

Observations and Mechanisms of a Simple Stochastic Dynamical Model

Capturing El Niño Diversity

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ABSTRACT

18 The El Niño-Southern Oscillation (ENSO) has significant impact on glob-
19 al climate and seasonal prediction. Recently, a simple modeling framework
20 was developed that automatically captures the ENSO diversity, where state-
21 dependent stochastic wind bursts and nonlinear advection of sea surface tem-
22 perature are coupled to a simple ocean-atmosphere model that is otherwise
23 deterministic, linear and stable. In this article, the coupled model is compared
24 with observations using reanalysis data during the last 34 years, where the ob-
25 served non-Gaussian statistics and the overall mechanisms of ENSO are both
26 captured by the model. Then the formation mechanisms of different types of
27 El Niño based on the model are systematically studied. First, ocean Rossby
28 waves induced by mean easterly trade wind anomalies facilitate the heat con-
29 tent buildup, which together with the reflected ocean Kelvin waves and the
30 nonlinear advection create the central Pacific (CP) El Niño. Secondly, two
31 formation mechanisms are revealed for the traditional El Niño, including the
32 super El Niño. The first mechanism indicates an optimal wind structure with
33 easterly wind bursts (EWBs) leading westerly wind bursts (WWBs) before
34 the event peak, where the EWBs build up heat content that is transported east-
35 ward by the WWBs. The second mechanism links the two types of El Niño,
36 where the CP El Niño prior to a traditional El Niño is responsible for the heat
37 content buildup of the latter. This article also highlights the mechanisms of
38 La Niña formation and El Niño breaking down as well as the importance of
39 the nonlinear advection and boundary reflections.

40 **1. Introduction**

41 The El Niño-Southern Oscillation (ENSO) is the most prominent interannual climate variability
42 on earth with large ecological and societal impacts. It consists of a cycle of anomalously warm El
43 Niño conditions and cold La Niña conditions with considerable irregularity in amplitude, duration,
44 temporal evolution and spatial structure. The well-known traditional El Niño (or eastern Pacific
45 (EP) El Niño) involves anomalous warm sea surface temperature (SST) in the equatorial eastern
46 Pacific ocean, where its atmospheric response is the eastward shift of the anomalous Walker cir-
47 culation with strong convection occurring near the west coast of America (Clarke 2008). In recent
48 decades, a different type of El Niño has been frequently observed, which is called the central Pa-
49 cific (CP) El Niño (Lee and McPhaden 2010; Kao and Yu 2009; Ashok et al. 2007b; Kug et al.
50 2009; Larkin and Harrison 2005; Guilyardi 2006). The CP El Niño is characterized by warm SST
51 anomalies confined to the central Pacific, flanked by colder waters to both east and west. Such
52 zonal SST gradients result in an anomalous two-cell Walker circulation over the tropical Pacific,
53 with strong convection located in the central Pacific (Kug et al. 2009). In addition to the distinct
54 climate patterns in the equatorial Pacific region, different types of El Niño and La Niña also have
55 different teleconnections that affect the global climate and seasonal prediction (Ashok et al. 2007b;
56 Weng et al. 2009; Capotondi et al. 2015).

57 The significant impact of ENSO requires a comprehensive understanding of the underlying for-
58 mation mechanisms of different types El Niño as well as the development of dynamical models
59 that reproduce the ENSO diversity. However, most of the current climate models have biases in
60 simulating the ENSO diversity. Some general circulation models (GCMs) are able to reproduce
61 only one single type of El Niño (Ham and Kug 2012; Kug et al. 2012; Ault et al. 2013; Capotondi
62 et al. 2015). Other climate models, despite their ability to reproduce the observed diversity of

63 ENSO to some extent, typically overestimate the amplitude of the ENSO interannual variability
64 and misrepresent the simulated frequency and duration of the El Niño events (Wittenberg 2009;
65 Kug et al. 2010). In addition, these climate models are too complicated to be employed for under-
66 standing the fundamental mechanisms of ENSO diversity.

67 Recently, a simple modeling framework was developed that automatically captures the statistical
68 diversity of ENSO (Thual et al. 2016; Chen and Majda 2016a,b). This simple modeling framework
69 is physically consistent and amenable to detailed analysis, which facilitates the understanding of
70 the formation mechanisms of ENSO diversity.

71 In this simple modeling framework, the starting model involves a coupled ocean-atmosphere
72 model that is deterministic, linear and stable (Kleeman 2008; Kleeman and Moore 1997; Moore
73 and Kleeman 1999). Then systematic strategies are developed for incorporating several major
74 causes of the ENSO diversity into the coupled system. First, a stochastic parameterization of the
75 wind bursts including both westerly and easterly winds is coupled to the simple ocean-atmosphere
76 system, where the amplitude of the wind bursts depends on the strength of SST in the western
77 Pacific warm pool through a Markov jump stochastic process. Such a coupled model is funda-
78 mentally different from the Cane-Zebiak (Zebiak and Cane 1987) and other nonlinear models (Jin
79 1997; Timmermann et al. 2003), for which the internal instability rather than the external wind
80 bursts maintains the ENSO cycle. It is shown that (Thual et al. 2016) in addition to recovering
81 traditional moderate El Niño and super El Niño as well as La Niña events, the coupled model is
82 able to capture key features of the observational record in the eastern Pacific. Secondly, a simple
83 nonlinear zonal advection with no ad-hoc parameterization of the background SST gradient is in-
84 troduced that creates a coupled nonlinear advective mode of SST. In addition, due to the recent
85 multidecadal strengthening of the easterly trade wind (England et al. 2014; Sohn et al. 2013; Mer-
86 rifield and Maltrud 2011), a mean easterly trade wind anomaly is incorporated into the stochastic

parameterization of the wind activity. The combined effect of the nonlinear zonal advection, the enhanced mean easterly trade wind anomaly and the effective stochastic noise facilitates the intermittent occurrence of the CP El Niño (Chen and Majda 2016a). Then, a three-state Markov jump stochastic process is developed to drive the stochastic wind bursts. It emphasizes the distinct properties of the wind activity at different ENSO phases and describes the state-dependent transition mechanisms in a simple and effective fashion. This simple stochastic switching process allows the coupled model to simulate different types of ENSO events with realistic features (Chen and Majda 2016a,b).

In this article, the formation mechanisms of ENSO diversity based on this simple coupled model are systematically studied and are compared with observational record using reanalysis data during the last 34 years. To begin with, the statistics in the observational data associated with different variabilities are presented. It is shown that both the non-Gaussian statistical features and the overall ENSO formation mechanisms of the coupled model are highly consistent with nature. Then the formation mechanisms of different types of El Niño events based on the coupled model are explained with detailed analysis and concrete examples, where the significant roles of the underlying ocean Kelvin and Rossby waves as well as the wind burst activities are highlighted. First, deterministic nonlinear advective modes are utilized to understand the formation mechanism of the CP El Niño (Chen and Majda 2016a). It is shown that the mean easterly trade wind anomalies are essential in triggering ocean Rossby waves that not only facilitate the heat content buildup in the western Pacific but induce weak ocean Kelvin waves due to the imperfect reflection at the western Pacific boundary as well. Such consecutive Kelvin waves with weak amplitudes together with the nonlinear advection are crucial in transporting the anomalous warm SST to the central Pacific and maintaining the climate patterns there. By incorporating effective random wind bursts noise into the coupled system, more realistic features of CP El Niño events are reproduced. Sec-

111 on only, two distinct mechanisms are revealed by the coupled model in the formation of traditional
112 El Niño and super El Niño events. In addition to the bursts of westerly winds preceding the event
113 peak that drive anomalous warm SST to the eastern Pacific (Harrison and Vecchi 1997; Vecchi
114 and Harrison 2000; Tziperman and Yu 2007; Hendon et al. 2007), a series of strong easterly wind
115 bursts (EWBs) (Hu and Fedorov 2016) are observed prior to those westerly wind bursts (WWBs)
116 that contribute to the buildup of the heat content in the western Pacific. Such EWBs-WWBs wind
117 structure provides the first formation mechanism of a traditional El Niño event. The second forma-
118 tion mechanism links the CP El Niño and traditional El Niño. Since the heat content has already
119 been transported to the central Pacific region during a CP event (Kug et al. 2009), a single WWB
120 right after the CP event is sufficient to push the anomalous warm water to the eastern Pacific and
121 creates a traditional El Niño event. Distinct from the first mechanism, no pronounced EWB is
122 involved in this second formation mechanism. Next, according to the lagged correlations, La Niña
123 is shown to be the discharge phase following its preceding traditional El Niño and no significant
124 wind burst activity is observed at La Niña phases.

125 In addition to the formation of different ENSO events, the mechanism of El Niño breaking down,
126 such as the one observed in year 2014 (Hu and Fedorov 2016), is also explained. Furthermore,
127 examples show that strong wind bursts are necessary but not sufficient conditions for triggering
128 El Niño events (Fedorov et al. 2015; Hu et al. 2014; Roulston and Neelin 2000; Levine and Jin
129 2010). Finally, the deficiencies in various simplified versions of the coupled model highlight the
130 important roles of both the nonlinear advection and the eastern Pacific boundary reflections in the
131 coupled model. The former is crucial in preventing the anomalous warm water being transported to
132 the eastern Pacific at CP El Niño phases while the latter is essential in obtaining realistic durations
133 of the traditional El Niño and particularly the super El Niño.

134 The remainder of this article is as follows. The coupled model is presented in Section 2, along
 135 with a description of the processing of observational data. Section 3 shows the statistical features
 136 and the overall ENSO formation mechanism of the coupled model, as compared with the obser-
 137 vations. The formation mechanisms of different types of El Niño and La Niña, in light of both
 138 conditional statistics and case studies, are demonstrated in Section 4. In the same section, the
 139 underlying reasons for El Niño breaking down and some strong wind bursts not triggering any
 140 El Niño are also revealed. The crucial roles of both the nonlinear advection and the reflection
 141 boundary condition in the eastern Pacific are emphasized in Section 5. Summary conclusions are
 142 included in Section 6.

143 **2. Model and Observational Data**

144 *a. Coupled ENSO Model*

145 The ENSO model consists of a non-dissipative atmosphere coupled to a simple shallow-water
 146 ocean and SST budget in the interannual time scale (Thual et al. 2016; Chen and Majda 2016a,b):

147 *Atmosphere model:*

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

148 *Ocean model:*

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

149 *SST model:*

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

150 with

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

151 In (1)–(4), x is zonal direction and τ is interannual time, while y and Y are meridional direction
 152 in the atmosphere and ocean, respectively. The u, v are zonal and meridional winds, θ is potential
 153 temperature, U, V , are zonal and meridional currents, H is thermocline depth, T is SST, E_q is
 154 latent heating, and τ_x is zonal wind stress, where u_p represents the wind bursts (see hereafter). All
 155 variables are anomalies from an equilibrium state, and are non-dimensional. The coefficient c_1
 156 is a non-dimensional ratio of time scales, which is of order $O(1)$. The term u_p in (4) describes
 157 stochastic wind burst activity. The atmosphere extends over the entire equatorial belt $0 \leq x \leq L_A$
 158 with periodic boundary conditions, while the Pacific ocean extends over $0 \leq x \leq L_O$ with reflection
 159 boundary conditions for the ocean model and zero normal derivative at the boundaries for the SST
 160 model. The parameter values of the coupled model are shown in Appendix A, and are the same as
 161 those in (Chen and Majda 2016a,b).

162 The above model retains a few essential processes that model the ENSO dynamics in a simple
 163 fashion. Latent heating E_q , proportional to SST, is depleted from the ocean and forces an atmo-
 164 spheric circulation. The resulting zonal wind stress τ_x in return forces an ocean circulation that
 165 imposes feedback on the SST through thermocline depth anomalies H . This thermocline feedback
 166 η is more significant in the eastern Pacific, as shown in Figure 1.

167 The coupled model introduces unique theoretical elements such as a non-dissipative atmosphere
 168 consistent with the skeleton model for the Madden-Julian oscillation (MJO) (Majda and Stech-
 169 mann 2009, 2011), valid here on the interannual timescale and suitable to describe the dynamics

170 of the Walker circulation (Majda and Klein 2003; Stechmann and Ogrosky 2014; Stechmann and
171 Majda 2015). In addition, the meridional axis y and Y are different in the atmosphere and ocean
172 as they each scale to a suitable Rossby radius. This allows for a systematic meridional decompo-
173 sition of the system into the well-known parabolic cylinder functions (Majda 2003), which keeps
174 the system low-dimensional (Thual et al. 2013) and will be discussed hereafter.

175 The coupled system (1)–(4) without the nonlinear zonal advection in (3) has been systematically
176 studied in (Thual et al. 2016). It succeeds in recovering the traditional El Niño with occasional
177 super El Niño as well as capturing the observed ENSO statistics in the eastern Pacific. Note
178 that if the stochastic wind burst u_p is further removed, the resulting coupled system is linear,
179 deterministic and stable (Kleeman 2008; Kleeman and Moore 1997; Moore and Kleeman 1999).
180 Such a coupled model is fundamentally different from the Cane-Zebiak (Zebiak and Cane 1987)
181 and other nonlinear models (Jin 1997; Timmermann et al. 2003) where the internal instability
182 rather than the external wind bursts plays the role of maintaining the ENSO cycles.

183 The observational significance of the zonal advection has been shown for the CP El Niño (Kug
184 et al. 2009; Su et al. 2014). Different from the previous works (Jin and An 1999; Dewitte et al.
185 2013) where the advection is mostly linear and requires ad hoc parameterization of the background
186 SST gradient, a simple nonlinear advection is adopted in (3) that contributes significantly to the
187 SST tendency. Such nonlinear advection provides the mechanism of transporting anomalous warm
188 water to the central Pacific region by the westward anomalous ocean zonal current. Importantly,
189 when stochasticity is included in the wind activity u_p , this nonlinear zonal advection involves the
190 contribution from both mean and fluctuation, the latter of which is usually ignored in the previous
191 works. The combined effect of this nonlinear advection, a mean easterly trade wind anomaly and
192 effective stochastic noise was shown to facilitate the intermittent occurrence of the CP El Niño
193 with realistic features (Chen and Majda 2016a).

1) MERIDIONAL TRUNCATION

For the purpose of solving and understanding the coupled system and retaining the key dynamical features, a simple meridional truncation is applied to the coupled model (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 $\phi_2 = (4\pi)^{-1/4} (2y^2 - 1) \exp(-y^2/2)$. The oceanic parabolic cylinder functions $\psi_m(Y)$ are identical to the expressions of 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 the zeroth-order parabolic cylinder function ϕ_0 . This is known to excite only the Kelvin and first Rossby atmospheric equatorial waves, of amplitude 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 , $\tau_x = \tau_x \psi_0$. This is known to excite only the Kelvin and first Rossby oceanic equatorial waves, of amplitude K_O and R_O . Similarly, for the SST model we assume a truncation ψ_0 , $T = T \psi_0$. With these truncations, the coupled ENSO model (1)–(3) becomes:

Atmosphere model:

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

Ocean model:

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

213 *SST model:*

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

214 where χ_A and χ_O are the projection coefficients from ocean to atmosphere and from atmosphere to
 215 ocean, respectively, because of the different extents in their meridional bases. Due to the absence
 216 of dissipation in the atmosphere, the solvability condition requires a zero equatorial zonal mean
 217 of latent heating forcing $\langle E_q \rangle$ (Majda and Klein 2003; Stechmann and Ogrosky 2014). Note that
 218 when meridional truncation is implemented, a projection coefficient $\chi \approx 0.65$ appears in front of
 219 the nonlinear term (Majda and Stechmann 2011), which is absorbed into the nonlinear advection
 220 coefficient μ for the notation simplicity.

221 Periodic boundary conditions are adopted for the atmosphere model (5). Reflection boundary
 222 conditions are adopted for the ocean model (6),

$$K_O(0, t) = r_W R_O(0, t), \quad R_O(L_O, t) = r_E K_O(L_O, t), \quad (8)$$

223 where $r_W = 0.5$ representing partial loss of energy in the west Pacific boundary across Indonesian
 224 and Philippine and $r_E = 0.5$ representing partial loss of energy due to the north-south propagation
 225 of the coast Kelvin waves along the eastern Pacific boundary. For the SST model, no normal
 226 derivative at the boundary of T is adopted, i.e. $dT/dx = 0$.

227 Now instead of solving the original system (1)–(3), we solve the system with meridional trunca-
 228 tion (5)–(7). The physical variables can be easily reconstructed in the following way.

$$\begin{aligned} u &= (K_A - R_A)\phi_0 + (R_A/\sqrt{2})\phi_2, \\ \theta &= -(K_A + R_A)\phi_0 - (R_A/\sqrt{2})\phi_2, \\ v &= (4\partial_x R_A - \bar{H}A - S^\theta)(3\sqrt{2})^{-1}\phi_1, \\ U &= (K_O - R_O)\psi_0 + (R_O/\sqrt{2})\psi_2, \\ H &= (K_O + R_O)\psi_0 + (R_O/\sqrt{2})\psi_2. \end{aligned} \quad (9)$$

For the convenience of demonstration, all the variables shown below are at equator. See (Majda and Stechmann 2011; Thual et al. 2016; Chen and Majda 2016a,b) for more details.

2) STOCHASTIC WIND BURST PROCESS

Stochastic parameterization of the wind activity is added to the model that represents several important ENSO triggers such as the WWBs (Harrison and Vecchi 1997; Vecchi and Harrison 2000; Tziperman and Yu 2007), the EWBs (Hu and Fedorov 2016), as well as the convective envelope of the MJO (Hendon et al. 2007; Puy et al. 2016). It also includes the recent multidecadal strengthening of the easterly trade wind anomaly. The wind bursts u_p reads:

$$u_p = a_p(\tau)s_p(x)\phi_0(y), \quad (10)$$

with amplitude $a_p(\tau)$ and fixed zonal spatial structure $s_p(x)$ shown in Figure 1. Here, $\phi_0(y)$ equals to the first parabolic cylinder function of the atmosphere. Both the wind burst perturbations (Tziperman and Yu 2007) and the strengthening of the trade wind anomaly (England et al. 2014; Sohn et al. 2013) are localized over the western equatorial Pacific according to the observations and for simplicity they share the same zonal extent.

The evolution of wind burst amplitude a_p reads:

$$\frac{da_p}{d\tau} = -d_p(a_p - \hat{a}_p(T_W)) + \sigma_p(T_W)\dot{W}(\tau), \quad (11)$$

where d_p is dissipation and $\dot{W}(\tau)$ is a white noise source, representing the intermittent nature of the wind bursts at interannual timescale. The amplitude of the wind burst noise source σ_p depends on T_W (See Eq. (12)), which is the average of SST anomalies in the western half of the equatorial Pacific ($0 \leq x \leq L_O/2$). Note that this state-dependent wind amplitude is fundamentally different from those in previous works (Jin et al. 2007; Levine and Jin 2015, 2010) that rely on the eastern Pacific SST. The term $\hat{a}_p < 0$ represents the mean strengthening of the easterly trade

249 wind anomaly. Corresponding to $\hat{a}_p < 0$, the direct response of the surface wind associated with
 250 the Walker circulation at the equatorial Pacific band is shown in Panel (c) of Figure 1, which is
 251 similar to the observed intensification of the Walker circulation in recent decades (England et al.
 252 2014; Sohn et al. 2013).

253 3) A THREE-STATE MARKOV JUMP STOCHASTIC PROCESS

254 Due to the fact that the ENSO diversity is associated with the wind activity with distinct features
 255 (Thual et al. 2016; Chen and Majda 2016a), a three-state Markov jump stochastic process (Gar-
 256 diner et al. 1985; Lawler 2006; Majda and Harlim 2012) is adopted to describe the wind activity.
 257 Here, State 2 primarily corresponds to the traditional El Niño and State 1 to the CP El Niño while
 258 State 0 represents discharge and quiescent phases. The following criteria are utilized to determine
 259 the parameters in (11) in each state. First, strong wind bursts play an important role in triggering
 260 the traditional El Niño (Vecchi and Harrison 2000; Tziperman and Yu 2007; Hendon et al. 2007),
 261 which suggests a large noise amplitude σ_p in State 2. Secondly, the observational fact that an
 262 enhanced easterly trade wind accompanies with the CP El Niño since 1990s indicates a negative
 263 (easterly) mean $\hat{a}_p$ in State 1. To obtain the CP El Niño, the amplitude of $\hat{a}_p$ and the stochastic
 264 noise must be balanced (Chen and Majda 2016a). This implies a moderate noise amplitude in
 265 State 1, which also agrees with observations (Chen et al. 2015). Finally, only weak wind activity
 266 is allowed in the quiescent state and the discharge phase with La Niña (State 0). Thus, the three
 267 states are given by

$$\begin{aligned}
 \text{State 2:} \quad & \sigma_{p2} = 3.75, \quad d_{p2} = 5.1, \quad \hat{a}_{p2} = -0.25, \\
 \text{State 1:} \quad & \sigma_{p1} = 1.2, \quad d_{p1} = 5.1, \quad \hat{a}_{p1} = -0.25, \\
 \text{State 0:} \quad & \sigma_{p0} = 0.5, \quad d_{p0} = 5.1, \quad \hat{a}_{p0} = 0,
 \end{aligned} \tag{12}$$

268 respectively, where $d_p = 5.1$ represents a relaxation time around 6.67 days. Note that the same
 269 mean easterly trade wind anomaly as in State 1 is adopted in State 2 due to the fact that both
 270 the traditional and the CP El Niño occurred during the last 25 years. Since the amplitude of the
 271 stochastic noise dominates the mean easterly wind in State 2, this mean state actually has little
 272 impact on simulating the traditional El Niño events. On the other hand, to guarantee no El Niño
 273 event occurring in the quiescent phase, no mean trade wind anomaly is imposed in State 0. With
 274 such choice of the parameters, both the amplitude and the timescale of the wind burst activity are
 275 similar to nature.

276 The local transition probability from State i to State j with $i \neq j$ for small $\Delta\tau$ is defined as
 277 follows

$$P(\sigma_p(\tau + \Delta\tau) = \sigma_{pj} | \sigma_p(\tau) = \sigma_{pi}) = v_{ij}\Delta\tau + o(\Delta\tau), \quad (13)$$

278 and the probability of staying in State i is given by

$$P(\sigma_p(\tau + \Delta\tau) = \sigma_{pi} | \sigma_p(\tau) = \sigma_{pi}) = 1 - \sum_{j \neq i} v_{ij}\Delta\tau + o(\Delta\tau). \quad (14)$$

279 Importantly, the transition rates v_{ij} (with $i \neq j$) depend on T_w , implying the state-dependence of
 280 the wind bursts (Levine et al. 2016; Lopez and Kirtman 2013; Lopez et al. 2013). A transition v_{ij}
 281 (with $i < j$) from a less active to a more active state is more likely when $T_w \geq 0$ and vice versa.
 282 This allows for example a rapid shutdown of wind burst activity followed by extreme El Niño
 283 events, as in nature.

284 The transition rates are chosen in accordance with the observational record. A higher transition
 285 probability from State 2 to State 0 is adopted compared with that to State 1, representing the
 286 situation that the traditional El Niño is usually followed by the La Niña rather than the CP El Niño
 287 (e.g., years 1963, 1965, 1972, 1982, 1987 and 1998.) Likewise, starting from the quiescent phase,
 288 the model has a preference towards the occurrence of the CP El Niño rather than the eastern Pacific

289 super El Niño, as observed in years 1968, 1990 and 2001. See Appendix B for details. It is shown
290 in (Chen and Majda 2016b) that the model statistics is robust with respect to the perturbation of the
291 parameters, which indicates that a crude estimation of the transition rates is sufficient for obtaining
292 the ENSO diversity with realistic features.

293 The parameterization of the wind activity in (10)–(14) emphasizes the distinct properties of the
294 wind activity at different ENSO phases and describes effective state-dependent transitions in a
295 simple and effective fashion. It is different from those adopted in previous works that involve
296 many complicated and detailed structures, such as the central location and peak time of each wind
297 burst event (Gebbie and Tziperman 2009; Gebbie et al. 2007; Tziperman and Yu 2007) and the
298 separation of the linear and nonlinear parts of the wind activity (Levine and Jin 2015).

299 *b. Definitions of Different El Niño Events in the Coupled Model*

300 In order to compare the model simulations with the observational record, we make use of the
301 three well-known Nino SST indices: Nino 3 (150W-90W), Nino 3.4 (170W-120W) and Nino 4
302 (160E-150W).

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

308 Here, simple criteria are proposed to distinguish different El Niño events are proposed in light
309 of the distinct roles of each of the three states in the wind activity model. Specifically, situating
310 in State 1 is one of the necessary conditions for identifying the CP El Niño. Other conditions for
311 recognizing a CP El Niño event are $Nino\ 4 > Nino\ 3 > 0$. The reason to pick up $Nino\ 3 > 0$ here

312 is to exclude the La Niña-like events and such positive value of the averaged SST over Nino 3
 313 region isn't incompatible with the CP El Niño feature that cooling occurs near the eastern Pacific
 314 boundary. Note that replacing Nino 4 by Nino 3.4 for identifying the CP El Niño doesn't lead
 315 to qualitative difference in any of the statistical features shown below thanks to the constraint of
 316 situating in State 1. On the other hand, although strong wind bursts (State 2) trigger the traditional
 317 El Niño, the decaying phase of each traditional El Niño event from anomalously warm SST back to
 318 the normal condition corresponds to State 0. In addition, the ENSO discharge phase with La Niña
 319 also typically lies in State 0 with occasional occurrences in State 2. Thus, traditional El Niño and
 320 La Niña requires locating in either State 0 or 2. Other conditions for traditional El Niño are $Nino\ 3 > 0.5$
 321 and $Nino\ 3 > Nino\ 3.4$, where the latter serves to exclude CP-like events that occasionally
 322 occur in State 2, and those for La Niña are $Nino\ 3 < -0.5$. These criteria are summarized in Table
 323 1.

324 *c. Observational Data*

325 In this work, observational data is utilized to assess the realism of the model solutions. The
 326 following three observational datasets are utilized: 1) daily zonal winds at 850hPa from the
 327 NCEP/NCAR reanalysis (Kalnay et al. 1996) (<http://www.esrl.noaa.gov/psd/>), 2) daily
 328 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
 329 (Behringer et al. 1998) (<http://www.esrl.noaa.gov/psd/>). Thermocline depth is computed
 330 from potential temperature as the depth of the 20C isotherm. All datasets are averaged meridion-
 331 ally within 5N-5S in the tropical Pacific (120E-80W) and cover the period from January 1982 to
 332 September 2016.

From the data, we obtain fields of zonal winds $u_{OBS}(x, t)$ in $m.s^{-1}$, sea surface temperature $T_{OBS}(x, t)$ in K , and thermocline depth $H_{OBS}(x, t)$ in m that depend only on zonal position x (deg lon) and time t (days). In addition, we define $u_{OBS}^W(t)$ as the amplitude of zonal winds anomalies u_{OBS} in the western Pacific, computed from an average in the region 140E-180E. This average is roughly equivalent to a projection on the zonal wind burst structure $s_p(x)$ of the ENSO model defined in Figure 1.

Each field is decomposed into climatology, interannual anomalies and intraseasonal anomalies. For example, zonal winds is decomposed into:

$$u_{OBS}(x, t) = u_{SC}(x, t) + u_A(x, t) + u_{HF}(x, t) \quad (15)$$

where u_{SC} is climatology, u_A is interannual anomalies to the climatology, and u_{HF} is intraseasonal anomalies. For this, a 90-days 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 using seasonal averages. 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 winds anomalies $u_A + u_{HF}$, sea surface temperature anomalies $T_A + T_{HF}$ and thermocline depth anomalies $H_A + H_{HF}$ 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 interannual and intraseasonal anomalies utilizing the same method as described above. The only difference is that there is no seasonal cycle in the model.

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 years are shown in Figure 2. The general circulation in the equatorial Pacific region consists of strong trade winds in the central Pacific as well as strong zonal gradients of SST and thermocline depth (See Panel (d), (f) and (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 Panel (c), (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 1982/1983, 1997/1998, 2006/2007 and 2015/2016 while prominent CP El Niño events are observed in years 1987/1988, 1990/1991, 1992/1993, 1994/1995, 2002/2003, 2004/2005 and 2009/2010. Note that a traditional El Niño is usually followed by a La Niña with reverse conditions. In addition, Figure 2 shows the details of wind bursts activity according to the intraseasonal zonal winds 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.

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 there (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 (11). On the other hand, both SST and thermocline depth show a gradual increase during the last 34 years.

377 *b. Statistical Properties of the Model and Observations*

378 We start with exploring the statistical properties of both the coupled model and observations.
379 The statistics of the model is based on a 5000-year-long simulation. Figure 4 and 5 show the
380 probability density functions (PDFs) for interannual anomalous SST, zonal winds and thermocline
381 depth associated with the coupled model and observations, respectively.

382 First, consistent with observations, the PDFs of SST associated with the coupled model in Ni-
383 no 4 and Nino 3 regions show negative and positive skewness, respectively. The presence of a
384 fat tail together with the positive skewness in Nino 3 indicates the extreme El Niño events in
385 the eastern Pacific (Burgers and Stephenson 1999). Note that, despite the correct skewed di-
386 rection, the skewness of T_A of the model in Nino 3 region seems to be underestimated com-
387 pared with that of the observations. Yet, the observations only contain a 34-year-long record
388 (1982-2016), which may not be sufficient to form unbiased statistics. In fact, the single su-
389 per El Niño event during 1997-1998 accounts for a large portion of the skewness in Nino 3.
390 On the other hand, despite a slight overestimation, the variance in all the three Nino region-
391 s associated with the model almost perfectly match those with observations. Particularly, the
392 fact that the variance of SST in Nino 4 region being roughly half as much as that in the oth-
393 er two regions is captured by the coupled model. Note that, as described in the previous work
394 (Thual et al. 2016; Chen and Majda 2016b), the parameters in the coupled model are calibrated
395 within physically reasonable ranges to match the SST variances with those provided by NOAA
396 (<https://www.ncdc.noaa.gov/teleconnections/enso/indicators/sst.php>). The Nino
397 indices in different regions are all slightly larger than the observational values utilized here. The
398 reason for this is that the climatology computed here is based on the whole 34-year period 1982-
399 2016 while only the data before 2000 is utilized in NOAA’s version. As shown in Figures 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 averaged over three different regions: the western Pacific (140E-180), the central Pacific (180-140W) and the eastern Pacific (140E-100W). The PDFs of the observed zonal surface winds in the central and eastern Pacific are both positively skewed and fat-tailed. Such non-Gaussian features are successfully captured by the coupled model. Although the associated variance from the model simulation as shown in Figure 4 seems to be underestimated, decreasing of both the wind stress coefficient γ in the ocean model and the latent heat feedback coefficient $\bar{Q}$ in the atmosphere model within a physically reasonable range serve to enhance the variance of the winds while leaving other variabilities unaffected. On the other hand, the variances of the observed thermocline depth in all the three regions are well captured by the coupled model.

In addition to the PDFs, the autocorrelation function of the total wind bursts anomalies a_{all} including both the interannual and the intraseasonal components is shown in Panel (a) of Figure 6 and that containing only the interannual wind bursts anomalies a_A resulting from the 90-day running average is shown in Panel (b). A short memory of a_{all} , around 6 to 7 days, is revealed, which is affirmed by the damping coefficient d_p in (12). On the other hand, the interannual anomalies of the wind bursts has a longer memory, around 2-3 months. Both values are consistent with observations (Levine and Jin 2015; Tziperman and Yu 2007; Yu et al. 2003). Finally, Panel (c) in Figure 6 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 years), as in the observations (Kleeman 2008).

All these findings indicate that the statistical features of the model are quite consistent with nature.

424 *c. Overall Mechanism of ENSO formation*

425 To understand the overall mechanism of ENSO formation, lagged correlation between Nino 3.4
426 SST index and different variabilities are shown in Figure 7 for the observations and in Figure 8
427 for the coupled model. The lagged correlations with SST T_A , thermocline H_A , zonal wind u_A and
428 the zonal wind averaged over the western Pacific u_A^W are all shown in both figures. Making use
429 of the advantage of the model that the wind bursts can be extracted from the overall zonal winds,
430 the lagged correlation with the overall wind bursts a_{all} and its interannual component a_A are also
431 shown in Figure 8. Note that except Panel (e) in Figure 8, all the variabilities utilized here are
432 the interannual with a 90-day running average as discussed in Section 2c. Here, focus is on the
433 overall formation mechanism that includes all events while the formation mechanisms conditioned
434 on different types of El Niño will be discussed in Section 4.

435 As shown in Figure 7, El Niño events are typically preceded by a buildup phase around 1 to
436 2 years prior to the event peak, during which SST and thermocline depth gradually enhance in
437 the western Pacific. Then, during the trigger phase around 0.5 year prior to the event peak, strong
438 westerly zonal surface winds, positive SST and thermocline depth anomalies all develop and prop-
439 agate from the western to the central-eastern Pacific. Importantly, in addition to a strong positive
440 correlation between Nino 3.4 index and the zonal wind averaged over the western Pacific u_A^W which
441 is less than 1 year prior to it, a weak negative correlation is found at lag times about 1-3 years. All
442 these features are captured by the coupled model with an exceptionally high skill.

443 Apparently, the positive correlation with u_A^W is mainly due to the reversal of anomalous Walker
444 circulation at the onset of an El Niño event as well as the presence of WWBs that serve to trigger
445 the El Niño. On the other hand, the appearance of the weak negative correlation with u_A^W remains
446 ambiguous by simply looking at the observational data. Nevertheless, due to the advantage of

the model that the wind bursts can be completely extracted from the total atmosphere wind, the lagged correlations between the Nino 3.4 and the wind bursts are shown in Panel (e) and (f) of Figure 8. These lagged correlations provide a significant evidence that the EWBs contribute to the negative correlation in the total winds. In fact, such EWBs facilitate the buildup of heat content in the western Pacific, which is crucial in the formation of El Niño events. Details will be shown in Section 4.

4. Mechanisms for the Formations of Different Events and Case Studies

In Figure 8, the overall mechanism of ENSO formation was revealed and the results from the coupled model are highly consistent with nature (Figure 7). 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-year short observational period, each type of El Niño events appear 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 is qualitatively similar as nature, is able to provide simulations with much longer period, which facilitates the understanding of the formation mechanisms of ENSO diversity within this simple coupled model.

Below, a 5000-year-long simulation from the coupled dynamical model is utilized for understanding the formation mechanisms of different events. Here is the summary of what we present below. The lagged correlations conditioned on traditional El Niño, La Niña and CP El Niño phases, respectively, are shown in Figure 9. Figure 10 illustrates the deterministic advective modes from the coupled nonlinear model associated with the CP El Niño (Chen and Majda 2016a). Figuring out the interactions between different waves and their link with physical variables in these

deterministic modes facilitates the understanding of more complicated stochastic model. Figure 11 shows properties of the waiting time between the occurrence of a CP El Niño event and its previous and adjacent traditional El Niño, which improves the understanding of the transition mechanism between these two types of events. Finally, Figure 12–15 include case studies of the formation and development of different El Niño events and are linked with the general discussions of the statistical features. A description of each case study is summarized in Table 2.

Same as in the previous section, the statistical features shown in Figure 9 and in Panel (a)-(d) of Figure 11 are all computed based on the interannual variables (with subscript $\cdot_A$). On the other hand, in order to see the structure of the wind bursts activity and the evolution of different types of waves, the raw output from the model that includes both interannual and intraseasonal components is shown in different case studies.

a. CP El Niño

1) STATISTICAL FEATURES

Row 3 in Figure 9 shows the conditional lagged correlations between the Nino 4 SST index and different fields conditioned on CP El Niño phases, where the criteria for identifying the CP El Nino phases was discussed in Section 2b. The Nino 4 SST index is utilized here due to the fact that CP El Niño is more related to the anomalous warm SST near dateline rather than the eastern Pacific.

First, different from the overall events (Figure 8) and the eastern Pacific El Niño/La Niña (Row 1-2 in Figure 9) in which significant correlations ($|\text{Corr}| > 0.5$) appear only at lag times less than 1 year, the Nino 4 SST index conditioned on CP El Nino phases has strong correlations with different fields even at 2- to 3-year lag times. This reveals the intrinsic difference between CP and traditional events that a CP El Niño episode can have a longer duration up to 4-5 years. Next, the

493 maximum correlation of Nino 4 index with both SST and thermocline appears in the central Pa-
 494 cific, as expected. This is consistent with the positive and negative correlations of the atmosphere
 495 surface wind u_A in the western and eastern Pacific, respectively, which indicates the structure of the
 496 anomalous Walker circulation that the surface winds converge in the central Pacific region. One
 497 noticeable feature for the CP El Niño is the flux divergence $-\mu\partial_x(UT)$. The lagged correlation
 498 structure in Panel (e) illustrates the role of the flux divergence in warming and cooling the SST in
 499 the central-western and central-eastern Pacific, respectively. Therefore, the flux divergence serves
 500 to transport anomalous warm SST from the eastern Pacific to the central region that is crucial in
 501 the formation of CP El Niño. Finally, corresponding to the mean easterly trade wind anomaly at
 502 the CP El Niño phases, a negative lagged correlation is found between the Nino 4 index and the
 503 wind bursts a_A .

504 2) DETERMINISTIC ADVECTION MODES

505 To see the development of waves and different physical variables in the formation of CP El Niño
 506 events, we start with the deterministic nonlinear advection modes (Chen and Majda 2016b). Here,
 507 the stochastic noise in (11) is removed and the wind activity $a_p \equiv \hat{a}_p$ becomes a constant. Based
 508 on different values of the nonlinear advection μ and the amplitude of the easterly mean trade wind
 509 anomaly a_p , three dynamical regimes are found in Panel (a) of Figure 10. In regime I, the steady-
 510 state solution has constant values at each longitude. Particularly, with a suitably strong a_p and
 511 even without the nonlinear advection, the anomalous warm SST is shifted to the central-eastern
 512 Pacific region. The corresponding anomalous ocean zonal current is westward and the anomalous
 513 atmospheric surface winds converge in the central-eastern Pacific as well. On the other hand,
 514 when both μ and a_p are sufficiently large (Regime III), the steady-state solution shows regular

oscillation patterns with period around 1.6 years. Within each period, warm water is transported westward and the maximum of anomalous warm SST is at the central and central-eastern Pacific.

The most important dynamical regime (Regime II) is shown in Panel (c), 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 becomes time-periodic, and the period is much longer than 2 years. 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, due to the fact that atmosphere winds force ocean waves (See Eq. (6)), the westerly atmosphere winds in the western Pacific push the reflected weak 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 have the wind stress which develops ocean Rossby waves, with amplitudes that peak at the central Pacific. Since thermocline depth is directly linked to the amplitude of ocean Kelvin and Rossby waves (See Eq. (9)), an increase of the thermocline depth occurs in the central Pacific region, which leads to the build-up of anomalous warm SST. Note that due to the fact that ocean Rossby waves have stronger amplitudes than ocean Kelvin waves, a westward anomalous ocean zonal current appears in the central Pacific region (See Eq. (9)), which together with the profile of SST anomalous results in a westward transport of anomalous warm water via the nonlinear advection

539 of SST, and brings about the eastern Pacific cooling. These anomalous warm SST then stays and
540 gradually develops in the central Pacific for a few years ($t = 95$). As the westward ocean zonal
541 current and the atmosphere easterly winds in the central and eastern Pacific continue to strength-
542 en, the anomalous warm SST is transported to the western pacific with a deeper thermocline there.
543 Such strong heat content storage provides the condition of triggering a strong traditional El Niño
544 ($t = 108$), followed by which is a discharge phase of La Niña that drives all the fields back to a
545 nearly quiescent state.

546 3) CASE STUDIES

547 Now we go back to the full model with effective state-dependent stochastic noise in the wind
548 activity (11)–(14). Case study (IV) in Figure 13 shows a period of 2.5-year CP El Niño from
549 $t = 413.5$ to $t = 416$.

550 First, the interannual wind time series in Panel (f) clearly indicates a mean easterly trade wind
551 anomaly. Starting from $t = 413$, moderate and weak ocean Rossby waves are triggered continu-
552 ously by this mean trade wind anomaly together with the effective stochastic wind bursts noise,
553 and heat content starts to accumulate in the western Pacific. Due to the imperfect reflection at the
554 western Pacific boundary $r_W = 0.5$, the reflected Kelvin waves have only weak amplitudes. Such
555 weak Kelvin waves are able to transport heat content eastward and adjust the atmosphere zonal
556 wind structure (Compare $t = 413$ and 414 in Panel (a)). However, due to the sea-air interaction, the
557 easterly atmosphere zonal surface winds in the eastern surface Pacific also weaken these Kelvin
558 waves (See Eq. (6)). Eventually, a balance is established in the central Pacific region. On the
559 other hand, this atmosphere zonal wind structure also induces ocean Rossby waves that start from
560 the eastern Pacific and peak at the location where easterly and westerly surface winds converge

(See Eq. (6)). Thus, according to the link between ocean waves and thermocline depth (9), 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.

Budget analysis is illustrated in Figure 15. Panel (d) shows the flux divergence $-\mu\partial_x(UT)$, which at the CP El Niño phases serves to increase and decrease SST anomalies in the central-western and central-eastern Pacific, respectively. On the other hand, the combined effect of the damping $-c_1\zeta E_q$ and thermocline feedback $c_1\eta H$ as shown in Panel (e) 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 (Panel (g)). In fact, as will be shown in Section 5, the thermocline feedback in the coupled model without the nonlinear zonal advection only warms the eastern Pacific and that model fails to reproduce CP El Niño.

b. Traditional El Niño Including Super El Niño

1) TWO FORMATION MECHANISMS

Row 1 of Figure 9 shows the lagged correlation between Nino 3 SST index and different fields conditioned on the traditional El Niño phases. The lagged correlations up to 1-year lag times for all fields are quite similar to those of the overall events (Figure 8). Particularly, in addition to the positive correlation with the WWBs occurring a few months prior to the event peak, a significant negative correlation between Nino 3 SST and wind bursts is found around 1 year before the event

584 peak, which corresponds to the appearance of the EWBs. Such EWBs are crucial for inducing and
585 accumulating heat content in the western Pacific that facilitates the following WWBs to trigger El
586 Niño. This provides the first formation mechanism of the traditional El Niño.

587 **Formation mechanism 1:** EWBs appear around 1 year before a traditional El Niño event, which
588 induce the buildup of anomalous warm SST and heat content in the western Pacific. Such EWBs
589 are followed by a series of WWBs occurring a few months prior to the event peak, which lead to
590 the eastward propagation of the anomalous warm SST and heat content to the eastern Pacific and
591 simultaneously alter the zonal circulation of the anomalous atmosphere winds.

592 In addition to the strong lagged correlation at a lag time around 1 year as discussed above,
593 Row 1 of Figure 9 also reveals moderate correlations between Nino 3 index and both SST and
594 thermocline in the central and western Pacific at time lag up to 3.5 years. Meanwhile, moderate
595 correlations with westerly and easterly zonal atmosphere winds in the western and eastern Pacific
596 are also noticeable. These lagged correlation structures resemble those in the CP El Niño phase.
597 In fact, a second formation mechanism of the traditional El Niño is linked to its previous CP El
598 Niño.

599 **Formation mechanism 2:** In the presence of a CP El Niño event, both the anomalous SST and
600 heat content have been transported to the central Pacific. Thus, instead of triggering heat content
601 in the western Pacific by a series of EWBs, a single strong WWBs right after a CP El Niño event
602 is sufficient to push the anomalous warm water and heat content from the central Pacific to the
603 eastern Pacific and forms the traditional El Niño. No significant EWBs is observed during this
604 process.

605 2) CASE STUDIES

606 Figure 12 shows two case studies that belong to formation mechanisms 1 and 2, respectively.

Case study (I) 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 led 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 Eq. (9)). 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, Case study (SII) in Section 5 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 EWBs-WWBs mechanism. 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 Figure 15, 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 the dominant factor.

Case study (II) in Figure 12 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 ($t = 365$ - 368), accompanied by which is a mean easterly trade wind anomaly. Different from Case study (I) that EWBs trigger heat content in the western Pacific that prepares the formation of a traditional El Niño, the heat

631 content in Case study (II) has already been transported to the central Pacific region due to the
632 CP El Niño (Kug et al. 2009). Then, with the presence of a series of WWBs ($t = 368$), both the
633 heat and anomalous warm SST are transported from the central Pacific to the eastern Pacific and
634 a super El Niño is formed. Clearly, the fact that no obvious EWB appears prior to the WWBs in
635 Case study (II) distinguishes the two mechanisms.

636 In Figure 13, two more case studies are included. Case study (III) is similar to Case study (II).
637 But the heat content in the central Pacific during CP El Niño phase in Case study (III) is weaker
638 than that in (II) and therefore the subsequent WWBs cause a moderate traditional El Niño rather
639 than a super El Niño. In Case Study (IV), there is a short quiescent phase between the CP and the
640 traditional El Niño events. Therefore, the two events should be treated as separate events despite
641 that the overall profiles look similar to those in (III), and thus the EWBs-WWBs structure is needed
642 in Case (IV) to form a traditional El Niño event.

643 3) WAITING TIME BETWEEN A CP EL NIÑO EVENT AND ITS PREVIOUS ADJACENT TRADI- 644 TIONAL EL NIÑO

645 As was seen in the previous subsection, a traditional El Niño event is likely to follow a series of
646 CP El Niño events, which is partly due to the prescribed high transition rate from State 1 to State
647 2 based on the observational evidence. Yet, understanding the reverse situation, i.e., the waiting
648 time between a CP El Niño and its previous adjacent traditional El Niño, is equally crucial in
649 interpreting the mutual effect between these two types of El Niño.

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

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

657 In Panel (a) of Figure 11, the standard definition of the traditional El Niño is adopted, i.e.,
658 $Nino\ 3 > 0.5$, which leads to a bimodal distribution of the waiting time. The major mode peaks
659 around 3.5 years while the peak of the minor mode is around 1 years. Such bimodality implies
660 two different scenarios. In the first scenario, a traditional El Niño is followed by a La Niña, and
661 then after 1-2 years quiescent period, a CP El Niño appears. Examples are shown in Panel (e). In
662 fact, the observed episodes during 1987-1990 and 1998-2002 both belong to the first situation. In
663 the second scenario, a CP El Niño occurs right after a traditional El Niño, where an example is
664 shown in Panel (f). Clearly, this second situation occurs less frequently than the first one in the
665 model simulation. It is also observed only in years 1977-1978 during the last 50 years (Chen et al.
666 2015). Particularly, with the enhancing of the threshold of the Nino 3 index for the traditional El
667 Niño, the minor mode disappears and the resulting distribution of waiting time is unimodal (Panel
668 (d)), which implies that the second situation happens only when the traditional El Niño has weak
669 amplitude. In fact, a La Niña usually follows a moderate or strong traditional El Niño in order to
670 discharge the heat content.

671 *c. La Niña*

672 The formation mechanism of La Niña is relatively simple. Lagged correlations in Row 2 of
673 Figure 9 show that both the SST anomalies and the heat content propagate from western Pacific to
674 eastern Pacific. A negative correlation between Nino 3 index and SST anomalies at a lag time of
675 2 years in Row 2 is consistent with that at 2 years lead time in Row 1 that is conditioned on the
676 traditional El Niño phases. Both indicate that La Nina serves as the discharge phase of El Niño.

677 This argument is further validated by the lagged correlation between Nino 3 index and wind bursts
678 a_A , where WWBs occur 1.5 prior to the La Nina event peak and trigger the previous recharge
679 phases that corresponds to a traditional El Niño event.

680 In Case study (I)-(IV), a traditional El Niño is always followed by a La Niña, where only little
681 wind bursts are found during all the La Niña phases. In addition, the anomalous Walker circulation
682 at La Niña phases is nearly opposite to that at traditional El Niño phases.

683 *d. An El Niño breaking down by EWBs*

684 As was seen in Case study (I) that a specific wind burst structure, where EWBs lead WWBs,
685 triggers a traditional El Niño. Yet, different wind burst structures are observed in nature. Actually,
686 a reverse wind structure, namely EWBs following WWBs, occurred during year 2014, where a
687 strong EWB stalled the development of a potential traditional El Niño that was triggered by its
688 previous WWBs (Hu and Fedorov 2016). Case (V) in Figure 14 exhibits an example with such
689 wind burst structure and provides a simple mechanism for EWBs' impeding the development of a
690 traditional El Niño.

691 In Case (V), a series of WWBs around $t = 480.3$ induces ocean Kelvin waves, which propa-
692 gate eastward and bring anomalous warm SST to the eastern Pacific. These WWBs also cool the
693 western Pacific and lead to a negative amplitude of Rossby waves in the western Pacific. Due to
694 such anomalous cool water, the subsequent EWBs around $t = 480.8$ are unable to build up heat
695 content in the western Pacific. Instead, they induce ocean Kelvin waves with negative amplitude.
696 These negative Kelvin waves propagate eastwards and suppress the positive Rossby waves reflect-
697 ed by the previous positive Kelvin waves. Therefore, the eastern Pacific becomes cooler and the
698 development of the traditional El Niño is stalled.

699 *e. Strong Wind Bursts with No/Weak El Niño*

700 Traditional El Niño events are typically triggered by wind bursts activities (Harrison and Vecchi
701 1997; Vecchi and Harrison 2000; Tziperman and Yu 2007; Hendon et al. 2007). However, there are
702 many examples that wind burst activity builds up without triggering any El Niño event, implying
703 that wind burst activity in the model is a necessary but non-sufficient condition to the El Niño
704 development (Hu et al. 2014; Fedorov et al. 2015; Roulston and Neelin 2000; Levine and Jin
705 2010).

706 Case studies (VI) in Figure 14 illustrate two wind structures, in which strong wind bursts lead to
707 no or very weak El Niño. In (VI.a), both WWBs and EWBs appear with large amplitudes and the
708 interannual component is even not close to zero. However, the total wind profile switches between
709 these two types of wind bursts with a high frequency. In (VI.b), the interannual wind bursts stay
710 near zero and the intraseasonal wind bursts resemble white noise. The common feature in the two
711 scenarios is that no enough heat content is able to build up in the western Pacific and no warm
712 water is able to accumulate in the eastern Pacific. Therefore, wind bursts with long durations and
713 strong amplitudes are not sufficient to trigger El Niño events. The buildup of the heat content
714 resulting from the wind bursts is the essence that facilitates the El Niño formation.

715 **5. Various Simplifications of the Coupled Model and their Dynamical Deficiencies**

716 To understand the role of several important components in the coupled ENSO model (1)–(3),
717 different simplified versions of the coupled model are studied in this section.

718 The first simplification involves dropping the nonlinear ocean zonal advection ($\mu = 0$) in the
719 SST equation (3), where a case study is shown in Figure 16. This simplified model retains the
720 traditional El Niño and La Niña ($t = 275$ – 278). In fact, as was shown in the previous work (Thual
721 et al. 2016), traditional El Niño was successfully simulated by the coupled model even in the

722 absence of the nonlinear advection. However, without the nonlinear zonal advection, the coupled
723 model has difficulties in reproducing the CP El Niño with realistic features. The major issue is that
724 the location of the anomalous warm SST at State 1 shifts from the central Pacific to the central-
725 eastern or eastern Pacific ($t = 268-269$ and $t = 278-281$). In fact, budget analysis in Figure 15
726 already indicates the key role of the nonlinear advection, without which the combined effect of the
727 thermocline feedback and cooling only leads to the anomalous warm SST in the eastern Pacific.
728 In addition, the deterministic modes in Figure 10 (Regime I) implies that the strengthening of the
729 trade wind anomaly itself is insufficient to transport the anomalous warm SST to the central Pacific
730 region. It is the combined effect of the strengthening of the trade wind and the nonlinear zonal
731 advection that develops and maintains the CP El Niño.

732 Another important element in the coupled model is the ocean reflection condition at the eastern
733 Pacific boundary r_E , which hasn't been emphasized yet. In fact, this ocean boundary reflection is
734 essential for obtaining the realistic features of the traditional El Niño and particularly the super El
735 Niño. To see the role of r_E , a reduced model with $r_E = 0$ is utilized, where a case study is shown
736 in Figure 17. Despite that the reduced model is able to reproduce the CP El Niño as in nature
737 ($t = 256-260$ and $t = 266-267$), the duration of the traditional El Niño ($t = 261, 264.5$ and 269.5)
738 is much shorter than the observed ones (years 1982, 1998 and 2016 in Figure 2). Particularly,
739 the duration of the super El Niño is only about 3 months ($t = 264.5$), which is much shorter than
740 the observations. Note that the thermocline depth, which is the dominant contributor to the SST
741 tendency at the traditional El Niño phase, is given by the linear superposition of ocean Kelvin
742 and ocean Rossby waves (See (7) and (9)). Since the wave speed of the Rossby wave is only $1/3$
743 as much as that of the Kelvin wave, the Rossby wave plays an important role in maintaining the
744 anomalous warm SST at the traditional El Niño phase. However, with $r_E = 0$, there is no reflected
745 ocean Rossby waves, which results in the short duration of the traditional El Niño events.

6. Discussion and Conclusions

Understanding the formation mechanisms of ENSO diversity has significant impact on global climate and seasonal prediction. Recently, a simple modeling framework has been developed that automatically captures the statistical properties and key features 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.

In this article, the formation mechanisms of ENSO diversity based on this simple coupled model are systematically studied. In Section 3, the statistics in the observational data associated with different variabilities are illustrated. The non-Gaussian statistical properties in nature are all successfully captured by the coupled model and the overall ENSO formation mechanism of the coupled model are highly consistent with that of the real observations. In Section 4, the formation mechanisms of different types of El Niño and La Niña based in the coupled model are discussed. It is shown in the model with only the deterministic nonlinear advective modes that the mean easterly trade wind anomaly is essential in triggering ocean Rossby waves that not only facilitate the heat content buildup in the western Pacific but also induce reflected ocean Kelvin waves with weak amplitude. The reflected ocean Kelvin waves, combined with the nonlinear advection, are especially crucial in maintaining the climate patterns associated with the CP El Niño (Chen and Majda 2016a). Incorporating effective random wind bursts into the nonlinear advective modes results in more realistic features of the CP El Niño. Two mechanisms are revealed for the formation of traditional El Niño events. The first mechanism involves an EWBs-WWBs structure of the wind

769 activity that corresponds to the buildup of the heat content in the western Pacific and the transport
770 of the anomalous warm water to the eastern Pacific, respectively. The second mechanism links the
771 CP El Niño and traditional El Niño, where a single WWB occurring right after a CP El Niño is
772 sufficient to push the anomalous warm water to the eastern Pacific. On the other hand, La Niña is
773 proved to be the discharge phase of traditional El Niño. In addition to the ENSO formation, the
774 underlying reasons for some strong wind bursts not triggering any El Niño and specific EWBs’
775 stalling the development a traditional El Niño event are also explained in Section 4. In Section 5,
776 several simplified versions of the coupled model are utilized to highlight the crucial roles of both
777 the nonlinear advection and the boundary reflections in the eastern Pacific that contribute to the
778 realistic features of CP and traditional El Niño, respectively.

779 As mentioned at the end of Section 2, seasonal cycle, which allows most of the El Niño events
780 peak at the end of the years, hasn’t been incorporated into the coupled model. Therefore, including
781 seasonal cycle into the simple modeling framework will be one of the future works. Another future
782 work is the El Niño prediction, which requires developing an effective data assimilation scheme
783 for the coupled model as initialization and designing practical prediction algorithms.

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787 APPENDIX A

788 Choices of parameter values

Two tables are included below. Table A1 summarizes the variables in the coupled model and lists the associated units and the typical unit values. Table A2 shows the nondimensional values of the parameters that are utilized in the meridional truncated model (5)–(7).

APPENDIX B

Details of the transition rates in the three-state Markov jump stochastic process

Recall in (11)–(14), a three-state Markov jump stochastic process is adopted for describing the wind activity. The criteria of transitions between different states were briefly mentioned there and the details are listed as follows. These transition rates are determined in accordance with the observational facts in Figure 2 and reference (Chen et al. 2015).

- The transition rates from State 2 to State 1 and from State 2 to State 0 are given by respectively

$$v_{21} = \frac{1}{10} \cdot \frac{1 - \tanh(2T_W)}{4}, \quad (\text{B1})$$

$$v_{20} = \frac{9}{10} \cdot \frac{1 - \tanh(2T_W)}{4}. \quad (\text{B2})$$

Starting from State 2, the probability of switching to State 0 is much higher than that to State 1. This comes from the fact that a traditional El Niño is usually followed by a La Niña rather than a CP El Niño (e.g., year 1963, 1965, 1972, 1982 and 1998). Typically, the La Niña event has a weaker amplitude and a longer duration compared with the preceding El Niño. This actually corresponds to a discharge phase of the ENSO cycle with no external wind bursts (State 0).

- The transition rates from State 1 to State 0 and from State 1 to State 2 are given by respectively

$$v_{10} = \frac{1 - \tanh(2T_W)}{12}, \quad (\text{B3})$$

$$v_{12} = \frac{1 + \tanh(2T_W)}{40}, \quad (\text{B4})$$

Although the denominator of v_{10} is smaller than that of v_{12} , quite a few CP El Niño events are associated with a slight positive T_W in the model, which means the transition rate v_{12} is not necessarily smaller than v_{10} . In fact, with the transition rates given by (B3)–(B4), the results show in the main article that more events are transited from state 1 to 2 than from state 1 to 0. This is consistent with the observations (e.g., year 1981 and 1995), implying that the CP El Niño is more likely to be followed by the classical El Niño than the quiescent phase.

- The transition rates from State 0 to State 1 and State 2 are given by

$$v_{01} = \frac{2}{3} \cdot \frac{1 + \tanh(2T_W)}{7}, \quad (\text{B5})$$

$$v_{02} = \frac{1}{3} \cdot \frac{1 + \tanh(2T_W)}{7}. \quad (\text{B6})$$

Again, the transition rates to State 1 and 2 are different. This is due to the fact that after a quiescent period or discharge La Niña phase, more events are prone to become CP El Niño as an intermediate transition instead of directly forming another traditional El Niño (e.g., year 1969, 1977, 1990 and 2002).

Note that in (B1)–(B6), the transition rate v_{ij} from a more active state to a less active state (with $i > j$) is always proportional to $1 - \tanh(2T_W)$ while that from a less active state to a more active state (with $i < j$) is always proportional to $1 + \tanh(2T_W)$. These are consistent with the fact stated in the main article that a transition from a less active to a more active state is more likely when $T_W \geq 0$ and vice versa.

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Table A1. Definitions of model variables and units in the meridional truncated model.

Variable	unit	unit value
x zonal axis	$[y]/\delta$	$15000km$
y meridional axis atmosphere	$\sqrt{c_A/\beta}$	$1500km$
Y meridional axis ocean	$\sqrt{c_O/\beta}$	$330km$
t time axis intraseasonal	$1/\delta\sqrt{c_A\beta}$	$3.3days$
τ time axis interannual	$[t]/\epsilon$	$33days$
u zonal wind speed anomalies	δc_A	$5ms^{-1}$
v meridional wind speed anomalies	$\delta[u]$	$0.5ms^{-1}$
θ potential temperature anomalies	15δ	$1.5K$
q low-level moisture anomalies	$[\theta]$	$1.5K$
a envelope of synoptic convective activity	1	
$\overline{H}a$ convective heating/drying	$[\theta]/[t]$	$0.45K.day^{-1}$
E_q latent heating anomalies	$[\theta]/[t]$	$0.45K.day^{-1}$
T sea surface temperature anomalies	$[\theta]$	$1.5K$
U zonal current speed anomalies	$c_O\delta_O$	$0.25ms^{-1}$
V zonal current speed anomalies	$\delta\sqrt{c}[U]$	$0.56cms^{-1}$
H thermocline depth anomalies	$H_O\delta_O$	$20.8m$
τ_x zonal wind stress anomalies	$\delta\sqrt{\beta/c_A}H_O\rho_Oc_O^2\delta_O$	$0.00879N.m^{-2}$
τ_y meridional wind stress anomalies	$[\tau_x]$	$0.00879N.m^{-2}$

Table A2. Nondimensional values of the parameters.

Parameter	description	Nondimensional values
c	ratio of ocean and atmosphere phase speed	0.05
ε	Froude number	0.1
c_1	ratio of c/ε	0.5
χ_A	Meridional projection coefficient from ocean to atmosphere	0.31
χ_O	Meridional projection coefficient from atmosphere to ocean	1.38
L_A	Equatorial belt length	8/3
L_O	Equatorial Pacific length	1.16
γ	wind stress coefficient	6.529
r_W	Western boundary reflection coefficient in ocean	0.5
r_E	Eastern boundary reflection coefficient in ocean	0.5
ζ	Latent heating exchange coefficient	8.5
α_q	Latent heating factor	0.3782
$\bar{Q}$	mean vertical moisture gradient	0.9
μ	nonlinear zonal advection coefficient	0.08
d_p	dissipation coefficient in the wind burst model	3.4

TABLE 1. Criteria of defining different ENSO events.

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

TABLE 2. Summary of case studies.

<i>Case</i>	Figure #.	Event type
(I)	12	An isolated moderate traditional El Niño and an isolated super El Niño
(II)	12	A series CP El Niño followed directly by a super El Niño
(III)	13	A series CP El Niño followed directly by a moderate traditional El Niño
(IV)	13	A traditional El Niño following a series of CP El Niño with a short quiescent phase in between
(V)	14	An EWB stalling the development of a traditional El Niño
(VI)	14	Strong wind bursts but no/weak El Niño
(SI)	16	Model without zonal advection: Failure in simulating CP El Niño
(SII)	17	Model no reflection in the eastern Pacific boundary: Failure in simulating traditional El Niño with realistic features

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982	wind anomaly $\hat{a}_p$, i.e. $\hat{a}_p \cdot s_p(x) \cdot \phi_0(y)$	53
983	Fig. 2. Observed ENSO variability. (a) Zonal winds averaged over the western Pacific (140E-180),	
984	with a superposition of interannual u_A^W (black) and intraseasonal u_{HF}^W (blue for $u_{HF}^W < 0$,	
985	red for $u_{HF}^W > 0$) anomalies. (b) Hovmoller of strong intraseasonal zonal winds u_{HF} (blue	
986	for $u_{HF} \leq -5 \text{ m.s}^{-1}$, red for $u_{HF} \geq 5 \text{ m.s}^{-1}$). (c) Hovmoller of zonal winds anomalies u_A .	
987	(d) Hovmoller of zonal winds anomalies and climatology $u_A + u_{SC}$. (e) Hovmoller of sea	
988	surface temperature anomalies T_A . (f) Hovmoller of sea surface temperature anomalies and	
989	climatology $T_A + T_{SC}$. (g) Hovmoller of thermocline depth anomalies H_A (m). (h) Homvoller	
990	of thermocline depth anomalies and climatology $H_A + H_{SC}$	54
991	Fig. 3. Linear trends in time of observed (a) zonal winds u_{OBS} , (b) sea surface temperature T_{OBS} and	
992	(c) thermocline depth H_{OBS} . The trends are obtained from a polynomial curve fit of the form	
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998	terannual anomalies H_A averaged over the same regions as zonal winds. The variance (v),	
999	skewness (s) and kurtosis (k) are listed below each panel.	56
1000	Fig. 5. Same as Figure 4 but for the model surrogates.	57
1001	Fig. 6. Panel (a): autocorrelation function of the total wind bursts a_{all} . Panel (b): autocorrelation	
1002	function of the interannual wind burst anomalies a_A . Panel (c): Power spectrum of the SST	
1003	averaged over the eastern half of the Pacific.	58
1004	Fig. 7. ENSO long-term memory of observations. Lagged correlations of the Nino 3.4 SST index	
1005	with (a) zonal winds anomalies averaged in western Pacific u_A^W , (b) zonal winds anomalies	
1006	u_A , (c) sea surface temperature anomalies T_A , and (d) thermocline depth anomalies H_A	59
1007	Fig. 8. ENSO long-term memory of the coupled model. Lagged correlations between the Nino	
1008	3.4 index and (a) zonal wind anomalies averaged in the western Pacific u_A^w , (b) zonal wind	
1009	anomalies u_A , (c) SST anomalies T_A , (d) thermocline depth anomalies H_A , (e) the total wind	
1010	bursts a_{all} and (f) the interannual wind burst anomalies a_A . Except the total wind burst a_{all}	
1011	which contains both the interannual and high-frequency components, all the other variables	
1012	including the Nino 3.4 index contain only the interannual anomalies.	60
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1014	phases. respectively. Panel (e) shows the lagged correlation between SST index and the flux	
1015	divergence $\partial_x(UT)$ interannual anomalies. The Nino 3 index is utilized for computing the	
1016	lagged correlations with traditional El Niño and La Niña while Nino 4 index is adopted for	
1017	the CP El Niño. Note that since Nino 3 is negative at La Niña phases, a negative correlation	
1018	between Nino 3 and a_A corresponds to positive (westerly) wind bursts.	61
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1020	functions of the nonlinear advection μ and the easterly mean trade wind anomaly $\hat{a}_p$. Panel	
1021	(b)-(d): Different fields (atmosphere wind including the mean trade wind anomaly $u + \hat{u}_p$,	

1022	ocean zonal current U , thermocline depth H , SST T , ocean Kelvin waves and ocean Ross-	
1023	by waves) corresponding to the three dynamical regimes, where the parameter values are	
1024	marked by red dots in (a).	62
1025	Fig. 11. Panel (a)-(d): Distribution of the waiting time $\tau_{E \rightarrow C}$ between a CP event and its previous	
1026	traditional El Niño event with four different threshold values of counting the traditional El	
1027	Niño before each CP event. Panel (e)-(f): Examples of CP El Niño following traditional one	
1028	with two different forms. Here, each blacking bar indicates the waiting periods between one	
1029	CP El Niño event and its previous traditional El Niño.	63
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1031	followed by a La Niña. (II) shows a series of CP El Niño followed directly by a super El	
1032	Niño and then a La Niña. Different panels show (a) atmosphere wind including the mean	
1033	trade wind anomaly $u + \hat{u}_p$, (b) thermocline depth H , (c) SST T , (d) ocean Kelvin waves, and	
1034	(e) ocean Rossby waves. All these fields are the direct model output and contain both the	
1035	interannual variability and the high frequency component. Panel (f) shows the time series of	
1036	the wind activity at the peak x^* of the zonal profile (see Panel (b) of Figure 1) and its 90-day	
1037	running average (brown curve). All variables are shown at equator.	64
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1039	El Niño and then a La Niña. (IV) shows a series of CP El Niño followed by a few months	
1040	quiescence ($t = 416.5$) and then a moderate El Niño and a La Niña. The wind burst structure	
1041	associated with (III) and (IV) is slightly different.	65
1042	Fig. 14. Case study (V) and (VI). (V) shows a moderate El Niño ($t = 478$) and an El Niño event	
1043	being stalled by an EWB ($t = 480.8$) that resembles to the observational event during year	
1044	2014. (VI) shows two examples that strong wind bursts lead to no or very weak El Niño,	
1045	where a clear switching between WWBs and EWBs is found in (VI.a) but the durations for	
1046	both WWBs and EWBs are short while the wind bursts in (VI.b) behave like white noise.	66
1047	Fig. 15. Budget analysis of Case study (I) in Figure 12 and (IV) in Figure 13. Panel (a) and (b) show	
1048	the SST T and ocean zonal current U . Panel (c) shows the SST tendency dT/dt , which is	
1049	the summation of the flux divergence $-\mu \partial_x(UT)$ (Panel (d)) and the combined effect of the	
1050	damping $-c_1 \zeta E_q$ and the thermocline feedback $c_1 \eta H$ (Panel (e)). The budget components	
1051	of the damping $-c_1 \zeta E_q$ and the thermocline feedback $c_1 \eta H$ are shown in Panel (f) and (g),	
1052	respectively.	67
1053	Fig. 16. Case study (SI): model without zonal nonlinear advection ($\mu = 0$) in the SST equation.	
1054	Without the zonal nonlinear advection the anomalous warm SST ($t = 268$ -269 and $t = 278$ -	
1055	281) corresponding to State 1 shifts eastward and occurs in the traditional El Niño regions.	
1056	68	
1057	Fig. 17. Case study (SII): model with zero reflection coefficient in the eastern Pacific boundary $r_E =$	
1058	0. Without the reflection in the eastern Pacific boundary, the traditional El Niño is mainly	
1059	dominated by Kelvin waves and the duration of these traditional El Niño, including the super	
1060	El Niño, is much shorter than the realistic ones.	69

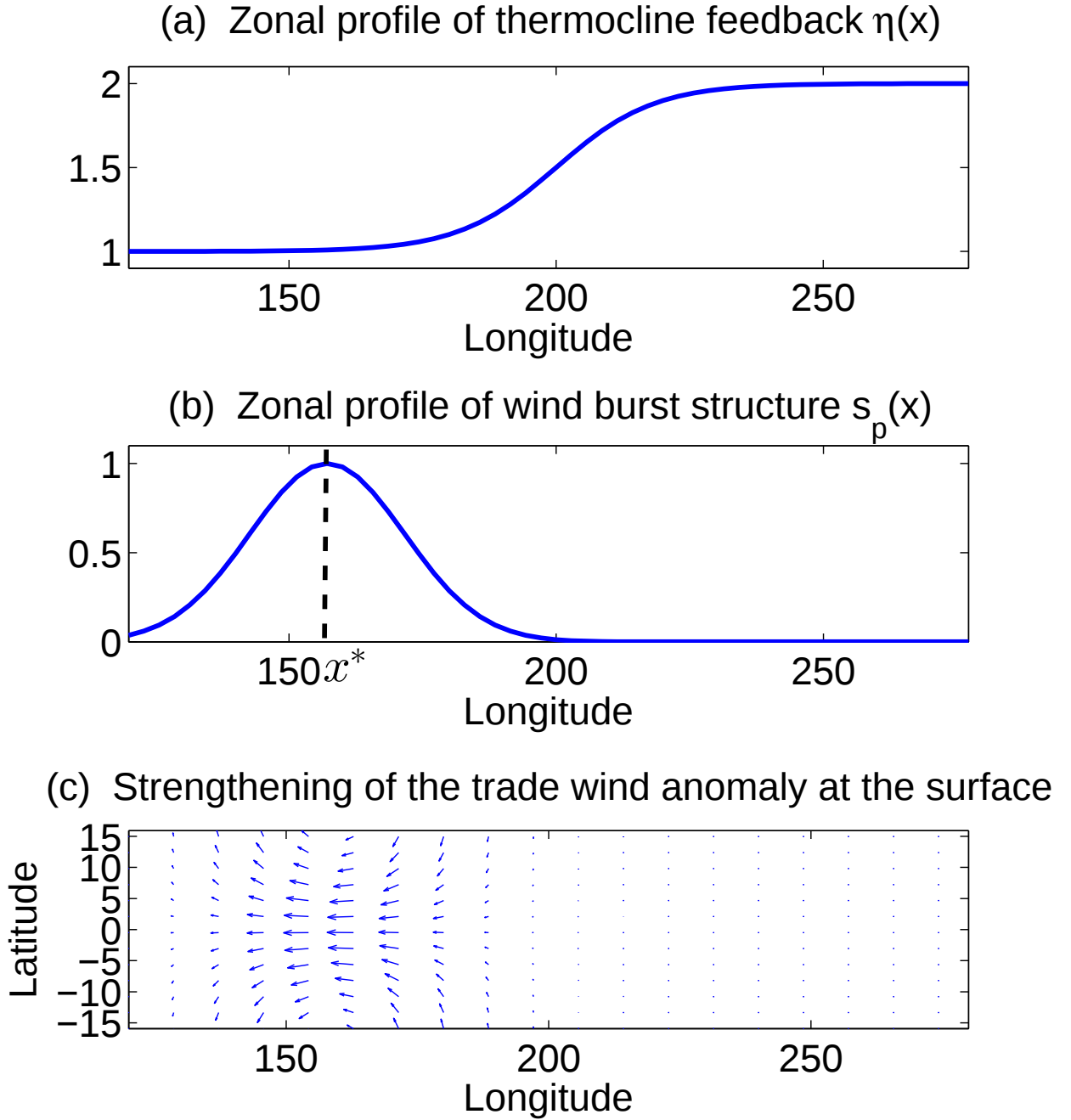


FIG. 1. (a) Zonal profile of thermocline feedback $\eta(x)$. (b) Zonal profile of wind burst structure $s_p(x)$ in the western Pacific. (c) Surface wind response directly to the mean easterly trade wind anomaly $\hat{a}_p$, i.e. $\hat{a}_p \cdot s_p(x) \cdot \phi_0(y)$.

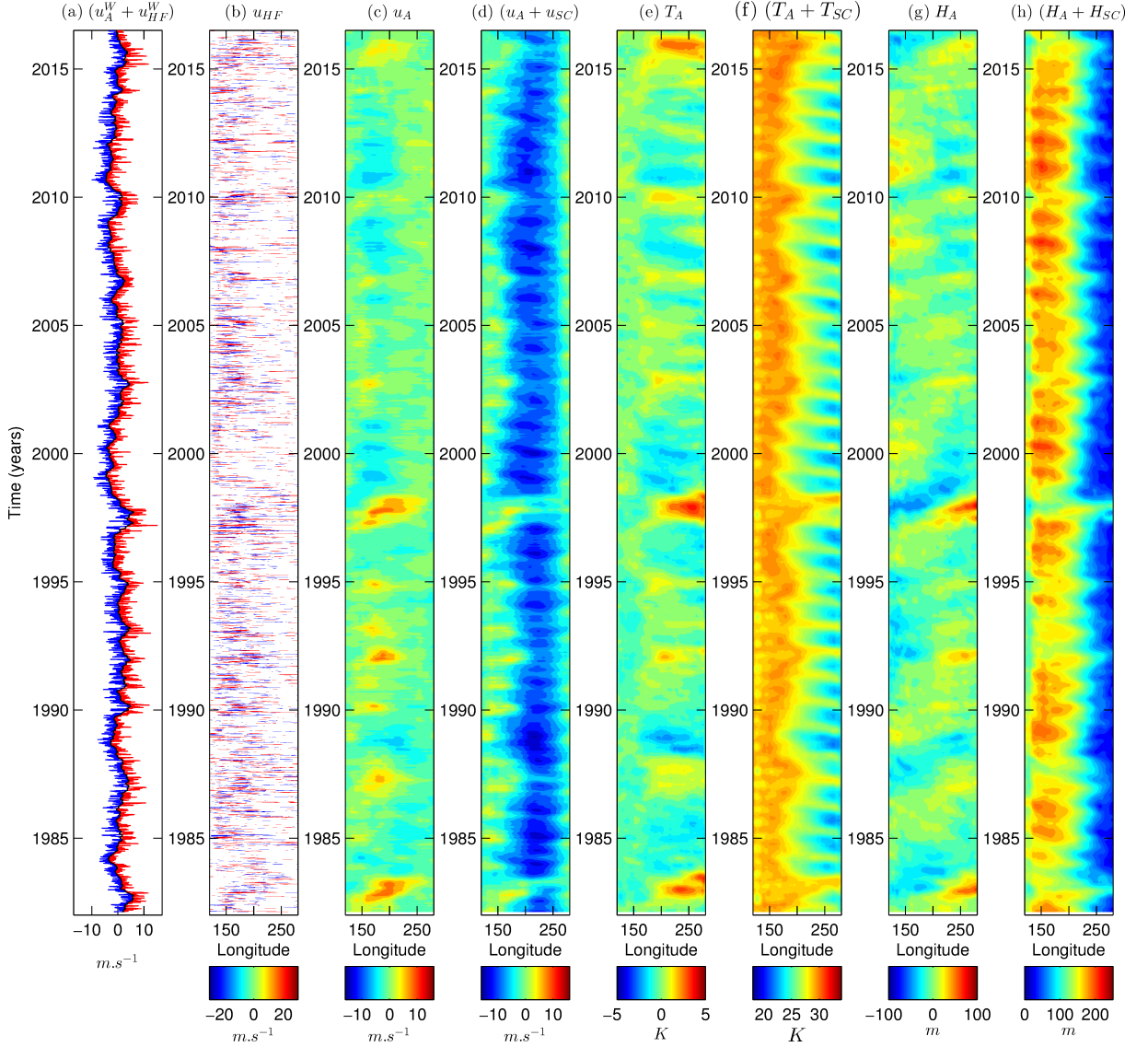


FIG. 2. Observed ENSO variability. (a) Zonal winds averaged over the western Pacific (140E-180), with a superposition of interannual u_A^W (black) and intraseasonal u_{HF}^W (blue for $u_{HF}^W < 0$, red for $u_{HF}^W > 0$) anomalies. (b) Hovmoller of strong intraseasonal zonal winds u_{HF} (blue for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

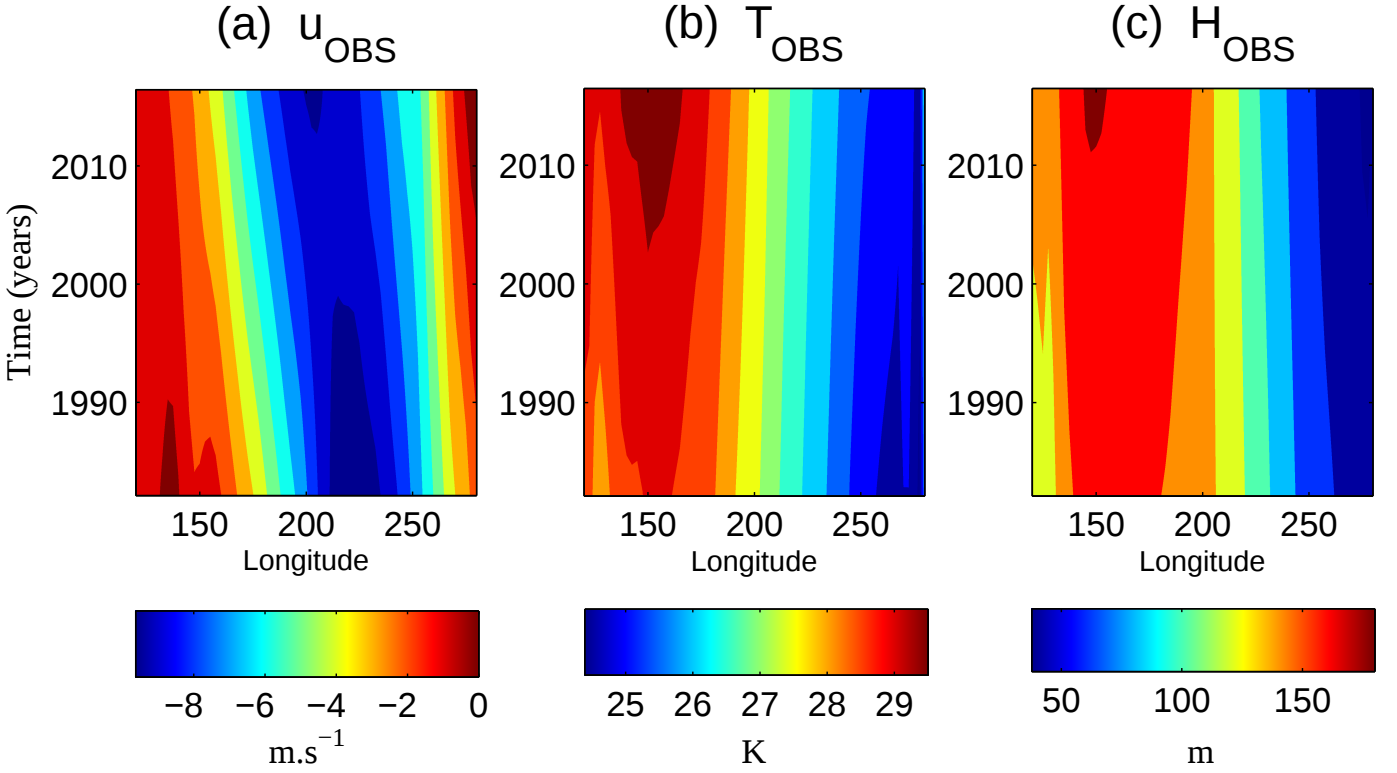


FIG. 3. Linear trends in time of observed (a) zonal winds u_{OBS} , (b) sea surface temperature T_{OBS} and (c) thermocline depth H_{OBS} . The trends are obtained from a polynomial curve fit of the form $u_{OBS}(x,t) = p_0(x) + p_1(x)t + o(1)$, where x is longitude, t is time and p_0, p_1 are coefficients.

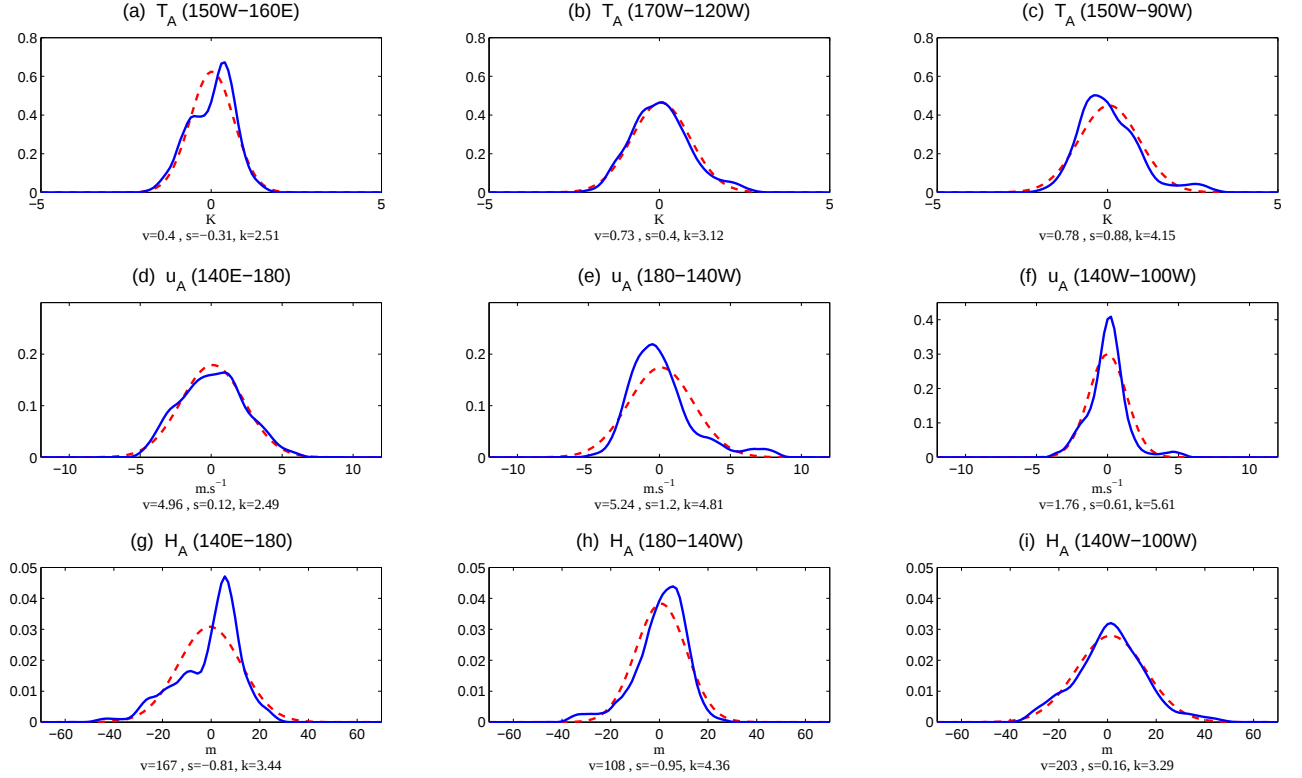


FIG. 4. Probability density functions (PDFs) of observations. Panel (a)-(c): SST interannual anomalies T_A averaged over the Nino 4, Nino 3.4 and Nino 3 regions. Panel (d)-(f): Zonal winds interannual anomalies u_A averaged over the western Pacific (140E-180), the central Pacific (180-140W), and the eastern Pacific (140W-100W). Panel (g)-(i): Thermocline depth interannual anomalies H_A averaged over the same regions as zonal winds. The variance (v), skewness (s) and kurtosis (k) are listed below each panel.

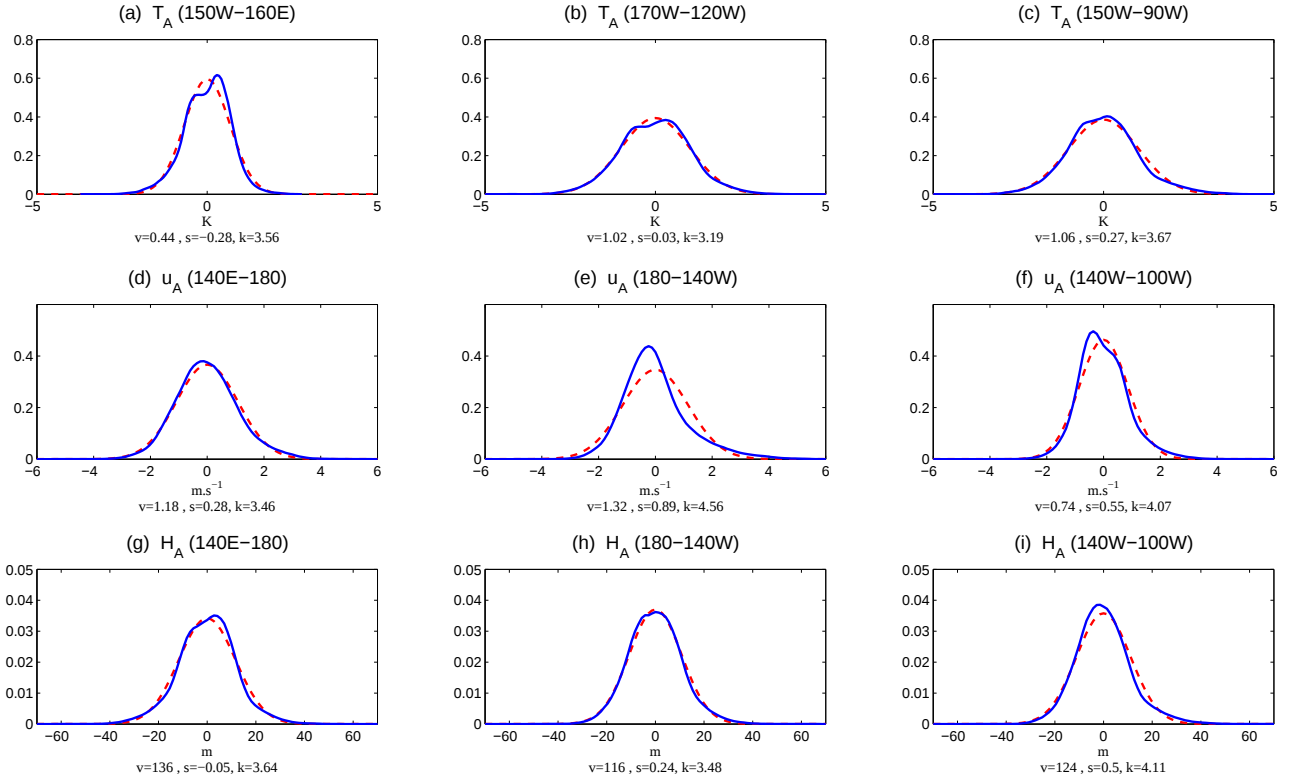


FIG. 5. Same as Figure 4 but for the model surrogates.

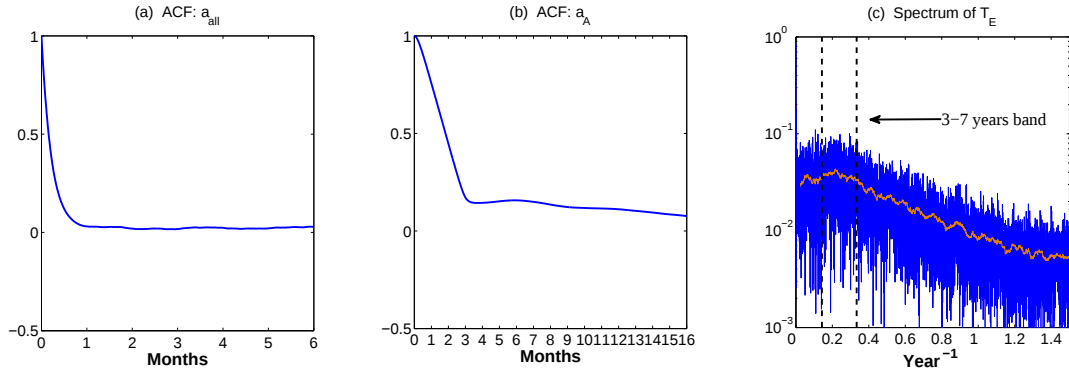


FIG. 6. Panel (a): autocorrelation function of the total wind bursts a_{all} . Panel (b): autocorrelation function of the interannual wind burst anomalies a_A . Panel (c): Power spectrum of the SST averaged over the eastern half of the Pacific.

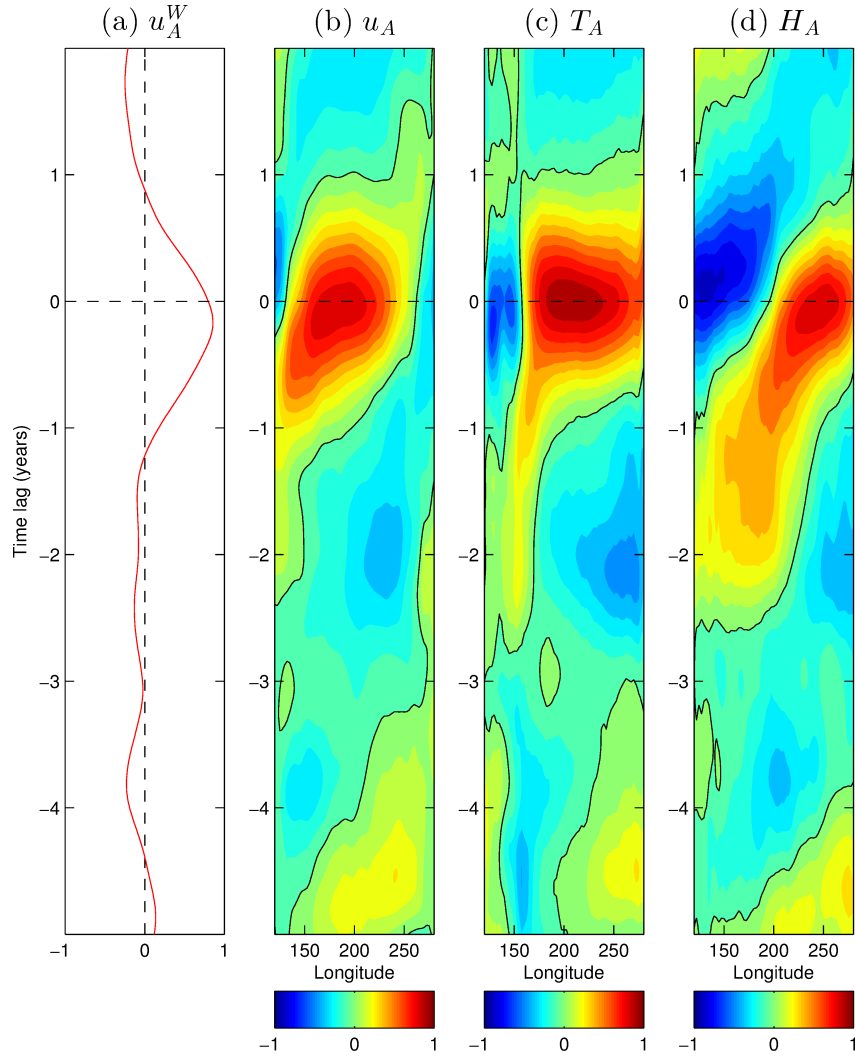


FIG. 7. ENSO long-term memory of observations. Lagged correlations of the Niño 3.4 SST index with (a) zonal winds anomalies averaged in western Pacific u_A^W , (b) zonal winds anomalies u_A , (c) sea surface temperature anomalies T_A , and (d) thermocline depth anomalies H_A .

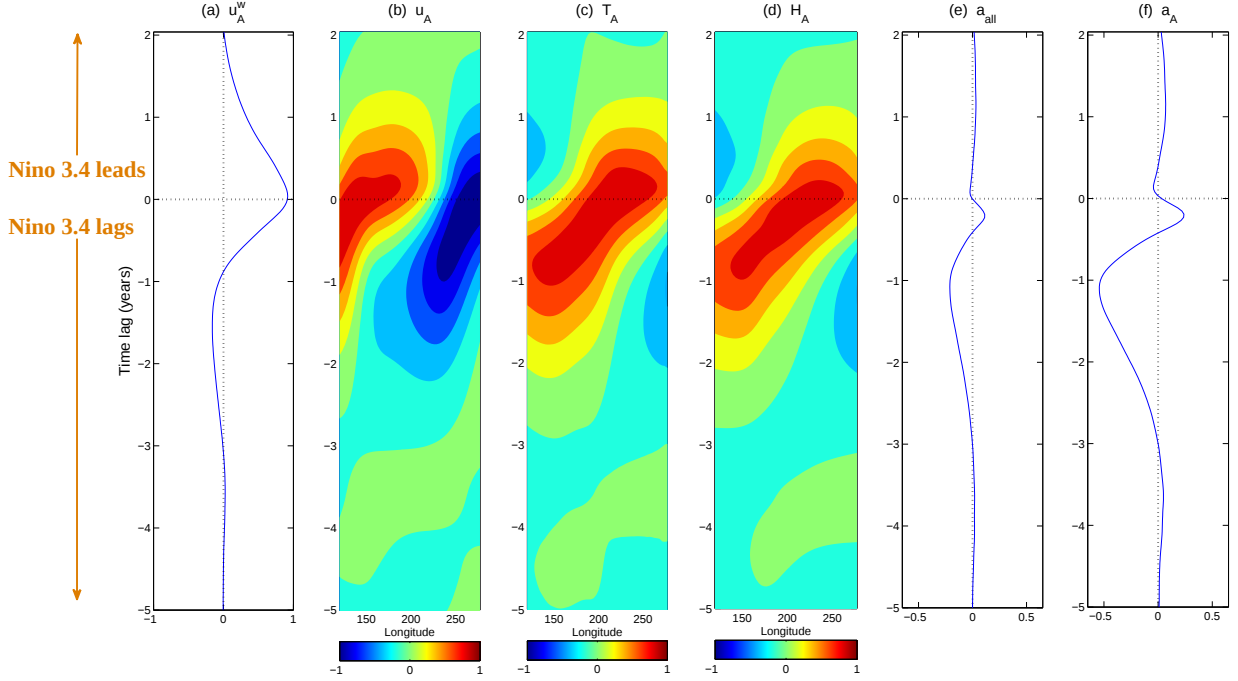


FIG. 8. ENSO long-term memory of the coupled model. Lagged correlations between the Nino 3.4 index and (a) zonal wind anomalies averaged in the western Pacific u_A^w , (b) zonal wind anomalies u_A , (c) SST anomalies T_A , (d) thermocline depth anomalies H_A , (e) the total wind bursts a_{all} and (f) the interannual wind burst anomalies a_A . Except the total wind burst a_{all} which contains both the interannual and high-frequency components, all the other variables including the Nino 3.4 index contain only the interannual anomalies.

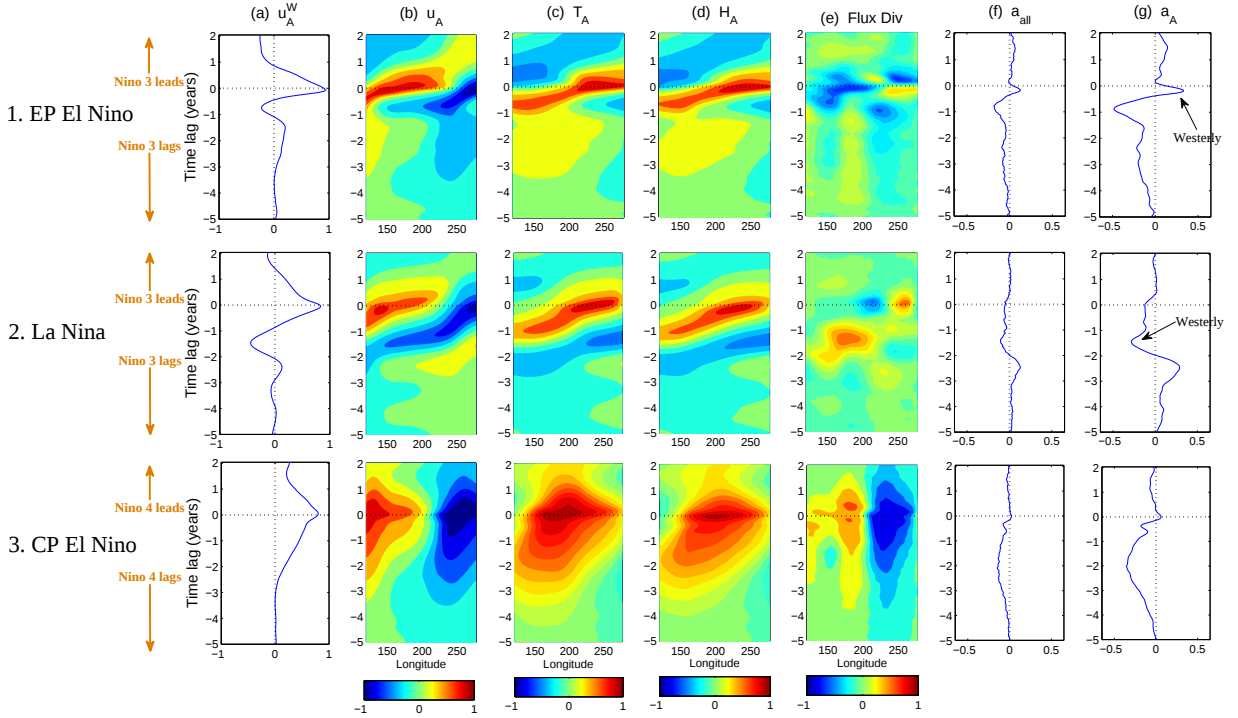


FIG. 9. Same as Figure 8 but conditioned on the EP (traditional) El Niño, La Niña and CP El Niño phases. respectively. Panel (e) shows the lagged correlation between SST index and the flux divergence $\partial_x(UT)$ interannual anomalies. The Nino 3 index is utilized for computing the lagged correlations with traditional El Niño and La Niña while Nino 4 index is adopted for the CP El Niño. Note that since Nino 3 is negative at La Niña phases, a negative correlation between Nino 3 and a_A corresponds to positive (westerly) wind bursts.

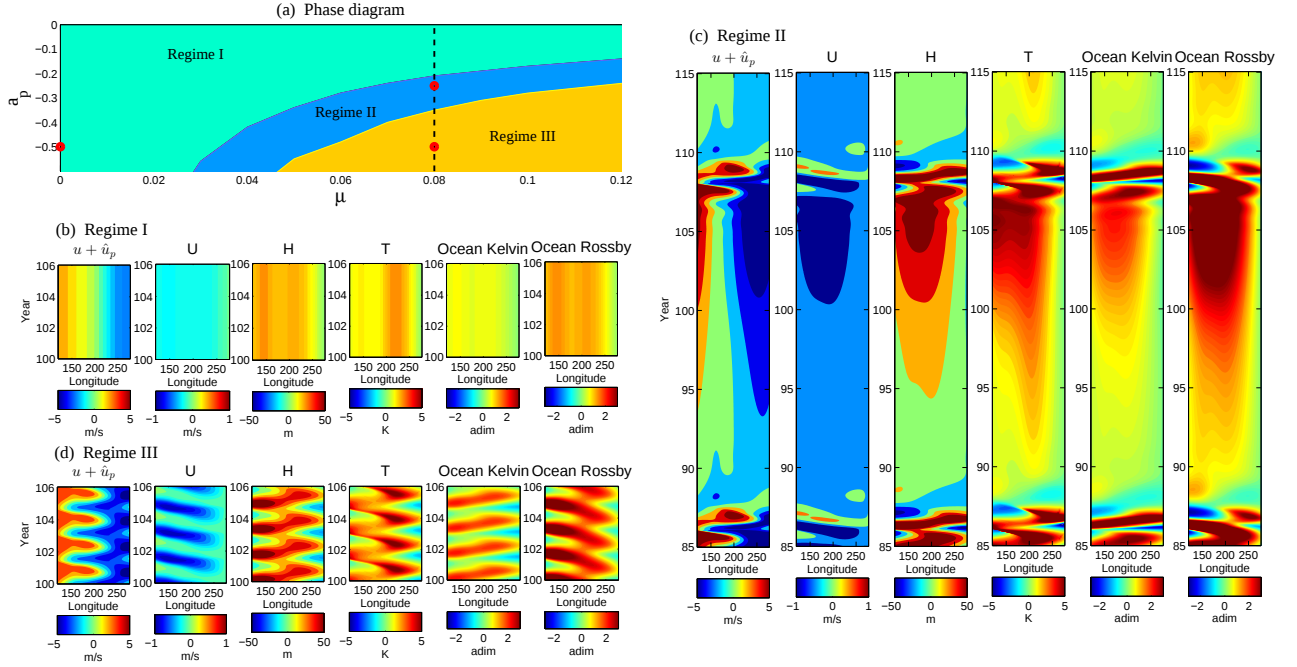


FIG. 10. Deterministic advection modes from the model. Panel (a): Three dynamical regimes as functions of the nonlinear advection μ and the easterly mean trade wind anomaly $\hat{a}_p$. Panel (b)-(d): Different fields (atmosphere wind including the mean trade wind anomaly $u + \hat{u}_p$, ocean zonal current U , thermocline depth H , SST T , ocean Kelvin waves and ocean Rossby waves) corresponding to the three dynamical regimes, where the parameter values are marked by red dots in (a).

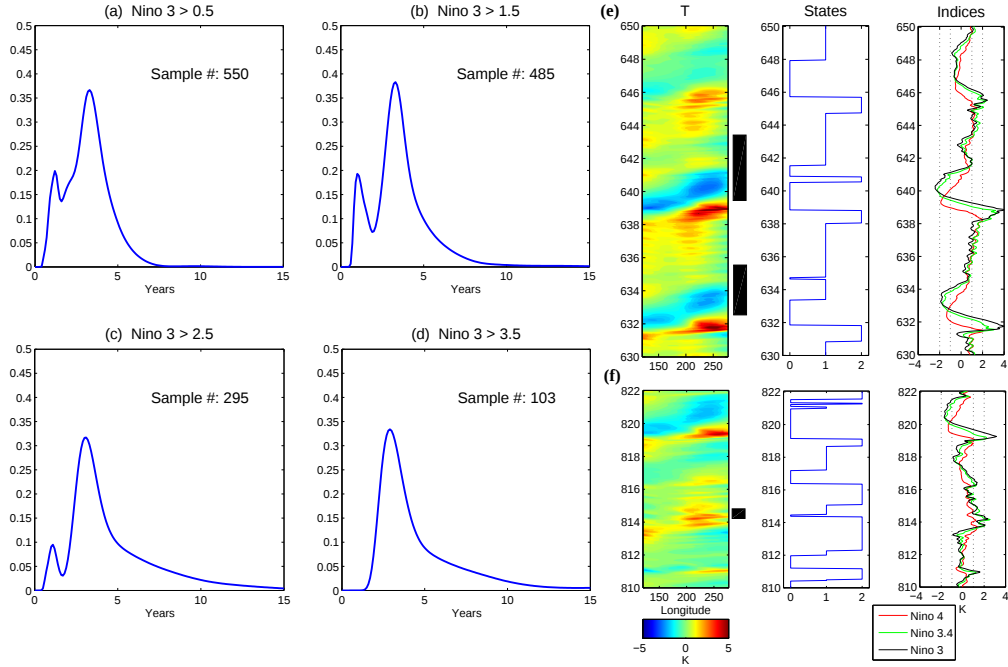


FIG. 11. Panel (a)-(d): Distribution of the waiting time $\tau_{E \rightarrow C}$ between a CP event and its previous traditional El Niño event with four different threshold values of counting the traditional El Niño before each CP event. Panel (e)-(f): Examples of CP El Niño following traditional one with two different forms. Here, each blacking bar indicates the waiting periods between one CP El Niño event and its previous traditional El Niño.

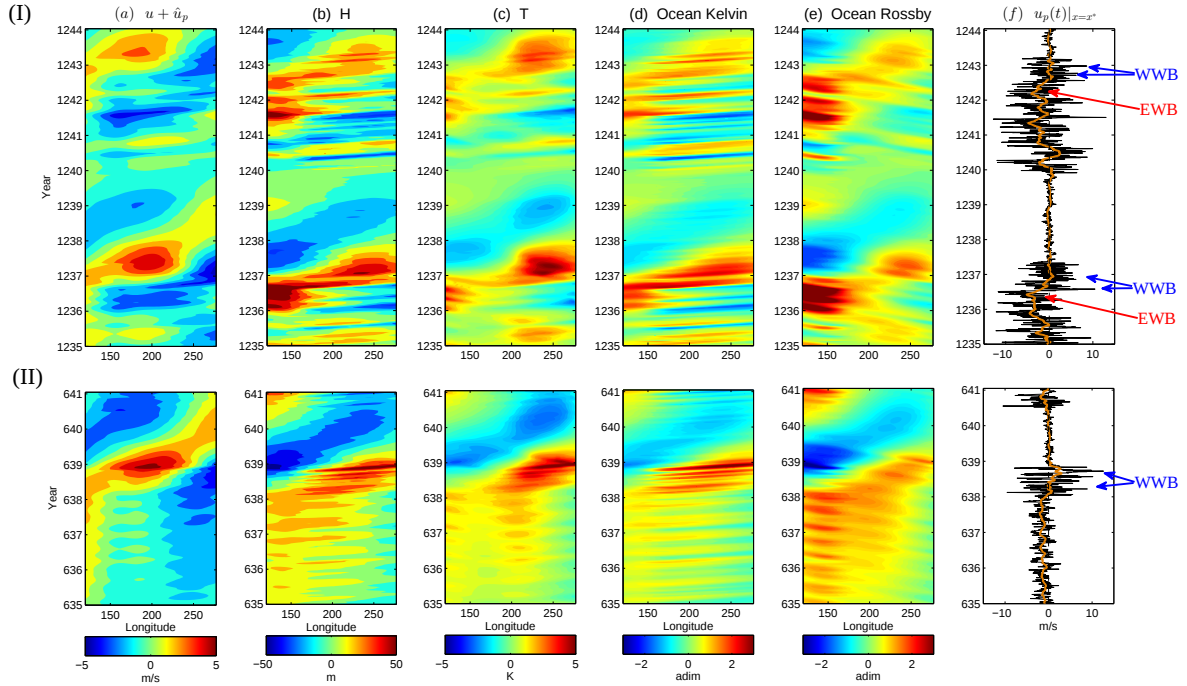


FIG. 12. Case study (I) and (II). (I) shows one moderate traditional El Niño and one super El Niño followed by a La Niña. (II) shows a series of CP El Niño followed directly by a super El Niño and then a La Niña. Different panels show (a) atmosphere wind including the mean trade wind anomaly $u + \hat{u}_p$, (b) thermocline depth H , (c) SST T , (d) ocean Kelvin waves, and (e) ocean Rossby waves. All these fields are the direct model output and contain both the interannual variability and the high frequency component. Panel (f) shows the time series of the wind activity at the peak x^* of the zonal profile (see Panel (b) of Figure 1) and its 90-day running average (brown curve). All variables are shown at equator.

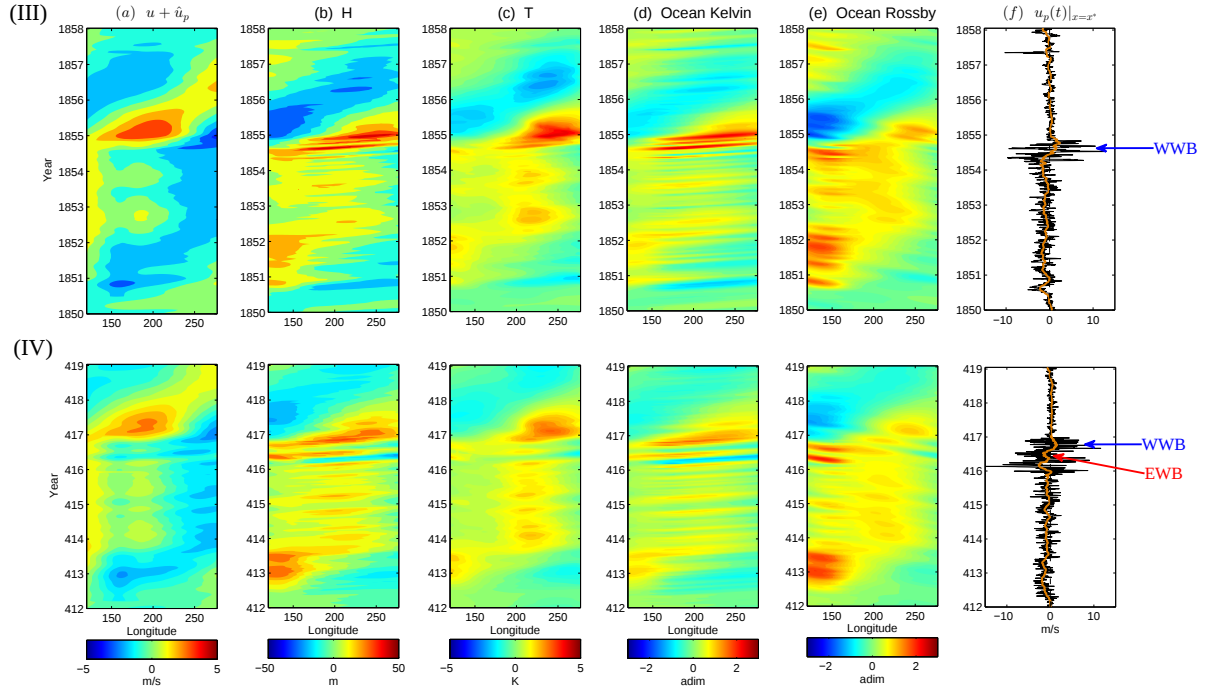


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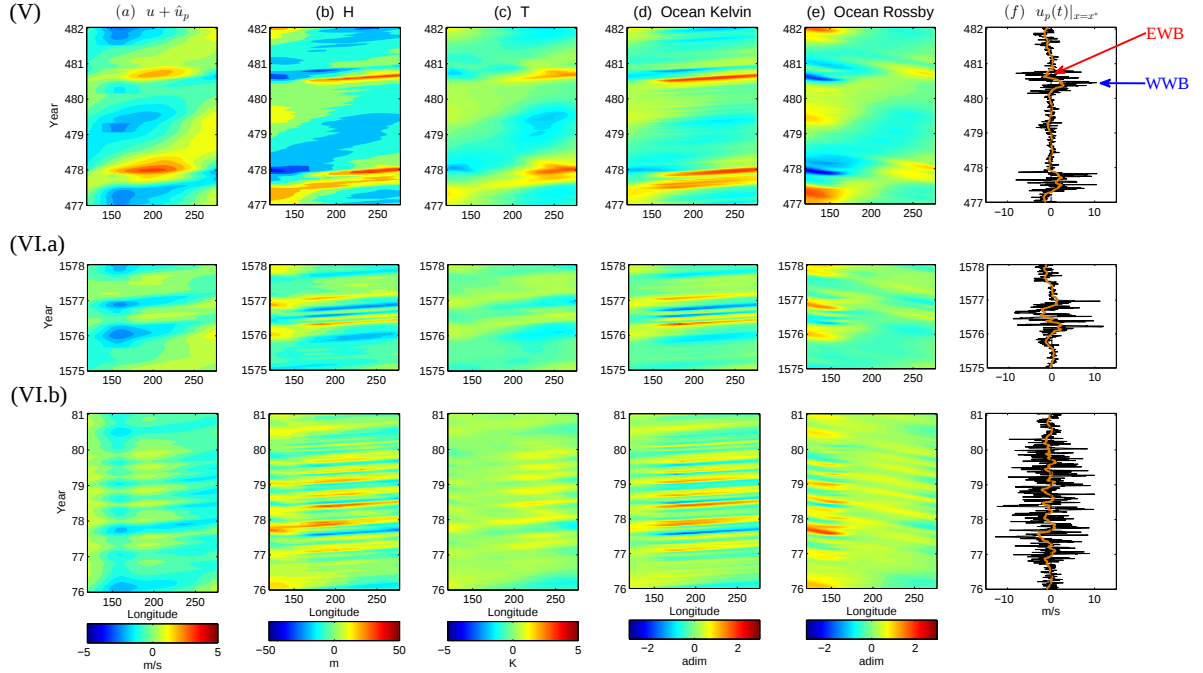


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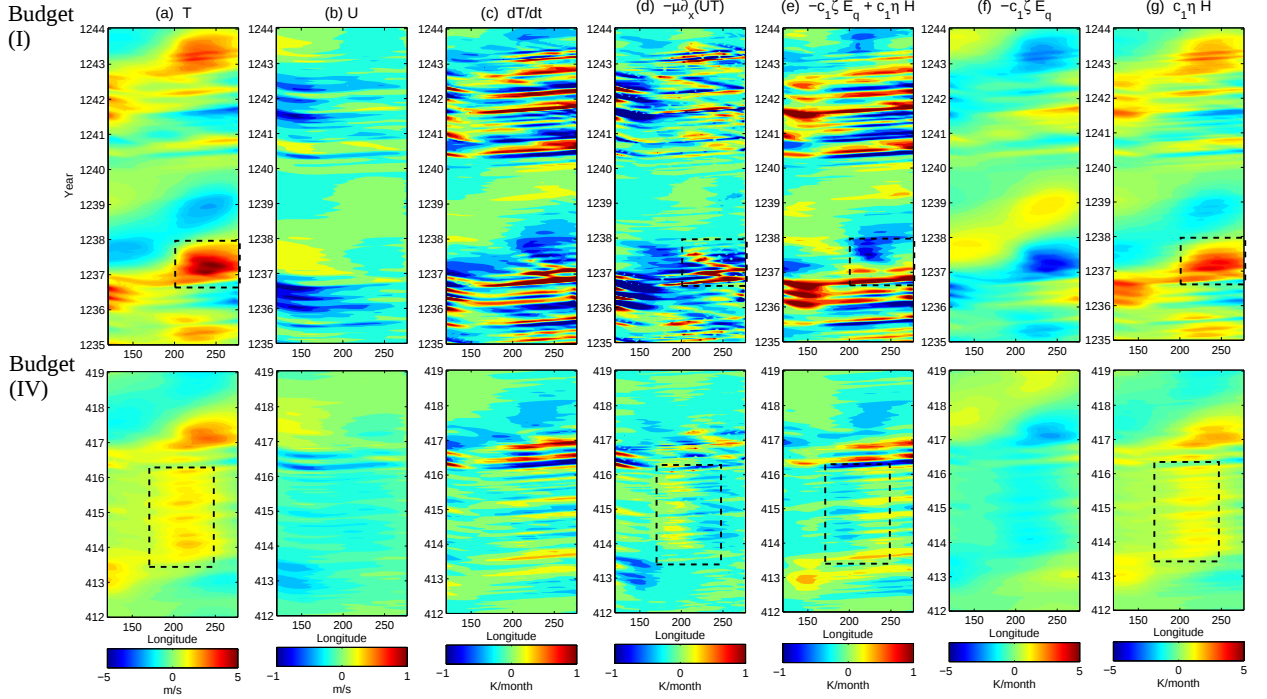


FIG. 15. Budget analysis of Case study (I) in Figure 12 and (IV) in Figure 13. Panel (a) and (b) show the SST T and ocean zonal current U . Panel (c) shows the SST tendency dT/dt , which is the summation of the flux divergence $-\mu\partial_x(UT)$ (Panel (d)) and the combined effect of the damping $-c_1\zeta E_q$ and the thermocline feedback $c_1\eta H$ (Panel (e)). The budget components of the damping $-c_1\zeta E_q$ and the thermocline feedback $c_1\eta H$ are shown in Panel (f) and (g), respectively.

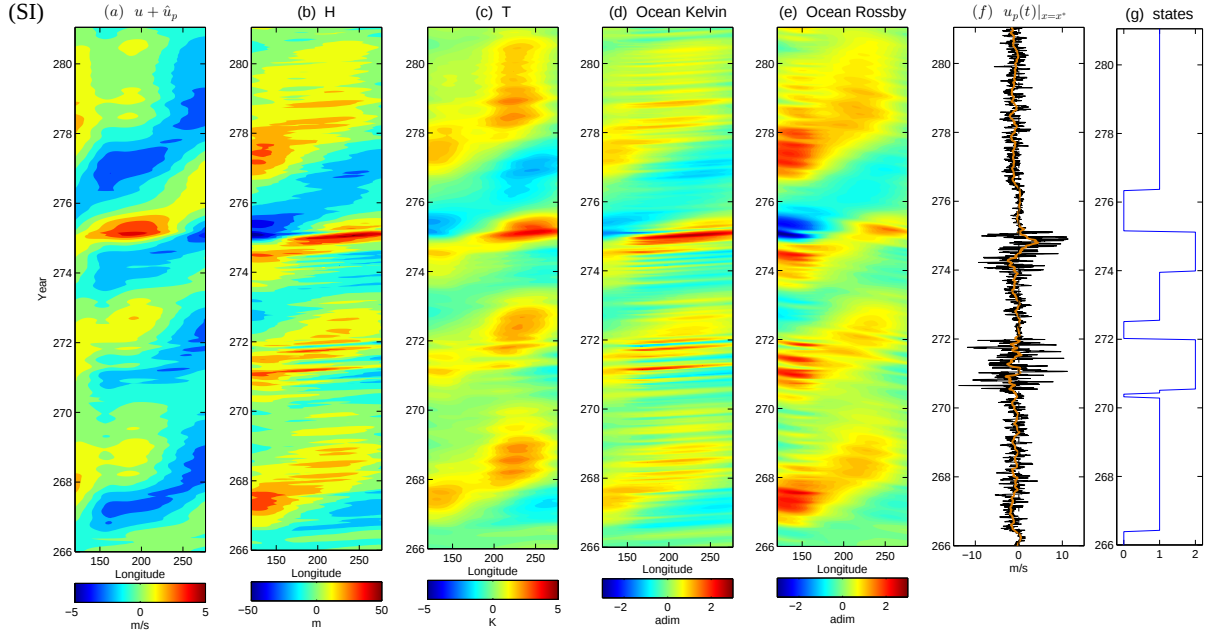


FIG. 16. Case study (SI): model without zonal nonlinear advection ($\mu = 0$) in the SST equation. Without the zonal nonlinear advection the anomalous warm SST ($t = 268-269$ and $t = 278-281$) corresponding to State 1 shifts eastward and occurs in the traditional El Niño regions.

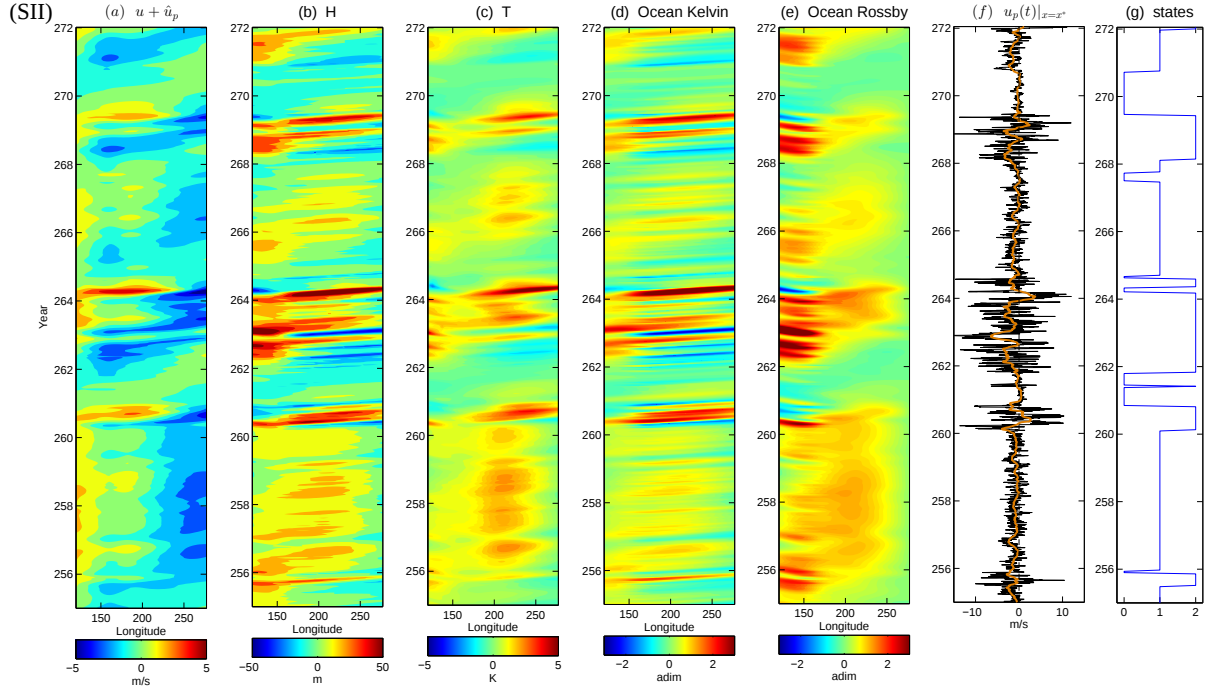


FIG. 17. Case study (SII): model with zero reflection coefficient in the eastern Pacific boundary $r_E = 0$.

Without the reflection in the eastern Pacific boundary, the traditional El Niño is mainly dominated by Kelvin waves and the duration of these traditional El Niño, including the super El Niño, is much shorter than the realistic ones.