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# Impact of different El Niño types on the El Niño/IOD relationship

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**Abstract** Previous studies reported that positive phases of the Indian Ocean Dipole (IOD) tend to accompany El Niño during boreal autumn. Here we show that the El Niño/IOD relationship can be better understood when considering the two different El Niño flavors. Eastern-Pacific (EP) El Niño events exhibit a strong correlation with the IOD dependent on their magnitude. In contrast, the relationship between Central-Pacific (CP) El Niño events and the IOD depends mainly on the zonal location of the sea surface temperature anomalies rather than their magnitude. CP El Niño events lying further west than normal are not accompanied by significant anomalous easterlies over the eastern Indian Ocean along the Java/Sumatra coast, which is unfavorable for the local Bjerknes feedback and correspondingly for an IOD development. The El Niño/IOD relationship has experienced substantial changes due to the recent decadal El Niño regime shift, which has important implications for seasonal prediction.

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## 28    **1. Introduction**

29        The El Niño-Southern Oscillation (ENSO) is the dominant low-frequency climate  
30    phenomenon resulting from coupled ocean-atmosphere interactions in the tropical  
31    Pacific [e.g., Philander et al. 1990; Wallace et al. 1998]. Although ENSO originates in  
32    the tropical Pacific, its impacts can be detected in remote oceans through the so-called  
33    atmospheric bridge mechanism [e.g., Klein et al. 1999; Alexander et al. 2002; Lau and  
34    Nath 2003]. Especially during the mature (boreal winter) and decaying phases (boreal  
35    spring) of El Niño, a basin-wide sea surface temperature (SST) warming appears in  
36    the tropical Indian Ocean (IO) due to the ENSO-induced surface heat flux anomalies  
37    [Klein et al. 1999]. In contrast, during the preceding boreal summer and autumn  
38    seasons, a dipole structure of SST anomalies tends to occur in the tropical IO, usually  
39    described as the Indian Ocean Dipole (IOD) [Saji et al. 1999; Webster et al. 1999]. A  
40    positive IOD event features SST cooling along the Java-Sumatra coast and SST  
41    warming in the western tropical IO. A positive correlation between the ENSO and  
42    IOD during boreal autumn suggests that IOD events are closely related to ENSO  
43    (positive and negative IOD events usually co-occurred with El Niño and La Niña  
44    events, respectively) [e.g., Allan et al. 2001; Baquero-Bernal et al. 2002; Xie et al.  
45    2002; Annamalai et al. 2003]. However, this argument was challenged by other  
46    studies [e.g., Saji et al. 1999; Webster et al. 1999; Saji and Yamagata 2003; Meyers et  
47    al. 2007], which argued that the IOD is an independent mode of coupled  
48    ocean-atmosphere climate variability in the tropical IO. Although the ENSO/IOD  
49    relationship still remains open to debate, observational and modeling results generally

50 suggest that the IOD seems to be a relatively weak natural mode, which can be  
51 excited by external forcings such as the ENSO variability [e.g., Li et al. 2003; Scott et  
52 al. 2009].

53 ENSO exhibits a considerable degree of complexity in its zonal SST anomaly  
54 structure. The Central-Pacific, or CP El Niño has occurred more frequently in recent  
55 decades, which differs considerably from traditional El Niño events (Eastern-Pacific,  
56 or EP El Niño) that are characterized by maximum SST anomalies over the eastern  
57 equatorial Pacific [e.g., Ashok et al. 2007; Kao and Yu 2009; Kug et al. 2009]. The CP  
58 type of El Niño has become more common while the EP El Niño has occurred less  
59 frequently since the 1990s [e.g., Yeh et al. 2009; Xiang et al. 2013; Zhang et al. 2014].

60 Whether the IOD experienced changes along with the El Niño regime shift deserves  
61 attention as the IOD can cause substantial climate anomalies over the  
62 Asian-Australian monsoon regions [e.g., Saji and Yamagata 2003; Meyers et al. 2007;  
63 Cai et al. 2009]. Another study further separated the CP El Niño into two different  
64 sub-types based on different SST anomalies over the subtropical northeastern Pacific  
65 and argued that these two CP sub-types exhibit different relationships with the IOD  
66 [Wang and Wang 2014]. At present, the exact relationship between the IOD and the  
67 two types of El Niño (EP and CP) is still not well understood. Here we discuss the  
68 different dynamical linkages between these two types of El Niño and the IOD. We  
69 conclude that the relationship between EP El Niño events and the IOD is mainly  
70 governed by El Niño event amplitude. In contrast, the CP El Niño/IOD relationship is  
71 predominantly governed by the zonal location of El Niño SST anomalies.

## 72    **2. Data and Methods**

73        The tropical Indo-Pacific SST anomalies were analyzed to demonstrate the  
74 ENSO/IOD relationship based on the Hadley Centre sea ice and SST dataset  
75 [HadISST; Rayner et al., 2003]. The associated atmospheric circulation was  
76 investigated using the National Center for Environmental Prediction/National Center  
77 for Atmospheric Research reanalysis data [Kalnay et al., 1996]. We also used  
78 sea-surface height data from the Simple Ocean Data Assimilation (SODA 2.2.4)  
79 reanalysis [Carton et al., 2000]. The anomalies are defined as a departure from the  
80 climatological mean of the entire study period (1951-2013) for all datasets, except for  
81 the SODA dataset over the period 1951-2010. A 6-120-month Butterworth band-pass  
82 filter is applied to each dataset since inter-annual variability is our focus and we wish  
83 to remove the effects of intraseasonal variability such as the Madden-Julian  
84 Oscillation, as well as variability on multi-decadal time scales. The datasets were  
85 analyzed for the boreal autumn season (September-November: SON), when the IOD  
86 usually reaches its peak and El Niño is still developing towards its peak.

87        The EP and CP El Niño indices (EPI and CPI) were calculated based on a simple  
88 transformation [Ren and Jin 2011] (also see the auxiliary material) using Niño3 (SST  
89 anomalies averaged over 5°S–5°N and 90°–150°W) and Niño4 (SST anomalies  
90 averaged over 5°S–5°N and 160°E–150°W) from the Climate Prediction Center  
91 (CPC). El Niño events are identified when the EPI or CPI exceeds 0.6 standard  
92 deviation during SON (Fig. 1). All these selected events are also identified as El Niño  
93 events by the CPC, except for the 1990 warming event, which has been identified as

94 an El Niño by many other studies [e.g., Ashok et al., 2007; Kug et al. 2009].  
95 Furthermore, the El Niño events with EPI significantly greater than CPI are  
96 considered as EP El Niño events, while those with EPI significantly less than the CPI  
97 are defined as CP El Niño events. Here, “significance” means a clear separation of  
98 respective error bars for the two El Niño flavors (Fig. 1). Therefore, there are eight EP  
99 El Niño events (1951, 1957, 1965, 1972, 1976, 1979, 1982, 1997) and eight CP El  
100 Niño events (1977, 1986, 1990, 1991, 1994, 2002, 2004, 2009), which is mostly  
101 consistent with previous studies [e.g., Ashok et al. 2007; Zhang et al., 2011]. The  
102 other five years (1963, 1969, 1987, 2003, 2006) are classified as mixed type El Niño  
103 events, which will not be discussed in the remainder of the paper considering the  
104 uncertainty of the classification. Our qualitative conclusions remain the same if we  
105 use other CP El Niño indices, such as the index defined by Ashok et al. (2007).

106

### 107 **3. Results**

108 We first examine the El Niño/IOD linkage during boreal autumn (Fig. 2a). Here,  
109 the Niño3.4 index (SST anomalies averaged over 5°S–5°N and 120°–170°W) is used  
110 to measure El Niño intensity. The IOD intensity is captured by the dipole mode index  
111 (DMI, after Saji et al., 1999), which represents the SST anomaly zonal gradient  
112 between the western equatorial (10°S–10°N and 50°–70°E) and southeastern  
113 equatorial IO (10°S–0° and 90°–110°E). A strong positive correlation ( $r=0.67$ )  
114 indicates that a positive IOD usually coincides with El Niño events and becomes  
115 stronger as the intensity of El Niño increases. However, this relationship appears to be

116 caused by the EP El Niño events rather than the CP El Niño events (Fig. 2a), which is  
117 further confirmed when separating El Niño into the two different flavors (Fig. 2b,c).  
118 For the EP El Niño events, the correlation coefficient between the EPI and IOD  
119 attains value as high as 0.96 (statistically significant at the 99% level even though  
120 there are only 8 samples), indicating a nearly perfect linear relationship. In contrast,  
121 no significant linear correlation is found for the CP El Niño events ( $r=0.16$ ).

122 Previous studies demonstrated that the atmospheric response is very sensitive to  
123 the CP El Niño's SST anomaly zonal location due to the climatological basic state of  
124 the Western Pacific Ocean [Zhang et al. 2013, 2015]. Inspired by these works, we  
125 examine possible effects of CP El Niño's SST anomaly zonal location on the IOD.  
126 The longitude of the maximum zonal gradient of the equatorial ( $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ ) mean SST  
127 anomalies is used to measure the zonal location of the CP El Niño following the  
128 definition of Zhang et al. [2013]. This definition captures well the location of  
129 anomalous rising motion in the atmosphere west of the warm SST anomaly center.  
130 Here we find a strong linear relationship ( $r=0.93$ ) between the CP zonal location and  
131 the IOD intensity, significant at the 99% confidence level. The IOD tends to be  
132 weaker as the CP El Niño shifts further westward. We also test if the EP El Niño's  
133 zonal location has an impact on the IOD, but we find no robust indication for this (Fig.  
134 S1 in the auxiliary material,  $r=0.17$ ).

135 As seen above, different El Niño flavors exhibit very different linkages with the  
136 IOD: the relationship for EP events depends on the SST anomaly intensity, while the  
137 relationship for CP events depends on the SST anomaly zonal location. Next, we use a



138 composite analysis to explore possible physical mechanisms responsible for the  
139 varying El Niño/IOD relationship. We can separate the EP El Niño events with respect  
140 to their intensity during boreal autumn. We composite 3 strong EP events (SEP: 1972,  
141 1982, 1997) and 5 weak EP events (WEP: 1951, 1957, 1965, 1976, 1979). The SEP  
142 event composite exhibits the typical SST anomaly pattern of traditional El Niño  
143 events over the tropical Pacific, which is characterized by strong warm SST  
144 anomalies in the eastern tropical Pacific and cold SST anomalies in the western  
145 tropical Pacific (Fig. 3a). The atmospheric response occurs mainly over the tropical  
146 Pacific with strong surface westerly anomalies over the central and eastern Pacific.  
147 Simultaneously, the Walker Circulation weakens with anomalous large-scale  
148 ascending motion east of the dateline and anomalous descending motion over the  
149 Indo-Pacific region near 120°E (Fig. 3c). Associated with the anomalous sinking  
150 motion, a strong anomalous divergence is located over the Indo-Pacific region in the  
151 lower troposphere. The surface easterly anomalies near Java-Sumatra are effective in  
152 enhancing oceanic upwelling and thermocline tilting in the eastern tropical IO, which  
153 brings colder subsurface water to the surface and leads to negative SST anomalies.  
154 These cold SST anomalies can further enhance the surface easterly anomalies through  
155 the positive “Bjerknes feedback” loop, which favors the development and  
156 maintenance of the IOD. In comparison, the WEP event composite shows a similar  
157 SST anomaly pattern over the tropical Pacific but with a much weaker intensity (Fig.  
158 3b). Thus, we also find that the associated atmospheric response is weaker for the  
159 WEP composite (Fig. 3b,d). Over the Indo-Pacific region, we find much weaker

160 sinking motion and surface easterly anomalies over the tropical IO, which are not  
161 effective in initiating the IOD.

162 Similarly, the CP El Niño events are also separated into two groups: eastward CP  
163 El Niño events (ECP: 1991, 1994, 2002) and westward CP El Niño events (WCP:  
164 1977, 1986, 1990, 2004, 2009) according to their SST anomaly zonal locations. The  
165 SST anomalies associated with the CP El Niño events are confined to the central  
166 tropical Pacific (Fig. 4a,b), very different from the EP El Niño (Fig. 3a,b). For the two  
167 groups of CP El Niño events, the WCP composite is located about 15 degrees further  
168 westward compared to the ECP composite. In agreement, the atmospheric response to  
169 the WCP composite is also located further westward compared to the ECP composite  
170 (Fig. 4a-d). For example, the surface westerly anomalies appear over the central  
171 equatorial Pacific for the ECP events, while they are located over the western and  
172 central equatorial Pacific for the WCP events (Fig. 4a, b). For the Walker Circulation,  
173 the center of anomalous rising air is located east of the dateline for the ECP events,  
174 whereas it is located west of the dateline for the WCP events (Fig. 4c,d). There is no  
175 large difference in the location of the anomalous sinking air between the two groups  
176 over the equatorial Indo-Pacific region (Fig. 4c,d), however they exhibit different  
177 intensities, which seems inconsistent with the observed difference in surface easterly  
178 anomalies over the eastern IO (Fig. 4a,b). The zonal wind anomalies are usually  
179 located south of the equator over the eastern IO, which is the upwelling-favoring  
180 region off Java-Sumatra. To depict the zonal structure more clearly, we show the  
181 surface zonal wind anomalies averaged over the southern equatorial IO (0°-10°S) and

182 the equatorial Pacific ( $5^{\circ}\text{S}$ - $5^{\circ}\text{N}$ ) to examine the associated atmospheric response (Fig.  
183 S2 in the auxiliary material). Consistent with the surface wind anomalies in Figure 4a  
184 and b, the zonal wind anomaly center is clearly shifted westward for the WCP events  
185 over the tropical Pacific in comparison with the ECP events, and a slight westward  
186 displacement is found over the IO. However, over the southeastern equatorial IO,  
187 significant easterly anomalies occur near the Java-Sumatra coast during the ECP  
188 events while insignificant wind anomalies are found in this region during the WCP  
189 events. Away from this key upwelling region, the Bjerknes positive feedback  
190 mechanism is weak and cannot effectively produce strong negative SST anomalies  
191 over the eastern equatorial IO. Thus, the IOD is not well developed for the WCP event  
192 composite. In contrast, the ECP associated easterly anomalies are strong off  
193 Java-Sumatra, which favors the establishment of a positive IOD.

194 The atmospheric responses to the WCP and ECP SST anomaly patterns display a  
195 large difference in amplitude in addition to the zonal location (Fig. 4 and S2), which  
196 may contribute to the differences in the surface wind anomalies over the southeastern  
197 IO. The anomalous response associated with the WCP events is only about half the  
198 amplitude of that associated with the ECP events. The interesting question that  
199 remains to be addressed is why the atmospheric responses exhibit such a large  
200 difference in amplitude between the WCP and ECP event composites despite a similar  
201 magnitude of SST anomaly forcing. One possible reason is that the negative SST  
202 anomalies over the far-western Pacific during the ECP events are stronger than those  
203 during the WCP (Fig. 4a,b). The larger SST anomaly gradient could give rise to a

204 stronger local atmospheric response. A previous theoretical study has demonstrated  
205 that the growth rate and period of ENSO over the tropical Pacific decreases as the  
206 surface wind anomaly center is displaced westward [Cane et al. 1990]. The upwelling  
207 Kelvin wave reflected by the upwelling Rossby wave at the western Pacific boundary  
208 during El Niño is more effective at returning the anomalous thermocline to its normal  
209 state when the center of the anomalous air/sea interaction is located further westward.  
210 To confirm this hypothesis, we used the zonal wind anomaly associated with the WCP  
211 and ECP to perform a linear regression on the sea-surface height (SSH) anomalies  
212 (Fig. S3 in the auxiliary material). Here the regions of 5°S–5°N, 160°–180°E and  
213 5°S–5°N, 150°–170°E are selected as the key areas of anomalous westerly activity for  
214 the ECP and WCP events, respectively, according to their surface wind anomaly  
215 patterns (Fig. 4a,b). We see that the zonal gradient of the anomalous SSH for the  
216 WCP-related surface westerly anomalies is weaker than that for the ECP. Especially  
217 over the western Pacific, the negative equatorial-mean SSH anomalies are much  
218 stronger for the ECP associated surface westerly anomalies than those for the WCP.  
219 The stronger IOD for the ECP events also contributes to a stronger Walker Circulation  
220 response and thus stronger divergence anomalies over the Indo-Pacific region and  
221 zonal surface wind anomalies compared to WCP events. Additionally, it is notable that  
222 stronger negative cloud-radiation feedback could also play a certain role on the  
223 weaker atmospheric response during the WCP than that during the ECP, due to a  
224 different background SST pattern.

225

## 226 **4. Conclusions and Discussion**

227 A large positive correlation ( $r=0.67$ ) is found between the intensity of the El  
228 Niño and IOD phenomena during the boreal autumn season for 1951-2013. However,  
229 this linkage is attributed to Eastern-Pacific (EP) El Niño events rather than  
230 Central-Pacific (CP) El Niño events. Considering different El Niño flavors, their  
231 relationships with the IOD exhibit very different characteristics. For the EP El Niño  
232 type, a near perfect linear correlation ( $r=0.96$ ) is detected between the El Niño  
233 intensity and the simultaneous IOD intensity. Compared to the strong EP (SEP) events,  
234 the weak EP (WEP) events are usually accompanied by a weaker atmospheric  
235 response and thus weaker surface easterly anomalies over the eastern IO. These weak  
236 easterly anomalies are not able to induce a strong local air-sea interaction and thus are  
237 not efficient in causing an IOD event. However, the zonal location of CP El Niño  
238 events is highly correlated with the IOD intensity ( $r=0.93$ ). Along with the westward  
239 movement of the westward CP (WCP) compared to the eastward CP (ECP), the  
240 associated atmospheric anomalies are shifted westward over the tropical Pacific as  
241 well. Over the upwelling-favoring region off Java-Sumatra, significant easterly  
242 anomalies occur during the ECP events while insignificant wind anomalies are found  
243 during the WCP events. Thus, the Bjerknes positive feedback in the IO cannot  
244 effectively be perturbed during the WCP, resulting in only a weak IOD. It can be seen  
245 that the El Niño/IOD relationship experienced a remarkable change due to the ENSO  
246 regime shift. Especially for recent decades when the CP type dominates the El Niño  
247 phenomenon, the zonal location of El Niño events need be emphasized to examine the

248 ENSO/IOD relationship.

249 A previous study [Wang and Wang, 2014] further separated the CP El Niño into  
250 two sub-types and argued, based on composite analysis that one sub-CP type  
251 co-occurs with positive phases of the IOD while the other sub-CP type accompanies  
252 negative phases of the IOD. However, no negative values of the DMI (and thus no  
253 negative IOD events) are found in this study (Fig. 2d). The difference between their  
254 and our results can be explained by the choice of selected El Niño events and the  
255 differing methodologies. For instance, the CP region exhibits pronounced decadal  
256 variability [Zhang et al. 2014], which we remove in our study, as the interaction  
257 between the interannual ENSO phenomenon and the IOD is our focus. It is also noted  
258 that they used normalized IOD values, while raw values are used here. We emphasize  
259 the importance of considering the zonal location of the El Niño events in addition to  
260 its amplitude when assessing the interaction of El Niño with the tropical Indian  
261 Ocean.

262 Another previous study displayed a high consistency between ENSO amplitude  
263 and the ENSO/IOD correlation, especially exhibiting a simultaneous decadal  
264 enhancement around the late 1970s [Santoso et al. 2012], which is consistent with the  
265 EPI/IOD relationship in this study. However, this consistency may be weakened due  
266 to more frequent occurrences of CP El Niño events in the recent decade. This study  
267 also sets a further challenge for forecast models to accurately predict both the  
268 amplitude and location of El Niño – our earlier work suggests that impactful  
269 teleconnections greatly depend on whether a CP or EP El Niño occurs [e.g., Zhang et

270 al., 2014] – but here we go further to suggest that the location of the CP events  
271 themselves causes a great variation in connections to the Indian Ocean. Further efforts  
272 are thus required to more realistically capture different El Niño features in coupled  
273 climate models although considerable process has been made [e.g., Guilyardi et al.  
274 2009; Bellenger et al. 2014].

275

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281

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372

### 373 **Figure Captions**

374 **Figure 1.** Normalized EPI (red) and CPI (green) during all El Niño boreal autumn  
 375 (SON) seasons for the 1951-2013 period (note that El Niño events only are shown,  
 376 not all years). Error bars represent 0.5-standard deviation error estimates for EPI and  
 377 CPI. EP and CP indicate different types of El Niño as described in the text. MIX  
 378 denotes mixed El Niño events that cannot be clearly separated into the two types.  
 379 Units are □.

380 **Figure 2.** Scatter diagrams of DMI (□) with (a) the Niño3.4 (□) for both EP (circle)  
 381 and CP (square) El Niño events, (b) the intensity (EPI in □) of EP El Niño events, (c)  
 382 the intensity (CPI in □) of CP El Niño events, and (d) longitude (Xt; °E) of CP El  
 383 Niño events during autumn. The longitudinal position is defined as the longitude of

384 the maximum zonal gradient of the equatorial ( $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ ) mean SST anomalies. The  
385 correlation coefficients in (a), (b), and (d) exceeds the 99% confidence level, while  
386 correlation coefficient in (c) is not statistically significant at the 80% confidence level.

387 **Figure 3.** Composite SST (shading in  $\square$ ) and surface wind (vector in  $\text{m/s}$ ) anomalies  
388 for strong (a) and weak (b) El Niño events; (c) and (d) are the same as the (a) and (b)  
389 except for the anomalous vertical pressure velocity (shading in  $10^{-2} \text{ Pa s}^{-1}$ ), Walker  
390 Circulation (vector in  $\text{m s}^{-1}$ ; the anomalous vertical velocity being multiplied by a  
391 factor of -100), and velocity potential (contour in  $10^6 \text{ m}^2 \text{ s}^{-1}$ ) averaged over  $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ .  
392 The shading and vector are only shown when the values are significant at the 90%  
393 significance level from a two-tailed Student t-test.

394 **Figure 4.** Same as Figure 3, but for the East and West CP El Niño event composite.  
395 The green dot in (a) and (b) marks the zonal location of the East and West El Niño  
396 event composite (based on the maximum zonal SSTA gradient), respectively.







