Seasonal Modulations of El Niño–Related Atmospheric Variability: Indo–Western Pacific Ocean Feedback

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ABSTRACT
The spatial structure of atmospheric anomalies associated with El Niño–Southern Oscillation varies with season because of the seasonal variations in sea surface temperature (SST) anomaly pattern and in the climatological basic state. The latter effect is demonstrated using an atmospheric model forced with a time-invariant pattern of El Niño warming over the equatorial Pacific. The seasonal modulation is most pronounced over the north Indian Ocean to northwest Pacific where the monsoonal winds vary from northeasterly in winter to southwesterly in summer. Specifically, the constant El Niño run captures the abrupt transition from a summer cyclonic to winter anticyclonic anomalous circulation over the northwest Pacific, in support of the combination mode idea that emphasizes nonlinear interactions of equatorial Pacific SST forcing and the climatological seasonal cycle. In post–El Niño summers when equatorial Pacific warming has dissipated, SST anomalies over the Indo–northwest Pacific Oceans dominate and anchor the coherent persisting anomalous anticyclonic circulation. A conceptual model is presented that incorporates the combination mode in the existing framework of regional Indo–western Pacific Ocean coupling.

1. Introduction
El Niño–Southern Oscillation (ENSO) is the dominant mode of interannual variability, arising from ocean–atmosphere interaction in the tropical Pacific. El Niño occurs every few years. It is strongly phase locked to the annual cycle, growing in boreal summer JJA(0), peaking in winter ND(0)J(1), and decaying rapidly in the following spring MAM(1) (Fig. 1a, black curve) (Rasmusson and Carpenter 1982; Deser et al. 2010). Here the numerals denote the El Niño developing (0) and decaying (1) years, respectively. Hereafter, seasons refer to those of the Northern Hemisphere.

The equatorial Pacific Ocean is the center of action for Bjerknes’s feedback that gives rise to ENSO. Although sea surface temperature (SST) there (e.g., in the Niño-3.4 region) is widely used to track ENSO, SST elsewhere responds to ENSO (Klein et al. 1999) and can be important in forcing atmospheric anomalies. For example, an anomalous anticyclonic circulation (AAC) develops in the lower troposphere over the tropical northwest Pacific (NWP) in post–El Niño summers (Fig. 1a, red curve). The focus on summer in the literature is motivated by large socioeconomic impacts as it is the rainy season for South and East Asia with active typhoons over the northwest Pacific. El Niño–induced warming over the tropical Indian Ocean (Yang et al. 2007; Xie et al. 2009) and cooling over the northwest Pacific (Wang et al. 2003) anchor the atmospheric AAC in JJA(1) as equatorial Pacific SST anomalies are vanishingly small statistically. In fact, the Indo–west Pacific SST anomalies are coupled with the AAC through wind–evaporation–SST (WES) feedback, and the regional coupled mode explains why the center of action of El Niño–related anomalies shifts to the Indo–northwest Pacific Oceans in JJA(1) when the region is covered by the southwest monsoonal winds (Kosaka et al. 2017).
This Indo–western Pacific Ocean capacitor (IPOC) mode brings about predictability to East Asian climate variability in summer rainfall and landfalling typhoons through the NWP AAC (Chowdary et al. 2010; Kubota et al. 2016; Zhang et al. 2016).

Atmospheric anomalies may experience rapid changes despite rather smooth variations in Niño-3.4 SST. Over the NWP, atmospheric anomalies are weak in October(0) but in three months, the AAC is fully developed in January(1) (Fig. 1a, red curve) (Zhang et al. 1996; Weisberg and Wang 1997; Wang et al. 1999). The AAC develops rapidly despite little changes in equatorial Pacific SST anomalies from October(0) to February(1) at the peak of El Niño (Fig. 1a). Wang et al. (2000) suggest that WES feedback under the winter northeast trades over the NWP is important for the AAC in winter and spring but the rapid development of AAC remains to be explained. The prevailing winds switch to northeasterly over the NWP around November, a mean state change that is in favor of WES feedback and coincides with the rapid onset of AAC.

Recently, Stuecker et al. (2013, 2015) proposed a new idea that nonlinear interactions of the annual cycle and ENSO give rise to the NWP AAC. The time evolution of their combination (C) mode, symbolically cast as

\[ C = \cos(\omega_a t - \phi)N(\omega t) \]  

includes additional frequencies of \((\omega_a + \omega)/2\) and \((\omega_a - \omega)/2\), where \(\omega_a\) and \(\omega\) are the angular frequencies of the annual cycle and ENSO, \(\phi\) is a constant, and \(N\) is the Niño-3.4 index with enhanced power in a broad band of \(\omega\) at interannual frequencies. Since the combination tones of \((\omega_a + \omega)/2\) and \((\omega_a - \omega)/2\) are of much higher frequencies, they can cause sharper changes in time than the slow interannual ENSO frequency \(\omega\) may imply. Specifically, Eq. (1) explains the rapid transition from negative to positive anomalies of sea level pressure (SLP) over the NWP at the peak phase of El Niño.

Stuecker et al. (2013) detected spectral peaks associated with the C-mode of ENSO, although this result is being contested (Li et al. 2016) because the broadband nature of ENSO complicates the interpretation of spectral analysis (Stuecker et al. 2016). To resolve the debate of whether the C-mode exists, it is important to go beyond the mathematical expression of Eq. (1) and probe the physical mechanism. Using an atmospheric general circulation model (GCM), Stuecker et al. (2015) showed that the combination mode arises in response to a sinusoidal interannual variation in eastern Pacific SST due to the seasonal modulation by the mean state. The mechanism for this seasonal modulation and the relative importance between eastern Pacific SST forcing and regional ocean feedback from the Indo–western Pacific remain unclear.

The present study investigates the physical identity of the C-mode and casts it in the context of a large body of published literature. We show that the leading mode from seasonally stratified empirical orthogonal function (EOF) analysis (Wang et al. 2003) contains the C-mode..
of Stuecker et al. (2015) (section 3). The C-mode represents the seasonal modulation of atmospheric manifestations of ENSO over the NWP. To isolate the seasonal modulation by the mean state, we conduct an idealized experiment with an atmospheric GCM by setting SST forcing of El Niño constant in time (section 4). The same atmospheric GCM is also used to explore the relative importance of SST anomalies between the equatorial Pacific and the rest of the World Ocean. Our results show that the seasonal modulation by the mean state is most pronounced over the north Indian Ocean to NWP because of the climatological monsoon. Consistent with the C-mode, the direct forcing by eastern Pacific SST anomalies is instrumental in the NWP AAC onset at the peak phase of El Niño, but SST anomalies outside the equatorial Pacific are essential to sustain the NWP AAC through the post–El Niño summer. Section 5 is a summary and presents a conceptual model that reconciles the C-mode and regional IPOC effect.

2. Datasets and methods

a. Observations

We use the Hadley Centre Sea Ice and Sea Surface Temperature, version 1 (HadISST1, 1870–2016) dataset (Rayner et al. 2003), the National Oceanic and Atmospheric Administration’s (NOAA) Precipitation Reconstruction (PREC, 1948–2016) dataset (Chen et al. 2002), and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (1948–2016) surface winds and sea level pressure (SLP) datasets (Kalnay et al. 1996). We have regridded these datasets onto a common 2.5° latitude × 2.5° longitude grid for a 68-yr period from 1948 to 2016.

The present study focuses on near-annual to interannual variability associated with ENSO. Monthly anomalies are derived relative to the climatological mean over the whole period (1948–2016) after removing the linear trend. To reduce the effect of intraseasonal variability, we perform a 3-month running average. A 9-yr running mean is then applied (separately for each calendar month) to remove decadal and longer variations. Our conclusions remain unchanged when using the unfiltered data. We use SST averaged over the eastern Pacific (Niño-3.4: 5°S–5°N, 120°W–170°W) to track ENSO, which is referred to as the ENSO index.

Composite and linear regression analyses are employed to extract impacts of ENSO and C-mode. For composite analysis, we choose the following three strong El Niño events: 1972/73, 1982/83, and 1997/98. The main conclusions remain unchanged even if we use all El Niño events for our composite analysis, albeit with a reduced atmospheric response.

We perform two types of EOF analysis. The conventional EOF analysis projects data onto eigenvectors that are I × J long, where I and J are the numbers of grid points in the zonal and meridional directions, respectively. Each eigenvector is associated with a monthly time series called the principal component (PC) that is 12 months × N years long. A typical El Niño event lasts for one year from June to the following May. Since the structure of atmospheric response (e.g., NWP AAC) evolves through an ENSO year, one can alternatively perform EOF analysis by stacking monthly data for an ENSO year from June to the following May (e.g., Wang et al. 2003). Each eigenvector is 12 × I × J long and represents a set of 12 sequential monthly spatial patterns that share the same yearly PC. Since the 12 monthly patterns are generally not the same, a seasonally stratified EOF mode projects onto more than one mode obtained from the conventional EOF analysis without seasonal stratification.

b. GCM

We use the Community Atmosphere Model, version 5.3 (CAM5.3), the atmospheric component of the Community Earth System Model, version 1.2.2 (CESM1.2.2). The model is developed at NCAR with significant community collaboration. CAM5.3 includes some modifications to the spectral element dynamical core based on CAM5 (Neale et al. 2012). We choose a 1.9° × 2.5° grid (‘‘f19_f19’’) in the horizontal and 30 sigma levels in the vertical.

The model is forced with the observed monthly climatology of SST and sea ice. We analyze a 20-yr period of the control (CTL) simulation. We carry out three additional experiments. In the constant El Niño run, we impose El Niño warming over the equatorial Pacific. The pattern of SST anomalies is obtained by regressing SST against the Niño-3.4 SST index without distinguishing seasons, and only the region of positive values from 170°E to the coast of South America within the tropical Pacific (25°S–25°N) is retained (Stuecker et al. 2015). We impose this SST pattern of a constant amplitude (Niño-3.4 SST anomalies = 1.32°C). The constant El Niño run lasts for 21 years and the results of the last 20 years are analyzed.

A seasonally evolving composite of El Niño for 24 months from January(0) to December(1) is obtained from three major events of 1972/73, 1982/83, and 1997/98. Two ensemble experiments are performed as follows. In the tropical Pacific ocean–global atmosphere (POGA) run, the 24-month SST composite was imposed in the equatorial Pacific region defined above while the
SST climatology is prescribed elsewhere. In the global ocean–global atmosphere (GOGA) run, the anomalous SST forcing is imposed globally. Both the POGA and GOGA runs are repeated 10 times with slightly different initial conditions. The 10-member ensemble-mean difference from CTL is analyzed for atmospheric response to seasonally evolving El Niño.

3. Physical identity of C-mode

The tropical Indo–Pacific Oceans feature strong annual cycles. Over the Indo–western Pacific, the rainfall maximum is displaced into the summer hemisphere and the prevailing winds change from northeasterly in winter to southwesterly in summer. Furthermore, ENSO is strongly phase locked onto the annual cycle (Fig. 1). Here we define an ENSO year that starts in June \((m = 1)\) and ends in the following May \((m = 12)\).

We perform a seasonally stratified EOF (S-EOF) analysis to extract the seasonal dependency of atmospheric anomalies associated with ENSO, following Wang et al. (2003). The analysis is applied to zonal wind variability in the equatorial Pacific \((10°S–10°N, 100°E–60°W)\), the same domain as in Stuecker et al. (2015). The first S-EOF mode captures 42.3% of the total variance in the domain, and the PC is highly correlated with the ND(0)J(1) Niño-3.4 index at 0.96. The most prominent feature is the westerly wind anomalies near the equator in the western half of the equatorial Pacific (Fig. 2), which are a key element of Bjerknes’s feedback flattening the equatorial thermocline. A close look reveals that this patch of anomalous westerlies gradually migrates southeastward from June(0) to May(1). The southward displacement away from the equator from December(0) onward contributes to the rapid decay of ENSO by weakening the equatorial thermocline response (Harrison and Vecchi 1999; Ohba and Ueda 2009; McGregor et al. 2012). The seasonal dependency of ENSO’s poleward teleconnection is also apparent: a midlatitude anomalous cyclone develops over the South Pacific in JJA(0) and North Pacific in DJF(1) (Fig. 2) because the intensified westerly jet serves a waveguide for barotropic Rossby waves in the winter hemisphere.

Over the NWP, atmospheric circulation variability in S-EOF1 is highly seasonal. In the El Niño developing summer JJA(0), an anomalous cyclone develops. At the peak of El Niño DJF(1), the circulation anomalies become anticyclonic, and the AAC is fully developed at the decay phase of El Niño MAM(1). The S-EOF1 captures this rapid transition from cyclonic to anticyclonic circulation to El Niño within one ENSO year (Fig. 1a, blue curve).

While Eq. (1) is used as the C-mode index (Stuecker et al. 2013, 2015), it is never formally derived. Here we show that the linear S-EOF1 mode contains the combination mode in Eq. (1), which involves nonlinear interactions in frequency and time. S-EOF analysis finds 12 monthly spatial patterns \((R_m)\) that share the same yearly evolution \(N(t)\):

\[
R = R_m(x, y)N(t), \quad m = 1, 2, \ldots, 12. \tag{2}
\]

We can decompose \(R_m\) into an annual-mean pattern \(\overline{R}(x, y)\) and the monthly deviation \(R'_m(x, y)\) as

\[
R_m(x, y) = A_m\overline{R}(x, y) + R'_m(x, y), \tag{3}
\]

where \(A_m\) is the monthly Niño-3.4 SST in the El Niño composite (black curve in Fig. 1a), and \(R = \sum R/\sum A_m\).
denoting the 12-month sum. The first term on the right-hand side of Eq. (3) describes the seasonal amplitude modulation of the response pattern common to all seasons $R(x, y)$; e.g., the westerly anomalies in the central equatorial Pacific, Fig. 2, while the second term represents seasonal variations in patterns of atmospheric anomalies (e.g., the southeastward displacement of the equatorial westerly anomalies, and the abrupt onset of the AAC over the NWP, Fig. 2).

Combining Eqs. (2) and (3) yields $R = [A_m R(x, y) + R'_m(x, y)]N(t)$, or

$$R = A_1(\omega_t t)N(t)R(x, y) + A_2(\omega_t t)N(t)R_2(x, y),$$

where $A_1$ is an annual periodic function with a nonzero annual mean, and $A_2$ is the annual periodic function with a zero annual mean ($\sum R'_m = 0$). Here we have decomposed $R'_m(x, y)$ into a spatial pattern $R_2(x, y)$ and phase evolution $A_2(\omega_t t)$, say by EOF analysis.

One may choose to perform EOF analysis without distinguishing seasons, as in Stuecker et al. (2013, 2015). The first PC is highly correlated with Niño-3.4 SST ($r = 0.9$). Thus, with our choice of weight $A_m$ in Eq. (3), the first term on the right-hand side of Eq. (4) approximates the variability of the nonseasonal EOF1 (N-EOF1). Indeed, when we perform nonseasonally stratified EOF analysis on S-EOF1 variability, the first two N-EOF modes closely resemble those of the raw variability (Fig. 3, first vs second row). The first N-EOF mode represents the zonal wind variability in the western half of the equatorial Pacific and is nearly identical to the weighted annual mean of S-EOF1, $R(x, y)$ in Eq. (3) (Fig. 3e), while the second mode is tied to the AAC over the NWP (Stuecker et al. 2013). As the first two N-EOF modes together explain almost all (98%) of the S-EOF1 variance, the second term on the right-hand side of Eq. (4) represents the second N-EOF mode, and the PC, $A_2(\omega_t t)N(t)$, is indeed the combination mode of Stuecker et al. (2013, 2015) as expressed in Eq. (1).

ENSO-related atmospheric variability shows strong seasonal dependency. Equation (2) views ENSO variability in seasonally stratified analysis while Eq. (4) represents the same variability continually in time without seasonal stratification. Equation (4) decomposes the seasonal variations in ENSO anomalies into a component due to the seasonal modulation of Niño-3.4 SST amplitude (first term on the right-hand side) and one associated with the seasonal modulation of the spatial pattern (second term). The latter is the combination mode of Stuecker et al. (2013), which is embedded in the seasonally stratified view in Eq. (2) (e.g., Wang et al. 2003). The combination mode offers an explanation for the abrupt transition of ENSO-related atmospheric variability (e.g., the rapid onset of the NWP AAC from late fall to early
winter), even though the evolution of Niño-3.4 SST is much smoother. Our derivation of the C-mode index in Eq. (4) from S-EOF1 resolves the debate that questions the mode’s physical existence and identity (Li et al. 2016; Stuecker et al. 2016).

4. Seasonal pattern modulation

While Stuecker et al. (2015) emphasized the SST forcing from the equatorial Pacific for the combination mode, it does not need to be limited to equatorial Pacific SST variability but can come from other ocean basins. For example, many studies have identified the tropical Indian Ocean warming as a major driver for the summer NWP AAC as JJA(1) Niño-3.4 is not significantly correlated with that in the preceding winter (see the review of Xie et al. 2016).

While the C-mode view Eq. (1) is vague on how the annual cycle physically interacts with ENSO, the seasonal stratified view Eq. (2) explicitly states that the atmospheric response is in quasi-equilibrium with interannual SST variations. This quasi-equilibrium approximation allows us to identify SST forcing from various regions by using a realistic atmospheric GCM. Generally, the seasonality of atmospheric response to ENSO is due to seasonal variations in the magnitude and spatial structure of SST anomalies, and to the seasonal cycle in the climatological background.

a. Constant El Niño run

To isolate the last effect of the seasonally varying mean state, we carry out an idealized experiment using CAMS by imposing an El Niño SST pattern over the equatorial Pacific that does not vary in time. The SST pattern is based on the N-EOF1 for tropical Pacific SST variability and retains only the region of positive SST anomalies over the equatorial Pacific (Fig. 4a) as in Stuecker et al. (2015). While Stuecker et al. (2015) vary the amplitude of this pattern in sinusoidal functions of time, we further simply by setting the amplitude constant in time (Niño-3.4 = 1.32°C). The simplification isolates the seasonal modulation effect by the climatology, and the seasonal variations in atmospheric response are due entirely to the interaction with the seasonally varying background state.
Figure 4 shows the CAM5 response to the time-constant El Niño in precipitation and 925-hPa wind velocity. While the SST anomalies are confined to the equatorial Pacific, marked atmospheric response is found over the entire Indo-Pacific oceans. Precipitation increases over the equatorial Pacific as a direct response to the ocean warming while decreasing over the equatorial Indian Ocean and Maritime Continent where there is no local SST forcing. The surface wind response over the equatorial Indo-Pacific represents a slowdown of the Walker circulation, with anomalous westerlies in the Pacific that act to amplify the ocean warming by flattening the thermocline and zonal advection, and anomalous easterlies in the Indian Ocean that would lift the thermocline off the west coast of Indonesia and excite the Indian Ocean dipole (IOD; Saji et al. 1999) mode. These equatorial wind anomalies display considerable seasonal variations even though the SST forcing is constant. Southeasterly alongshore wind anomalies develop in March–August off the west coast of Indonesia (Figs. 4b,e) in favor of IOD development (Yang et al. 2015).

Over the equatorial Pacific, the anomalous westerlies are centered on the equator for July–November but displaced south from December to May (Fig. 4). This southward displacement reduces the efficiency of forcing oceanic equatorial waves, contributing to the rapid decay of ENSO during these seasons (Harrison and Vecchi 1999; McGregor et al. 2012). Our experiment shows that this change in atmospheric feedback onto El Niño warming is due to the seasonal variations in the background (Spencer 2004; Stuecker et al. 2015). The anomalous westerlies in the equatorial Pacific show an eastward extension from JJA to MAM much as in the observed S-EOF1 (Fig. 2) as the annual climatological ocean warming brings the cool eastern equatorial Pacific closer to the convective threshold (Zheng et al. 2016). Indeed, the precipitation increase over the eastern Pacific is centered on the equator in MAM while displaced north in other seasons because of the latitudinal asymmetry in

**FIG. 6.** Observed El Niño anomalies composites: (left) SST (shading, °C) and (right) rainfall (shading, mm day$^{-1}$) and 925-hPa wind (vectors, m s$^{-1}$). (left) The black contours denote the boundary of tropical Pacific SST anomalies imposed in the constant El Niño and POGA runs (Fig. 4a).
the mean SST. In reality, El Niño begins to decay in MAM(1) so the eastward extension of precipitation and westerly wind anomalies into the eastern equatorial Pacific takes place only for extremely strong El Niño events (Cai et al. 2014). This could change as ocean warming under increasing greenhouse gas forcing is projected to be locally enhanced in the eastern equatorial Pacific so El Niño events with extreme eastward intrusion of wind and rainfall anomalies might become more frequent (Power et al. 2013; Zheng et al. 2016).

We can derive the contributions of eastern equatorial Pacific SST to N-EOF1 and N-EOF2 (C-mode) from the constant El Niño experiment. From Eq. (4), we obtain

$$R = A_1(\omega, t)N(t) \left\{ \overline{R}(x, y) + \frac{A_1(\omega, t)}{A_1(\omega, t)} R_2(x, y) \right\}.$$

(5)

With $A_1(\omega, t)N(t) =$ constant in our constant El Niño experiment, the annual-mean response gives rise to N-EOF1 $\overline{R}(x, y)$ while the monthly deviations (Fig. 5, color shading) yield the C-mode $R_2(x, y)$ with appropriate scaling in time. The annual-mean wind-velocity response over the Pacific indeed resembles the N-EOF1 (Fig. 5b vs Fig. 3 left column). Over the Indian Ocean, the annual-mean response in the constant El Niño run is nearly antisymmetric about the equator in low-level wind and precipitation (Figs. 5a,b), despite a nearly symmetric SST forcing over the Pacific. While rainfall increases off the south coast of Asia, it decreases on the other side of the equator. The annual-mean wind anomalies near the surface feature a C-shaped pattern with southerly cross-equatorial flow over the Indian Ocean.

The largest seasonal variations in atmospheric response to the constant equatorial Pacific warming are found over the monsoonal oceans from the north Indian Ocean to NWP (Fig. 5), where the climatological winds vary from northeasterly in winter to southwesterly in summer. In fact, the atmospheric response in this region is not uniquely defined but varies from one season to another. In JJA, an anomalous cyclonic circulation occupies the north Indian Ocean–NWP region with increased rainfall (Fig. 4b), possibly because active convective feedback (Xie et al. 2009) amplifies the warm Rossby wave response to El Niño (Fig. 5c; Gill 1980). In MAM, the response over this region is replaced with decreased rainfall and a weak anomalous

![Rainfall (shading, mm day$^{-1}$) and 925-hPa wind (vectors, m s$^{-1}$) response in (left) POGA and (middle) GOGA runs, and (right) their difference, which measures the SST forcing effect outside the equatorial Pacific.](image)
anticyclonic circulation as the climatological rainband and hence convective feedback weakens north of the equator (Fig. 4e).

Figure 1b shows the evolution of SLP response over the NWP to a constant El Niño. Without temporal variations in El Niño SST forcing, our result shows a rapid transition of NWP SLP from negative to positive values in November due to the seasonal modulation by the mean state. In support of Stuecker et al. (2015), the C-mode of the atmospheric response to equatorial Pacific SST anomalies indeed contributes to the abrupt onset of AAC over the NWP at the peak phase of El Niño. The AAC in the response to El Niño itself (equatorial Pacific SST) is too weak, however, indicating that SST anomalies elsewhere are important to boost AAC to a realistic magnitude.

b. SST anomalies beyond the equatorial Pacific

SST anomalies associated with El Niño are most pronounced over but not confined to the equatorial Pacific (Fig. 6). For example, the tropical Indian Ocean and South China Sea warm following El Niño (Klein et al. 1999; Du et al. 2009). El Niño–induced atmospheric response $R(x, y, t)$ may be decomposed into

$$R = R_P + R_C,$$

where $R_P$ and $R_C$ are the response to SST variability in and outside the equatorial Pacific, respectively. Results from the GOGA and POGA runs approximate $R$ and $R_P$, respectively. As SST anomalies outside the equatorial Pacific are ultimately induced by El Niño, $R_C$ may be viewed as the ocean capacitor effect. The C-mode of Stuecker et al. (2015) is part of $R_P$, which vanishes in post–El Niño summer JJA(1) with equatorial Pacific SST anomalies (Fig. 1a, black).

The GOGA run captures major features of observed evolution of atmospheric anomalies, including the transition from the JJA(0) anomalous cyclonic to anticyclonic circulation from DJF(1) onward through JJA(1) over the tropical NWP (Fig. 7, middle column). The simulated AAC in GOGA is somewhat weak and displaced northward compared to observations (Fig. 6). In the metric of NWP SLP, both the POGA and GOGA produce the abrupt onset of high pressure anomalies from September(0) to January(1) (Fig. 8). The transition is slightly larger in magnitude in GOGA. Much of the AAC from the South China Sea to NWP is due to SST anomalies outside the equatorial Pacific during JJA(1) (Fig. 7, right column).

The capacitor effect $R_C$ outlasts the equatorial Pacific SST effect and persists into post–El Niño summers. In JJA(1) when equatorial Pacific and NWP SST anomalies are both weak (Fig. 8a), the AAC shows a secondary intensification in both observations and GOGA run (Fig. 8b), likely due to SST variability over the north Indian Ocean that also features a second peak at the same time. The JJA(1) AAC extends well into the Indian Ocean sector, and the anomalous easterlies on the south flank help sustaining the SST warming over the north Indian Ocean and South China Sea via WES feedback by reducing the southwest monsoon winds (Du et al. 2009; Kosaka et al. 2013). This supports the idea...
that IPOC is a regional coupled mode arising from positive feedback between AAC and regional SST anomalies over the Indo–NWP region (Xie et al. 2016). Indeed, the analysis of intermember spread in ensemble seasonal forecast shows that the equatorial Pacific SST and IPOC modes \([R_P \text{ and } R_C \text{ in Eq. (6)}]\) are distinct, both contributing to the uncertainty in predicting JJA atmospheric circulation anomalies over the NWP (Ma et al. 2017).

5. Summary and discussion

It is well known that both the amplitude and spatial pattern of atmospheric anomalies associated with ENSO vary in season. The structural variations from one season to another are most pronounced from the north Indian Ocean to NWP, as often represented by seasonally stratified EOF modes (e.g., Wang et al. 2003). We show that the C-mode in Eq. (4) represents an alternative, nonseasonally stratified view of structural variations in atmospheric anomalies of ENSO. Thus, the C-mode is entirely consistent with and in fact embedded in the seasonally stratified view of ENSO.

In an extreme case, the structural variations with season can be due entirely to the seasonal variations in the climatological background as demonstrated in our constant El Niño run. A remarkable prediction of the C-mode is that the interaction of slowly varying interannual variations in equatorial Pacific SST with the background seasonal cycle can lead to the abrupt onset of the AAC over the NWP at the peak of El Niño. In the constant El Niño experiment, the rapid AAC transition takes place even though the atmosphere is in quasi-equilibrium with the SST forcing that does not vary in time. The seasonal migration of deep convection and convective feedback on the circulation response appear to be important for the onset of AAC. From the constant El Niño experiment, we can reconstruct the C-mode for a hypothetical El Niño in which equatorial Pacific warming peaks in summer instead (see also Stuecker et al. 2015). The SLP anomaly over the NWP still displays a sharp transition at the peak of this hypothetical El Niño, but from anticyclonic to cyclonic (Fig. 1b, green curve), because the Northern Hemisphere is in warm and convective season during JJA (Fig. 4b).

Figure 9 is a schematic of El Niño evolution that relates the C-mode to other ocean–atmospheric processes discussed in the literature. The C-mode forced directly by El Niño warming in the equatorial Pacific triggers the onset of an AAC over the NWP at the peak of El Niño in ND(0)J(1). WES feedback under the northeast trades then amplifies the AAC over the NWP (Wang et al. 2000). The WES feedback itself does not favor either anticyclonic or cyclonic SLP anomalies, and indeed an anomalous cyclone might be possible over the NWP should El Niño peak in summer as in the above hypothetical case. The C-mode view offers a new perspective on the AAC onset by explaining why an AAC emerges in ND(0)J(1) and highlighting the importance of seasonal modulation by the mean state (Stuecker et al. 2015; see our Figs. 1b and 4).

The Walker circulation slowdown is another mechanism for the eastern Pacific warming to trigger the IPOC mode (Fig. 9). Over the Indian Ocean, the anomalous easterlies on the equator trigger the IOD mode of Bjerknes’s feedback (Saji et al. 1999) and force downwelling ocean Rossby waves that cause the ocean mixed layer to warm south of the equator (Xie et al. 2002). The Rossby wave–induced ocean warming reinforces the

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**Fig. 9.** Schematic of ocean–atmospheric processes that trigger the regional coupled IPOC mode.
anomalous easterly winds on the south flank of the NWP AAC. The wind anomalies and SST are coupled in the subsequent summer to generate positive feedback and form the regional IPOC mode in the monsoon westerly regime from the north Indian Ocean to NWP (Kosaka et al. 2013). Thus, regional SST anomalies over the Indo–western Pacific are important for the atmospheric response to El Niño, especially the AAC in the post–El Niño summer JJA (1) when SST anomalies have dissipated over the equatorial Pacific (Figs. 7, 8). Thus, recent C-mode research adds an important element to the regional IPOC framework in the literature (e.g., Wang et al. 2000; Xie et al. 2009) by highlighting the role of the seasonally varying mean state in triggering the AAC.

To the extent that the atmosphere is in quasi-equilibrium with SST variability, the success of seasonal prediction hinges on predicting ENSO and regional modes it excites (e.g., the IOD and IPOC). This requires a solid physical understanding of ENSO as well as regional modes outside the equatorial Pacific that bring about additional predictability. It remains to be explored how the C-mode view from a nonseasonally stratified perspective can offer additional predictability beyond the current practice of seasonal prediction.

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REFERENCES


