Combination Mode Dynamics of the Anomalous Northwest Pacific Anticyclone*

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ABSTRACT

Nonlinear interactions between ENSO and the western Pacific warm pool annual cycle generate an atmospheric combination mode (C-mode) of wind variability. The authors demonstrate that C-mode dynamics are responsible for the development of an anomalous low-level northwest Pacific anticyclone (NWP-AC) during El Niño events. The NWP-AC is embedded in a large-scale meridionally antisymmetric Indo-Pacific atmospheric circulation response and has been shown to exhibit large impacts on precipitation in Asia. In contrast to previous studies, the authors find the role of air–sea coupling in the Indian Ocean and northwestern Pacific only of secondary importance for the NWP-AC genesis. Moreover, the NWP-AC is clearly marked in the frequency domain with near-annual combination tones, which have been overlooked in previous Indo-Pacific climate studies. Furthermore, the authors hypothesize a positive feedback loop involving the anomalous low-level NWP-AC through El Niño and C-mode interactions: the development of the NWP-AC as a result of the C-mode acts to rapidly terminate El Niño events. The subsequent phase shift from retreating El Niño conditions toward a developing La Niña phase terminates the low-level cyclonic circulation response in the central Pacific and thus indirectly enhances the NWP-AC and allows it to persist until boreal summer. Anomalous local circulation features in the Indo-Pacific (e.g., the NWP-AC) can be considered a superposition of the quasi-symmetric linear ENSO response and the meridionally antisymmetric annual cycle modulated ENSO response (C-mode). The authors emphasize that it is not adequate to assess ENSO impacts by considering only interannual time scales. C-mode dynamics are an essential (extended) part of ENSO and result in a wide range of deterministic high-frequency variability.

1. Introduction

The northwest Pacific (NWP) is one of the key regions controlling climate variability in Asia (Wang et al. 2000; Lau and Nath 2006, 2009; Wang et al. 2013; Kosaka et al. 2013; Chowdary et al. 2013). A major regional climate feature is a large-scale anomalous anticyclonic surface circulation near the Philippine Sea and the South China Sea (Harrison and Larkin 1996; Wang et al. 1999, 2000; Wang and Zhang 2002; Li and Wang 2005). This anomalous low-level northwest Pacific anticyclone (NWP-AC; also referred to as Philippine Sea anticyclone or western North Pacific subtropical high) usually develops during the peak phase of El Niño events in December–February (DJF) and persists until Northern Hemisphere spring [March–May (MAM); Harrison and Larkin 1996] and summer [June–August (JJA)] of the following year (see NWP-AC circulation index in Fig. 1c).
Additional to its impact on Asian climate variability, it plays an important role in the phase transition process of the El Niño–Southern Oscillation (ENSO) (e.g., Weisberg and Wang 1997; Wang et al. 1999; Kug et al. 2006; Ohba and Ueda 2009; Okumura et al. 2011; McGregor et al. 2012; Stuecker et al. 2013; McGregor et al. 2013). The NWP-AC is usually associated with a meridional shift of zonal westerly surface wind anomalies from the equator to the Southern Hemisphere (e.g., McGregor et al. 2012; Stuecker et al. 2013). This southward wind shift (e.g., Harrison 1987) can contribute to the rapid termination of El Niño events by weakening the anomalous equatorial westerlies and thus generate eastward-propagating upwelling oceanic Kelvin waves (e.g., Harrison and Vecchi 1999; Spencer 2004; Lengaigne et al. 2006; Vecchi and Harrison 2006; Ohba and Ueda 2009) and more importantly by a meridionally antisymmetric modulation of the equatorial heat content recharge/discharge process (e.g., Kug et al. 2003; McGregor et al. 2012; Stuecker et al. 2013; McGregor et al. 2014a).

Several mechanisms have been proposed to explain the development and persistence of the anomalous NWP-AC during and after an El Niño event. Wang et al. (1999) proposed that the El Niño sea surface temperature (SST) anomalies cause suppressed convection over the Maritime Continent and thus lead to an anomalous low-level anticyclonic wind stress curl in the northwestern Pacific (meridionally symmetric Gill Rossby wave response; Gill 1980), which acts to deepen the thermocline locally via Ekman pumping. Furthermore, the annual cycle of the winds exhibits an important impact on the anomalous NWP-AC circulation (Wang and Zhang 2002). It has been argued that, during the development of an El Niño event, the symmetric Gill response (Gill 1980) creates anomalous low-level anticyclonic vorticity over southern Asia (Rossby wave response), which is advected by the mean southwesterly boreal summer monsoon winds to the northwestern Pacific region, thereby establishing the anomalous NWP-AC (Wang and Zhang 2002). Additional, the importance of extratropical cold air outbreaks due to anomalous low-level anticyclonic circulation over Asia to the anomalous NWP-AC development is highlighted by Wang and Zhang (2002).

When air–sea interactions are considered, the symmetric anticyclonic Rossby wave Gill response (Gill 1980) can lead to the following positive thermodynamic feedback mechanism (Fig. 16 in Wang et al. 2000): During boreal winter the established anomalous NWP-AC interacts with the climatological northeasterly trade winds, causing enhanced (decreased) wind speed on the eastern (western) flank of the anticyclone, thus inducing a local negative (positive) SST anomaly tendency. Hence, Wang et al. (2000) argue that negative SST anomalies to the east allow an eastward expansion and persistence of the NWP-AC. Furthermore, the baroclinic structure of the anomalous NWP-AC is highlighted, suggesting that its origin is likely due to anomalous tropical diabatic forcing (Wang et al. 2000).

In addition to the Maritime Continent subsidence and local air–sea interaction in the northwestern Pacific region, several other studies have suggested that the delayed tropical Indian Ocean warming after the El Niño peak phase (also known as capacitor effect) contributes to the development of the NWP-AC (Watanabe and Jin 2002, 2003; Kug et al. 2006; Xie et al. 2009; Wu et al. 2010b; Okumura et al. 2011; Yuan et al. 2012; Wang et al. 2013). According to these studies, the delayed Indian Ocean warming forces a baroclinic atmospheric Kelvin wave, which results in anomalous equatorial surface easterlies and suppressed convection in the western Pacific, thereby establishing the anomalous low-level NWP-AC circulation.

A few things are important to note: (i) the anticyclonic Rossby wave response due to the El Niño–induced subsidence over the Maritime Continent is largely symmetric with respect to the equator (Gill 1980) and centered in the Indian Ocean (Fig. 1a); (ii) the eastward extension of this Indian Ocean anticyclonic circulation is limited to approximately 140°E (Fig. 1a) because of the convectively driven equatorial central Pacific cyclonic Rossby wave pair (Fig. 1a); and (iii) the observed western Pacific anomalous circulation response is highly meridionally antisymmetric (Fig. 1b). The delayed Indian Ocean warming atmospheric Kelvin wave response is symmetric about the equator and is hence not able to explain this meridional antisymmetry.

The seasonal cycle of solar insolation provides the key meridionally antisymmetric component to the system. Furthermore, two factors affect the deviation of the annual mean state from total symmetry: (i) the difference in land–sea distribution among the hemispheres in the western Pacific and (ii) the amplitude of the annual cycle of solar insolation, which for the current phase of precession and eccentricity is considerably smaller in the Northern Hemisphere than in the Southern Hemisphere. It is important to note that the interhemispheric land–sea distribution and the annual cycle of insolation act together to create the annual cycle of winds.

Here we set out to show that the required antisymmetric mechanism is provided by the tropical warm pool annual cycle and is characterized by a combination mode (C-mode) of near-annual variability (Stuecker et al. 2013). The C-mode, which is further discussed in section 2, emerges from the seasonal modulation of ENSO-related atmospheric anomalies. The results
presented here suggest that the C-mode is the main contributor to the time evolution of the antisymmetric anticyclonic and cyclonic circulation anomalies in the Indo-Pacific during both El Niño and La Niña events. The associated near-annual time scale of the C-mode (Stuecker et al. 2013) has been overlooked in previous studies of Indo-Pacific climate. The persistence of the anomalous low-level NWP-AC until JJA of year 1 after El Niño events [JJA(1)] can only occur if the cyclonic Rossby wave response in the central and eastern Pacific has already weakened sufficiently. Local air–sea coupling and remote Indian Ocean forcing only act to amplify these processes and contribute to the persistence of the anomalous NWP-AC until JJA(1).

In this paper, we provide a combination mode dynamics explanation for the observed temporal evolution of the anomalous antisymmetric Indo-Pacific circulation. The rest of this article is organized as follows: In section 2, we discuss the observed features of the NWP-AC and the implications for its dynamics. In section 3, we explain our experiments, and the results are presented in section 4. Our conclusions are discussed in section 5.

2. The northwest Pacific anticyclone in observations

For our observational analysis we use the Niño-3.4 (N3.4) SST anomaly index (shading in Fig. 1c), which is calculated as a box average of anomalous HadISST1 (Rayner et al. 2003) SST in the equatorial central Pacific (5°S–5°N, 170°–120°W). Furthermore, we define a NWP-AC circulation index as the area-averaged anomalous ERA-40 (Uppala et al. 2005) surface wind streamfunction \( \psi \) [with \( u = \partial \psi / \partial y; v = - (\partial \psi / \partial x) \)] in the northwestern Pacific (5°–20°N, 120°–160°E; Fig. 2). In this study, we focus on a 22-yr epoch with strong ENSO variability and high quality observational data coverage (1980–2001) to improve the signal-to-noise ratio in our analysis (Fig. 1c).

Previous studies suggest that the mechanism responsible for the development of the anomalous low-level NWP-AC is closely related to the ENSO phenomenon. However, linear methods such as pattern regression (Fig. 1a) and time series cross correlations (Fig. 1c) with the N3.4 index are unable to capture its pattern and time evolution correctly. For instance, no simultaneous significant correlation (\( R = 0.04 \)) can be found between

![Fig. 1. ENSO and combination mode impacts on surface wind circulation and precipitation. Anomalies are calculated for the 1980–2001 period for HadISST1 sea surface temperatures, GPCP V2.2 precipitation, and ERA-40 surface winds. Areas where the anomalous circulation regression coefficient is significant above the 95% confidence level are nonstippled. (a) Precipitation anomalies regressed onto normalized N3.4 index (shading; mm day\(^{-1}\)) and anomalous surface wind streamfunction \( \psi \) regressed onto normalized N3.4 index (contours; 10\(^5\) m\(^2\) s\(^{-1}\)). (b) Precipitation anomalies regressed onto normalized theoretical C-mode (N3.4 \( \times \) annual cycle) index (shading; mm day\(^{-1}\)) and anomalous surface wind streamfunction \( \psi \) regressed onto normalized theoretical C-mode index (contours; 10\(^5\) m\(^2\) s\(^{-1}\)). (c) N3.4 index (shading; °C), NWP-AC \( \psi \) index (black line; \( \psi \) anomalies averaged over 5°–20°N, 120°–160°E; 10\(^5\) m\(^2\) s\(^{-1}\)) and normalized theoretical C-mode index (orange line; no units). The cross-correlation coefficients (R) at zero lag for these time series are given in the figure. El Niño and La Niña events that exhibit a relatively clear combination mode phase transition during boreal winter are indicated by vertical gray lines.]
N3.4 and the NWP-AC circulation index (Fig. 1c). Only seasonally stratified regressions with N3.4 are able to capture the correct pattern evolution (not shown).

Recently, Stuecker et al. (2013) showed that nonlinear atmospheric interactions between ENSO and the annual cycle generate a previously overlooked mode of climate variability, which is referred to as the C-mode. This C-mode is in essence an amplitude modulation of the annual cycle by ENSO (or vice versa) and it exhibits characteristic time scales (near-annual combination tones) that are clearly distinct from interannual ENSO variability (Stuecker et al. 2013). The C-mode exhibits profound impacts on tropical climate (Stuecker et al. 2013), ENSO phase locking (Stuecker et al. 2013; Stein et al. 2014), and interhemispheric sea level variability in the Pacific (Widlansky et al. 2014). The first-order approximation of the theoretical C-mode can be written as (Stuecker et al. 2013)

\[
C_{\text{mode}}(t) = ENSO(t) \cos(\omega_f t - \varphi) \tag{1}
\]

Here, \(\omega_f\) denotes the angular frequency of the annual cycle and \(\varphi\) represents a 1-month phase shift. This captures the observed phase as the warm pool annual cycle has its maximum in February and minimum in August. We use the N3.4 index to represent ENSO(t) in this study.

A regression of the anomalous ERA-40 (Uppala et al. 2005) surface wind streamfunction \(\psi\) onto the first-order approximation of the theoretical C-mode (Fig. 1b) shows that the C-mode index ["PC2 SIMPLE" in Stuecker et al. (2013)] captures the anomalous low-level NWP-AC pattern much better than the N3.4 regression (Fig. 1a). Moreover, the C-mode index correlates reasonably well with the NWP-AC index \((R = 0.46; \text{Fig. 1c})\). The correlation between the theoretical C-mode index and the anomalous NWP-AC circulation is high for active ENSO periods and goes by definition to zero for ENSO neutral years (e.g., 1980/81 and 1990/91; Fig. 1c), during which the system is characterized by a low signal-to-noise ratio. Unforced internal variability with a relatively high variance occurs during these ENSO neutral conditions (Fig. 1c), thus resulting only in a moderate correlation \((R = 0.46)\) between the theoretical C-mode index and the NWP-AC circulation when considering the whole time series.

The areas in which the linear regression coefficients of the circulation anomalies are significant above the 95% confidence level (using a two-tailed Student’s \(t\) test) are nonstippled (Figs. 1a,b). The equivalent degrees of freedom for each grid point are determined using the decorrelation time scale of the observed anomalies.

Consistent with combination mode dynamics, the C-mode index exhibits only variability at the near-annual combination tones (approximately 9- and 15-month periods; for details, refer to Stuecker et al. 2013); the NWP-AC \(\psi\) index shows power at the near-annual difference tone (corresponding to a period of approximately 15 months) (black line in Fig. 3a). However, the NWP-AC \(\psi\) index also displays power in the interannual frequency band (black line in Fig. 3a), indicating that this index is also influenced directly by ENSO. Near-annual peaks in the northwest Pacific thermocline and wind stress curl have been identified first by Wang et al. (1999). The origin of this time-scale variability, however, was explained only recently (Stuecker et al. 2013) and can be attributed to combination mode dynamics. The combination tone frequencies have been overlooked in previous studies because of the use of methods like interannual bandpass filtering (e.g., Chung et al. 2011). However, the C-mode is essential in understanding the dynamics and ascertaining the evolution of the anomalous low-level NWP-AC.

We calculate the same NWP-AC \(\psi\) index for the 500-yr preindustrial control run of the GFDL CM2.1 (Delworth et al. 2006) coupled ocean–atmosphere general circulation model (CGCM). The NWP-AC \(\psi\) spectrum in GFDL CM2.1 shows power at interannual frequencies, as well as for both the ENSO–annual cycle sum \((1 + f_E)\) and difference \((1 - f_E)\) tones (gray line in Fig. 3a). However, when we truncate the time series in 25-yr-long time slices (similar length to the observational time series), we find that the spectra of the individual time slices exhibit sometimes both tones, sometimes only one of them, and sometimes none (Fig. 3b). This indicates that the missing sum tone in the ERA-40 NWP-AC \(\psi\) spectrum might be attributed to the relatively low signal-to-noise ratio in the observational epoch compared to a long CGCM experiment with relatively strong ENSO amplitude and hence large signal-to-noise ratio and...
maybe also to the representation of nonlinear processes in the AGCM (Stuecker et al. 2013).

To assess the precipitation pattern related to the different observed modes of variability in the Indo-Pacific, we utilize the GPCP V2.2 globally gridded dataset (Adler et al. 2003). The ENSO precipitation response is predominantly characterized by positive anomalies in the tropical central and eastern Pacific and negative anomalies over the Maritime Continent, Australia, and tropical South America (Fig. 1a). In contrast, the combination mode is associated with a different precipitation response. We observe a band of enhanced precipitation on the equator, in the eastern Indian Ocean, and a positive precipitation anomaly south of the equator in the central Pacific (Fig. 1b). This precipitation anomaly pattern is associated with so-called zonal South Pacific convergence zone (SPCZ) events (Cai et al. 2012; Stuecker et al. 2013).

Previous studies of the NWP-AC have focused predominantly on the positive ENSO phase (El Niño events) when discussing the NWP-AC feature. However, the NWP-AC is not limited to the Pacific basin. The Indian Ocean is characterized by an anomalous low-level anticyclonic circulation (southern Indian Ocean anticyclone (SIO-AC); e.g., Yamanaka et al. 2009) that is stronger than its counterpart in the northern Indian Ocean (Fig. 1b). This anomalous SIO-AC circulation brings extratropical air to northwestern Australia and results in reduced local precipitation.

3. Model experiments

We design a suite of general circulation model (GCM) experiments, which includes idealized ENSO SST forcings, to investigate the impact of nonlinear atmospheric interaction between ENSO and the annual cycle on the anomalous Indo-Pacific circulation (including but not limited to the NWP-AC) as well as the quantitative role of air–sea coupling.

We utilize both atmospheric general circulation model (AGCM) experiments (SST boundary conditions are prescribed everywhere) as well as AGCM experiments that are partially coupled (PARCP) in some regions to
a thermodynamic slab ocean model (SOM). A number of previous studies employed the partial coupling technique (sometimes referred to as a pacemaker) to ascertain circulation and precipitation responses to a given forcing (e.g., ENSO). For this type of experiment, fixed SSTs or heat fluxes are prescribed in certain regions, while either a thermodynamic SOM (Lau et al. 2005; Lau and Nath 2006; Lee et al. 2008; Lu et al. 2011; Lee et al. 2013; McGregor et al. 2014b) or a full dynamical ocean model (Kucharski et al. 2007; Kosaka et al. 2013; Kosaka and Xie 2013) is used in the remaining model domain.

The idealized forcing spatial pattern for these experiments is identified by simply regressing the HadISST1 (Rayner et al. 2003) anomalies onto the normalized N3.4 index to obtain the linear ENSO spatial pattern (Fig. 2). Only the positive loading in the forcing region from 170°E to the coast of South America situated in the tropical band (25°S–25°N) is utilized. We remove the negative “boomerang” loading and any ENSO associated anomalies outside of this forcing region. The ENSO pattern is then multiplied by a simple sinusoidal time series to obtain the time evolution of the ENSO-related SST anomalies. Thus, only the tropical eastern Pacific positive (negative) SST anomalies are retained for an El Niño (La Niña) event. By using a perfect sinusoidal ENSO forcing, we are able to unambiguously separate features of the circulation response in the frequency domain: The symmetric ENSO response has the same features of the circulation response in the frequency domain. The symmetric ENSO response will be characterized by near-annual combination tones (1/fE and 1 + fE) (Stuecker et al. 2013) and higher-order tones with frequencies such as 1 ± 2fE and 2 ± fE. The sinusoidal forcing will result in narrow combination tone peaks that are easy to identify. Note that the relatively broad combination tone spectral peaks in the observations result from the broad non-sinusoidal ENSO peak (Fig. 3a).

For all our experiments we use the atmospheric component CAM4 of the Community Earth System Model (CESM) version 1.0.3 (Neale et al. 2013) in a spectral T42 horizontal resolution and 26 vertical levels. The details of the experiments are described in the following subsections.

a. Atmospheric general circulation model experiments

The aforementioned SST anomalies are added in the forcing region (from 170°E to South America and 25°S–25°N) to an SST and sea ice climatology based on the 1982–2001 observations (Hurrell et al. 2008). We design the following AGCM experiments (also see Table 1):

(i) A 100-yr AGCM run using a sinusoidal 2.5-yr period ENSO SST anomaly forcing and modern day SST annual cycle: The forcing phase is chosen so that the peak El Niño phase will occur once aligned with December and once with June during any 5-yr cycle (the peak La Niña phase will occur during March and September). After the run, we remove the long-term monthly means to obtain the anomalies. This experiment allows us to analyze the sensitivity of the anomalous circulation and precipitation response in relation to both the ENSO phase (El Niño or La Niña) and the phase of the annual cycle. The 2.5-yr ENSO period is motivated by reproducing the observed N3.4 time evolution during an El Niño event (e.g., the SST anomalies peak during December and have decayed by the following boreal summer season) while still being able to clearly separate the combination tone peaks (1 − fE = 0.6 yr⁻¹ and 1 + fE = 1.4 yr⁻¹) from the forcing peak (fE = 0.4 yr⁻¹). The 100-yr integration gives us 40 idealized ENSO cycles. It is noted that, because the La Niña events do not peak at a similar time to that seen in the observations, the results may be less realistic. However, we still expect these experiment results to highlight the C-mode dynamics during La Niña events, albeit with shifted phase.

(ii) A 120-yr AGCM run using a sinusoidal 4-yr period ENSO SST anomaly forcing and modern day SST annual cycle: Because the ENSO forcing is exactly 4 yr, all El Niño and La Niña peak phases are aligned with the boreal winter season. This experiment allows us to test (i) if the anomalous NWP-AC can persists until JJA(1) if there are still large positive SST anomalies present in the equatorial eastern Pacific and (ii) the C-mode dynamics when La Niña events have the correct peak phase. The 120-yr integration will result in 30 ENSO cycle ensemble members.

(iii) A 30-yr AGCM run in which we force the model with repeated idealized 1997–99 eastern tropical Pacific SST anomalies (hence 10 realizations of the ENSO cycle) and modern day SST annual cycle everywhere: Note that these simulations utilize the

<table>
<thead>
<tr>
<th>Model</th>
<th>ENSO forcing period</th>
<th>Integration time</th>
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<tbody>
<tr>
<td>AGCM</td>
<td>2.5-yr sinusoidal</td>
<td>100 yr (40 cycles)</td>
</tr>
<tr>
<td>PARCP</td>
<td>2.5-yr sinusoidal</td>
<td>100 yr (40 cycles)</td>
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<tr>
<td>AGCM</td>
<td>4-yr sinusoidal</td>
<td>120 yr (30 cycles)</td>
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<tr>
<td>PARCP</td>
<td>4-yr sinusoidal</td>
<td>120 yr (30 cycles)</td>
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<tr>
<td>AGCM</td>
<td>Observed 1997–99 SST anomalies</td>
<td>30 yr (10 cycles)</td>
</tr>
<tr>
<td>PARCP</td>
<td>Observed 1997–99 SST anomalies</td>
<td>30 yr (10 cycles)</td>
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same idealized ENSO spatial pattern (Fig. 2) as the previous experiments. The time evolution of the SST anomalies is obtained by multiplying the idealized spatial pattern with the observed N3.4 index for 1997–99. The anomalous low-level circulation response will be compared to the ERA-40 observations.

b. Partially coupled experiments

Atmospheric models have significantly improved in capturing the anomalous circulation and precipitation response to interannual forcing (e.g., Song and Zhou 2014). However, substantial biases between modeled and observed variables still persist; the parameterization of convection is a major contribution in determining model skill. Furthermore, sea surface temperatures are treated as a fixed boundary condition in traditional AGCM runs and hence are not able to adjust to the heat fluxes. Because the observed SSTs are partly forcing and partly solution in the coupled system, a number of previous studies (Lau et al. 2005; Lau and Nath 2006; Lee et al. 2008; Kucharski et al. 2007; Lu et al. 2011; Lee et al. 2013; Kosaka et al. 2013; Kosaka and Xie 2013) utilized the PARCP technique, in which SSTs (or associated heat fluxes) are prescribed in parts of the model domain while a coupling with an underlying ocean (thermodynamic and/or dynamic) is allowed in the rest of the domain.

In our experimental design, we utilize the same idealized ENSO spatial pattern as in the previous experiments and prescribe SSTs from the date line (180°) to the South American coast within a narrow tropical band (15°S–15°N); in the rest of the domain, we use a thermodynamic SOM. Further, we use a “buffer zone” (170°E–180°, 25°–15°S and 15°–25°N), in which we gradually merge the prescribed and calculated SSTs to avoid unphysical sharp SST gradients (Fig. 2).

We couple the SOM framework of Kiehl et al. (2006) to CAM4 to obtain heat fluxes based on the observed SST climatology. CAM4 is first run in spectral T42 resolution only with the modern SST climatology (Hurrell et al. 2008) as boundary conditions for 30 yr to calculate the climatological net heat fluxes $F_{\text{net}}$. Next, we calculate the climatological $Q_{\text{clim}}^\text{flx}$ term, which includes the SST damping time scale (Kiehl et al. 2006),

$$\frac{d\text{SST}_{\text{clim}}}{dt} = \frac{(F_{\text{net}} - Q_{\text{flx}}^\text{clim})}{\rho c_p h_{\text{mix}}},$$

where $F_{\text{net}}$ is the surface net energy balance at each time step. We design the following PARCP experiments (also refer to Table 1) to test the effect of air–sea coupling on the anomalous Indo-Pacific response:

(i) A 100-yr PARCP run using a sinusoidal 2.5-yr period ENSO SST anomaly forcing and modern day SST annual cycle: The phase is chosen so that the peak El Niño phase will occur once aligned with December and once with June during any 5-yr cycle (the peak La Niña phase will occur during March and September).

(ii) A 120-yr PARCP run using a sinusoidal 4-yr period ENSO SST anomaly forcing and modern day SST annual cycle: As the ENSO forcing is exactly 4 yr, all El Niño and La Niña peak phases are aligned with the boreal winter season.

(iii) A 30-yr PARCP run in which we force the model with repeated idealized 1997–99 eastern tropical Pacific SST anomalies and modern day SST annual cycle everywhere.

Throughout this article, SST denotes sea surface temperatures for the observations (Fig. 1) and the radiative surface temperatures (over ocean and land) for the model experiment output. Additional information about our PARCP framework can also be found in McGregor et al. (2014b).

We use a two-tailed Student’s $t$ test to test our composite analysis against the null hypothesis that the anomalous circulation is not caused by combination mode dynamics. To sample the anomalous internal circulation variance, we integrate the model for 50 yr each for both the AGCM and PARCP setup without ENSO forcing (only the annual cycle is prescribed). The areas where the anomalous circulation is statistically significant above the 95% confidence level are nonstippled in the following composite map figures. Further, solid lines (in contrast to dashed lines) for the circulation anomaly indices in the following line plot composites indicate statistical significance above the 90% confidence level. We also use a two-tailed Student’s $t$ test for the model experiment regression coefficient maps. The areas where the respective
regression coefficients are statistically significant above the 95% confidence level are again nonstippled.

4. Model results

a. 2.5-yr sine ENSO experiments

We composite the time evolution of the N3.4 SST time series (solid black line), the normalized theoretical C-mode time series (orange line), the NWP-AC $\psi$ index for the AGCM experiment (blue line), and the NWP-AC $\psi$ index for the PARCP experiment (red line) for the following four cases (20 member averages each):

(i) El Niño events with the peak phase occurring in December (Fig. 4a);

(ii) El Niño events with the peak phase occurring in June (Fig. 4b);

(iii) La Niña events with the peak phase occurring in March (Fig. 5a); and

(iv) La Niña events with the peak phase occurring in September (Fig. 5b).

First, it is evident that the time series of the first-order approximation to the theoretical C-mode (N3.4 × annual cycle; orange line) captures the time evolution of the NWP-AC circulation indices very well for both experiments (AGCM and PARCP) and all four cases (Figs. 4, 5). These three time series exhibit a high correlation, while no or very little correlation exists between the NWP-AC circulation indices and the direct N3.4 SST forcing (Table 2).

![Fig. 4. Index time evolution composites (20 members each) for the 2.5-yr sine ENSO experiments. Shown are the N3.4 SST forcing index (solid black; °C), the normalized theoretical C-mode (orange), the NWP-AC $\psi$ index for the AGCM experiment (blue; $10^6$ m$^2$ s$^{-1}$), and the NWP-AC $\psi$ index for the PARCP experiment (red; $10^6$ m$^2$ s$^{-1}$). Solid lines indicate when the circulation is statistically significant above the 90% confidence level. Because of the sinusoidal 2.5-yr ENSO period, the events can be separated and composited into the following groups: (a) composite for the El Niño events with a peak in December and (b) composite for the El Niño events with a peak in June.](image_url)

![Fig. 5. Index time evolution composites (20 members each) for the 2.5-yr sine ENSO experiments. Shown are the N3.4 SST forcing index (solid black; °C), the normalized theoretical C-mode (orange), the NWP-AC $\psi$ index for the AGCM experiment (blue; $10^6$ m$^2$ s$^{-1}$), and the NWP-AC $\psi$ index for the PARCP experiment (red; $10^6$ m$^2$ s$^{-1}$). Solid lines indicate when the circulation is statistically significant above the 90% confidence level. Because of the sinusoidal 2.5-yr ENSO period, the events can be separated and composited into the following groups: (a) composite for the La Niña events with a peak in March and (b) composite for the La Niña events with a peak in September.](image_url)
For El Niño events with a winter peak (Fig. 4a), as generally observed (Fig. 1c), the theoretical C-mode time series predicts an anomalous low-level NWP-C during the developing boreal summer [JJA(0)] and fall [SON(0)] and an anomalous low-level NWP-AC circulation during the peak phase (DJF) and decaying spring [MAM(1)]. Negative La Niña–related SST anomalies during the following late boreal summer combined with the Northern Hemisphere summer annual cycle phase (as seen in observations: Fig. 1c) result in a return to the NWP-AC circulation. Both the AGCM (blue line) and the PARCP (red line) experiment NWP-AC indices follow the theoretical model (orange line) very well. The internal unforced low-level circulation variability in the NWP region is largest during boreal winter, thus resulting in a low deterministic signal-to-noise ratio during DJF for the AGCM experiment. Also note that the air–sea interaction results in an amplification of the anomalous NWP-AC. However, the rapid phase transition from cyclonic to anticyclonic northwest Pacific surface circulation is clearly initiated by the atmospheric response alone, as is apparent in the AGCM experiment. The skill of the theoretical C-mode as a predictor for the anomalous NWP-AC circulation is smallest when the theoretical C-mode is near zero. The model indices are statistically insignificant (dashed blue and red lines) when either or both the theoretical C-mode is smallest and/or the internal variability is largest (Fig. 4). This is exactly what we expect from our combination mode hypothesis.

We also find strong agreement of the regression pattern for the normalized theoretical C-mode time series and the normalized NWP-AC indices (Fig. 6 for the PARCP experiment). Only the regression coefficients for the PARCP experiment are shown (Fig. 6), as the AGCM experiment pattern exhibits a very similar spatial structure but with slightly reduced amplitude.

The antisymmetric surface circulation (Figs. 6b,c) is accompanied by enhanced low-level convergence (positive velocity potential χ) and positive precipitation anomalies over both the equatorial Indian Ocean and the Southern Hemisphere part of the Maritime Continent and the South Pacific convergence zone (Figs. 6e,f). This is contrasted by negative precipitation anomalies and low-level divergence (negative velocity potential χ) over the Northern Hemisphere part of the Maritime Continent and the northwestern Pacific region (Figs. 6e,f). In contrast, the normalized N3.4 regressions show a mostly quasi-symmetric circulation pattern, resembling the classic Gill response to an anomalous equatorial central Pacific heating source (Figs. 6a,d).

However, the model experiments exhibit a negative precipitation anomaly associated with the C-mode centered near the date line and ~10°S (Figs. 6e,f), which we do not find in the observations (Fig. 1b). While the maximum C-mode positive precipitation anomaly is found east of the date line in the observations, this maximum is shifted westward, such that it overlays Papua New Guinea in all the experiments. The longitudinally shifted heating source explains minor differences in the associated circulation pattern. This disparity between observations and model may be attributed to our experimental setup and model biases in the mean state and more important in the annual cycle of precipitation and circulation fields. The area close to the date line and the equator is part of the forcing region and the SST anomalies are set to zero here (Figs. 2, 7, 8), thereby reducing the antisymmetric C-mode response.

Nevertheless, the model is able to capture the characteristic interhemispheric precipitation anomaly of the C-mode [mostly positive (negative) anomalies in the Southern Hemisphere (Northern Hemisphere) part of the equatorial Pacific] and a realistic spatial and temporal NWP-AC evolution (Figs. 4, 5, 6b,c,e,f).

Seasonally stratified composites of the anomaly fields (shown here are SST anomalies and anomalous surface ψ) are the traditional way of visualizing the seasonal evolution of an El Niño event (e.g., Alexander et al. 2013) (Figs. 7, 8). We find a distinctively different circulation response to the SST anomaly forcing depending on the phase of the annual cycle (left versus right columns in Figs. 7, 8), as shown previously in the line composites (Figs. 4, 5). During SON(0) of the composite El Niño, we observe the quasi-symmetric anomalous low-level circulation response to the prescribed SST anomalies (Figs. 7a, 8a). In contrast, the circulation is highly antisymmetric during the DJF peak (Figs. 7b, 8b) and following MAM(1) (Figs. 7c, 8c). It features an enhanced low-level cyclonic circulation in the SPCZ, which acts to weaken its Northern Hemisphere counterpart and allows the anomalous NWP-AC to strengthen and expand eastward. The antisymmetric circulation during MAM(1) (Figs. 7c, 8c) is very similar to the observed C-mode circulation (Fig. 1b).
If El Niño events in nature would occur with a boreal summer peak (Fig. 4b), the temporal evolution of the northwest Pacific circulation response would be virtually the opposite of those events with a winter peak. In the El Niño summer peak case, the C-mode response features a rapid transition from NWP-AC to NWP-C circulation during the event peak phase (Fig. 4b).

We also find that the PARCP experiment (Fig. 8) reproduces the observed delayed warming of the Indian Ocean, the warming on the western flank of the NWP-AC and the cooling on its eastern flank. However, no anomalous NWP-AC develops during the peak and decaying phases of the El Niño boreal summer peak case (Figs. 8e–h), in contrast to the El Niño boreal winter peak case (Figs. 8a–d). This shows that the correct annual cycle phase is the necessary condition for the antisymmetric circulation. Importantly, while air–sea interactions act as an amplifier of the NWP-AC circulation, the formation of the NWP-AC can occur independently. In the AGCM experiment without any air–sea interactions, we observe nearly no anomalous NWP-AC circulation persisting until JJA(1) at the 95% significance level (Fig. 7d). During JJA(1), the ENSO forcing is near zero in this experiment; thus, atmospheric noise dominates the region.

Our theoretical model for the C-mode (orange line) also predicts the correct temporal evolution of the northwest Pacific circulation (blue and red lines) for the La Niña phase (Fig. 5). Our experiment shows that a negative annual cycle phase (cold Southern Hemisphere) combined with a negative ENSO phase (La Niña) results
in the NWP-AC circulation during Northern Hemisphere summer (Fig. 5), as we also see in observations (Fig. 1c).

These results have an important implication: The fact that we observe (Fig. 1c) a cyclonic northwest Pacific circulation during the developing summer of an El Niño event [JJA(0)] and the NWP-AC circulation during the decaying summer [JJA(1)] is augmented by the forcing N3.4 SST asymmetry between the summer of the developing and the boreal summer of the decaying year. For instance, observed El Niño events typically start to develop during Northern Hemisphere spring [MAM(0)] and by JJA(0) already exhibit a relatively strong positive N3.4 SST anomaly. Hence, the northwestern Pacific region is dominated by the western edge of the symmetric cyclonic Rossby wave response and the anticyclonic Rossby wave response to the Maritime Continent subsidence is mostly confined to the Indian Ocean basin. The anomalous cyclonic surface circulation is accurately captured by the experiments.
captured by the theoretical C-mode, as both the El Niño phase and annual cycle phase are given (Fig. 4a).

In contrast, during JJA(1) of an El Niño event with observed time evolution (boreal winter peak), we usually already have negative SST anomalies in the N3.4 region (Fig. 1c). Hence, the symmetric cyclonic Rossby wave response to the negative N3.4 SST anomalies, the antisymmetric combination mode response, the delayed Indian Ocean warming-induced atmospheric Kelvin wave, and local air–sea interaction all amplify the anticyclonic northwest Pacific circulation.

We identify a possible positive feedback loop that connects the combination mode, the expected rapid termination of an El Niño event and the anomalous low-level NWP-AC circulation: During DJF and MAM(1), the annual cycle of SST has its maximum in the Southern Hemisphere. The antisymmetric response to SST anomalies results in a weakening of the Northern Hemisphere wave.
anomalous central Pacific cyclone and hence allows the expansion of the anomalous anticyclonic circulation in the northern Indian Ocean to the northwestern Pacific region. In the Southern Hemisphere, the diabatic heating in the South Pacific convergence zone prevents the expansion of the SIO anticyclone to the east thereby limiting its impact to Western Australia. Once the NWP-AC is established during this season, it accelerates the El Niño event termination by triggering an upwelling oceanic Kelvin wave (e.g., Harrison and Vecchi 1999; Spencer 2004; Lengaigne et al. 2006; Vecchi and Harrison 2006; Ohba and Ueda 2009) and promoting antisymmetric zonally averaged upper-oceanic heat content discharge (e.g., Kug et al. 2003; McGregor et al. 2012; Stuecker et al. 2013; McGregor et al. 2014a). This results in a rapid decrease of SST anomalies in the N3.4 region, thereby weakening the atmospheric cyclonic Rossby wave pair (or even already establishing a symmetric anticyclone pair induced by negative N3.4 SST anomalies). As soon as the Northern Hemisphere anomalous cyclone has weakened sufficiently, both the delayed Indian Ocean warming and local air–sea interactions work together to prolong the NWP-AC and promote its westward expansion until JJA(1).

The combination of symmetric and antisymmetric diabatic forcing explains why we observe both the interannual forcing frequency and near-annual combination tone frequencies in the NWP-AC spectrum (Fig. 3). As we used a purely sinusoidal ENSO forcing frequency ($f_E$) in our experiments, we are able to see the combination tones as clear identifiable spectral peaks in the model experiments (Fig. 9). The NWP-AC index in the experiments is dominated by combination tone variability with only relatively weak power at the symmetric forcing frequency ($f_E$) and the first overtone ($2f_E$).
simulated NWP-AC circulation indices for our sinusoidal experiments exhibit not only the first-order combination tone frequencies ($1 - f_E$ and $1 + f_E$) but also higher-order combination tones (all statistically significant peaks are marked and labeled in Fig. 9b).

b. 4-yr sine ENSO experiments

The 4-yr sine ENSO experiment allows us to demonstrate that the observed persistence of the NWP-AC circulation until JJA(1) is not possible if strong positive SST anomalies are still present in the N3.4 region during that time, thereby supporting the identified positive feedback mechanism that accelerates El Niño event termination. The phase of the N3.4 SST forcing in these experiments is chosen such that both the El Niño and La Niña peak phases occur during Northern Hemisphere winter (Fig. 10). The rapid transition from anomalous low-level NWP-C to NWP-AC circulation associated with C-mode dynamics during DJF and MAM(1) is clearly demonstrated (Fig. 10a). However, the NWP-AC circulation rapidly terminates before the summer season of the decaying year [JJA(1)], as predicted by our theory (orange line). Even if air–sea coupling is present (red line), the NWP-AC circulation is only prolonged by approximately 1 month and returns to cyclonic circulation by the end of June. Again, most power for the NWP-AC indices is concentrated at combination tone frequencies (Fig. 9d). As in the 2.5-yr sine ENSO experiments, the theoretical C-mode time series exhibits high correlation with the NWP-AC indices and no or very little correlation with N3.4 (Table 3).

c. Repeated 1997–99 idealized forcing experiments

After introducing our theoretical framework with idealized sinusoidal ENSO experiments, we conduct additional AGCM and PARCP experiments in which we repeat an idealized 1997–99 ENSO cycle and compare the model anomalous low-level circulation with the anomalous low-level ERA-40 circulation observations (Fig. 11). We find very high correlations between the theoretical C-mode, the modeled circulation, and the observations (Table 4). It is clearly shown that the La Niña state in boreal summer 1998 contributes to the prolonged anomalous low-level NWP-AC circulation (Fig. 11).

5. Discussion and conclusions

Our study demonstrates that the Indo-Pacific circulation and rainfall response to ENSO can be separated into a meridionally quasi-symmetric part (independent of the annual cycle) and an antisymmetric part (resulting from ENSO interaction with the annual cycle). In the frequency domain, the symmetric response is characterized by the interannual forcing frequency ($f_E$) and its first overtone ($2f_E$). In contrast, the antisymmetric response is characterized by power at combination tone frequencies (Fig. 9). Anomalous local circulation features in the

![Fig. 10. Index time evolution composites (30 members each) for the 4-yr sine ENSO experiments. Shown are the N3.4 SST forcing index (solid black; °C), the normalized theoretical C-mode (orange), the NWP-AC $\psi$ index for the AGCM experiment (blue; $10^6$ m$^2$ s$^{-1}$), and the NWP-AC $\psi$ index for the PARCP experiment (red; $10^6$ m$^2$ s$^{-1}$). Solid lines indicate when the circulation is statistically significant above the 90% confidence level. Because of the 4-yr integer period, all El Niño and La Niña events peak during the Northern Hemisphere winter. (a) Composite for the El Niño events. (b) Composite for the La Niña events.](http://journals.ametsoc.org/jcli/article-pdf/28/3/1093/4050222/jcli-d-14-00225_1.pdf)

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Indo-Pacific can be considered as a superposition of the quasi-symmetric linear ENSO response (Fig. 1a) and the antisymmetric annual cycle modulated ENSO response (Fig. 1b; combination mode). We show in this study that the combination mode response is the dominant factor in the NWP region (e.g., Fig. 9). It is important to note that linear ENSO analysis alone fails to capture the essential pattern, time evolution and spectral characteristics associated with the anomalous low-level northwest Pacific anticyclone (NWP-AC) circulation.

The nonlinear interactions between ENSO and the annual cycle (C-mode) are essential in generating the anomalous observed NWP-AC and southern Indian Ocean anticyclone (SIO-AC) circulation. If the El Niño peak phase is aligned with Northern Hemisphere summer (JJA), the anomalous NWP-AC during the El Niño decaying phase cannot develop. An anomalous low-level cyclonic circulation develops in the northwestern Pacific instead.

Furthermore, we identify a possible positive feedback loop (Fig. 12) that links the initial combination mode triggered development of the anomalous low-level NWP-AC to the rapid El Niño event termination and the subsequent persistence of the circulation anomalies until JJA(1).

The atmospheric response to different ENSO phases is shown to be nonlinear (Hoerling et al. 1997), with McGregor et al. (2013), for instance, finding that in observations the response of equatorial wind anomalies are more sensitive to positive SST anomalies than to negative SST anomalies given a similar SST anomaly magnitude. This asymmetry is not very apparent in the northwest Pacific anomalous low-level circulation response (Fig. 10) in the AGCM simulations. In contrast, the PARCP experiments show an ENSO phase asymmetry in the magnitude of the northwest Pacific anomalous low-level circulation response prior to the event peak, with El Niño events exhibiting a stronger response than La Niña events (Fig. 10). While this simulated asymmetry does not appear to be as large as that reported in reanalysis products (McGregor et al. 2013), our simulations suggest a role for air–sea coupling in this asymmetry. It is also important to note that the positive feedback we identified for the anomalous northwest Pacific circulation through ENSO interaction with the C-mode does not result in a rapid La Niña event termination in observations, even though an anomalous low-level NWP-C is generated during boreal winter/spring of a La Niña event in our idealized experiments (Figs. 5b, 10b). Additional to the aforementioned nonlinear atmospheric response (Hoerling et al. 1997; McGregor et al. 2013), two additional mechanisms are contributing to the relatively weak circulation asymmetry between the two ENSO phases in our experiments. First, La Niña events exhibit a weaker SST anomaly amplitude compared to El Niño events in nature (positive ENSO skewness) and hence generate a weaker anomalous circulation response (e.g., Kang and Kug 2002; Wu et al. 2010a; Stuecker et al. 2013). This is not the case in our experiments, in which we prescribe the same SST anomaly forcing magnitude during the positive and negative ENSO phase. Second, the SST and wind anomalies during La Niña events are located farther to the west in nature (e.g., Kang and Kug 2002; Ohba and Ueda 2009; Wu et al. 2010a). Hence, the anomalous low-level NWP-C during boreal winter/spring in the observations is also shifted westward and is thus not
able to strongly affect the equatorial wind anomalies and trigger an oceanic response.

Observations show that usually during JJA(1) of an El Niño event we have already approached a neutral or negative ENSO phase. These negative SST anomalies combined with a negative phase of the annual cycle generate a positive C-mode index (orange line in Fig. 1c), which captures the time evolution of the NWP-AC $\phi$ index (black line in Fig. 1c) during JJA(1)-SON(1). Examples for anomalous low-level NWP-AC circulation generated by this La Niña–annual cycle mechanism can be found, for instance, in JJA 1989, JJA 1998, and JJA 1999 (Fig. 1c). During La Niña conditions in JJA the combination mode response creates anomalous anticyclonic surface circulation in the northwestern Pacific (Figs. 5b, 10b) as well as the symmetric Rossby wave response due to the negative N3.4 SST anomalies. Hence, both these processes act together to create a strong anomalous NWP-AC response. The importance of the La Niña SST anomalies on the boreal summer monsoon circulation was recently discussed by Fan et al. (2013). Note that these SST anomalies certainly contribute to the strong amplitude of the anomalous NWP-AC during JJA(1), but we even observe the anomalous NWP-AC in our experiments without any negative N3.4 anomalies present (Fig. 8d).

Moreover, in DJF during La Niña events we can observe an anomalous cyclonic circulation in the northwestern Pacific region (e.g., DJF 1988/89, DJF 1998/99, and DJF 1999/2000; Fig. 1c), which is the predicted C-mode response (Fig. 10b).

The northwest Pacific anomalous surface circulation response with given ENSO phase and annual cycle phase information can be summarized as follows:

(i) El Niño phase in DJF/MAM leads to anomalous northwest Pacific low-level anticyclonic circulation;
(ii) El Niño phase in JJA/SON leads to anomalous northwest Pacific low-level cyclonic circulation;
(iii) La Niña phase in DJA/MAM leads to anomalous northwest Pacific low-level cyclonic circulation; and
(iv) La Niña phase in JJA/SON leads to anomalous northwest Pacific low-level anticyclonic circulation.

Whether a GCM is able of capturing this combination mode response depends crucially on both the amplitude of the unforced internal variability in the NWP region and the amplitude of the forced signal. Increasing the number of ensemble members is an effective way to increase the signal-to-noise ratio and thus offers potential improvement for the NWP-AC predictability.

Our study confirms our previous results (Stuecker et al. 2013) that the two dominant modes of anomalous surface wind variability in the Indo-Pacific are related to ENSO and the combination mode (C-mode). In this study, we showed that the combination mode is the dominant cause for the formation of the anomalous meridionally antisymmetric circulation, which includes the low-level NWP-AC feature. This conclusion differs
from previous studies (Wang et al. 2013; Xiang et al. 2013) that explain the second most dominant Indo-Pacific mode as a local air–sea interaction mode. Our explanation is complementary to Kosaka et al. (2013), who propose an ENSO excited air–sea coupled mode as an explanation for the boreal summer circulation anomalies in the northwestern Pacific.

Despite some relatively small model biases, our experiments are able to reproduce the anomalous Indo-Pacific response reasonably well. It is interesting to note, however, that our simulations utilized observed climatological forcing of the AGCM and the SOM model. This may well have led to the underestimation of the effect of model biases on the representation of the C-mode. Thus, more research is needed to understand the representation of the C-mode in the fully coupled setting.

Furthermore, previous studies of the NWP-AC missed the important near-annual combination mode variability by using frequency domain filtering methods. This may lead to spurious results and unphysical interpretations of Indo-Pacific climate variability. We show that it is not adequate to assess ENSO impacts by considering only the interannual aspect of ENSO. ENSO–annual cycle combination mode dynamics are an essential (extended) part of ENSO and result in a wide range of clearly deterministic higher-frequency variability (Fig. 9).

Concluding, we presented a novel physical interpretation of the anomalous low-level NWP-AC in terms of ENSO–annual cycle interactions. The general C-mode framework outlined here may be applicable to other seasonally modulated climate phenomena, such as the Pacific–North America pattern (PNA), the Pacific–South America pattern (PSA), the North Atlantic Oscillation (NAO), and the Indian summer monsoon.

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