Niño-3.4 index [the SST averaged over the eastern equatorial Pacific (5°S–5°N, 120°–170°W)] is calculated for both the observational and model data to represent ENSO. A WNPSH index is defined as 850-hPa geopotential height averaged over the western Pacific (10°–30°N, 110°–150°E).

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Table 1. Details of the 15 CMIP5 models. HR denotes the horizontal resolution in grid points of the atmospheric model. The minimum and maximum of the 21-yr running correlation between the EASM and the Niño-3.4 index in the previous winter in each model during 1871–2005 are also listed in the model name column in parentheses. (Expansions of acronyms are available at http://www.ametsoc.org/PubsAcronymList.)
The factors influencing observational climate variability and change include both external forcing (such as greenhouse gases, aerosols, volcanic eruption, and solar variability) and internal variability (such as ENSO and IPO; IPCC 2014). Many studies assume that the observational anomalies are the sum of the forced component and the real-world realization of internal variability, so the real historical internal variability can be estimated as the residual after linearly removing the forced component (Dong et al. 2014; Steinman et al. 2015; Dai et al. 2015). Steinman et al. (2015) further justifies the reasonability of the linear removal method. To investigate the role of internal variability on the relationship in the 20CR data, the external forcing is also linearly removed in each grid according to the following formula (Dai et al. 2015):

$$y = ax + b.$$  

Here, $y$ is a variable in the 20CR data, and $x$ is the corresponding variable in the ensemble mean of historical simulations from the 15 CMIP5 models, representing the external forcing. Through linear regression, $ax$ represents the linear external forcing part of the variable in the 20CR data. Hence, $b$ represents the internal variability part of the variable in the 20CR data.

### 3. Results

#### a. The decadal variation of the EASM–ENSO relationship in the 20CR data

The 21-yr running correlation between the EASM and Niño-3.4 index in the previous winter in the 20CR data is shown in Fig. 1. The correlation shows significant decadal variation, ranging from 0.07 to 0.66. There is some uncertainty of the correlation due to the ensemble spread in the 20CR data. The uncertainty is relatively larger before the twentieth century and during the 1940s and 1950s. The larger uncertainty during the 1940s and 1950s may come from the discontinuity of the observation due to World War II, which is noted by Chowdary et al. (2012). However, the decadal variation of the EASM–ENSO relationship is still evident, so the influence of the spread is relatively low.

According to the running correlation maximum (minimum) with no overlap of the adjacent periods, two high- and two low-correlation periods are selected in the 20CR data: 1892–1912 and 1979–99 as high-correlation periods and 1914–34 and 1958–78 as low-correlation periods. Our selected periods are consistent with those of Chowdary et al. (2012), who used shipboard surface meteorological observations. The 1979–99 high- and 1958–78 low-correlation periods were previously revealed by Wang et al. (2008b) through analyzing National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996).

To further show the spatial patterns of 20CR’s uncertainty in the two high- and two low-correlation periods, the ensemble spread of SLP is shown in Fig. 2. It is found that the ensemble spread of SLP is relatively large in the early period but mainly located in the high latitudes. In the western Pacific and East Asia, the region this study focuses on, the ensemble spread is relatively low (Fig. 2).
Hence, the influence of the ENSO on the EASM is significant during the high-correlation periods but weakens or disappears during the low-correlation periods.

Many previous studies (Yang et al. 2007; Xie et al. 2009; Wu et al. 2010) have investigated how the Indian Ocean SST facilitates the ENSO’s teleconnection to the EASM in its decaying summer through an atmospheric Kelvin wave. The SST and upper-tropospheric temperature anomalies in the four periods are displayed in Fig. 4. In the high-correlation periods (Figs. 4a,d), both the Indian Ocean basinwide warming and western Pacific cooling are significant. The tropospheric temperature displays a prominent Matsuno–Gill pattern over the Indo-Pacific region (Matsuno 1966; Gill 1980): the off-equatorial maxima over the TIO are suggestive of baroclinic Rossby waves, and the warm wedge into the western Pacific is suggestive of the Kelvin wave. In the low-correlation periods (Figs. 4b,c), the Indian Ocean warming is weak and the western Pacific cooling almost disappears. The tropospheric temperature over the Indo-Pacific region is uniform, and the Matsuno–Gill pattern disappears.

b. The role of internal variability inferred from the 20CR data by removing external forcing

To investigate the role of internal variability in the decadal variation of the EASM–ENSO relationship, the influence of external forcing is removed using Eq. (2) in the 20CR data. The SLP and 850-hPa wind anomalies in the four periods after external forcing removal are shown in Fig. 5. Comparing Fig. 5 with Fig. 3, we find that the spatial patterns are almost unchanged and have similar spatial variation and magnitude. The spatial correlations for SLP exceed 0.99 in the four periods. This indicates that internal variability plays a substantial role on the EASM–ENSO relationship and its decadal variation.

The SST and upper-tropospheric temperature after removing the external forcing are shown in Fig. 6. The spatial patterns are also almost identical to those shown in Fig. 4, including the spatial variation and magnitude. The spatial correlations for SST are somewhat smaller than the SLP but also exceed 0.90 in the four periods. This indicates that the external forcing may also have little influence on the TIO SST anomalies and Matsuno–Gill patterns in the tropospheric temperature, which are important for the EASM–ENSO relationship.

The intensities of WPAC and the TIO SST anomaly in the El Niño–decaying summer are compared before and after removing the external forcing (Table 2). After removing the external forcing, the WPAC and TIO SST anomaly are mostly unchanged. The internal variability accounts for 103%–127% for WPAC (the external forcing cancels some influence), whereas it accounts for 78%–134% for TIO SST in the four periods. For the two decadal shifts, the internal variability contributes 102% for WPAC and 87% and 158%, for each decadal shift.
respectively, for TIO SST. Hence, the internal variability exerts large influences on the WPAC and TIO SST in the four periods and also the decadal shift.

What factors of internal variability are responsible for the decadal variation of the EASM–ENSO relationship? The dominant mode of global SST after removing the external forcing is shown in Fig. 7. The dominant mode shows a typical IPO pattern: the warm SST anomaly occupies the tropical eastern Pacific Ocean, and two cold SST anomalies are centered in the North and South Pacific. PC1 shows significant decadal variation, with two warming centers in the 1900s and after the mid-1970s, coinciding well with the two high-correlation periods. Hence, the IPO may exert an important influence on the decadal variation of the EASM–ENSO relationship.

Previous studies (Wang et al. 2008a; Xie et al. 2010; Chowdary et al. 2012) have suggested that ENSO variance is crucial for the decadal variation of the EASM–ENSO relationship. As shown in Fig. 7c, the EASM variance is consistent with the EASM–ENSO relationship’s variation, although there is some uncertainty in the EASM’s variance in the early twentieth century and during the 1940s and 1950s. In the high-correlation period, the ENSO’s variance is larger than that in the low-correlation period, consistent with the previous studies. However, the ENSO variance is much smaller than the EASM variance before the 1920s, and does not show a weakening tendency after 1915. This feature is inconsistent with the shift of the EASM–ENSO relationship. Further, the ENSO variance becomes larger after the 1960s, about 10 years ahead of the enhancement of EASM variance and the EASM–ENSO relationship. Thus, ENSO variance alone may not explain the decadal variation of the EASM–ENSO relationship. The spread of EASM variance, especially in the early twentieth century, does not affect the result, but uncertainties of the observed SST are not accounted for. Hence, some caution is needed in evaluating the ENSO’s role.

Meanwhile, the mean state in the western Pacific is important in determining the location and intensity of the WPAC (Wu et al. 2010). When the mean-state precipitation is stronger over the western Pacific, the Kelvin wave–induced interannual divergence can suppress more mean-state precipitation, leading to stronger
negative heating anomalies. The stronger negative heating anomalies further stimulate a stronger atmospheric Rossby wave response and a stronger WPAC. Hence, in addition to the ENSO variance, the change of the mean state related to the IPO may also be vital for the EASM–ENSO relationship.

To confirm this hypothesis, a composite analysis is conducted according to positive and negative IPO
phases. The mean-state changes of SST and 850-hPa geopotential height between positive and negative IPO phases are displayed in Fig. 8. The western Pacific is warmer west of 140°E, and the North Pacific is colder in the positive IPO phases compared to the negative IPO phases. In correspondence with the mean-state warmer SST over the western Pacific, the WNPSH weakens and shifts northeastward, consistent with the results of Song and Zhou (2014b). The mean-state convection and precipitation over the western Pacific is stronger because of the weaker WNPSH. Hence, the Kelvin wave–induced boundary layer divergence strengthens, and thus a stronger WPAC occurs. This relationship is also clearly shown in Fig. 8b. The 21-yr running mean of WNPSH is negatively correlated to the EASM–ENSO 21-yr running correlation. The correlation coefficient is −0.36 using the Spearman rank correlation, statistically significant at the 5% level. Thus, when the mean-state WNPSH is weaker, the WPAC is stronger because of the stronger Kelvin wave response, and the EASM is more closely connected to ENSO in the previous winter.

c. The role of internal variability inferred from a CCSM4 piControl experiment

The above analysis shows that internal variability plays a substantial role in the decadal variation of the EASM–ENSO relationship based on the 20CR data after removing the external forcing. It further indicates that the IPO is important for the relationship through weakening the WNPSH. However, these conclusions

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<tr>
<td>WPAC Before removal</td>
<td>0.311</td>
<td>0.108</td>
<td>0.128</td>
<td>0.290</td>
</tr>
<tr>
<td>After removal</td>
<td>0.333</td>
<td>0.133</td>
<td>0.135</td>
<td>0.296</td>
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<tr>
<td>Percentage in the period</td>
<td>107%</td>
<td>123%</td>
<td>105%</td>
<td>102%</td>
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<tr>
<td>Percentage in the shift</td>
<td>99%</td>
<td>102%</td>
<td>99%</td>
<td></td>
</tr>
<tr>
<td>TIO SST Before removal</td>
<td>0.144</td>
<td>0.071</td>
<td>0.033</td>
<td>0.123</td>
</tr>
<tr>
<td>After removal</td>
<td>0.170</td>
<td>0.055</td>
<td>0.044</td>
<td>0.122</td>
</tr>
<tr>
<td>Percentage in the period</td>
<td>118%</td>
<td>77%</td>
<td>133%</td>
<td>99%</td>
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<tr>
<td>Percentage in the shift</td>
<td>171%</td>
<td>87%</td>
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Fig. 6. As in Fig. 4, but for the internal variability after removing the external forcing based on the 15 models.
may be adversely impacted by the method used to extract the external forcing and the relatively short time span in the 20CR data. To further confirm these conclusions, we use a 501-yr piControl simulation from CCSM4 to investigate the possibility of a decadal variation of the EASM–ENSO relationship when the external forcing is prescribed at the preindustrial level. We also investigate whether the IPO is important for modulating the relationship’s decadal variation.

The 21-yr running correlation of the EASM and Niño-3.4 index in the previous winter in the simulation is shown in Fig. 9. It clearly shows that the relationship displays decadal and even centurial variation. The variation ranges from -0.15 to 0.65, similar to that in the 20CR data. The interannual variance of the EASM is well explained by the ENSO at a multidecadal or even a centurial time scale in CCSM4. But both the EASM and ENSO variances do not correspond well to the EASM–ENSO relationship, suggesting that other processes (such as the IPO) may play a role in addition to the ENSO’s variance. The IPO index in Fig. 9c shows significant decadal variation. Whether the decadal variation of the IPO contributes to the EASM–ENSO relationship is discussed later according to the composite analysis.

To investigate the spatial differences between the high- and low-correlation periods in the simulation, the

FIG. 7. The leading mode (EOF1) of low-pass filtered global SST in annual mean after removing the external forcing, showing (a) the spatial pattern and (b) the corresponding PC1. The EOF1 explains 20.3% of the variance. Also shown in (b) is the 21-yr running correlation between the EASM and Niño-3.4 index in the previous winter (red line). (c) The 21-yr running variance of the standardized EASM (black line) and Niño-3.4 (blue line) index in the previous winter. The gray shading in (b) and (c) indicates the 5th and 95th percentiles of the running correlation and EASM variance, respectively.

FIG. 8. (a) The composite differences of SST (shaded; K) and geopotential height at 850 hPa (contours; gpm) between positive and negative IPO phases after removing the external forcing. The white dots (thick contours) indicate that the SST (850-hPa geopotential height) difference is significant at the 5% level by Student’s t test. (b) The scatterplot between 21-yr running mean of WNPSH index (gpm) and the 21-yr running correlation between the EASM and Niño-3.4 index in the previous winter after removing the external forcing.
SLP and 850-hPa winds were regressed on the Niño-3.4 index in the previous winter as shown in Fig. 10. In the three high-correlation periods (Figs. 10a,c,e), the WPAC is strong and a dipolar SLP pattern is evident, comparable to the high-correlation periods in the 20CR data. In the three low-correlation periods (Figs. 10b,d,f), the WPAC and dipolar SLP pattern are absent or weak, similar to the low-correlation periods in the 20CR data.

To further consider the role of the Indian Ocean, the SST and tropospheric temperatures associated with the ENSO are displayed in Fig. 11. In the high-correlation periods (Figs. 11a,c,e), the Indian Ocean warming is slightly stronger than in the low-correlation periods (Figs. 11b,d,f). The tropospheric temperature displays an evident Matsuno–Gill pattern only in the high-correlation period. Therefore, Indian Ocean warming related to the ENSO plays an important role in the decadal variation of the EASM–ENSO relationship.

Besides the Indian Ocean warming, the IPO’s role in the decadal variation of the EASM–ENSO relationship is confirmed in Fig. 12. The SLP and 850-hPa winds regressed on the Niño-3.4 index in the previous winter are shown in Fig. 12a and Fig. 12b for positive and negative IPO phases, respectively. In both positive and negative IPO phases, the WPAC and dipolar SLP pattern are evident, but they shift farther westward and strengthen in the positive IPO phases. Hence, eastern China is more heavily influenced by WPAC in the positive IPO phases.

The climatological WNPSH and precipitation differences between positive and negative IPO phases are shown in Figs. 12c and 12d. The WNPSH is much weaker in the positive IPO phases than that in the negative IPO phases. Hence, the convection and precipitation are heavier in the positive IPO phases, so the Kelvin wave–induced divergence and WPAC are stronger than those in the negative IPO phases, suggestive of the IPO’s role.

4. Summary

This study investigated the role of internal variability on the decadal variation of the EASM–ENSO relationship using 20CR data and CMIP5 simulations. By removing external forcing signals based on historical simulations of 15 CMIP5 models, internal variability was found to play a crucial role in the relationship’s decadal variation. The IPO was also shown to play a crucial role in modulating the relationship by changing the mean
state over the western Pacific. The diagnoses of 20CR data are further confirmed by an analysis of the 501-yr piControl simulation of CCSM4. The major findings are summarized below:

1) The EASM–ENSO relationship is not stable during the twentieth century in the 20CR data, with one peak around the 1900s and another after 1970s. Two high- and two low-correlation periods are selected in the 20CR data: 1892–1912 and 1979–99 as high-correlation periods and 1914–34 and 1958–78 as low-correlation periods. In the high-correlation periods, the WPAC and dipolar SLP pattern are evident, accompanied by significant Indian Ocean warming and a tropospheric temperature Matsuno–Gill pattern. In the low-correlation periods, all these features are weak or absent.

2) After removing the external forcing signals based on historical simulations of 15 CMIP5 models, the interannual responses of the EASM and the Indian Ocean to the preceding ENSO derived from 20CR are almost unchanged. Hence, the internal variability

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Fig. 10. The SLP (shaded; hPa) and 850-hPa wind (vectors; m s\(^{-1}\)) regressed on the Niño-3.4 index in the previous winter in the 501-yr piControl experiment of CCSM4 during three periods of (left) higher and (right) lower correlation between the EASM and Niño-3.4 index in the previous winter. The black dots indicate that the SLP is significant at the 5% level using Student’s t test. The 850-hPa wind is shown when the meridional or zonal component is significant at the 5% level using Student’s t test.
plays a substantial role in the decadal variation of the EASM–ENSO relationship in the twentieth century. This is also confirmed through an analysis of the 501-yr piControl simulation of CCSM4: the decadal variation of the EASM–ENSO relationship is evident with correlation ranging from −0.15 to 0.65, similar to that in the 20CR data. Composite analysis indicates that the magnitudes of the WPAC and dipolar SLP pattern are comparable to those in the 20CR data in the high-correlation periods, whereas they are weaker or absent in the low-correlation periods. In the Indo-Pacific region, the Indian Ocean warming is stronger in the high-correlation periods, and the tropospheric temperature Matsuno–Gill pattern is evident only in the high-correlation periods. Hence, the observed decadal variation of the EASM–ENSO and related features can be well reproduced without the external forcing changes, suggestive of the important role of internal variability.

3) Through the EOF analysis of observed global SST after removing the external forcing, the IPO is identified as the dominant mode. The time evolution of the IPO matches well with the decadal variation of the EASM–ENSO relationship, with one peak around the 1900s and another after the mid-1970s. In the positive IPO phases, the SST over the western Pacific is warmer, so the WNPSH is weaker, favoring enhanced convection and heavier precipitation. Therefore, the interannual response to the Kelvin wave–induced divergence is stronger because of the stronger mean-state precipitation. Hence, the weaker (stronger) WNPSH in the positive (negative) IPO phases is corresponding well to the enhanced (suppressed) EASM–ENSO relationship. In the piControl simulation of CCSM4, the WPAC and dipolar SLP pattern are also stronger in the positive IPO phases than those in the negative IPO phases and accompanied by a stronger WNPSH. Hence, the IPO
modulates the EASM–ENSO relationship via the WNPSH.

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