Observations and Mechanisms of a Simple Stochastic Dynamical Model Capturing El Niño Diversity

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ABSTRACT

El Niño–Southern Oscillation (ENSO) has significant impact on global climate and relevance for seasonal forecasts. Recently, a simple modeling framework was developed that captures the ENSO diversity, where state-dependent stochastic wind bursts and nonlinear advection of sea surface temperature are coupled to a simple ocean–atmosphere model that is otherwise deterministic, linear, and stable. In this article, the coupled model is compared with observations using reanalysis data over the last 34 yr, where the observed non-Gaussian statistics and the overall mechanisms of ENSO are both captured by the model. Then the formation mechanisms of both the central Pacific (CP) and the traditional El Niño in the model are systematically studied. First, ocean Rossby waves induced by easterly trade wind anomalies facilitate the heat content buildup. Then the reflected ocean Kelvin waves and the nonlinear advection lead to positive SST anomalies in the CP region and create a CP El Niño. Second, two formation mechanisms are revealed for the traditional El Niño, including the super (extreme) El Niño. The first mechanism indicates a preferred wind structure with easterly wind bursts (EWBs) leading westerly wind bursts (WWBs), where the EWBs build up heat content and then the WWBs trigger the El Niño. The second mechanism links the two types of El Niño, where a CP El Niño favors a heat content buildup and the advent of a traditional El Niño. This article also highlights the mechanisms of La Niña formation and El Niño termination.

1. Introduction

El Niño–Southern Oscillation (ENSO) is a naturally occurring phenomenon in the tropical Pacific, which has significant impact on global climate and relevance for seasonal forecasts. El Niño and La Niña are the warm and cool phases of the ENSO, and the pattern shifts back and forth every few years with irregularity in amplitude,

duration, temporal evolution, and spatial structure. The traditional El Niño [or eastern Pacific (EP) El Niño] has anomalous warm sea surface temperature (SST) in the equatorial eastern Pacific Ocean at event peak, and the corresponding anomalous Walker circulation is shifted eastward with strong convection occurring near the west coast of America (McPhaden et al. 2006; Clarke 2008). In recent decades, a different type of El Niño has been frequently observed, which is called the central Pacific (CP) El Niño (Lee and McPhaden 2010; Kao and Yu 2009; Ashok et al. 2007; Kug et al. 2009; Larkin and Harrison 2005; Guilyardi 2006). CP El Niños are characterized with the maximum warming in the central equatorial Pacific and an anomalous two-cell Walker

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circulation over the tropical Pacific (Kug et al. 2009). In addition to the distinct climate patterns in the equatorial Pacific region, different types of El Niño as well as La Niña also have different teleconnections that affect the global climate (Ashok et al. 2007; Weng et al. 2009; Capotondi et al. 2015; Chung and Li 2015).

The significant impact of ENSO requires a comprehensive understanding of the underlying formation mechanisms of different types of El Niño as well as the development of dynamical models that reproduce the ENSO diversity. Yet, most of the current climate models have biases in simulating the ENSO diversity. Some general circulation models (GCMs) are able to reproduce only one single type of El Niño (Ham and Kug 2012; Kug et al. 2012; Ault et al. 2013; Capotondi et al. 2015). Other climate models, despite their ability to reproduce the observed diversity of ENSO to some extent, typically overestimate the amplitude of the ENSO interannual variability and misrepresent the simulated frequency and duration of the El Niño events (Wittenberg 2009; Kug et al. 2010). Nevertheless, some consensus regarding the El Niño diversity have been reached. Westerly wind bursts (WWBs) are well known to be one of the crucial triggering effects of the traditional El Niño (Harrison and Vecchi 1997; Vecchi and Harrison 2000; Tziperman and Yu 2007). Zonal advection plays a crucial role in the CP El Niño phases (Kao and Yu 2009; Yu and Kao 2007; Yeh et al. 2009, 2014), which has a significant contribution to the SST tendency (Kug et al. 2009) and is particularly important during the initiation phase of CP El Niño (Su et al. 2014). Yet, many issues require further exploration. For example, whether ENSO is a self-sustained unstable and naturally oscillatory mode (Jin 1997; Timmermann et al. 2003) or a stable mode triggered by atmospheric random forcing (Penland and Sardeshmukh 1995; Kleeman and Moore 1997; Moore and Kleeman 1999; Kleeman 2008) is still under debate. In addition, the impacts of both easterly wind bursts (EWBs) and WWBs on ENSO are suggested by Chiodi and Harrison (2015) and Hu and Fedorov (2016), while Puy et al. (2016) argue that the EWBs have less potential to influence ENSO evolution. Furthermore, the connection between global warming and El Niño diversity remains an active ongoing research (Cai et al. 2014; Kohyama et al. 2017; Xiang et al. 2013). In contrast with operational models, simple dynamical models reveal the ENSO mechanisms in a more straightforward way and are usually computationally efficient, which allows detailed studies. Therefore, they are useful complementary tools for the more complex operational models in understanding the El Niño diversity.

Recently, a simple modeling framework was developed that captures the statistical diversity of ENSO (Thual et al.

2016; Chen and Majda 2016, 2017). This simple modeling framework is physically consistent and amenable to detailed analysis, which facilitates the understanding of the formation mechanisms of ENSO diversity. In this simple modeling framework, the starting model is a coupled ocean–atmosphere model that is deterministic, linear, and stable (Kleeman 2008; Kleeman and Moore 1997; Moore and Kleeman 1999). Then systematic strategies are developed for incorporating several major causes of the ENSO diversity into the coupled system. First, a stochastic parameterization of the wind bursts including both westerly and easterly winds is coupled to the simple ocean–atmosphere system, where the amplitude of the wind bursts depends on the SST in the western Pacific warm pool. Such a coupled model is fundamentally different from the Cane–Zebiak (Zebiak and Cane 1987) and other nonlinear models (Jin 1997; Timmermann et al. 2003), in which the internal instability rather than the external wind bursts maintains the ENSO cycle. It is shown that (Thual et al. 2016), in addition to simulating traditional moderate El Niño and super (extreme) El Niño and La Niña events, the coupled model is able to reproduce key observational features in the eastern Pacific, such as the non-Gaussian probability density function and spectrum. Second, a simple nonlinear zonal advection and a mean easterly trade wind anomaly are both incorporated into the coupled model that facilitate the intermittent occurrence of the CP El Niño (Chen and Majda 2016). Here the mean easterly trade wind anomaly represents the recent acceleration of the trade winds (England et al. 2014; Sohn et al. 2013; Merrifield and Maltrud 2011). The nonlinear advection involves the contribution from both mean and fluctuation in the appearance of the stochastic wind bursts, which was not emphasized in previous work. Then, a three-state Markov jump stochastic process is developed to drive the stochastic wind bursts, which describes the state-dependent transition mechanisms of the wind activity in a simple and effective fashion and allows the coupled model to simulate the El Niño diversity with realistic features (Chen and Majda 2017).

In this article, the formation mechanisms of ENSO diversity based on this simple coupled model are systematically studied and are compared with the observational record using reanalysis data over the last 34 yr. First, the non-Gaussian statistics of different variables of the coupled model are compared with those in the observational data in different regions ranging from the western to eastern Pacific. This is followed by a comparison of the overall ENSO formation mechanisms via lagged correlation analysis. Then the formation mechanisms of traditional El Niño, CP El Niño, and La Niña in the coupled model are illustrated respectively with detailed analysis and concrete examples. The significant roles of the underlying ocean Kelvin and Rossby

waves, the imperfect reflections, and the wind burst activities are all highlighted. In addition to the formation of different ENSO events, the mechanism of El Niño termination through ocean waves is also explained.

The coupled model is presented in [section 2](#), along with a description of the processing of observational data. [Section 3](#) shows the statistical features and the overall ENSO formation mechanism of the coupled model, as compared with the observations. The formation mechanisms of different types of El Niño and La Niña, in light of both conditional statistics and case studies, are demonstrated in [section 4](#). A summary discussion of the model advantages and shortcomings is included in [section 5](#). The supplementary material includes sensitivity tests that indicate the important roles of both the nonlinear advection and the eastern Pacific boundary reflections in the coupled model.

2. Model and observational data

a. Coupled ENSO model

The ENSO model consists of a nondissipative atmosphere coupled to a simple shallow-water ocean and SST budget in the interannual time scale. The coupled model was extensively discussed in [Thual et al. \(2016\)](#) and [Chen and Majda \(2016, 2017\)](#), and we summarize the key features below.

Atmosphere model:

$$\begin{aligned} -yv - \partial_x \theta &= 0, \\ yu - \partial_y \theta &= 0, \\ -(\partial_x u + \partial_y v) &= E_q / (1 - \bar{Q}), \end{aligned} \quad (1)$$

Ocean model:

$$\begin{aligned} \partial_\tau U - c_1 YV + c_1 \partial_x H &= c_1 \tau_x, \\ YU + \partial_y H &= 0, \\ \partial_\tau H + c_1 (\partial_x U + \partial_y V) &= 0, \end{aligned} \quad (2)$$

SST model:

$$\partial_\tau T + \mu \partial_x (UT) = -c_1 \zeta E_q + c_1 \eta H, \quad (3)$$

with

$$E_q = \alpha_q T \quad \text{and} \quad \tau_x = \gamma(u + u_p). \quad (4)$$

Here, τ is time and x is zonal direction, while y and Y are meridional directions in the atmosphere and ocean, respectively. The variables $u, v, \theta, U, V, H, T, E_q$, and τ_x are, respectively, zonal wind, meridional wind, potential temperature, zonal current, meridional current, thermocline depth, SST, latent heat, and zonal wind stress. The

oceanic variables are meant to represent the layer above the thermocline, and the atmospheric ones correspond to the lower troposphere. All variables are anomalies from an equilibrium state and are non-dimensional. The term u_p in (4) describes stochastic wind burst activity and will be discussed below. The atmosphere extends over the entire equatorial belt $0 \leq x \leq L_A$, while the Pacific Ocean extends over $0 \leq x \leq L_O$. The meaning of the parameters and their values are shown in the supplementary material.

The coupled model introduces unique theoretical elements such as a nondissipative atmosphere consistent with the skeleton model for the Madden–Julian oscillation (MJO) ([Majda and Stechmann 2009, 2011](#)), valid here on the interannual time scale and suitable to describe the dynamics of the Walker circulation ([Majda and Klein 2003](#); [Stechmann and Ogrosky 2014](#); [Stechmann and Majda 2015](#)). In addition, the meridional axes y and Y are different in the atmosphere and ocean as they each scale to a suitable Rossby radius. This allows for a systematic meridional decomposition of the system into the well-known parabolic cylinder functions ([Majda 2003](#)), which keeps the system low-dimensional ([Thual et al. 2013](#)) as will be discussed hereafter.

The coupled system in (1)–(4) without the nonlinear zonal advection in (3) was systematically studied in ([Thual et al. 2016](#)) for simulating the traditional El Niño events. Note that if the stochastic wind burst u_p is further removed, then the resulting coupled system is linear, deterministic, and stable ([Kleeman 2008](#); [Kleeman and Moore 1997](#); [Moore and Kleeman 1999](#)). Therefore, the external wind bursts rather than the internal instability play the role of maintaining the ENSO cycles in the coupled system. On the other hand, the simple nonlinear advection in (3) involves the contribution from both mean and fluctuation, which has a significant impact on the SST tendency and is crucial for the intermittent occurrence of the CP El Niño ([Chen and Majda 2016](#)). The nonlinear advection is also related to the compensation of zonal, meridional, and vertical divergences in the simple modeling framework. Note that there is no ad hoc parameterization of the background SST in the coupled model, which is different from previous works ([Jin and An 1999](#); [Dewitte et al. 2013](#)).

1) MERIDIONAL TRUNCATION

To solve the coupled system while retaining the key dynamical features, a simple meridional truncation is applied to (1)–(3) ([Thual et al. 2013, 2016](#)). Different parabolic cylinder functions are utilized in the ocean and atmosphere due to the difference in their deformation radii. The zeroth-order atmospheric parabolic cylinder

function has a Gaussian profile that is centered at the equator and reads $\phi_0(y) = (\pi)^{-1/4} \exp(y^2/2)$, and the second-order one, which will be utilized as the reconstruction of solutions, reads $\phi_2(y) = (4\pi)^{-1/4} (2y^2 - 1) \exp(y^2/2)$. The oceanic parabolic cylinder functions $\psi_m(Y)$ are similar to the atmospheric ones except that they depend on the Y axis. In the atmosphere we assume a truncation of moisture, wave activity, and external sources to ϕ_0 . This is known to excite only the Kelvin and first Rossby atmospheric equatorial waves K_A and R_A (Majda and Stechmann 2009, 2011). In the ocean (Clarke 2008), we assume a truncation of zonal wind stress forcing to ψ_0 , where for the notation simplicity, we still use τ_x as the projected coefficient. This is known to excite only the Kelvin and first Rossby oceanic equatorial waves K_O and R_O . Similarly, for the SST model we assume a truncation ψ_0 , where again T is the projected coefficient. With these truncations, the coupled ENSO model in (1)–(3) is given by the following.

Atmosphere model:

$$\begin{aligned} \partial_x K_A &= \chi_A (E_q - \langle E_q \rangle) (2 - 2\bar{Q})^{-1}, \\ -\partial_x R_A / 3 &= \chi_A (E_q - \langle E_q \rangle) (3 - 3\bar{Q})^{-1}, \end{aligned} \quad (5)$$

Ocean model:

$$\begin{aligned} \partial_\tau K_O + c_1 \partial_x K_O &= \chi_O c_1 \tau_x / 2, \\ \partial_\tau R_O - (c_1/3) \partial_x R_O &= -\chi_O c_1 \tau_x / 3, \end{aligned} \quad (6)$$

SST model:

$$\partial_\tau T + \mu \partial_x [(K_O - R_O)T] = -c_1 \zeta E_q + c_1 \eta H, \quad (7)$$

where χ_A and χ_O are the projection coefficients from ocean to atmosphere and from atmosphere to ocean, respectively, because of the different extents in their meridional bases.

Periodic boundary conditions are adopted for the atmosphere model in (5). Reflection boundary conditions are adopted for the ocean model in (6):

$$K_O(0, t) = r_W R_O(0, t) \quad \text{and} \quad R_O(L_O, t) = r_E K_O(L_O, t), \quad (8)$$

where $r_W = 0.5$ represents partial loss of energy in the western Pacific boundary across Indonesia and the Philippines and $r_E = 0.5$ represents partial loss of energy due to the north–south propagation of the coastal Kelvin waves along the eastern Pacific boundary. The reflection boundary conditions were discussed in Jin (1997), and the choices of these parameters are based on the observations (Boulanger and Menkes 1995, 1999). For the SST model, no normal derivative at the boundary is adopted (i.e., dt/dx).

Now instead of solving the original system in (1)–(3), we solve the system with meridional truncation in (5)–(7). The physical variables can be easily reconstructed by the linear combination of the waves. See the supplementary material for the formulas.

2) STOCHASTIC WIND BURST PROCESS

Stochastic parameterization of the wind activity is added to the model that represents several important ENSO triggers and terminations including the WWBs (Harrison and Vecchi 1997; Vecchi and Harrison 2000; Tziperman and Yu 2007), the EWBs (Chiodi and Harrison 2015; Hu and Fedorov 2016; Puy et al. 2016), and the convective envelope of the MJO (Hendon et al. 2007; Puy et al. 2016). It also includes the strengthening of the easterly trade wind anomaly as observed since the 1990s, when the occurrences of the CP El Niño become more frequently (Lee and McPhaden 2010; Kao and Yu 2009). The wind bursts u_p have the following structure:

$$u_p = a_p(\tau) s_p(x) \phi_0(y), \quad (9)$$

where $a_p(\tau)$ is a time series representing the amplitude of wind activity with positive for WWB and negative for EWB and the evolution of $a_p(\tau)$ will be defined below. In (9), $s_p(x)$ is a fixed zonal spatial structure in the western Pacific (see Fig. 1) while $\phi_0(y)$ is the meridional basis. The localization of $s_p(x)$ in the western Pacific is consistent with observations for both the WWB perturbations (Tziperman and Yu 2007) and the strengthening of the trade wind anomaly (England et al. 2014; Sohn et al. 2013). Note that the EWBs in nature can extend more eastward and contain off-equator structures in the eastern Pacific (Puy et al. 2016; Hu and Fedorov 2016). Since the coupled system here is truncated to a Gaussian meridional basis without an asymmetric component and the convective envelope of the MJO in particular features both westerly and easterly wind in the western Pacific, we adopt the same basis for EWBs as the WWBs for simplicity. The setup here emphasizes the influence of the wind bursts on both triggering and terminating ENSO in the western Pacific. Note that the interactions between ENSO and EWBs are the same as those between ENSO and WWBs on a short time scale but have different cumulation effects on time scales of up to 6 months in the current model.

The evolution of wind burst amplitude a_p reads as follows:

$$\frac{da_p}{d\tau} = -d_p(a_p - \hat{a}_p) + \sigma_p(T_W) \dot{W}(\tau), \quad (10)$$

where d_p is dissipation and $\dot{W}(\tau)$ is a white noise source, representing the intermittent nature of the wind bursts

FIG. 1. (a) Zonal profile of thermocline feedback $\eta(x)$. (b) Zonal profile of wind burst structure $s_p(x)$ in the western Pacific. (c) Surface wind response directly to the mean easterly trade wind anomaly [i.e., $\hat{a}_p \times s_p(x) \times \phi_0(y)$]. This figure is taken from [Chen and Majda \(2016, 2017\)](#).

at interannual time scale. The amplitude of the wind burst noise source σ_p depends on T_W [see (11)], which is the average of SST anomalies in the western half of the equatorial Pacific ($0 \leq x \leq L_O/2$). A transition from the quiescent (small σ_p) to active state (large σ_p) is more likely when $T_W > 0$ owing to the fact that wind burst activity is usually favored by warmer SST in the western Pacific.

Note that this state-dependent wind amplitude is fundamentally different from those in previous works ([Jin et al. 2007](#); [Levine and Jin 2017, 2010](#)) that rely on the eastern Pacific SST. Instead of using a gradual increased function to describe the trend of the acceleration of the easterly trade wind as observed in nature since the 1990s ([England et al. 2014](#); [Sohn et al. 2013](#); [Merrifield and Maltrud 2011](#)), we adopt a constant value $\hat{a}_p < 0$ to represent the mean acceleration of the trade wind anomaly in the model for simplicity. Corresponding to $\hat{a}_p < 0$, the direct response of the surface wind

associated with the Walker circulation in the equatorial Pacific band is shown in [Fig. 1c](#), which is similar to the observed intensification of the Walker circulation in recent decades ([Sohn et al. 2013](#)). Although there were occasional occurrences of CP El Niño before the 1990s, the goal here is to include the link between the frequent occurrences of the CP El Niño and the acceleration of the trade wind since the 1990s into the coupled model. It is shown in [Chen and Majda \(2016\)](#) that this easterly trade wind anomaly is crucial in reproducing CP El Niño events in the simple model framework.

3) A THREE-STATE MARKOV JUMP STOCHASTIC PROCESS

A three-state Markov jump stochastic process ([Gardiner et al. 1985](#); [Lawler 2006](#); [Majda and Harlim 2012](#)) is adopted for the wind activity to reproduce the ENSO diversity. This three-state process essentially combines the previous two-state models for the traditional

El Niño (Thual et al. 2016) and the CP El Niño (Chen and Majda 2016). Here, state 2 primarily corresponds to the traditional El Niño and state 1 to the CP El Niño, while state 0 represents the La Niña (discharge) and quiescent phases. Since strong wind bursts play an important role in triggering the traditional El Niño (Vecchi and Harrison 2000; Tziperman and Yu 2007; Hendon et al. 2007), a large noise amplitude σ_p is chosen in state 2. On the other hand, an enhanced easterly trade wind has been observed since the 1990s (England et al. 2014; Sohn et al. 2013), during which period CP El Niño occurred more frequently. To make the link between the CP El Niño and the acceleration of the easterly trade wind anomaly, a negative (easterly) mean $\hat{a}_p$ is imposed in state 1. To obtain the CP El Niño, the amplitude of $\hat{a}_p$ and the stochastic noise must be balanced (Chen and Majda 2016). This implies a moderate noise amplitude in state 1, which also agrees with observations (Chen et al. 2015). Finally, only weak wind activity is allowed in the quiescent state and the discharge phase with La Niña (state 0). Thus, the three states are given by

$$\begin{aligned} \text{State 2: } \sigma_{p2} &= 3.75, \quad d_{p2} = 5.1, \quad \hat{a}_{p2} = -0.25, \\ \text{State 1: } \sigma_{p1} &= 1.2, \quad d_{p1} = 5.1, \quad \hat{a}_{p1} = -0.25, \quad \text{and} \\ \text{State 0: } \sigma_{p0} &= 0.5, \quad d_{p0} = 5.1, \quad \hat{a}_{p0} = 0, \end{aligned} \quad (11)$$

respectively, where $d_p = 5.1$ is slightly different from the values used in the previous works (Chen and Majda 2017), which represents a relaxation time around 6.67 days to meet the observational facts (see section 3). Note that the same mean easterly trade wind anomaly as in state 1 is adopted in state 2 because both the traditional and the CP El Niño occurred during the last 25 yr. Since the amplitude of the stochastic noise dominates the mean easterly wind in state 2, this mean state actually has little impact on simulating the traditional El Niño events. On the other hand, no mean trade wind anomaly is imposed in state 0 to guarantee no El Niño event occurring in the quiescent phase. With such choice of the parameters, both the amplitude and the time scale of the wind burst activity are similar to nature.

The transition rates between the three states are chosen with guidelines from the observational record over the last a few decades. See supplementary material for details. First, the traditional El Niño is usually followed by the La Niña and therefore a higher transition probability from state 2 to state 0 than to state 1 is set in the model. Second, starting from the quiescent state, the model is allowed to develop both types of El Niño afterward. In addition, after a series of CP El Niño, the model has a chance to switch to traditional El Niño as observed in the 1990s, which provides a direct link

between the two types of El Niño in the model and is the basis for one of the formation mechanisms of the traditional El Niño (see section 4 hereafter). These facts are all taken into account for setting the transition rates. Although the observational period used for determining the transition rates is short (see section 2c), a long simulation of the model is able to provide results that are statistically significant and discover some new features that are not easily found in the limited observational record.

b. Definitions of different El Niño events in the coupled model

To compare the model simulations with the observational record, we make use of the three well-known Niño SST indices: Niño-3 (150°–90°W), Niño-3.4 (170°–120°W), and Niño-4 (160°E–150°W). Since the meridional truncation only includes the leading meridional basis, the off-equator asymmetric features are not captured by the model. Therefore, unlike the observations where the statistics is computed within 5°N–5°S, the model statistics at the equator is used for comparison with the observations in this work.

The definitions of the traditional El Niño and the La Niña are quite simple, with anomalous SST above 0.5 K and below –0.5 K in the Niño-3 region. On the other hand, the identification of a CP El Niño event requires the combination of different Niño indices and a universal definition is still under debate (Ashok et al. 2007; Trenberth and Stepaniak 2001; Kug et al. 2009; Yeh et al. 2009; Ren and Jin 2011).

Here, simple criteria are proposed to distinguish different El Niño events in light of the distinct roles of each of the three states in the wind activity model. Specifically, situating in state 1 is one of the necessary conditions for identifying the CP El Niño. Other conditions for recognizing a CP El Niño event are Niño-4 > 0.5 and Niño-4 > Niño-3 > 0. The reason to pick up Niño-3 > 0 here is to exclude the La Niña-like events, and such positive value of the averaged SST over the Niño-3 region is not incompatible with the CP El Niño feature that cooling occurs near the eastern Pacific boundary. Note that replacing Niño-4 by Niño-3.4 for identifying the CP El Niño does not lead to qualitative difference in any of the statistical features shown below thanks to the constraint of situating in state 1. On the other hand, although strong wind bursts (state 2) trigger the traditional El Niño, the decaying phase of each traditional El Niño event from anomalously warm SST back to the normal condition corresponds to state 0. In addition, the ENSO discharge phase with La Niña also typically lies in state 0 with occasional occurrences in state 2. Thus, traditional El Niño and La Niña requires locating in

TABLE 1. Criteria of defining different ENSO events.

	Condition 1	Condition 2	Condition 3
Traditional El Niño	State 0 or 2	Niño 3 > 0.5	Niño 3 > Niño 3.4
La Niña	State 0 or 2	Niño 3 < -0.5	—
CP El Niño	State 1	Niño 4 > 0.5	Niño 4 > Niño 3 > 0

either state 0 or 2. This is consistent with the discharge–recharge theory in Jin (1997) and the observational analyses in Meinen and McPhaden (2000). Other conditions for traditional El Niño are Niño-3 > 0.5 and Niño-3 > Niño-3.4, where the latter serves to exclude CP-like events that occasionally occur in state 2, and those for La Niña are Niño 3 < -0.5. These criteria are summarized in Table 1.

c. Observational data

In this work, observational data are utilized to assess the realism of the model solutions. The following three observational datasets are utilized: 1) daily zonal winds at 850 hPa from the NCEP–NCAR reanalysis (Kalnay et al. 1996) (<http://www.esrl.noaa.gov/psd/>), 2) daily sea surface temperatures from the OISST reanalysis (Reynolds et al. 2007) (<https://www.ncdc.noaa.gov/oisst>), and 3) monthly thermocline depth from the NCEP GODAS reanalysis (Behringer et al. 1998) (<http://www.esrl.noaa.gov/psd/>). Thermocline depth is computed from potential temperature as the depth of the 20°C isotherm. All datasets are averaged meridionally within 5°N–5°S in the tropical Pacific (120°E–80°W) and cover the period from January 1982 to September 2016.

From the data, we obtain fields of zonal winds $u_{\text{OBS}}(x, t)$ in meters per second, sea surface temperature $T_{\text{OBS}}(x, t)$ in kelvin, and thermocline depth $H_{\text{OBS}}(x, t)$ in meters that depend only on zonal position x (degrees longitude) and time t (days). In addition, we define u_{OBS}^w as the amplitude of zonal wind anomalies $u_{\text{OBS}}^w(t)$ in the western Pacific, computed from an average in the region 140°E–180°. This average is roughly equivalent to a projection on the zonal wind burst structure $s_p(x)$ of the ENSO model defined in Fig. 1.

Each field is decomposed into climatology, interannual anomalies, and high-frequency anomalies on monthly time scale. For example, zonal winds are decomposed into

$$u_{\text{OBS}}(x, t) = u_{\text{SC}}(x, t) + u_A(x, t) + u_{\text{HF}}(x, t), \quad (12)$$

where u_{SC} is climatology, u_A is interannual anomalies to the climatology, and u_{HF} is intraseasonal anomalies. For this, a 90-day-centered running mean is applied to u_{OBS} from which u_{HF} is extracted as residual. The running mean signal is then decomposed into climatology u_{SC} and anomalies u_A . A similar decomposition is utilized for T_{OBS} , H_{OBS} , and u_{OBS}^w .

The observed fields presented above are the potential surrogates for the variables in the simple ENSO model described in section 2a. Observed zonal wind anomalies $u_A + u_{\text{HF}}$, sea surface temperature anomalies $T_A + T_{\text{HF}}$, and thermocline depth anomalies $H_A + H_{\text{HF}}$ are direct surrogates for $u + u_p$, T , and H in the model, respectively.

In the comparison of the statistics below, the model variables are also decomposed into the same components as described above. The only difference is that there is no seasonal cycle in the model. Although the lack of the seasonal cycle results in a uniform distribution of the ENSO peaks throughout the year, it will not influence the long-term statistics of the anomalies as well as the understanding of the ENSO mechanisms via the interactions between atmosphere wind bursts, ocean waves, and boundary reflections.

3. Observations, statistical properties, and overall mechanisms of ENSO

a. Observed ENSO variabilities

The observed ENSO variabilities in the equatorial Pacific during the last 34 yr are shown in Fig. 2. The general circulation in the equatorial Pacific region consists overall of strong trade winds in the central Pacific as well as strong zonal gradients of SST and thermocline depth (see Figs. 2d,f,h). Such a general circulation is destabilized during El Niño events as positive SST anomalies develop in the central-eastern Pacific along with enhanced eastward zonal winds and strong thermocline depth anomalies (see Figs. 2c,e,g). Particularly, the traditional El Niño with maximum SST anomalies in the eastern Pacific is distinguished from the CP El Niño with maximum SST anomalies in the central Pacific. Major traditional El Niño events occur in years 1982/83, 1997/98, 2006/07, and 2015/16, while prominent CP El Niño events are observed in years 1987/88, 1990/91, 1992/93, 1994/95, 2002/03, 2004/05, and 2009/10. Note that a traditional El Niño is usually followed by a La Niña with reverse conditions. In addition, Fig. 2 shows the details of wind bursts activity according to the zonal wind anomaly over the entire tropical Pacific u_{HF} or averaged in the western Pacific u_{HF}^w . Wind burst activity is highly irregular and intermittent with both westerly and easterly wind bursts.

FIG. 2. Observed ENSO variability. (a) Zonal winds averaged over the western Pacific (140°E – 180°), with a superposition of interannual u_A^W (black) and intraseasonal u_{HF}^W (blue for $u_{HF}^W < 0$, red for $u_{HF}^W > 0$) anomalies. Hovmöller diagrams of: (b) strong intraseasonal zonal winds u_{HF} (blue for $u_{HF} \leq -5 \text{ m s}^{-1}$, red for $u_{HF} \geq 5 \text{ m s}^{-1}$); (c) zonal winds anomalies u_A ; (d) zonal wind anomalies and climatology $u_A + u_{SC}$; (e) sea surface temperature anomalies T_A ; (f) sea surface temperature anomalies and climatology $T_A + T_{SC}$; (g) thermocline depth anomalies H_A (m); and (h) thermocline depth anomalies and climatology $H_A + H_{SC}$.

Figure 3 shows linear trends in time for the observed zonal winds, SST, and thermocline depth over the period 1982–2016. A displacement of the trade winds from the central-eastern to the central Pacific is clearly observed. Particularly, the decreased averaged value of the observed zonal wind over the western Pacific indicates a multidecadal strengthening of the easterly trade wind in that region (England et al. 2014; Sohn et al. 2013; Merrifield and Maltrud 2011). This justifies imposing the mean easterly trade wind anomaly $\hat{a}_p$ into the coupled

model in (10) for representing the recent period. On the other hand, both SST and thermocline zonal gradient show a gradual increase during the last 34 yr (Amaya et al. 2015; Kim et al. 2014).

b. Statistical properties of the model and observations

We start with exploring the statistical properties of both the coupled model and observations. Figure 4 shows the probability density functions (PDFs) for interannual anomalous SST, zonal winds, and thermocline

FIG. 3. Linear trends in time of observed (a) zonal winds u_{OBS} , (b) sea surface temperature T_{OBS} , and (c) thermocline depth H_{OBS} . The trends are obtained from a polynomial curve fit of the form $u_{\text{OBS}}(x, t) = p_0(x) + p_1(t) + O(1)$, where x is longitude, t is time, and p_0, p_1 are coefficients.

depth associated with the coupled model and observations, respectively, and the statistics are recorded in Table 2. The PDFs of the observations (red) are based on the 34-yr period, while those of the model are based on a 5000-yr-long simulation (blue). The Gaussian fits of the model PDFs with the same variance are shown in dashed black curves. In addition, the 95% confidence interval of the model simulation based on 100 simulations of nonoverlapped 34-yr period (gray) are illustrated to assess the statistical consistency between the model and observations.

First, consistent with observations, the PDFs of SST associated with the coupled model in the Niño-4 and Niño-3 regions show negative and positive skewness, respectively. The presence of a fat tail together with the positive skewness in Niño-3 indicates the extreme El Niño events in the eastern Pacific (Burgers and Stephenson 1999). Note that, despite the correct skewed direction, the skewness of T_A of the model in the Niño-3 region seems to be underestimated compared with that of the observations. Yet, given the short observational period (34 yr), a single super-El Niño event during 1997/98 accounts for a large portion of the skewness in Niño-3, and therefore the statistics may contain bias. Nevertheless, the PDF of the observed Niño-3 index lies within the confidence interval of the model simulation, indicating the model capability in generating a similar 34-yr period as observations. On

the other hand, despite a slight overestimation, the variance in all the three Niño regions associated with the model almost perfectly match those with observations. Particularly, the fact that the variance of SST in the Niño-4 region being roughly half as much as that in the other two regions is captured by the coupled model. Note that, as described in the previous work (Thual et al. 2016; Chen and Majda 2017), the parameters in the coupled model are calibrated within physically reasonable ranges to match the SST variances with those provided by NOAA (<https://www.ncdc.noaa.gov/teleconnections/enso/indicators/sst.php>). The Niño indices in different regions are all slightly larger than the observational values utilized here. The reason for this is that the climatology computed here is based on the whole 34-yr period 1982–2016 while only the data before 2000 are utilized in NOAA's version. As shown in Figs. 2 and 3, the SST has an increased trend especially after year 2000, which implies a larger amplitude of climatology and in turn a smaller variance in the SST anomalies of the observations.

Next, the zonal winds and thermocline depth are both averaged over three different regions: the western Pacific (140°E–180°), the central Pacific (180°–140°W), and the eastern Pacific (140°E–100°W). The PDFs of the observed zonal surface winds in the central and eastern Pacific are both positively skewed and fat tailed. Such non-Gaussian

FIG. 4. PDFs of observations over the last 34 yr (1982–2016; red) and model output (5000 yr; blue). The gray areas show the 95% confidence interval of the model statistics computed from 100 nonoverlapped 34-yr periods, and the black dashed curves are the Gaussian fit of the model statistics (5000 yr). (a)–(c) SST interannual anomalies T_A averaged over the Niño-4, Niño-3.4, and Niño-3 regions. (d)–(f) Zonal wind interannual anomalies u_A averaged over the western Pacific (140°E–180°), the central Pacific (180°–140°W), and the eastern Pacific (140°–100°W). (g)–(i) Thermocline depth interannual anomalies H_A averaged over the same regions as zonal winds. The variance, skewness, and kurtosis are listed in Table 2.

features are successfully captured by the coupled model. Although the zonal wind variance from the model simulation as shown in Figs. 4d–f is underestimated, this could be avoided through rescaling of model parameters within reasonable range (e.g., decreasing both γ and $\bar{Q}$). In addition, including more meridional bases and computing the averaged value within an equatorial band may help correct this underestimation, which remains as a future work. On the other hand, the PDFs of the observed thermocline depth in all the three regions are within the 95% confidence interval of the model with 34-yr simulations. Although the observed skewness of the PDFs of the thermocline depth in the central and western Pacific are not well captured by the long simulation in the model, consistent profiles of the PDF for the model and observations in the eastern Pacific are found. The latter is crucial since the thermocline in the eastern Pacific has a significant and direct contribution to the traditional El Niño events including super El Niños.

In addition to the PDFs, the autocorrelation function of the total wind burst anomalies a_{all} , including

both a_{SC} and a_A , is shown in Fig. 5a, and that containing only the interannual wind bursts anomalies a_A resulting from the 90-day running average is shown in Fig. 5b. A short memory of a_{all} , around 6–7 days, is revealed, which is affirmed by the damping coefficient d_p in (11). On the other hand, the interannual anomalies of the wind bursts have a longer memory, around 2–3 months. Both values are consistent with observations (Levine and Jin 2017; Tziperman and Yu 2007; Yu et al. 2003). Finally, Fig. 5c shows the power spectrum of the interannual SST anomalies averaged over the eastern Pacific ($L_O/2 < x < L_O$). The peak is at the interannual band (3–7 yr) as in the observations and previous analysis (Kleeman 2008; Li 1997), and the spectrum of the observations lies within the [2.5%, 97.5%] percentile interval of the model with the same 34-yr length.

c. Overall mechanism of ENSO growth and decay

To compare the overall mechanism of ENSO formation that includes all events, lagged correlations between

TABLE 2. Statistics of the interannual variables T_A , u_A and H_A in different regions corresponding to those in Fig. 4. Both observations (34 yr) and model output (5000 yr) are shown. Since all variables here stand for anomalies, the mean values are zero.

	Observations (34 yr)			Model output (5000 yr)		
	$T_A(160^\circ\text{E}–150^\circ\text{W})$	$T_A(170^\circ\text{E}–120^\circ\text{W})$	$T_A(150^\circ\text{W}–90^\circ\text{W})$	$T_A(160^\circ\text{E}–150^\circ\text{W})$	$T_A(170^\circ\text{E}–120^\circ\text{W})$	$T_A(150^\circ\text{W}–90^\circ\text{W})$
Variance	0.40	0.73	0.78	0.44	1.02	1.06
Skewness	–0.31	0.40	3.12	–0.28	0.03	3.19
Kurtosis	2.51	3.12	4.15	3.56	3.19	3.67
	$u_A(140^\circ\text{E}–180^\circ)$	$u_A(180^\circ–140^\circ\text{W})$	$u_A(140^\circ\text{W}–100^\circ\text{W})$	$u_A(140^\circ\text{E}–180^\circ)$	$u_A(180^\circ–140^\circ\text{W})$	$u_A(140^\circ\text{W}–100^\circ\text{W})$
Variance	4.96	5.42	1.76	1.18	1.32	0.74
Skewness	0.12	1.20	0.61	0.28	0.89	0.55
Kurtosis	2.49	4.81	5.61	3.46	4.56	4.07
	$H_A(140^\circ\text{E}–180^\circ)$	$H_A(180^\circ–140^\circ\text{W})$	$H_A(140^\circ\text{W}–100^\circ\text{W})$	$H_A(140^\circ\text{E}–180^\circ)$	$H_A(180^\circ–140^\circ\text{W})$	$H_A(140^\circ\text{W}–100^\circ\text{W})$
Variance	167	108	203	136	116	124
Skewness	–0.81	–0.95	0.16	–0.05	0.24	0.50
Kurtosis	3.44	4.36	3.29	3.64	3.48	4.11

the Niño-3.4 SST index and different variabilities are shown in Fig. 6 for both the observations and the coupled model. The lagged correlations with SST T_A , thermocline H_A , zonal wind u_A , and the zonal wind averaged over the western Pacific u_A^W are all shown in the figure. Note that the amplitude of the signals was already compared in Fig. 4, and therefore the lagged correlation rather than the lagged regression coefficients are used here to address the underlying formation mechanism. The wind bursts in the coupled model have their own evolution equation [(10)], which means the wind bursts can easily be distinguished from the overall zonal winds. Therefore, the lagged correlation with the total wind bursts a_{all} and its interannual component a_A are also shown in Fig. 6. To highlight the direct contributions from WWBs and EWBs in Fig. 6, the component of the mean easterly trade wind anomaly is removed from a_{all} and a_A by subtracting $-\hat{a}_p$ from the signals obtained from (10) when the corresponding Markov state is 1 or 2. In the next section, the full outputs a_{all} and a_A from (10), which contain $\hat{a}_p$, will be used to compute the conditional lagged correlation on different ENSO phases. Here, except for a_{all} in Fig. 6e, all the variabilities utilized here are interannual as they have been subjected to a 90-day running average as discussed in section 2c.

As shown in Figs. 6a–d, El Niño events are typically preceded by a buildup phase around 1–2 yr prior to the event peak, during which SST and thermocline depth gradually enhance in the western Pacific. Then, during the trigger phase around 6–9 months prior to the event peak, strong westerly zonal surface winds, positive SST, and thermocline depth anomalies all develop and propagate from the western to the central-eastern Pacific. The positive correlation with u_A^W is mainly due to the reversal of anomalous Walker circulation at the onset of an El Niño event as well as the presence of WWBs that serve to trigger the El Niño. In addition to a strong positive correlation between the Niño-3.4 index and the zonal wind averaged over the western Pacific u_A^W , which is less than 1 year prior to it, a weak negative correlation is found at lag times about 1–3 yr. All these features are captured by the coupled model.

Next, we look at the lagged correlations between the Niño-3.4 and the wind bursts a_A (or a_{all}) of the model as shown in Fig. 6e. WWBs occur around 6 months before the event peak, which is qualitatively similar to the observations although the starting time is slightly later than nature (6–9 months in advance) (Harrison and Vecchi 1997; Vecchi and Harrison 2000; Tziperman and Yu 2007). In addition to the WWBs, lagged correlation also indicates strong EWBs at lead times from 6 months to 2 yr before the event, which as will be

FIG. 5. (a) Autocorrelation function of the total wind bursts a_{all} . (b) Autocorrelation function of the interannual wind burst anomalies a_A . (c) Power spectrum of the SST averaged over the eastern half of the Pacific, where the red curve shows the spectrum of the observations over the last 34 yr and the gray area covers from 2.5% to 97.5% of 100 model runs with nonoverlapped 34-yr periods with its mean being shown by the blue curve.

shown in section 4 is responsible for the heat content buildup in the coupled model. The underlying discharge–recharge mechanism has some common features as that described in earlier work (Wang and Picaut 2004), although the EWBs in nature are much

weaker during the recharge process. In the case studies below, the role of the state-dependent wind bursts in triggering the ENSO cycles and the buildup and shut-down of wind burst activity relying on the ENSO events are both emphasized.

FIG. 6. ENSO long-term memory. Lagged correlations of the Niño-3.4 SST index with (a) zonal wind anomalies averaged in western Pacific (u_A^w), (b) zonal wind anomalies (u_A), (c) sea surface temperature anomalies (T_A), and (d) thermocline depth anomalies (H_A). (e)–(h) As in (a)–(d), but for the model output. In (e), the total wind bursts (a_{all} , blue) and the interannual wind burst anomalies (a_A , black) are also included with the u_A^w (red). Here the component of the mean easterly trade wind is removed by subtracting $-\hat{a}_p$ from the signals a_{all} and a_A obtained from (10) when the Markov state is 1 or 2. Except the total wind burst a_{all} , which contains both the interannual and high-frequency components, all the other variables including the Niño-3.4 index contain only the interannual anomalies.

FIG. 7. As in Fig. 6 for the model output, but conditioned on 1) CP El Niño, 2) EP (traditional) El Niño, and 3) La Niña phases. Here, a_{all} and a_A are directly taken from (10) without removing $\hat{a}_p$. (e) The lagged correlation between SST index and the flux divergence $-\partial_x(UT)$ interannual anomalies. The Niño-3 index is used for computing the lagged correlations with traditional El Niño, while the Niño-4 index is adopted for the CP El Niño. Note that the minus of Niño-3 is used for the La Niña such that the negative correlation in the eastern Pacific at the event peak is consistent with the occurrence of a La Niña event and the positive (negative) correlation with the wind bursts stands for westerly (easterly).

4. Mechanisms for the formations of different events and case studies

In Fig. 6, the overall mechanism of ENSO formation was revealed. Nevertheless, each type of El Niño has its unique formation mechanism that is distinct from the overall behavior. Understanding the difference in these formation mechanisms is of importance. However, within the 34-yr short observational period, each type of El Niño event appears only a few times. Such a small number of samples is not sufficient for arriving at any unbiased conclusions. On the other hand, the simple coupled dynamical model, which has been shown to possess the statistical features that are qualitatively similar as nature, is able to provide simulations with a much longer period, which facilitates the understanding of the formation mechanisms of ENSO diversity within this simple coupled model.

Below, a 5000-yr-long simulation from the coupled dynamical model is utilized for understanding the formation mechanisms of different events. The lagged correlations conditioned on CP El Niño and traditional El Niño and La Niña phases as well as the corresponding case studies are studied. As in the previous section, the

statistics are all computed based on the interannual variables (with subscript A). On the other hand, in order to see the structure of the wind burst activity and the evolution of different types of waves, the raw output from the model is shown in different case studies.

a. CP El Niño

1) STATISTICAL FEATURES

Row 1 in Fig. 7 shows the conditional lagged correlations between the Niño-4 SST index and different fields conditioned on CP El Niño phases, where the criteria for identifying the CP El Niño phases was discussed in section 2b. The Niño-4 SST index is utilized here due to the fact that CP El Niño is more related to the anomalous warm SST near the date line rather than the eastern Pacific.

First, different from the overall events (Fig. 6) and the eastern Pacific El Niño–La Niña (rows 2 and 3 in Fig. 7) in which significant correlations ($|\text{corr}| > 0.5$) appear only at lag times less than 1 year, the Niño-4 SST index conditioned on CP El Niño phases has strong correlations with different fields even at 2–3-yr lag times. This reveals the intrinsic difference between CP and

FIG. 8. Deterministic advection modes from the model. (a) Three dynamical regimes as functions of the nonlinear advection μ and the easterly mean trade wind anomaly $\hat{a}_p$. (b)–(d) Different fields (from left to right) atmosphere wind including the mean trade wind anomaly u , ocean zonal current U , thermocline depth H , SST, ocean Kelvin waves, and ocean Rossby waves, corresponding to dynamic regimes I, II, and III, respectively, where the parameter values are marked by red dots in (a).

traditional events that a CP El Niño episode can have a longer duration up to 4–5 yr (Trenberth and Hoar 1996; Chen et al. 2015) (e.g., the long mild central Pacific warming during late 1990 to 1995). Next, the maximum correlation of the Niño-4 index with both SST and thermocline appears in the central Pacific, as expected. This is consistent with the positive and negative correlations of the atmosphere surface wind u_A in the western and eastern Pacific, respectively, which indicates the structure of the anomalous Walker circulation that the surface winds converge in the central Pacific region. One noticeable feature for the CP El Niño is the flux divergence $-\mu\partial_x(UT)$. The lagged correlation structure in Fig. 7e illustrates the role of the flux divergence in warming and cooling the SST in the central-western and central-eastern Pacific, respectively. This is in contrast with the thermocline feedback, which has a positive contribution in the eastern Pacific. Therefore, the flux divergence is crucial in the formation of CP El Niño. Finally, corresponding to the mean easterly trade wind anomaly at the CP El Niño phases, a negative lagged correlation is found between the Niño-4 index and the wind bursts a_A .

2) DETERMINISTIC ADVECTION MODES

The formation of the CP El Niño involves complicated interactions between the nonlinear advection and the stochastic wind activity. To see the development of

waves and different physical variables due to the nonlinear advection alone, we start with the deterministic nonlinear advection modes (Chen and Majda 2017). Here, the stochastic noise in (10) is removed and the wind activity $a_p \equiv \hat{a}_p$ becomes a constant. These deterministic nonlinear advection modes are computed by solving the coupled system in (1)–(4) with this constant $a_p \equiv \hat{a}_p$. Based on different values of the nonlinear advection μ and the amplitude of the easterly mean trade wind anomaly a_p , three dynamical regimes are found in Fig. 8a. In regime I, the steady-state solution has constant values at each longitude. Particularly, with a suitably strong a_p and even without the nonlinear advection, the anomalous warm SST is shifted to the central-eastern Pacific region. The corresponding anomalous ocean zonal current is westward and the anomalous atmospheric surface winds converge in the central-eastern Pacific as well. On the other hand, when both μ and a_p are sufficiently large (regime III), the steady-state solution shows regular oscillation patterns with period around 1.6 yr. Within each period, warm water is transported westward and the maximum of anomalous warm SST is at the central and central-eastern Pacific. Note that the strong ocean currents due to the strong wind stress together with the boundary reflections in both sides create an advective–reflective oscillator (Picaut et al. 1997) and the oscillation structure results from the nonlinear advection term. Without the nonlinear

advection (e.g., regime I), the anomalous warm water simply stays in the same location and does not display an oscillation pattern. The patterns in this regime have potentially implications if the easterly trade winds keep accelerating in the future.

The most important dynamical regime (regime II) is shown in Fig. 7c, which is directly associated with the CP El Niño when random wind bursts are included. This dynamical regime requires a nonzero zonal advection and a suitably strong easterly trade wind anomaly a_p , such that all of the fields become time periodic, and the period is much longer than 2 yr. Starting from a nearly quiescent phase, the easterly mean trade wind anomaly triggers ocean Rossby waves that propagate westward ($t = 88$). When these Rossby waves arrive at the western Pacific boundary, the reflection boundary condition induces ocean Kelvin waves that propagate eastward, where the amplitude of the reflected Kelvin waves is weaker than the Rossby waves due to the energy loss at the boundary. The combined effect of these Rossby and Kelvin waves results in an increase of the thermocline depth and SST anomalies in the western Pacific. As a direct response to the latent heat, the atmosphere winds become stronger and the westerly and easterly surface winds converge in the central Pacific region ($t = 90$). Then, because atmosphere winds force ocean waves [see (6)], the westerly atmosphere winds in the western Pacific force the reflected weak downwelling ocean Kelvin waves to the central Pacific, while the easterly winds in the eastern Pacific prevent these weak waves arriving at the eastern Pacific boundary. Meanwhile, these easterly winds force ocean Rossby waves, with amplitudes that peak in the central Pacific. Since thermocline depth is directly linked to the amplitude of ocean Kelvin and Rossby waves [see (S1) in supplementary material], an increase of the thermocline depth occurs in the central Pacific region, which leads to the buildup of anomalous warm SST. Note that because ocean Rossby waves have stronger amplitudes than ocean Kelvin waves, a westward anomalous ocean zonal current appears in the central Pacific region [see (S1)], which together with the profile of SST anomalies results in a westward transport of anomalous warm water via the nonlinear advection of SST and brings about the eastern Pacific cooling. These anomalous warm SST then stays and gradually develops in the central Pacific for a few years ($t = 95$). As the westward ocean zonal current and the atmosphere easterly winds in the central and eastern Pacific continue to strengthen, the anomalous warm SST is transported to the western Pacific with a deeper thermocline there. Such strong heat content storage provides the condition of triggering a strong traditional El Niño ($t = 108$), followed by a discharge phase of La Niña that drives all the

fields back to a nearly quiescent state. Note that the long period in the deterministic advection mode is due to nonlinearity, which is different from the ENSO cycle from the linear solution. In addition, the random wind burst forcing as will be shown in the next subsection is able to break the structure of such regular deterministic pattern. In particular, the initiation, termination, and duration of the CP mode will become more stochastic. See the case studies below and as well as the illustrations in the supplementary material of Chen and Majda (2016).

3) CASE STUDIES

Now we go back to the full model with the effective state-dependent stochastic noise in the wind activity in (10). The case study in Fig. 9 shows a period of 2.5-yr CP El Niño from $t = 413.5$ to $t = 416$.

First, the interannual wind time series in Fig. 9f clearly indicates a mean easterly trade wind anomaly. Starting from $t = 413$, moderate and weak ocean Rossby waves are triggered continuously by this mean trade wind anomaly together with the effective stochastic wind burst noise, and heat content starts accumulating in the western Pacific. Because of the imperfect reflection at the western Pacific boundary $r_W = 0.5$, the reflected Kelvin waves have only weak amplitudes. Such weak Kelvin waves are able to transport heat content eastward and adjust the atmosphere zonal wind structure (cf. $t = 413$ and 414 in Fig. 9a). However, because of the atmospheric forcing along the ocean Kelvin wave track, the easterly atmosphere zonal surface winds in the eastern surface Pacific also weaken these Kelvin waves [see (6)]. Eventually, a balance is established in the central Pacific region. On the other hand, this atmosphere zonal wind structure also induces ocean Rossby waves that start from the eastern Pacific and peak at the location where easterly and westerly surface winds converge [see (6)]. Thus, according to the link between ocean waves and thermocline depth [(S1)], the maximum of the heat content appears in the central Pacific and results in the CP El Niño.

Note that the reflected weak Kelvin waves created by the mean easterly trade anomaly are crucial in the formation of the CP El Niño. As will be seen in section 4b, strong Kelvin waves induced by the WWBs are able to overcome the barriers from atmosphere easterly zonal winds and therefore transport the heat content and anomalous warm SST to the eastern Pacific and eventually change the atmosphere wind structure. The overall formation mechanism here is similar to that in regime II of the deterministic advection mode. But the stochasticity modifies the deterministic mode in terms of the initiation, termination, and duration of the event.

FIG. 9. Case studies of a series of CP El Niño and a traditional El Niño. Shown are (a) atmosphere wind including the mean trade wind anomaly $u + \hat{u}_p$, (b) ocean zonal current U , (c) thermocline depth H , (d) SST, (e) ocean Kelvin waves, and (f) ocean Rossby waves. All these fields are the direct model output and contain both the interannual variability and the high-frequency component. (g) The time series of the wind activity at the peak x^* of the zonal profile (see Fig. 1b) and its 90-day running average (brown curve). The budget terms, where (h) the time tendency equals (i) the summation of the flux divergence and (j) the damping plus the thermocline feedback, with (h)–(n) each of the four contributions. The flux divergence (blue) and the combined effect of damping and thermocline feedback (red) in the (o) Niño-4 and (p) Niño-3 regions. The dashed rectangles indicate the spatiotemporal region of the CP El Niño in this case. All variables are shown at equator.

Budget analysis is illustrated in Figs. 9h–p. The total SST tendency equals the summation of the flux divergence $-\mu\partial_x(UT)$ and the combined effect of the damping by latent heat fluxes $-c_1\zeta E_q$ and the thermocline feedback $c_1\eta H$. The flux divergence in Fig. 9i serves to increase SST in the central-western and decrease SST in the central-eastern Pacific at the CP El Niño phases, where its two components $-\mu\partial_x(U)$ and $-\mu\partial_x(T)$ are shown in Figs. 9k,l. Here, $-\mu T\partial_x(U)$ mainly heats central Pacific while $-\mu U\partial_x(T)$ heats both central and western Pacific. On the other hand, the combined effect of the damping and thermocline feedback in Fig. 9j indicates an opposite structure with an increased SST tendency in the eastern Pacific. Note that, even if the cooling $-c_1\zeta E_q$ is excluded from the combined effect, the thermocline feedback still tends to warm the central-eastern Pacific (Fig. 9n) and it slightly cools the area near the east boundary. In Figs. 9o,p, we show the budget of the flux divergence $-\mu\partial_x(UT)$ and that of the combined effect from the damping and thermocline feedback $-c_1\zeta E_q + c_1\eta H$ averaged over the Niño-4 and Niño-3 regions, respectively. The positive contribution of $-\mu\partial_x(UT)$ and negative contribution of $-c_1\zeta E_q + c_1\eta H$ to the total SST

tendency are nearly balanced in the Niño-4 region at the CP El Niño phases, indicating the significant role of the nonlinear advection in the central Pacific. On the other hand, the combined effect from damping and thermocline feedback dominates the flux divergence in the eastern Pacific Niño-3 region. As is shown in Fig. S2 of the supplementary material, the thermocline feedback in the simplified model without the nonlinear zonal advection only warms the eastern Pacific and that model fails to reproduce realistic CP El Niño.

b. Traditional El Niño including super El Niño

1) TWO FORMATION MECHANISMS

Row 2 of Fig. 7 shows the lagged correlation between the Niño-3 SST index and different fields conditioned on the traditional El Niño phases. The lagged correlations up to 1-yr lag times for all fields are quite similar to those of the overall events (Fig. 6). Particularly, in addition to the positive correlation with the WWBs occurring a few months prior to the event peak, a significant negative correlation between Niño-3 SST and wind bursts is found around 1 year before the event peak, which corresponds

to the appearance of the EWBs. Such EWBs are crucial for inducing and accumulating heat content in the western Pacific that facilitates the following WWBs to trigger El Niño in the model, although these EWBs in real observations are weaker. This provides the first formation mechanism of the traditional El Niño.

(i) Formation mechanism 1

EWBs appear around 1 yr before a traditional El Niño event, inducing the buildup of anomalous warm SST and heat content in the western Pacific. Such EWBs (Levine and McPhaden 2016) are followed by a series of WWBs occurring a few months prior to the event peak, which lead to the eastward propagation of the anomalous warm SST and heat content to the eastern Pacific and simultaneously alter the zonal circulation of the anomalous atmosphere winds.

In addition to the strong lagged correlation at a lag time around 1 yr as discussed above, row 2 of Fig. 7 also reveals moderate correlations between the Niño-3 index and both SST and thermocline in the central and western Pacific at time leads up to 3.5 yr. Meanwhile, moderate correlations with westerly and easterly zonal atmosphere winds in the western and eastern Pacific are also noticeable. These lagged correlation structures resemble those in the CP El Niño phase. In fact, a second formation mechanism of the traditional El Niño is linked to its previous CP El Niño.

(ii) Formation mechanism 2

In the presence of a CP El Niño event, both the anomalous SST and heat content have been transported to the central Pacific. Thus, instead of triggering heat content in the western Pacific by a series of EWBs, a single strong WWB right after a CP El Niño event is sufficient to push the anomalous warm water and heat content from the central Pacific to the eastern Pacific and form the traditional El Niño. No significant EWB is observed during this process.

To summarize, the two mechanisms share the common feature that WWBs trigger a traditional El Niño event. The difference between the two mechanisms lies in the heat content accumulation prior to the WWBs, where EWBs have a main contribution in mechanism 1 while CP El Niño plays an important role in mechanism 2. The following case studies will characterize both the mechanisms.

2) CASE STUDIES

Figure 10 shows two case studies that belong to the formation mechanisms 1 and 2, respectively.

Case study 1 illustrates the formation mechanism 1, where a moderate traditional El Niño and a super

El Niño are shown. Both events are isolated traditional El Niño that are not preceded by CP El Niño. Starting from $t = 1236$, a series of EWBs occur, which induces strong ocean Rossby waves in the western Pacific. These ocean Rossby waves propagate westward and form Kelvin waves after they reflect at the western Pacific boundary. During this process, heat content represented by thermocline depth is accumulated in the western Pacific, due to the positive amplitudes of both Kelvin and Rossby waves [see (S1)]. Then around $t = 1237$ a series of WWBs arise and bring about significant ocean Kelvin waves, which transport strong heat content and SST to the eastern Pacific and form a super El Niño. Finally, the Kelvin waves reflect at the eastern Pacific boundary and induce Rossby waves. These reflected Rossby waves are crucial in obtaining the realistic duration of the El Niño event. In fact, a case study in the supplementary material (see Fig. S3) shows that in the absence of the reflected Rossby waves the durations of the simulated traditional El Niño events are much shorter than those in nature. Similarly, the moderate traditional El Niño around $t = 1243$ is formed by the EWB–WWB mechanism. Looking at the relationship between heat content, ocean waves, and the atmosphere wind forcing in (6) and (S1), it is clear that the most direct source of the heat content buildup in the western Pacific in the model is easterly winds. This explains the strong EWBs in the preconditioning phase of a traditional El Niño. Note that the accumulated heat content in the western Pacific induced by the ocean Rossby waves is one of the most crucial factors that determine the strength of the El Niño in the eastern Pacific. Other factors affecting the El Niño strength include the amplitudes of both the WWBs and EWBs as well as the profile of their interannual profile. In Fig. 9, budget analysis shows that thermocline feedback plays the most important role in the formation of a traditional El Niño event, which is different from the CP El Niño where the flux divergence is one of the dominant factors.

Case study 2 in Fig. 10 describes the formation of a super El Niño in the eastern Pacific based on mechanism 2. First, a series of CP El Niño is developed ($t = 635$ – 638), accompanied by a mean easterly trade wind anomaly. Different from case study 1 where EWBs trigger heat content in the western Pacific, which prepares the formation of a traditional El Niño, the heat content in case study 2 has already been increased in the central Pacific region due to the CP El Niño (Kug et al. 2009). Then, with the presence of a series of WWBs ($t = 638$), both the heat and anomalous warm SST move from the central Pacific to the eastern Pacific and a super El Niño is formed. Clearly, the fact that no obvious EWB

FIG. 10. Case studies of the two formation mechanisms for the traditional El Niño: 1) one super El Niño followed by a La Niña and one moderate traditional El Niño and 2) a series of CP El Niño followed directly by a super El Niño and then a La Niña. Shown are (a),(g) atmosphere wind including the mean trade wind anomaly $u + \hat{u}_p$; (b),(h) thermocline depth H ; (c),(i) SST; (d),(j) ocean Kelvin waves; and (e),(k) ocean Rossby waves. All these fields are the direct model output and contain both the interannual variability and the high-frequency component. (f),(l) The time series of the wind activity at the peak x^* of the zonal profile (see Fig. 1b) and its 90-day running average (brown curve). All variables are shown at equator.

appears prior to the WWBs in case study 2 distinguishes the two mechanisms.

3) WAITING TIME BETWEEN A CP EL NIÑO EVENT AND ITS PREVIOUS ADJACENT TRADITIONAL EL NIÑO

As was seen in the previous subsection, a traditional El Niño event is likely to follow a series of CP El Niño events, which is partly due to the prescribed high transition rate from state 1 to state 2 based on the observational evidence. Yet, understanding the reverse situation (i.e., the waiting time between a CP El Niño and its previous adjacent traditional El Niño) is equally crucial in interpreting the mutual interactions between these two types of El Niño.

The distribution of such waiting time can be obtained by searching both types of events from the long-term model simulations. Recall that the definition of a CP event was provided in Table 1. For the traditional El Niño events, in addition to the standard definition with $\text{Niño-3} > 0.5$, different threshold values of Niño-3 are adopted, which allows us to include from all traditional

El Niño events to only super-El Niño events. Note that when the threshold value becomes large, more moderate and weak traditional El Niño events will appear between a CP El Niño and its previous adjacent super El Niño.

In Fig. 11a, the standard definition of the traditional El Niño is adopted (i.e., $\text{Niño-3} > 0.5$), which leads to a bimodal distribution of the waiting time. The major mode peaks around 3.5 yr while the peak of the minor mode is around 1 year. Such bimodality implies two different scenarios. In the first scenario, a traditional El Niño is followed by a La Niña, and then after a 1–2-yr quiescent period, a CP El Niño appears. Examples are shown in Fig. 11e. In fact, the observed episodes during 1987–90 and 1998–2002 both belong to the first situation. In the second scenario, a CP El Niño occurs right after a traditional El Niño, where an example is shown in Fig. 11f. Clearly, this second situation occurs less frequently than the first one in the model simulation. It is also observed only in years 1977–78 during the last 50 yr (Chen et al. 2015). Particularly, with the enhancing of the threshold of the Niño-3 index for the traditional El

FIG. 11. (a)–(d) Distribution of the waiting time $\tau_{E \rightarrow C}$ between a CP event and its previous traditional El Niño event with four different threshold values of counting the traditional El Niño before each CP event. (e),(f) Examples of CP El Niño following traditional one with two different forms, showing from left to right the Hovmöller diagram of SST, the Markov states, and the Niño indices. Here, each black bar indicates the waiting periods between one CP El Niño event and its previous traditional El Niño.

Niño, the minor mode disappears and the resulting distribution of waiting time is unimodal (Fig. 11d), which implies that the second situation happens only when the traditional El Niño has weak amplitude. In fact, a La Niña usually follows a moderate or strong traditional El Niño in order to discharge the heat content.

c. La Niña

The formation mechanism of La Niña is relatively simple. Lagged correlations in row 3 of Fig. 7 show that both the SST anomalies and the heat content propagate from the western Pacific to eastern Pacific. A positive correlation between minus Niño-3 index and SST anomalies at a lag time of 2 yr in row 2 is consistent with that at 2 yr lead time in row 1 that is conditioned on the traditional El Niño phases. Both indicate that La Niña serves as the discharge phase of El Niño. This argument is further validated by the lagged correlation between the Niño-3 index and wind bursts a_A , where WWBs occur 1.5 yr prior to the La Niña event peak and trigger the previous recharge phases that corresponds to a traditional El Niño event.

In the case studies shown in Fig. 10, a traditional El Niño is always followed by a La Niña, and the anomalous Walker circulation at La Niña phases is nearly opposite to that at traditional El Niño phases.

In addition to the formation mechanisms of the two types of El Niños as well as La Niña, the simple modeling framework also allows us to study many other related issues. It is shown that strong wind bursts are only necessary but not sufficient conditions for triggering the El Niños (see Fig. S1 in the supplementary material) (Hu et al. 2014; Fedorov et al. 2015; Roulston and Neelin 2000; Levine and Jin 2010). In addition, the insufficiency of various simplifications in the coupled model are illustrated with concrete examples (see Figs. S2 and S3).

5. Discussion and conclusions

Understanding the formation mechanisms of ENSO diversity has significant impact on global climate and relevance for seasonal forecasts. Recently, a simple modeling framework has been developed that captures

the statistical properties and many key characteristics of ENSO diversity. This simple modeling framework is physically consistent and amenable to detailed analysis. The starting model in this framework is a simple ocean–atmosphere model that is deterministic, linear, and stable. Then several key features are incorporated into the coupled system to capture several major causes of the ENSO diversity. These key features are state-dependent stochastic wind bursts and nonlinear advection of SST, which allow effective transitions between different ENSO states (Thual et al. 2016; Chen and Majda 2016, 2017).

In this article, the formation mechanisms of ENSO diversity based on this simple coupled model are systematically studied. Robust statistics and dynamical features are obtained based on a long (5000 yr) simulation of the coupled model. In section 3, the statistics of atmosphere and ocean variables in the model are compared with the observations in different regions ranging from the western to eastern Pacific. The statistics of the observed SST and thermocline in different regions all lie within the confidence interval of the model, and the model succeeds in capturing the skewness directions of the three Niño SST indices. The non-Gaussian features in the atmosphere zonal winds are also reproduced by the model although the variance is underestimated. Lagged correlation shows that the overall ENSO formation mechanism of the model in terms of zonal winds, thermocline depth, and SST are all similar to that of the observations. In section 4, the formation mechanisms of different types of El Niño and La Niña based on the coupled model are discussed. Budget analysis (Fig. 9) indicates that the zonal advection is crucial in the formation of the CP El Niño, as is observed in nature (Su et al. 2014). The comparison of the heat budget in Niño-3 and Niño-4 regions justifies the fact that zonal advection plays an important role for the formation of the CP El Niño while the thermocline feedback is the dominant factor of the traditional El Niño (Kug et al. 2009). Next, two mechanisms are revealed for the formation of traditional El Niño events (Fig. 10). The first mechanism suggests a preferred profile of the wind activity, where EWBs up to 2 yr prior to the event peak are responsible for the heat content buildup and the following WWBs trigger El Niño. The underlying discharge–recharge mechanism in the coupled model has some common features as that described in earlier work (Wang and Picaut 2004), but here the role of the state-dependent wind bursts in triggering the ENSO cycles and the buildup and shutdown of wind bursts activity relying on the ENSO events are both emphasized. In addition, the wave propagation mechanisms in the model here are more explicit than those in the simple oscillators (Jin 1997; Weisberg

and Wang 1997). As is shown in the budget analysis (Fig. 9) and a simpler version of the coupled model in Thual et al. (2016), the advection is not necessary in this modeling framework for the formation of the traditional El Niño, which differs from the advective–reflective oscillator (Picaut et al. 1997). It is also shown in the supplementary material that the eastern boundary reflection is crucial for reaching the realistic duration of the El Niño events, which was not included in the simple delayed oscillator (Suarez and Schopf 1988; Battisti and Hirst 1989). The second mechanism links the CP El Niño and traditional El Niño, where WWBs occurring right after a CPEl Niño are sufficient to push the anomalous warm water to the eastern Pacific. This mechanism no longer involves an explicit oscillator structure, and the emphasis here is more toward the buildup and transportation of the heat content as well as the transitions between the two types of El Niño (Chen and Majda 2016). On the other hand, La Niña is shown to be the discharge phase of traditional El Niño as is widely accepted in the literature. In the supplementary material, it is also illustrated with example (Fig. S1) that strong wind bursts in the coupled model are only necessary but not sufficient conditions to trigger El Niño events, as observed in nature (Hu et al. 2014; Fedorov et al. 2015; Roulston and Neelin 2000; Levine and Jin 2010). It is worthwhile pointing out that the state-dependent multiplicative wind burst noise [(10)] in the simple modeling framework here has the potential to significantly improve the representation of ENSO in GCMs. In fact, in a recent work (Christensen et al. 2017), a simple stochastic parameterization with multiplicative noise was incorporated into NCAR CAM4. Compared with the additive noise, the multiplicative noise leads to a significant improvement of the power spectrum and the magnitude of ENSO in NCAR CAM4. In addition, consistent with the mechanisms shown in Fig. 10 here, the initiation and termination of El Niño events become more realistic in NCAR CAM4 in the presence of the multiplicative noise.

In addition to the dynamical and statistical features mentioned above that are consistent with observations, there are a few shortcomings due to the simple nature of the coupled model, which require improvements in the future. First, instead of using a constant $\hat{a}_p$ that represents the mean trade wind anomaly, incorporating a gradually increasing function of the trade wind anomaly that lasts for a few decades into the model is more consistent with nature. To represent this decadal variability in a more realistic way, the same value for the trade wind acceleration is also preferred for all the Markov states. In addition, the “effective background SST gradient” induced by the acceleration of the trade wind and its contribution to the CP El Niño events

compared with the observed background SST gradient also require further studies. Second, the first mechanism for the traditional El Niño (Fig. 10, top) involves a preferred structure of the wind bursts with EWB–WWB profile before the event peak. The EWBs induce downwelling Rossby waves in the western Pacific that account for the heat content buildup [see (6) and (S1)] and the WWBs trigger El Niño. This recharge–discharge mechanism is consistent with nature. Yet, the timing of the WWB occurrence is slightly delayed according to the lagged correlation analysis (6 months before the event peak) compared with nature (6–9 months before the event peak), and the EWBs in the model seem to be much stronger than those in nature. One obvious reason for the strong amplitude of the EWBs is that the EWBs in the coupled model is constrained in the western Pacific while EWBs in nature can extend to the eastern Pacific (Chiodi and Harrison 2015; Puy et al. 2016). If a wider zonal domain of the EWBs is used while the total zonal flux of the EWBs remains the same, then the strength of EWBs will decrease and resemble those in nature. Since EWBs may initiate from off-equator regions (Hu and Fedorov 2016), including more meridional basis functions can be a potential improvement of the model. More meridional structure will also provide alternative sources for the heat content buildup of both the traditional El Niño and the CP El Niño in addition to the one in the current model that results from the first ocean Rossby wave induced by the EWBs or easterly trade wind. This may also have potential to allow occasional occurrences of CP El Niño even without the appearance of strong easterly trade wind. In addition, the current model here does not include the seasonal synchronization. In a recent work, a cloud radiative feedback with seasonal variations of convective activity is included into the simple modeling framework that allows the ENSO to peak in boreal winter as in observations (Thual et al. 2017) while the other statistical and dynamical features of the model are unchanged. Finally, the coupled system has a relatively simple atmosphere model [(1)]. More refined structure that takes into account the details of the MJO and the convectively coupled waves can be incorporated into the model that can improve the modulations of the wind burst activity (Puy et al. 2016).

The transition rates in the three-state Markov process in this work are determined based on the observational period over the last a few decades, and the model simulation aims at capturing the features as observed in the same period. In the future for climate change scenarios, both the transition rates and parameter values in the simple model can be modified accordingly. Then the simple model may have the potential to reveal new phenomena and mechanisms. In fact, as is indicated in

Fig. 8, the patterns in regime III may become dominant if the easterly trade wind keeps accelerating. In fact, the link between the variation in a_p with the decadal sea level variations needs to be analyzed. The studies of these possible changes are left to future work. In addition, developing an effective data assimilation scheme for the coupled model that is used for the ENSO prediction is another future direction.

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