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## Direct ENSO impact on East Asian summer precipitation in the developing summer

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21 **Abstract:** In the developing stage of ENSO, the East Asia summer precipitation  
22 (EASP) shows a large variability that is significantly different from that in the  
23 decaying summer. In this study, we will focus on understanding the direct El Niño  
24 impact on the precipitation over East Asia in the developing summer in the  
25 observation. It is found that in its developing summer, the El Niño sea surface  
26 temperature anomaly affects the EASP directly from the eastern-central tropical  
27 Pacific, with little interference from the rest of the global ocean. The corresponding  
28 precipitation anomaly exhibits a tri-pole pattern, with two positive nodes in northeast  
29 and southeast China, sandwiched by a negative node in northern/central China. The  
30 tri-pole precipitation response is mainly attributed to the El Niño-induced cyclonic  
31 anomaly in Northeast Asia and the anticyclonic anomaly in the western North Pacific,  
32 which are part of the circulation anomalies of a circumglobal wave teleconnection in  
33 the subtropical jet in the Northern Hemisphere and a low-level meridional wave train  
34 along East Asia coast. These circulation anomalies are generated by the summer El  
35 Niño in three pathways: (1) the vertical motion-induced perturbation over the  
36 central-eastern tropical Pacific entering into the subtropical jet excites a circumglobal  
37 wave train propagation eastward along the jet; (2) the El Niño-induced dipole heating  
38 across the equatorial Maritime Continent is mainly responsible for the meridional  
39 wave propagation along East Asia coast; (3) the El Niño-induced indirect heating over  
40 Northwest India triggers another perturbation in the jet waveguide, all contributing to  
41 the precipitation variation in East Asia. Further demonstration indicates the  
42 atmospheric response to the El Niño direct heating and perturbation over the tropical  
43 Pacific has the major contribution to the El Niño-induced circulation anomaly. As to  
44 the El Niño indirect heating over Northwest India, a zonal wave train response in the  
45 upper midlatitude which is mainly confined in the Eurasia sector makes a competing  
46 contribution to the circulation anomaly over East Asia.

47 **Key words:** Direct impact of El Niño, East Asian Summer Precipitation,  
48 Developing stage summer, Physical Mechanisms of El Niño impact

49

## 50 **1. Introduction**

51 El Niño<sup>1</sup> is the most prominent climate variability in the tropical Pacific.  
52 Due to the phase-locking of its peak SST anomaly to boreal winter, El Niño SST  
53 anomaly usually undergoes a developing stage and a decaying stage in the preceding  
54 and following summers, respectively. Since the early 1980s, it has been noticed that  
55 the relationships between El Niño and the East Asian Summer Monsoon differ in  
56 different stages of ENSO (Fu 1987; Huang and Wu 1989; Liu and Ding 1992; Zhang  
57 et al. 1996). Most studies, however, have focused on the El Niño influence on East  
58 Asia summer precipitation (EASP) in the following summer, due to its great influence  
59 on the Mei-yu rainfall after El Niño and the precursor role of the sea surface  
60 temperature (SST) anomaly in winter providing a convenient predictor for EASP in  
61 the following summer. These studies seem to converge on the conclusion that, in the  
62 decay stage of El Niño, the climate impact of El Niño on East Asian monsoon in the  
63 following summer has to be produced indirectly through the El Niño-induced SST  
64 anomalies outside the tropical Pacific. For instance, the local air-sea coupling over the  
65 western North Pacific (WNP) maintains the anomalous Philippine Sea anticyclone to  
66 the ensuing summer and then conveys the impact of El Niño to the East Asian  
67 summer monsoon; the El Niño-induced tropical Indian Ocean SST anomaly persistent  
68 to the following summer feeds back on the atmosphere and extends the winter El Niño  
69 effect into the following summer (Wang and Zhang 2002; Yang et al. 2007; Xie et al.  
70 2009; Wu et al. 2012).

71 In comparison with the El Niño influence on the EASP in the following  
72 summer, the impact of El Niño on the EASP in the developing summer has been more  
73 controversial. Huang and Wu (1989) found that in the developing summer of El Niño,  
74 the Yangtze River and Huaihe Valleys are anomalously wet, while North and South  
75 China experience anomalously dry. It is hypothesized that the cold SST in the western  
76 tropical Pacific suppresses the atmospheric convection over South Asia, resulting in a  
77 southward shift of the subtropical high in the WNP and, in turn, abundant rainfall in

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<sup>1</sup> In this paper, unless otherwise specified, El Niño refers to both its warm and cold phase.

78 the Yangtze-Huaihe River Valley. Zhang et al. (1999, 2001) attributed the deficient  
79 rain over South China and abundant rain over the Yangtze-Huaihe Valley to the  
80 westward intensification of the western Pacific subtropical high, and the anomalous  
81 dryness over the northern part of China is due to the reduced moisture transport by the  
82 weakened Indian monsoon. Wu et al. (2003) related the East Asian precipitation  
83 anomaly with the evolution of ENSO, using the extended singular value  
84 decomposition (SVD) method. Their result indicates that in the El Niño developing  
85 summer, the major feature of the precipitation anomaly over East Asia is the negative  
86 precipitation over western North China, which is associated with an anomalous  
87 cyclone over East Asia. Wu et al. (2009) did similar analysis but using a  
88 season-dependent empirical orthogonal function (S-EOF) analysis. Their result  
89 revealed an increase of rainfall over southeastern China in El Niño-developing  
90 summer, which is associated with an intense cyclonic anomaly over the intertropical  
91 WNP and an anticyclone over the northern Indian Ocean. These controversial results  
92 motivate us to perform a comprehensive study on the relation between the EASP and  
93 ENSO in the developing summer.

94         Furthermore, from the perspective of atmospheric dynamics, previous works  
95 have suggested some teleconnection patterns in boreal summer that impact East Asian  
96 climate, with implicit relationship with ENSO. Statistically, Kripalani and Kulkarni  
97 (1997, 2001) found that the all Indian rainfall has a robust in-phase correlation with  
98 the rainfall over North China but an out-of-phase relation over South Japan. Wu (2002)  
99 furthermore pointed out that it is the mid-latitude Asian summer (MAS) pattern that is  
100 featured as barotropic anticyclones northwest of the Tibetan Plateau and over  
101 Northeast China, connecting the Indian summer monsoon with the East Asian  
102 monsoon. Hu et al. (2004) attempted to separate the internal variability of Indian  
103 rainfall influence on the atmosphere from that caused externally by ENSO. Their  
104 result indicated that the variability independent of Indian rainfall is connected to the  
105 zonal propagation over the mid-latitude Asian continent, while the variability related  
106 to ENSO prefers to propagate meridionally: one over the East Asian continent, and the  
107 other over the western Pacific Ocean. Covering the MAS pattern over the Eurasia

108 continent, also termed as the “Asian continent pattern” (Krishman and Sugi, 2001) or  
109 the “silk-road pattern” (Enomoto et al., 2003), Ding and Wang (2005) expanded the  
110 summer teleconnection around the globe, naming it as the Circumglobal  
111 Teleconnection (CGT) pattern. The CGT exhibits six prominent ‘centers of action’ over  
112 western Europe (+), European Russia(-), west-central Asia(+), east Asia(+), the  
113 North Pacific(+), and North America(+), with the sign ‘+/-’ in the bracket  
114 indicating the phase relation. Two mechanisms have been proposed for the generation  
115 of a CGT teleconnection. One is the abnormal Indian summer monsoon rainfall,  
116 which may excite an anomalous wave train over the Eurasia extending downstream to  
117 the North Pacific and North America. The other is the wave train excited in the jet  
118 region of the North Atlantic which may affect the western-central Asian High and thus  
119 the intensity of the ISM. Further investigation of Ding et al. (2011) indicated that the  
120 first leading SVD mode of the Northern hemisphere 200hPa geopotential height and  
121 the tropical rainfall in summer involves CGT pattern in midlatitude, and appears  
122 preferentially in the summer preceding the peak phase of ENSO. They suggested that  
123 ENSO forces a zonally symmetric response in the tropics and extratropics, while  
124 ENSO-related India summer monsoon is mainly responsible for the wave component  
125 of CGT.

126 In addition to the zonal teleconnection pattern in boreal summer, there is also  
127 a prominent feature of meridional propagation of Rossby wave over the western North  
128 Pacific, known as the Pacific-Japan (PJ) pattern by Nitta (1987). This PJ pattern has a  
129 significant influence on the climate in Japan through its impact on the subtropical  
130 Bonin High. Kosaka and Nakamura (2010a) further performed a dynamical analysis  
131 of the meridional dipole mode excited by the convection anomaly near the Philippines  
132 in the context of East Asian summer monsoon. They suggested that the PJ pattern can  
133 be regarded as a moist dynamical mode, which sustains itself via both dry energy  
134 conversion and the interaction with moisture process. Their study also identified  
135 another prominent meridional teleconnection pattern, which is slightly eastward of the  
136 PJ pattern with a quarter phase shift in the meridional direction, and is likely related  
137 with the tropical SST anomaly in the developing summer of ENSO. The

138 aforementioned teleconnection either focused on the zonal propagation pattern or  
139 meridional pattern, both of which seem to be related to the ENSO in the developing  
140 stage. One question addressed here is therefore whether these teleconnections  
141 occurred coherently or occurred independent from each other. If they are related, how  
142 are they organized coherently by El Niño?

143         Recently, in studying the direct impact of SST on the atmosphere, Wen et al.  
144 (2015) identified a robust climate response over East Asia to summer El Niño-pattern  
145 SST anomaly using the Generalized Equilibrium Feedback Assessment (GEFA). Their  
146 GEFA analysis implied a potential role of the El Niño SST anomaly forcing on EASP  
147 from the tropical Pacific in the developing summer of ENSO, because one of their  
148 SST forcing factors is dominated by the summer El Niño SST anomaly in the  
149 equatorial Pacific. The GEFA analysis, however, does not show the direct summer  
150 response explicitly from the time evolution perspective and therefore does not address  
151 the direct response in summer. As a follow-up research, this study will further  
152 investigate the relation between EASP and the El Niño SST anomaly in developing  
153 summer from the time evolution perspective using the composite analysis. We  
154 explicitly demonstrate that in the developing summer, the El Niño SST anomaly in the  
155 eastern-central tropical Pacific is the direct forcing factor on the EASP. The  
156 corresponding precipitation anomaly exhibits a tri-pole pattern, with an anomalous  
157 wetness in northeastern and southern China and an anomalous dryness in  
158 northern/central China. The tri-pole precipitation response is attributed mainly to the  
159 El Niño-induced cyclonic anomaly in Northeast Asia and the anticyclonic anomaly in  
160 the WNP. Furthermore, this study identifies three possible atmospheric  
161 teleconnections that enable the El Niño SST in the tropical Pacific to affect East Asia  
162 in summer: the zonal wave train across the globe (CGT), the meridional wave  
163 propagation along the East Asia coast (PJ) and wave guide triggered by the indirect  
164 heating from Northwest India. The paper is organized as follows. The data and  
165 method are described in Section 2. The direct impact of El Niño on the EASP in its  
166 developing stage and its associated circulation anomaly are documented in Section 3.  
167 Possible mechanisms of El Niño influence on the circulation are examined in Section

168 4. Further analysis of the atmospheric response to El Niño-induced direct / indirect  
169 heating is given in Section 5. A summary and discussion are given in Section 5.

## 170 **2. Data and method**

171 The monthly precipitation data are from 160 stations covering mainland  
172 China during the period of 1958 to 2012, which is provided by the Chinese  
173 Meteorological Data Center, CMA. The domain for investigation is east of 100E, with  
174 140 evenly distributed stations. To reduce the noise, the data is reconstructed by using  
175 the first leading 10 empirical orthogonal function (EOF) modes, which retains about  
176 60 percent of the total variance. Main conclusions remain unchanged when the  
177 original rainfall data are used. The precipitation variability is represented by the  
178 precipitation percentage, which is defined as the ratio of the precipitation anomaly to  
179 its seasonal climatological mean at each station. In addition to the regional  
180 precipitation data over East China, we also used the NOAA monthly precipitation  
181 reconstruction data (PREC) (Chen et al. 2004) and land precipitation data compiled  
182 by the Climatic Research Unit (CRU) (Michell and Jones 2005) in the same period  
183 (1958-2012) for the investigation of El Niño-induced heating at global scale.  
184 Furthermore, the combined precipitation data (GPCP) (Adler et al. 2003) and NOAA  
185 interpolated outgoing long-wave radiation (OLR) in a short period (1979-2012) are  
186 also used for consistent check. Other data used in this study includes monthly means  
187 of multiple variables from the National Centers for Environmental Prediction -  
188 National Center for Atmospheric Research (NCEP-NCAR) reanalysis data from 1958  
189 to 2012 (Kalnay et al. 1996). Variables include sea surface temperature, air  
190 temperature at 850 hPa, vertical velocity at 500 hPa, wind at 850 hPa and 500 hPa,  
191 and geopotential height at 850 hPa, 500 hPa and 200 hPa. Anomalies of all variables  
192 are defined as the deviations from the mean seasonal cycle, with the long-term trend  
193 removed by a third-order polynomial filter. Regional analysis across East Asia is  
194 performed using the reanalysis on the original grid ( $2.5^\circ \times 2.5^\circ$ ), while the global  
195 analysis is performed using a low-resolution version ( $7.5^\circ \times 7.5^\circ$ ) to reduce the spatial  
196 noise.



197 We mainly focus on the impact of El Niño on EASP in its developing stage.  
198 Physically, the time scale of the atmospheric response to underlying tropical SST  
199 forcing is within a few months (e.g. Gill 1980; Hoskins and Karoly 1981; Peng and  
200 Whitaker 1995; Li and Conil 2003; Liu and Alexander 2007). Any direct response of  
201 the East Asian summer monsoon to El Niño should be determined largely by the SST  
202 anomaly concurring in the same summer. Therefore, a summer El Niño index is  
203 employed for composite analysis in this study.

204 Summer is represented by the three-month mean of June, July and August  
205 (JJA). During the period of 1958 to 2012, a total of 36 El Niño/La Niña events are  
206 identified, according to the generally-accepted criterion that the three-month running  
207 mean SST anomalies in Nino-3.4 region ( $5^{\circ}\text{S} - 5^{\circ}\text{N}, 120^{\circ} - 170^{\circ}\text{W}$ ) exceeds the  
208 threshold of  $\pm 0.5^{\circ}\text{C}$  for a minimum of 5 consecutive over-lapping seasons  
209 ([http://www.cpc.ncep.noaa.gov/products/analysis\\_monitoring/ensostuff/ensoyears.shtml](http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml)  
210 ml). When the Nino-3 or Nino-3.4 SST anomaly in JJA(0) is above/below  $\pm 0.5^{\circ}\text{C}$ ,  
211 it is defined as a strong summer El Niño/La Niña case in the developing stage, with  
212 the exception of the 1998/1999 event because of an out-of-phase SST anomaly  
213 between Nino-3 and Nino-3.4 region. As shown in Fig. 1, 24 strong summer ENSO  
214 cases are selected with 11 El Niño events and 13 La Niña events. However, due to the  
215 complexity of the evolution of ENSO, La Niña often lasts a longer time than El Niño  
216 does, such as 1970-1972, 1973-1976 and 1998-2001. To avoid the mixture of the  
217 developing and decaying years for La Niña, the four events of 1971/1972, 1974/1975,  
218 1975/1976 and 2000/2001 are excluded. The final composite analysis is obtained with  
219 the 11 warm and 9 cold events (warm minus cold, and then divided by 2). The  
220 significance of the results is examined using a Monte Carlo method, in which the  
221 so-called ‘warm’ and ‘cold’ phases of ENSO are randomly scrambled among the 20  
222 summer ENSO events by 1000 times, effectively equivalent to the student “t” test  
223 (Wen et al. 2016). Notice that, when the four La Niña events are included, the results  
224 show little change from that presented here. In addition, we also did the high-pass  
225 filter analysis with the variability above 8-year periods removed, and the results

226 remain qualitatively unchanged.

227 To evaluate the wave energy propagation in a zonally asymmetric  
 228 climatological mean flow, the wave activity flux defined by Takaya and Nakamura  
 229 (2001) is used. The formula of its horizontal component in spherical coordinates is in  
 230 the following,

$$231 \quad W = \frac{p \cos \phi}{2|\bar{U}|} \left( \begin{array}{l} \frac{\bar{u}}{a^2 \cos^2 \phi} \left[ \left( \frac{\partial \phi'}{\partial \lambda} \right)^2 - \phi' \frac{\partial^2 \phi'}{\partial \lambda^2} \right] + \frac{\bar{v}}{a^2 \cos \phi} \left[ \frac{\partial \phi'}{\partial \lambda} \frac{\partial \phi'}{\partial \phi} - \phi' \frac{\partial^2 \phi'}{\partial \lambda \partial \phi} \right] \\ \frac{\bar{u}}{a^2 \cos \phi} \left[ \frac{\partial \phi'}{\partial \lambda} \frac{\partial \phi'}{\partial \phi} - \phi' \frac{\partial^2 \phi'}{\partial \lambda \partial \phi} \right] + \frac{\bar{v}}{a^2} \left[ \left( \frac{\partial \phi'}{\partial \phi} \right)^2 - \phi' \frac{\partial^2 \phi'}{\partial \phi^2} \right] \end{array} \right)$$

232 where  $\phi$  denotes the stream function,  $(u, v)$  is the horizontal velocity, and  $a, \lambda, \phi$   
 233 are the radius of the earth, and the latitude and longitude, respectively, with  $p$  for the  
 234 pressure level. Overbar and prime represent climatology mean and deviation. The flux  
 235 is independent of wave phase and equals to the group velocity of the stationary  
 236 Rossby wave in the WKB sense. Therefore, the calculation result for the composite  
 237 based on summer El Niño index could be regarded as the stationary Rossby wave  
 238 embedded in the mean flow. In addition, to reduce the spatial noise, a spherical  
 239 harmonic analysis is applied in the low level 850hPa, where the 95% information is  
 240 maintained.

241 To separate the atmospheric response to El Niño-induced direct heating over  
 242 the tropical Indo-Pacific and indirect heating over Northwest India, a partial  
 243 regression is used. In practice, to assess the atmospheric response to El Niño-induced  
 244 direct heating, we first remove the information of the heating variability over  
 245 Northwest India. We then perform the regression of the residual atmosphere on the  
 246 residual heating time series. It is the same as the partial regression on the indirect  
 247 heating over Northwest India. The direct heating index of El Niño is defined by the  
 248 normalized principle component of the first leading EOF mode of the PREC  
 249 precipitation over the region ( $70^\circ\text{E} \sim 80^\circ\text{W}$ ,  $20^\circ\text{S} \sim 10^\circ\text{N}$ ). The indirect heating  
 250 index over Northwest India is derived from the domain average for the PREC

251 precipitation over the region ( $60^{\circ}\text{E} \sim 88^{\circ}\text{E}$ ,  $15^{\circ}\text{N} \sim 35^{\circ}\text{N}$ ), normalized by its  
252 standard deviation.

### 253 **3. Direct Impact of El Niño on East Asian Summer Precipitation in** 254 **its Developing Summer**

255 The response of EASP differs significantly between the developing stage of  
256 El Niño and the decaying stage. This difference can be seen clearly in the Hovmöller  
257 diagram Fig. 2b, which shows the composite anomalous precipitation zonally  
258 averaged over East Asia for all El Niño events (as shown in Fig. 2a) from the  
259 preceding to the following year as a function of latitude. It's clearly seen that El Niño  
260 has a significant impact on the EASP, especially in summer when a large portion of  
261 the annual precipitation is concentrated (contours in Fig. 2b). In the developing  
262 summer, El Niño forces abundant rainfall in southern China and a deficiency of rain  
263 in northern China, while, in the decaying stage, the rainfall anomaly pattern is almost  
264 the opposite to that of the developing stage, with a slight northward or southward shift  
265 of the anomaly centers. This sharp contrast of EASP response between the developing  
266 and decaying stages of El Niño is consistent with early findings (Fu 1987; Huang and  
267 Wu 1989; Liu and Ding 1992; Zhang et al. 1996).

268 Actually, the response of the East Asian summer monsoon to El Niño must be  
269 forced directly by the sea surface temperature (SST) anomalies concurring in the  
270 summer, due to the rapidity of atmospheric response to SST forcings (within a few  
271 months) (e.g. Hoskins and Karoly 1981; Peng and Whitaker 1995; Li and Conil 2003).  
272 In the decay stage of El Niño, SST anomaly in the eastern-central equatorial Pacific  
273 diminishes (thick black line in Fig. 2a). The impact of El Niño on the East Asian  
274 monsoon in the following summer should be produced indirectly through the El  
275 Niño-induced SST anomalies in other oceans as indicated by previous studies (Wang  
276 and Zhang 2002; Yang et al. 2007; Xie et al. 2009). In contrast to the diminishing SST  
277 anomaly in the decaying summer, the SST anomalies in the eastern-central tropical  
278 Pacific in most cases are already well established in the developing summer of El  
279 Niño. As shown in Fig. 2a, the averaged SST anomaly over the Nino3.4 region is

280 around  $0.6^{\circ}\text{C}$  in July, which is half of the peak amplitude in winter. The asymmetric  
281 behaviors of the El Niño SST anomalies between the developing and decaying stages  
282 indicate that the direct effect of El Niño on EASP is different from that in the  
283 decaying stage.

### 284 **1) Direct impact of El Niño on EASP**

285 We now perform a composite analysis of the EASP response in the  
286 developing stage based on the summer El Niño index defined in Section 2. This  
287 summer index is highly correlated with ( $r=0.86$ , Fig.1), but not identical to, the winter  
288 index that has been generally used in most previous studies (Huang and Wu 1998; Wu  
289 et al. 2003; Yuan et al. 2012). This difference is caused by the complex temporal  
290 evolution pattern of El Niño, such that a strong El Niño in winter does not necessarily  
291 correspond to a strong El Niño SST anomaly in its developing summer. This can be  
292 seen in, for example, the later-developing El Niño in fall (e.g. 1968/1969, 1986/1987,  
293 1994/1995). Compared with the traditional winter index (as indicated by the  
294 supplementary figure), our summer index gives a clearer and more robust signal for  
295 the summer.

296 The global SST composite anomalies in the developing summer of El Niño  
297 (Fig. 3a) show that the El Niño SST anomaly has been well established, with the  
298 maximum anomaly around  $1.0^{\circ}\text{C}$  in the eastern-central tropical Pacific. Furthermore,  
299 it is important to note that there is little SST anomaly in the rest of the global ocean.  
300 This occurs because the heat capacity of the surface ocean requires  $\sim 2$ -3 months for  
301 SST anomalies to be established by the surface heat flux. In the developing stage of El  
302 Niño, no significant equatorial Pacific SST anomaly is present in the preceding  
303 months and therefore there is no SST response outside the tropical Pacific caused by  
304 the teleconnection from ENSO SST anomaly yet. This is in contrast to the decaying  
305 summer, when significant SST anomalies in other oceans are forced by peak ENSO  
306 SST anomaly in the preceding winter (Kumar and Hoerling, 2003). In other words, in  
307 the developing summer, the newly established SST anomaly in the tropical Pacific has  
308 not sufficient time to generate significant SST anomalies in other oceanic basins. As a

309 result, SST anomaly is significant only in the tropical Pacific. Therefore, if El Niño  
310 can influence EASP in the developing summer, the SST forcing should be dominated  
311 by that from the tropical Pacific.

312 Corresponding to the composite SST anomaly (Fig. 3a), there is a significant  
313 response in the precipitation over East Asia (Fig. 3b). The anomalously wet condition  
314 mainly occurs in northeastern and southern China, sandwiching a dry anomaly in  
315 northern China. The response amplitude for precipitation is around 10% (per 1°C  
316 change of SST in the eastern-central tropical Pacific) in most areas of East China,  
317 which passes the 90% significant test except for southern China. This composite  
318 analysis is consistent with the GEFA response of the EASP to the leading EOF mode  
319 of the summer SST anomaly in the tropical Pacific (Wen et al. 2015). That  
320 consistency is due to the resemblance of the two SST forcing patterns (Fig. 3a here  
321 and Fig. 1a in Wen et al. 2015). The amplitude of the composite precipitation anomaly  
322 is about half of the GEFA response, is the signal being proportional to the strength of  
323 the composite SST anomaly in the eastern-central tropical Pacific. Our results are  
324 consistent with those of Huang and Wu (1989) and Wu et al. (2003, 2009). The  
325 consistency of results implies a good robustness of the relationship between the  
326 summer El Niño SST anomaly in the tropical Pacific and the EASP anomaly in the  
327 developing summer of El Niño. It should be noted that the composite precipitation  
328 response over southern China is less robust than those over central and northern China,  
329 as indicated by the lack of significant precipitation anomaly there in Fig. 3b. This is  
330 mainly due to a combined effect of different El Niño-SST forcing from different  
331 locations in summer. The response of southern China precipitation in summer to El  
332 Niño-SST forcing from the eastern tropical Pacific is almost opposite to that from the  
333 central tropical Pacific. A detailed study on this issue of different types of ENSO is  
334 out of the scope of the current paper, and will be presented in a separate paper. In  
335 summary, it is clear that El Niño exerts impact on summer precipitation in East Asia  
336 in its developing stage, and predominantly through direct influences of the El Niño  
337 SST anomaly from the tropical Pacific. There is weak interference from other oceanic

338 basins.

## 339 **2) El Niño-induced circulation anomalies**

340 To understand the relation between El Niño and EASP in the developing  
341 summer, we examine the composite three-dimensional structure of atmospheric  
342 circulation anomalies over the globe.

343 First, in the tropics, the upper-level 200 hPa geopotential height anomaly  
344 shows a pair of Rossby wave ridges straddling across the central equatorial Pacific  
345 with an amplitude of  $\sim 10$  m (Fig. 4a). Compared with the equatorial Rossby wave  
346 response to El Niño in winter, which is only confined in the central and eastern  
347 equatorial Pacific (e.g. Liu et al., 2012), the response in summer extends far westward  
348 with two ‘wings’ deep into the northwestern and southwestern subtropical Pacific.  
349 Relative to the southern ridge, the northern ridge of the Rossby wave extends far more  
350 poleward (northwestward) in the subtropical Pacific. This upper atmospheric response  
351 is accompanied by a lower-level 850 hPa westerly anomaly in the equatorial Pacific  
352 and a strong ascending flow over the eastern-central Pacific in the middle troposphere  
353 (Fig. 4b). The tropical response is consistent with that to a deep equatorial heating  
354 (Gill, 1980).

355 Second, accompanying the anomalous high-pressure response over the entire  
356 tropics, there is a belt of low-pressure anomaly in the mid-latitude Northern  
357 Hemisphere (Fig. 4a), which reinforces the background circulation of the subtropical  
358 jet. Superimposed on the low-pressure belt, there is a circumglobal wave train with  
359 anomalous cyclone centers over the Mediterranean, the Iranian Plateau, Northeast  
360 Asia, the Central-North Pacific, off the California coast, the Great Lakes and the  
361 North Atlantic, as well as an anomalous anticyclone over Northwest Canada. The  
362 maximum response center is around 25° over Northeast Asia, with the others of  
363 amplitudes around 10~15m. This wave train pattern is similar to the first dominant  
364 SVD mode of 200hPa geopotential height associated with tropical heating (Ding et al.,  
365 2011), except over Western Europe where the CGT exhibits an out-of-phase anomaly  
366 over European Russia, upstream of the Iranian Plateau (Ding et al., 2005). The

367 difference occurs because the CGT mainly captures the geopotential height response  
368 to the heating of Indian summer monsoon, which is defined as the reference area to  
369 the Northwest of India ( $60^{\circ}\text{E}\sim 70^{\circ}\text{E}$ ,  $35^{\circ}\text{N}\sim 40^{\circ}\text{N}$ ), a point to be returned in section  
370 4.

371 Third, the low level response exhibits an anticyclone over the northern Indian  
372 Ocean and a wave train that propagates northward along the coast of East Asia, with  
373 an anomalous cyclone over the western tropical Pacific, an anomalous anticyclone  
374 over the western North Pacific and an anomalous cyclone over East Asia-Northwest  
375 Pacific (denoted as 'C - A -C' in Fig.4b). This wave train resembles closely to the  
376 'brother' Pacific-Japan (PJ) pattern, which is found as the second EOF mode of the  
377 850hPa vorticity in the area of  $100^{\circ}\text{E}\sim 160^{\circ}\text{E}$ ,  $0^{\circ}\text{N}\sim 60^{\circ}\text{N}$  in boreal summer  
378 (Kosaka and Nakanura, 2010a).

379 In summary, Fig. 4 reveals that there is a prominent atmospheric response  
380 over East Asia to the summer El Niño SST anomaly (in Fig.3a), which enables El  
381 Niño in its developing summer to impact the EASP.

382 We now focus on the sector of East Asia in Fig.5. It is clear that the  
383 composite precipitation anomaly in Fig.3b can be attributed to two circulation  
384 anomalies that are forced directly by El Niño: an anomalous low pressure in the upper  
385 atmosphere in Northeast Asia and an anomalous anticyclone in the western North  
386 Pacific in the lower atmosphere. At 500 hPa, a low pressure occurs in northeastern  
387 Asia, rendering the East Asian trough deepening and shifting southwestward into  
388 South China. Combined with the climatological air temperature gradient from  
389 Southwest to Northeast China (as indicated by the red contours in Fig.5a), this low  
390 pressure advects warm air in front of the trough from southern China into the East  
391 China Sea, and advects cold air behind the trough from Lake Baikal into southern  
392 China. As a result, air ascends in Southeast and Northeast China, and descends in  
393 North China. The 500 hPa vertical velocity anomalies, shown in Fig.5b, roughly  
394 coincide with the wet and dry regions shown in Fig.3b. In addition, the low-level  
395 anomalous wind provides a favorable moisture condition for precipitation.

396 As shown in Fig.5c, the anomalous anticyclone in the western North Pacific  
397 strengthens the western North Pacific subtropical high and makes it shift  
398 northeastward, which results in an enhanced moisture transport towards Southeast  
399 China. In Northeast China, the southerly wind anomalies over the southeastern tip of  
400 the anomalous cyclone in Northeast Asia vehicles abundant water vapor from its  
401 vicinity over the Japan Sea. The strong ascending and enhanced moisture supply from  
402 the South eventually intensify the rainfall in Southeast and Northeast China. To the  
403 contrary, the enhanced descending and the cold dry air from the North lead to dry  
404 conditions in northern China.

405 These composite results of atmospheric circulation response are consistent  
406 with those obtained with GEFA (see Fig.5 in Wen et al., 2015), in terms of both spatial  
407 pattern (spatial correlation 0.6 ~ 0.8) and amplitude of response. The circulation  
408 responses that are part of the wave train along the East Asian coast are also consistent  
409 to that found in Wu et al. (2003). The anomalous cyclonic vorticity over WNP  
410 identified by Wu et al. (2003) is actually an expansion of the anomalous cyclone in  
411 the East of the Philippines in Fig.5c. These consistent results suggest there is a good  
412 robustness for the two key anomalous circulations impacting climate in East Asia, the  
413 anomalous low pressure in Northeast Asia and the anomalous anticyclone in the  
414 western North Pacific. They play a direct role in linking the EASP anomalies to those  
415 of El Niño SST anomalies in the developing summer of El Niño.

#### 416 **4. Possible mechanisms**

417 The analysis above shows a clear anomalous circulation pattern that links the  
418 El Niño SST anomaly in the tropical Pacific with the East Asian climate in summer. It  
419 is mainly featured with a circumglobal wave train superimposed on a low-pressure  
420 belt in the midlatitude upper atmosphere, and a PJ-like wave train along the East  
421 Asian coast in the low level. The next question is how the El Niño SST anomaly in  
422 the tropical Pacific generates these circulation anomalies.

#### 423 **1) Associated heating sources**

424 To investigate the mechanism, we first examine the precipitation anomaly



425 stratified by the summer El Niño index to identify the forcings associated with latent  
426 heat release. Considering the small amplitude of precipitation anomalies in mid-high  
427 latitudes (not shown) and the large efficiency of tropical forcing on atmosphere, we  
428 mainly focus on anomalies of tropical precipitation and latent heat release. A common  
429 feature across all the four datasets is the presence of a tri-pole pattern, characterized  
430 by an anomalously abundant rainfall in the western equatorial Pacific elongated into  
431 the eastern equatorial Pacific and surrounded by deficient rainfalls over the maritime  
432 continent of Indonesia and over equatorial South America (Fig.6a-d). This tri-pole  
433 precipitation anomaly tends to shift the climatological precipitation toward the  
434 equatorial central Pacific, consistent with the observed precipitation change in the  
435 tropical Pacific in El Niño years (e.g. Wen et al., 2010). The precipitation anomalies  
436 also correspond well to the vertical velocity change in the middle troposphere, albeit  
437 with the disproportionately large amplitudes of vertical velocity over the  
438 eastern-central tropical Pacific relative to that over the western tropical Pacific (as  
439 shown in Fig.4b). The western tropical Pacific is located in the area of large  
440 climatological precipitation as shown in the black contour in Fig.6a. Thus, a small  
441 perturbation of the convection could lead to a substantial precipitation anomaly there.  
442 In contrast, the eastern-central Pacific lies in the relatively marine desert area with a  
443 low sea surface mean temperature, so that the ascending air has to reach a threshold to  
444 generate strong precipitation.

445         It should be noted that, in addition to the directly-induced precipitation  
446 anomaly, there is a robust negative precipitation anomaly over Northwest India, as  
447 shown in Fig.6. This negative precipitation anomaly is mainly related to the  
448 subsidence over the southwest maritime continent of Indonesia and its associated  
449 anticyclonic anomaly over the equatorial Indian Ocean. Such a situation can break the  
450 summer monsoon southwesterly wind and therefore reduce the Indian summer rainfall  
451 (as indicated in Fig.4b). This is consistent with previous studies on the failing Indian  
452 summer monsoon during the El Niño years (e.g. Kumar et al. 1997; Wang et al. 2003).

453         These heating sources and their associated vorticity perturbations may  
454 provide essential hints to understand how El Niño generates atmospheric circulation

455 anomalies. As shown in Fig.4, this can be realized through: 1) the vertical  
456 motion-induced vorticity perturbation in the central and eastern equatorial Pacific,2)  
457 the dipole heating anomaly across the equatorial maritime continent of Indonesia, and  
458 3) the suppressed heating over Northwest India that is indirectly induced by El Niño.

## 459 **2) Mechanism analysis**

460 To further explore the influence of the El Niño-induced heating sources on  
461 the atmospheric circulation anomaly in Fig.4, we calculate the vorticity, the wave  
462 activity flux and the velocity potential from the composite wind as shown in Fig.8 and  
463 9. Based on these analyses and the summer climatological background circulation  
464 (Fig.7), we can deduce three pathways of atmospheric teleconnection in boreal  
465 summer that enables El Niño to impact the atmospheric circulation and, in turn,  
466 rainfall over East Asia.

### 467 **(1) Pathway 1: Circumglobal Pathway**

468 The circumglobal wave anomaly in mid-latitude (Fig.4a) is primarily  
469 generated by the vertical motion-related vorticity perturbation from the eastern-central  
470 tropical Pacific. In boreal summer, a distinct feature in the upper troposphere is the  
471 presence of the westerly subtropical jet around  $40^{\circ}\text{N}$ , as indicated by the blue  
472 shading in Fig.7a. A wave train is outlined by the vorticity centers confined in the jet,  
473 most of which being statistically significant (as shown in Fig.8a). These zonally  
474 oriented vorticity anomalies are linked together by the distinct eastward wave activity  
475 flux in the belt of westerly jet.

476 The presence of the wave energy propagation along the westerly jet is not  
477 surprising since the jet acts as a waveguide (Branstator 2002). It is however unclear  
478 how the disturbances are excited by anomalies from the Pacific heating sources, given  
479 that the easterly mean flow in the tropics is unfavorable for the stationary Rossby  
480 wave to propagate into the extratropics. We believe that the key feature, is the  
481 secondary westerly jet and the corresponding westerly trough extending  
482 southwestward from  $\sim(120\text{W}, 40\text{N})$  towards the tropics of  $\sim(160\text{W}, 20\text{N})$ , in the  
483 subtropical eastern-central North Pacific, deep into the central tropical Pacific Fig.7a).

484 This westerly anomaly opens a ‘window’ of westerly duct in the eastern North Pacific  
485 double jet region, providing a potential way for the tropical Pacific ENSO signal to  
486 propagate into the mid-latitude and then to East Asia in the westerly wave guide  
487 (Webster and Holton, 1982).

488 As expected, Fig.8a shows a strong wave activity flux emanating from the  
489 negative vorticity center over the central subtropical Pacific, which corresponds to the  
490 El Niño-induced Rossby wave in the Northern Hemisphere (Fig.4a), to a positive  
491 vorticity off the California coast and then into the subtropical jet and wave guide over  
492 North America. This extratropical propagation occurs in the region of the mid-Pacific  
493 westerly trough, which provides several opportunities of the energy propagation from  
494 the tropics into the extratropics. First, the ‘window’ of westerly in the central tropical  
495 Pacific might play a role in communicating the Rossby waves from the tropics into  
496 the extratropics (Webster and Holton 1982). Second, the vorticity advection by the  
497 divergence flow in the central tropical Pacific (Fig.9a) could also help to transport  
498 perturbations into the westerly jet (Sardeshmukh and Hoskins 1988). Third, the  
499 southerly wind in front of the trough helps the vorticity anomaly in the central tropical  
500 Pacific to propagate into the subtropical jet (Wang et al., 2005).

## 501 **(2) Pathway 2: PJ-like wave train**

502 In contrast to the zonal wave train in the upper level, the low-level  
503 meridional wave propagation along the East Asian coast (Fig.4b) is mainly attributed  
504 to the dipole heating anomaly across the equatorial maritime continent of Indonesia.  
505 As shown in Fig.8b, a large positive vorticity anomaly over the western tropical  
506 Pacific emits energy flux into a negative vorticity center over the western North  
507 Pacific, and continues to propagate into the North Pacific. Corresponding to the  
508 northward wave activity flux in the low level, a southward flux in the upper level  
509 transmits from the positive vorticity over northern East Asia through the western  
510 North Pacific to the western-central tropical Pacific (Fig.8a).

511 These characters of the wave structure and energy propagation are similar to  
512 the ‘brother’ PJ pattern, which was found as the second EOF mode of the 850hPa

513 vorticity over the East Asian region in boreal summer (Figure15 in Kosaka and  
514 Nakamura, 2010). Their analysis suggested a high efficiency of energy conversion for  
515 the meridional mode under the basic state of the subtropical jet over East Asia, the  
516 South and East Asian monsoon system, and the subtropical anticyclone in the western  
517 North Pacific (Fig.7a and b). In their case studies, Kosaka and Nakamura (2010)  
518 attributed this pattern to an anomalous convection near the Bonin Island.

519 In our case of summer El Niño, Fig.8b shows that the dominant diabatic  
520 heating that triggers this mode of meridional wave train is located in the western  
521 tropical Pacific, coincident with the large precipitation anomaly over there (Fig.6).  
522 The anomalous cyclone/anticyclone over the western tropical Pacific in the low/upper  
523 level is therefore likely caused by the Rossby wave response to the summer El Niño  
524 SST anomaly in the eastern-central tropical Pacific. The baroclinic atmospheric  
525 response reinforces the convection, resulting in an abundant rainfall over the western  
526 tropical Pacific (Fig.6). The northwestward extension of the precipitation anomaly  
527 perturbs the western Pacific monsoon trough, and further excites the northward  
528 propagation of Rossby wave along the East Asia coast. The slight deficiency of the  
529 rainfall over Bonin Island (Fig.6) is probably a product of the dynamic process  
530 associated with the meridional mode. Besides the western tropical Pacific heating,  
531 there might also be a contribution from the large negative precipitation anomaly over  
532 the southwest maritime continent of Indonesia, where the subsidence is tightly  
533 connected with the air ascending across the equator over the maritime continent as  
534 indicated by the divergence and convergence marked in Fig.9b. Furthermore, the  
535 meridional pattern shown in Fig.4b might also involve an influence from the zonal  
536 wave propagation in mid-latitude, especially for the northern portion of the wave train  
537 over the North Pacific, which indicates the complicity of the meridional mode  
538 presented here, a point to be returned later.

### 539 **(3) Pathway3: South Asia Path**

540 In addition to the two direct effects of El Niño on the circulation anomaly  
541 discussed above, we can also see an indirect effect originating from the diabatic

542 heating over Northwest India. The robust negative precipitation anomaly over  
543 Northwest India (Fig.6) is caused by an anomalous anticyclone over the tropical  
544 Indian Ocean that breaks the Indian monsoon westerly (Fig.4b). The anticyclone is  
545 mainly induced by the subsidence over the southwest maritime continent of Indonesia.  
546 As demonstrated by Wang et al. (2003), the sinking branch of the anomalous Walker  
547 circulation induced heating centered on the equatorial Maritime Continent, which  
548 could generate an asymmetric Rossby wave response biased to the Northern  
549 Hemisphere, in the presence of the mean easterly vertical shear, as indicated in Fig.7a  
550 and b, a pronounced southwesterly monsoon over the North Indian Ocean  
551 corresponding to a strong upper-level easterly wind on the south edge of the South  
552 Asian High. In addition to the dominant influence of the subsidence over the Maritime  
553 Continent to the northwest of Australia, there may also be a contribution from its  
554 neighbor air ascending over the western tropical Pacific. As shown in Fig.9a and b,  
555 the suppressed heating over Northwest India induces a large convergence and  
556 divergence flow in the upper and low level, forming the second center of the velocity  
557 potential over there. The convergence flow in the upper level is so close to the  
558 subtropical jet such that it could be another potential perturbation source of the  
559 zonally propagating wave train in the westerly wave guide (Fig.8a).

## 560 **5. El Niño Direct and Indirect Heating Effects**

561 We now further separate the atmospheric responses to the direct and indirect  
562 heatings of the summer El Niño using the partial regression method. Here, the direct  
563 heating refers to those over the eastern-central tropical Pacific and the western Pacific  
564 (due to the strong co-variability of the two heating anomalies, it is difficult to do  
565 further separation of their impacts in observation), while the indirect heating refers to  
566 that over Northwest India (Fig.6).

567 The direct El Niño heating is represented by the so called Pacific heating  
568 index (PHI) that is defined as the normalized principle component of the first leading  
569 EOF mode of the PREC precipitation over the tropical Indo-Pacific region ( $70^{\circ}\text{E}\sim 80$   
570  $^{\circ}\text{W}$ ,  $20^{\circ}\text{S}\sim 10^{\circ}\text{N}$ ) (the red curve in Fig.10c). As shown in Fig.10a, a dipole

571 precipitation anomaly straddles across the equatorial Maritime Continent with the  
572 positive node tilting into the northern western-central tropical Pacific. This pattern is  
573 similar to the composite precipitation anomaly in both the spatial pattern and  
574 amplitude (Fig.6a).

575 For the indirect Indian heating index (IHI) over Northwest India, we use the  
576 domain averaged precipitation anomaly over Northwest India ( $60^{\circ}\text{E}\sim 88^{\circ}\text{E}$ ,  $15^{\circ}\text{N}$   
577  $\sim 35^{\circ}\text{N}$ ) (the blue curve in Fig.10c). As shown in Fig.10b, besides the primary  
578 negative precipitation anomaly over Northwest India with amplitude about one-third  
579 larger than that in Fig.6, the regression map on IHI also shows a moderate heating  
580 over the western and eastern tropical North Pacific. This indirect heating index IHI  
581 has relatively low, albeit statistically significant, correlation with the direct heating  
582 index of  $\sim 0.3$  (on PHI), indicating the substantial independence of the India summer  
583 monsoon heating. We note that our major conclusion below is insensitive to the  
584 choice of the size or shift of these domains for the definition of these indices of the El  
585 Niño-induced heating sources.

### 586 **1) Direct heating effect**

587 The atmospheric response to the direct heating is widespread (Fig. 11a, b).  
588 In the upper level, it shows a cross-equator Rossby wave response in the  
589 eastern-central tropical Pacific with a maximum amplitude around 10-m, and a wave  
590 train response superimposed on a low-pressure belt in the mid-latitudes. The response  
591 pattern is quite similar to the atmosphere anomaly in Fig.4a, indicating a major role of  
592 the direct heating there. The zonal contrast of the atmospheric response with the  
593 anomalous high (low) in the tropics (extratropics) is also consistent with the finding  
594 of Ding et al. (2011). Furthermore, in mid-latitude, the atmospheric response bears  
595 some resemblance to the PNA response over the Northeast Pacific and America sector,  
596 with an anomalous low pressure off the California coast, an anomalous high over  
597 Northwest Canada, and an anomalous low over Northeast America. Besides these  
598 anomaly centers; there are also low-pressure centers with amplitude  $\sim 10\text{m}$  over the  
599 North Atlantic, the Mediterranean and Caspian Sea, Northeast Asia and the Northwest

600 Pacific, forming a zonal wavenumber 5 wave train in the westerly jet. This is in line  
601 with the theoretical study on the preferential wavenumber of the zonally stationary  
602 Rossby wave in the waveguide of summer subtropical jet (Hoskins and Ambrizzi  
603 1993; Branstator 2002). In the lower level, besides the westerly anomaly in the  
604 equator Pacific in local response to the El Niño warming, the atmospheric response to  
605 the direct heating shows an anomalous anticyclone over the tropical India Ocean and  
606 a meridional wave train along the East Asian coast (as indicated by the markers in Fig.  
607 11b), which resembles the total low-level response in Fig.4b. This indicates the major  
608 contribution of the direct heating to the low level, consistent with the above  
609 discussion of the dynamic response of the South and East Asian circulation to the  
610 dipole heating across the equator Maritime Continent.

## 611 **2) Indirect heating effect**

612 In spite of the dominant global response to the direct heating, there is also a  
613 significant contribution from the indirect heating from Northwest India to the East  
614 Asia region (Fig. 11c, d). The local atmospheric response to the Indian summer  
615 rainfall anomaly is an anticyclone anomaly in the low level and a large cyclone  
616 anomaly over the northwest of the heating in the upper level with amplitude around  
617 10m, consistent with a Rossby wave response to an anomalous heating over the Indian  
618 region (Rodwell and Hoskins 1996). In mid-latitude, the upper atmospheric response  
619 also exhibits a circumglobal wave train along  $\sim 40^\circ$  N with a wave number 5 structure.  
620 The distinct feature of this wave train is that the most anomaly centers (with  
621 amplitude around 10~15m) are mainly confined within the Eurasia-North Pacific  
622 sector, with an anticyclone over European Russia, a broad cyclone over the Iranian  
623 Plateau, a smaller anticyclone over the southwest of Lake Baikal, and two linked  
624 cyclones over Northeast Asia and the central-North Pacific. The wave train pattern is  
625 identical to the MAS pattern (Wu, 2002), and bears some resemblance to the CGT  
626 pattern defined by Ding and Wang (2005). Although this wave train does not coincide  
627 with that generated by the direct heating (Fig.11c vs Fig.11a), over East Asia, both  
628 show a cyclone response, with the response to the indirect heating somewhat

629 southward, covering a large portion of eastern China. Therefore, although the indirect  
630 heating is much less important for the global response, it seems to have an impact on  
631 East Asia comparable to that of the direct heating.

### 632 **3) Combined response to El Niño-induced heatings**

633 Finally, we show that the total atmospheric response in Fig. 4 could be  
634 reproduced using the partial regression of the atmospheric response to the direct  
635 heating in the tropical Pacific combined with that to the indirect heating over  
636 Northwest India. Compared with the composite precipitation anomaly in Fig.6a, the  
637 regressed precipitation anomalies on these El Niño-induced heating indices show a  
638 comparable amplitude over the tropical Indo-Pacific in Fig. 10a (on PHI), and a  
639 one-third larger deficient rainfall over Northwest India in Fig. 10b (on IHI). Therefore,  
640 the combined atmospheric response is constructed with the partial regressions  
641 weighted by 1.0 and 0.7 for the El Niño-induced precipitation anomaly in the tropical  
642 Indo-Pacific (Fig. 11a and b) and the suppressed heating over Northwest India (Fig.  
643 11c and d), respectively. The result in Fig. 12 is comparable to the composite response  
644 in Fig. 4 both in the spatial pattern (the cross-equator Rossby wave response in  
645 tropical Pacific, the zonal wave train superimposed on the low-pressure belt in  
646 mid-latitude, and the meridional propagating wave along the East Asian coast) and the  
647 response amplitude. The pattern correlation of the combined 200hPa geopotential  
648 height (Fig. 12a) with the composite response in Fig. 4a is around 0.89, and the  
649 amplitude ratio of the absolute value over the observed one is about 0.80. The slight  
650 smaller atmospheric response both in the tropical Pacific and mid-latitude is probably  
651 attributable to the underestimation of the vertical perturbation over the eastern-central  
652 tropical Pacific, as indicated by the vertical velocity in Fig. 12b. That's because the  
653 PHI index-associated vertical motion mainly captures the El Niño  
654 precipitation-related moisture process, but underestimates the El Niño SST  
655 anomaly-induced dry dynamic process. In the low level, compared with the  
656 meridional wave in Fig. 11b, the combined response (Fig. 12b) is more consistent  
657 with observation in Fig. 4b. This indicates the role of the atmospheric response to the



658 India summer rainfall in maintaining and reshaping the pattern, especially for its  
659 northern portion of the wave train. The resemblance of the combined response to the  
660 total response demonstrates the dominant contribution of the El Niño-induced direct  
661 heating over the tropical Pacific to the global scale circulation anomaly, and the  
662 comparable effect of the indirect (negative) heating over Northwest India on the  
663 atmospheric anomaly over East Asia in the developing summer of El Niño.

## 664 **6. Conclusion and Discussions**

665 This paper is a further attempt to understand the impact of El Niño in its  
666 developing summer on the East Asian summer precipitation (EASP). Although El  
667 Niño is not yet at its fully-developed stage, our composite analysis shows that there is  
668 a significant direct impact of El Niño on EASP, with the SST anomaly in the  
669 eastern-central tropical Pacific as the dominant forcing. The corresponding  
670 precipitation anomaly is characterized by a pair of abnormal rainfall in northeastern  
671 and southeastern China, sandwiching a deficient rainfall in northern/central China.  
672 This tri-pole precipitation anomaly of EASP can be attributed to two key circulation  
673 anomalies induced by El Niño: an anomalous cyclone in the upper atmosphere in  
674 Northeast Asia and an anomalous anticyclone in the lower atmosphere in the western  
675 North Pacific. The anomalous cyclone provides a favorable condition for ascending  
676 over Southeast and Northeast China, and for descending over northern China, while  
677 the anomalous anticyclone provides the moisture supply into Southeast China.

678 These circulation anomalies are forced by El Niño in three teleconnection  
679 pathways, as indicated schematically in Fig. 13. First, the direct heating associated  
680 with the eastern-central Pacific SST anomaly induces the vertical motion and then the  
681 upper level vorticity perturbation over the eastern-central tropical Pacific; this upper  
682 level vorticity perturbation is located in the vicinity of the subtropical trough in the  
683 central tropical Pacific, enabling a stationary Rossby wave propagating into the  
684 waveguide of the subtropical jet and then to East Asia in the circumglobal  
685 teleconnection wave train (Branstator 2002; Ding and Wang 2005). Second, the direct  
686 heating of dipole structure across the equatorial Maritime Continent that is associated

687 with the high precipitation over the western tropical Pacific and the large subsidence  
688 over Northwest Australia could generate a northward propagation of Rossby wave  
689 along the east coast of East Asia under the background state of the East Asian  
690 climate (Kosaka and Nakanura 2010). Third, the indirect heating generated by El  
691 Niño with suppressed rainfall over Northwest India could also excite a perturbation in  
692 the subtropical jet due to its proximity to the jet, and in turn affect the East Asia  
693 through the wave guide (Wu 2002).

694 These mechanisms are also confirmed by a partial regression analysis of  
695 the atmospheric response to direct heating and indirect heating induced by El Niño  
696 In response to the El Niño-induced direct heating, a broad wave train response is  
697 generated that overlaps with a low pressure anomaly belt in the mid-latitude in  
698 the upper level; in addition, a meridional wave train is also generated along the East  
699 Asia coast in the low level. Both have major contributions to the summer El  
700 Niño-induced circulation anomaly. In response to the El Niño-induced indirect  
701 heating over Northwest India, the zonal wave train response in the upper mid-latitude  
702 is mainly confined in the Eurasian sector, which contributes to about one-third to the  
703 El Niño-induced circulation anomaly over these regions. Furthermore, the summer El  
704 Niño-induced circulation anomaly could be well reproduced by the linear combination  
705 of the atmospheric responses to the direct and indirect heatings.

706 A particularity of our study is the use of summer El Niño SST index to  
707 investigate impacts of developing El Niño on summer rainfall in East Asia, while  
708 most previous studies (Lin and Yu 1993; Gao and Wang 2007; Zong et al. 2010) used  
709 the winter El Niño SST signal for the prediction of EASP in the following summer. As  
710 shown in Fig.1, it is clear that EASP responses are both important, but different for  
711 the developing and decaying summers of El Niño. The EASP in the developing  
712 summer shows a robust interannual variability correlated with El Niño. For instance,  
713 in the summers of 1965, 1969, 1972, 1982, 1991, 1997 and 2002 (the developing  
714 stage of El Niño), northern China experienced droughts of various degrees,  
715 accompanied by unusually high precipitation in the Jianghuai River Basin and  
716 southeast China. However, in the decaying summers of La Niña (1964, 1970, 1973,

717 1988, 1995), the precipitation anomalies are roughly opposite over these regions. Our  
718 study here presents a more comprehensive analysis of the relation between the EASP  
719 and El Niño during its developing stage. Furthermore, we deduced a few mechanisms  
720 through which summer El Niño impacts the atmosphere.

721 Our findings should help to clarify some controversies in previous results  
722 (Huang and Wu 1989; Zhang et al. 1996; Wu et al. 2003; Wu et al. 2009; Zhu and Li  
723 2016). Our study provides a clearer picture on how the El Niño-induced atmospheric  
724 circulation anomaly forces the EASP anomaly in the developing summer. The  
725 understanding of the summer El Niño impact on the EASP implies some predictability  
726 of the EASP in the El Niño developing stage from the SST prediction from coupled  
727 climate models (<http://iri.columbia.edu/our-expertise/climate/forecasts/enso/current/>)  
728 and the observation-diagnosed response of the EASP anomaly to summer El Niño in  
729 this study.

730 It should be noted that our study here mainly distinguishes the ocean impact  
731 on the EASP in the developing stage of El Niño. There might be other factors such as  
732 heating over the Qinghai-Tibet Plateau, snow cover in Eurasia and land vegetation  
733 process and so on (Wu 2004; Wu et al. 2009; Chen et al. 2016). Comprehensive  
734 investigation and coupled model simulations on this topic will be addressed in future  
735 work. In addition, under the global warming, whether the relation of the EASP with  
736 summer El Niño will be changed is also an interesting topic to study.

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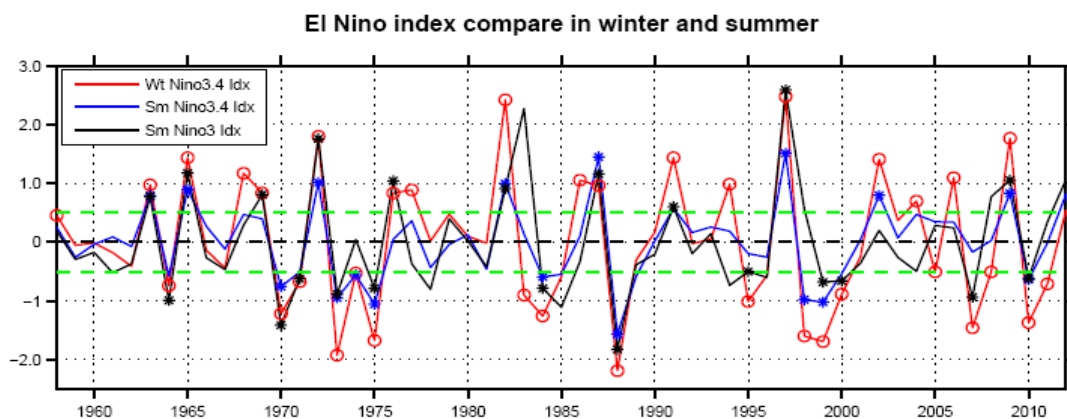
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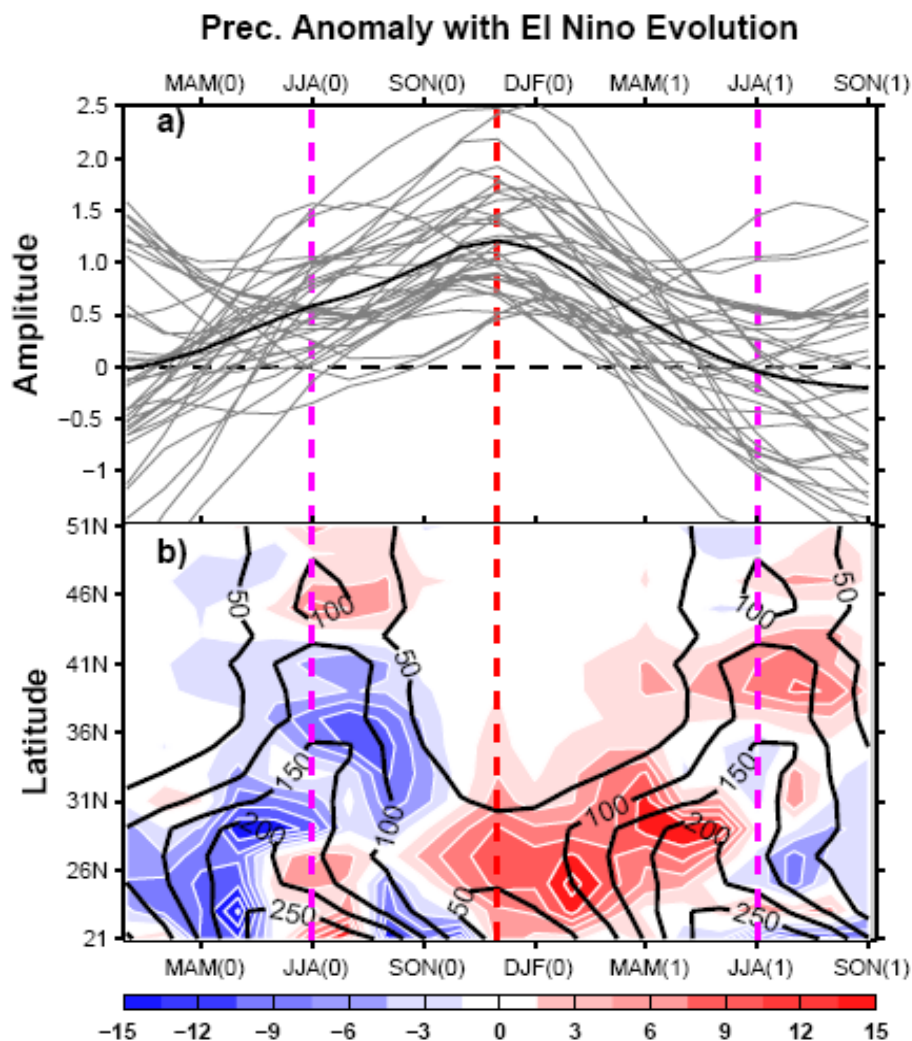




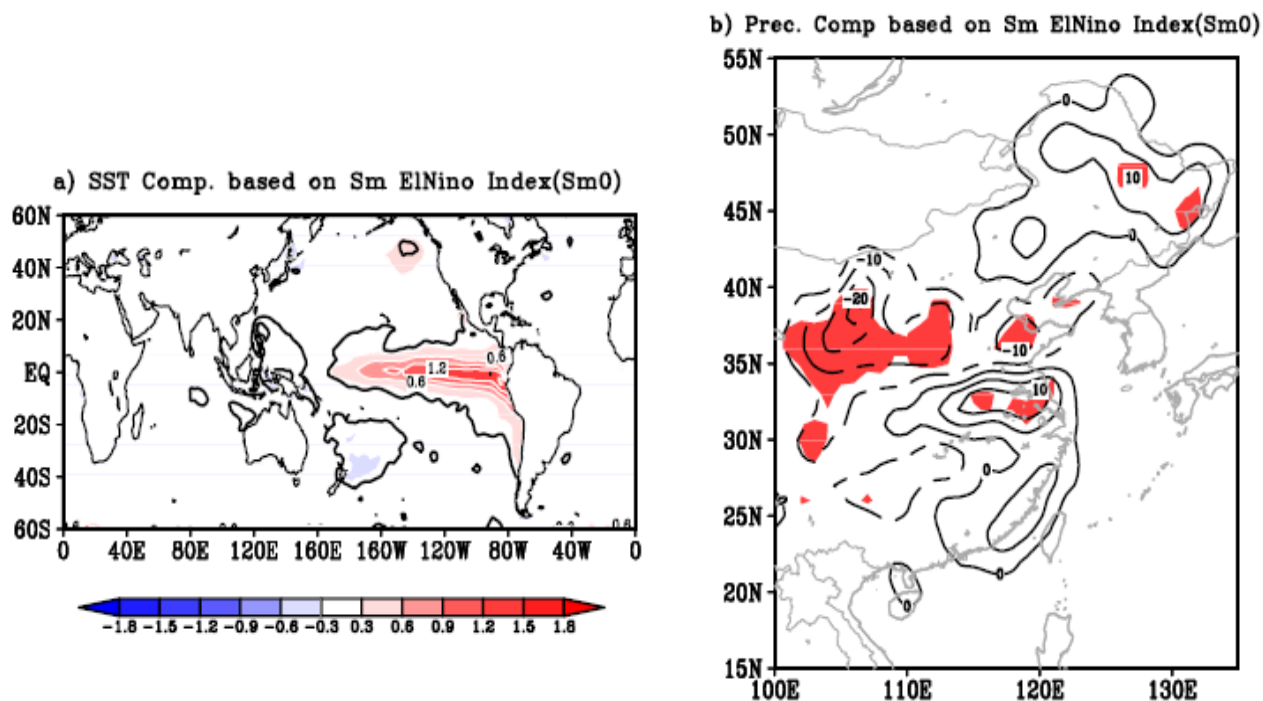
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880 **Figure 1** Comparison between the winter and summer El Niño indices. The red curve  
881 shows the Nino3.4 index in winter with red circles indicating El Niño/La Niña events.  
882 The blue and black curves show the developing summer Nino3.4 and Nino3 SST  
883 anomalies respectively with symbol 'star' marking values greater/less than  $\pm 0.5^{\circ}\text{C}$   
884 (as indicated by the green dash lines). As indicated, the strong El Niño cases are in the  
885 years of 1963/1964, 1965/1966, 1969/1970, 1972/1973, 1976/1977, 1982/1983,  
886 1987/1988, 1991/1992, 1997/1998, 2002/2003 and 2009/2010, and the strong La Niña  
887 cases are in the years of 1964/1965, 1970/1971, 1971/1972, 1973/1974, 1974/1975,  
888 1975/1976, 1984/1985, 1988/1989, 1995/1996, 1999/2000, 2000/2001, 2007/2008  
889 and 2010/2011, respectively.

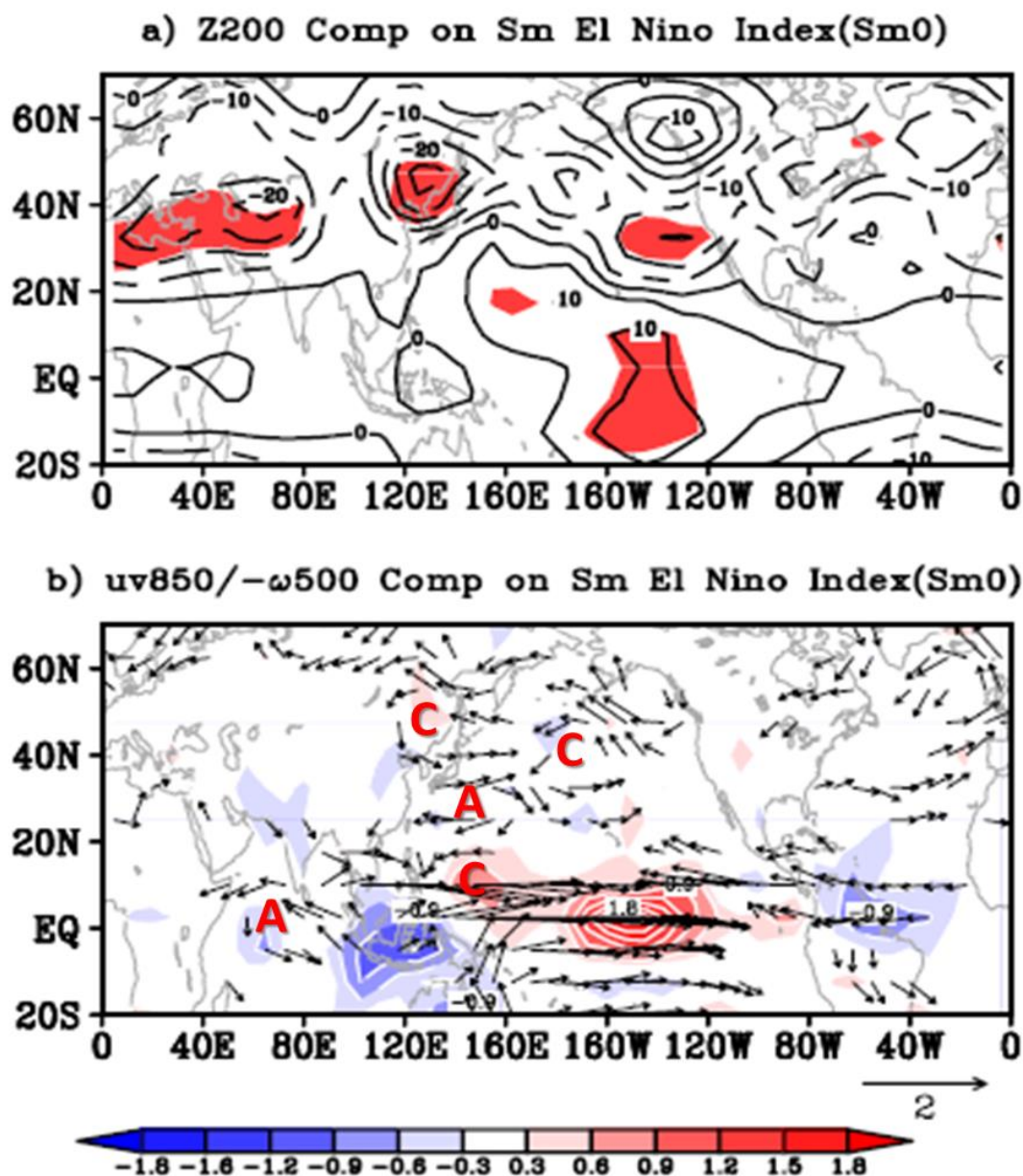
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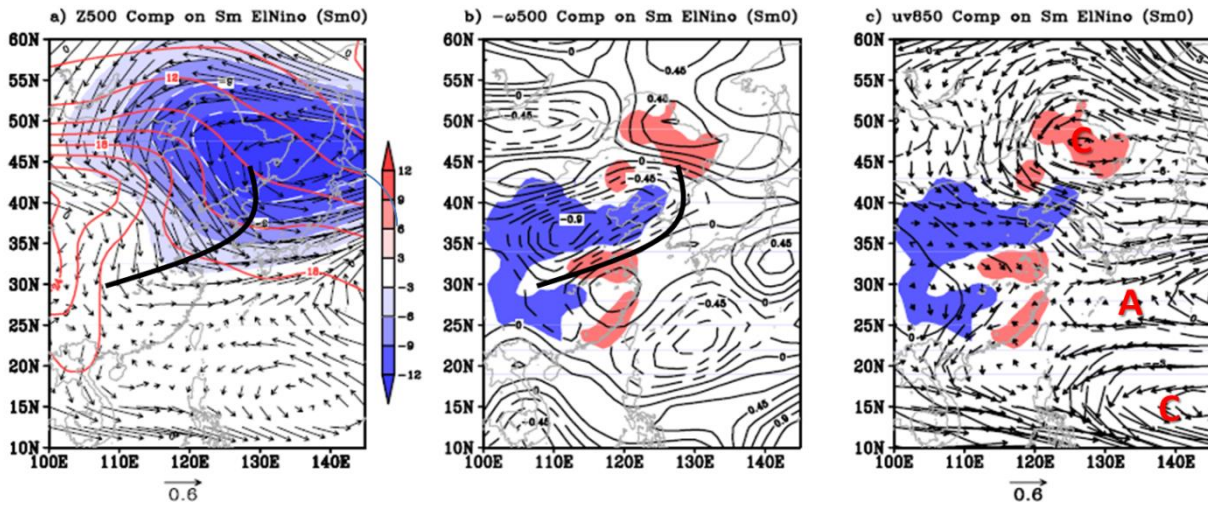
**Figure 2** Temporal evolution of El Niño and the corresponding zonal mean precipitation anomaly in East Asia. a) Three monthly mean SST anomalies (unit:  $^{\circ}\text{C}$ ) in Niño-3.4 region from February, March and April (FMA) of the developing year to September, October and November (SON) of the following year. The gray solid lines are for each individual El Niño event from 1958 to 2012. b) Zonal mean precipitations anomalies (unit: mm) in East Asia (East of 110E) corresponding to the ensemble mean of all El Niño events (thick black line in Fig.a). The shading indicates anomalies, and the black contours (contour interval of 50 mm/month) denote the seasonal cycle of mean climatology. The developing summer (JJA(0)), mature winter (NDJ(0)) and decaying summer (JJA(1)) of El Niño are highlighted by the three colored dash lines from left to right respectively. The numerals “0” and “1” in x-coordinate indicate the El Niño developing and decaying years respectively.



**Figure 3** a) Composite global SST anomalies based on summer El Niño index. The red (blue) shading with white solid (dashed) line is for the positive (negative) value with contour interval (CI) of  $0.3\text{ }^{\circ}\text{C}$ . b) Corresponding composite summer precipitation anomalies (relative variation, in percentage) over East Asia. The solid (dashed) line denotes the positive (negative) values with CI = 5%. The thick black contour in Fig. a and red shaded in Fig. b indicates the 90% confidence level.

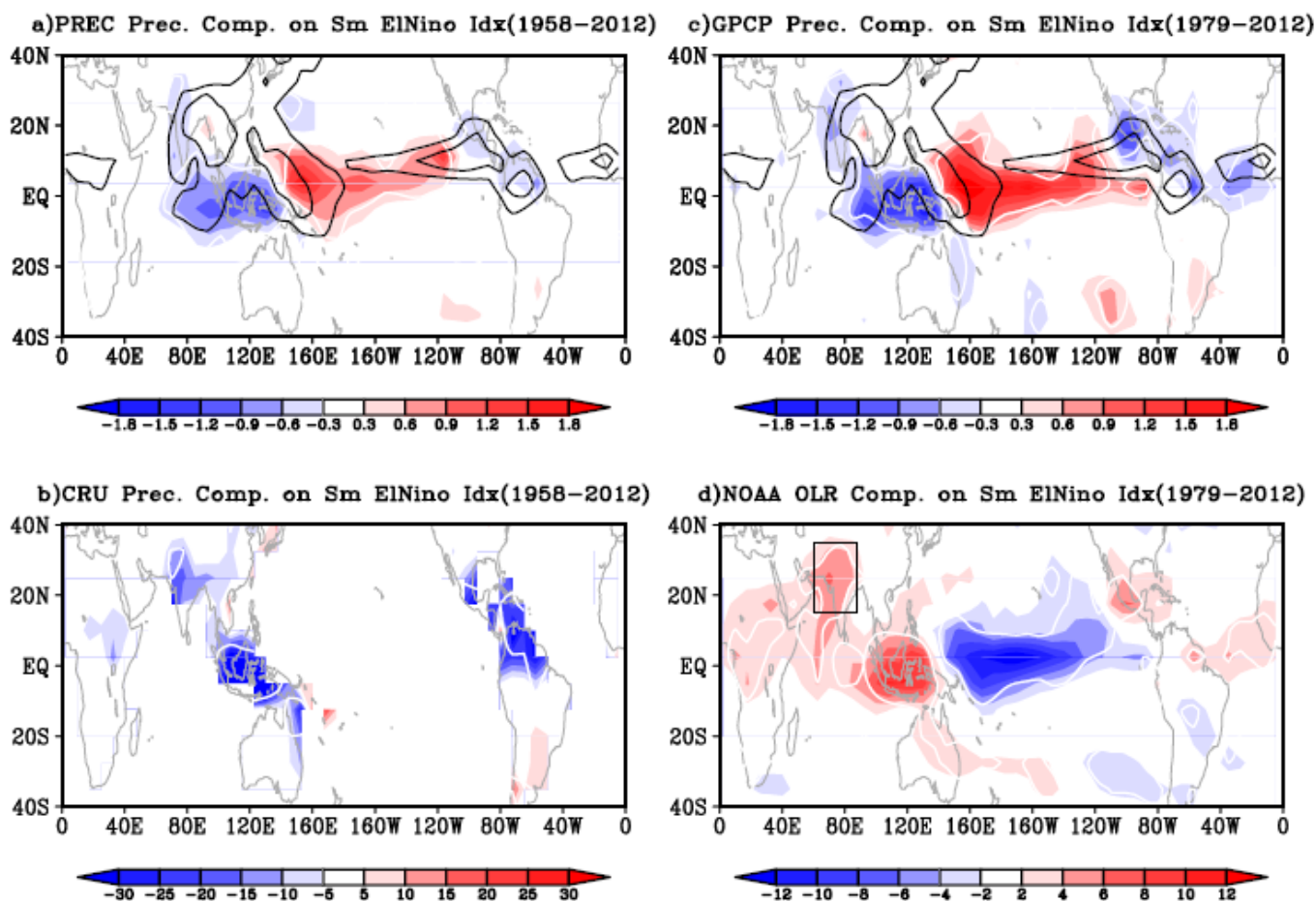


**Figure 4** Composite atmospheric anomalies in the tropical-northern hemisphere of a) 200 hPa geopotential height. Solid (dashed) line for positive (negative) value with contour interval (CI) of 5 m. The shaded indicates the 90% confidence level. b) 850 hPa wind (vector, unit: m/s) as well as 500 hPa vertical velocity ( $-\omega$ , shaded). The small value of the wind (magnitude less than 0.3 m/s) is omitted in the figure. Red letters “C” and “A” mark cyclone and anticyclone respectively. And, the red (blue) shading with white solid (dash) line (CI =  $0.3 \times 10^{-2}$  Pa/s) indicates the air ascending (descending).

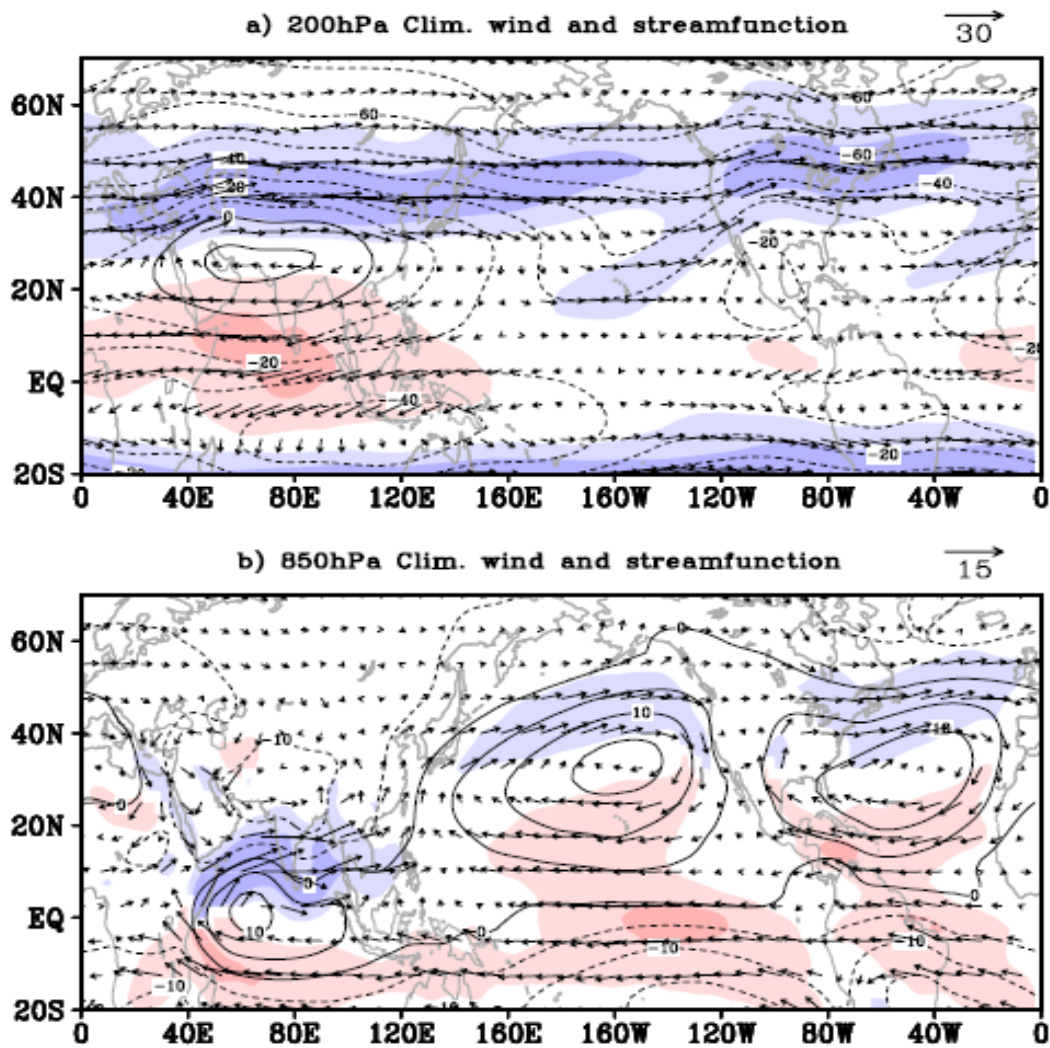


**Figure 5** Atmosphere anomalies composites over East Asia in the developing summer of El Niño for a) 500 hPa geopotential height (shaded, the blue (red) shading with white dash(solid) line for negative(positive) value, CI = 3 m) and 500 hPa wind (vector, unit: m/s) with superimposed climatological mean air temperature at 850 hPa (red contours, CI = 2 °C); b) 500 hPa vertical velocity ( $-\omega$ ) (contours, solid (dash) line for the positive (negative) value with CI of  $0.15 \times 10^{-2}$  Pa/s); c) 850 hPa geopotential height (black contours, solid (dash) line for the positive (negative) value with CI = 1 m) and 850 hPa wind (vector, unit: m/s) with red letters “C” and “A” marking the cyclone and anticyclone respectively. The shading in Fig.b and c, indicates the corresponding precipitation anomalies as in Fig.3b with red (blue) for the anomalous precipitation greater (less) than  $\pm 3\%$ . In Fig. a and b, the thick black solid line segments mark the strengthened 500 hPa East Asian trough.

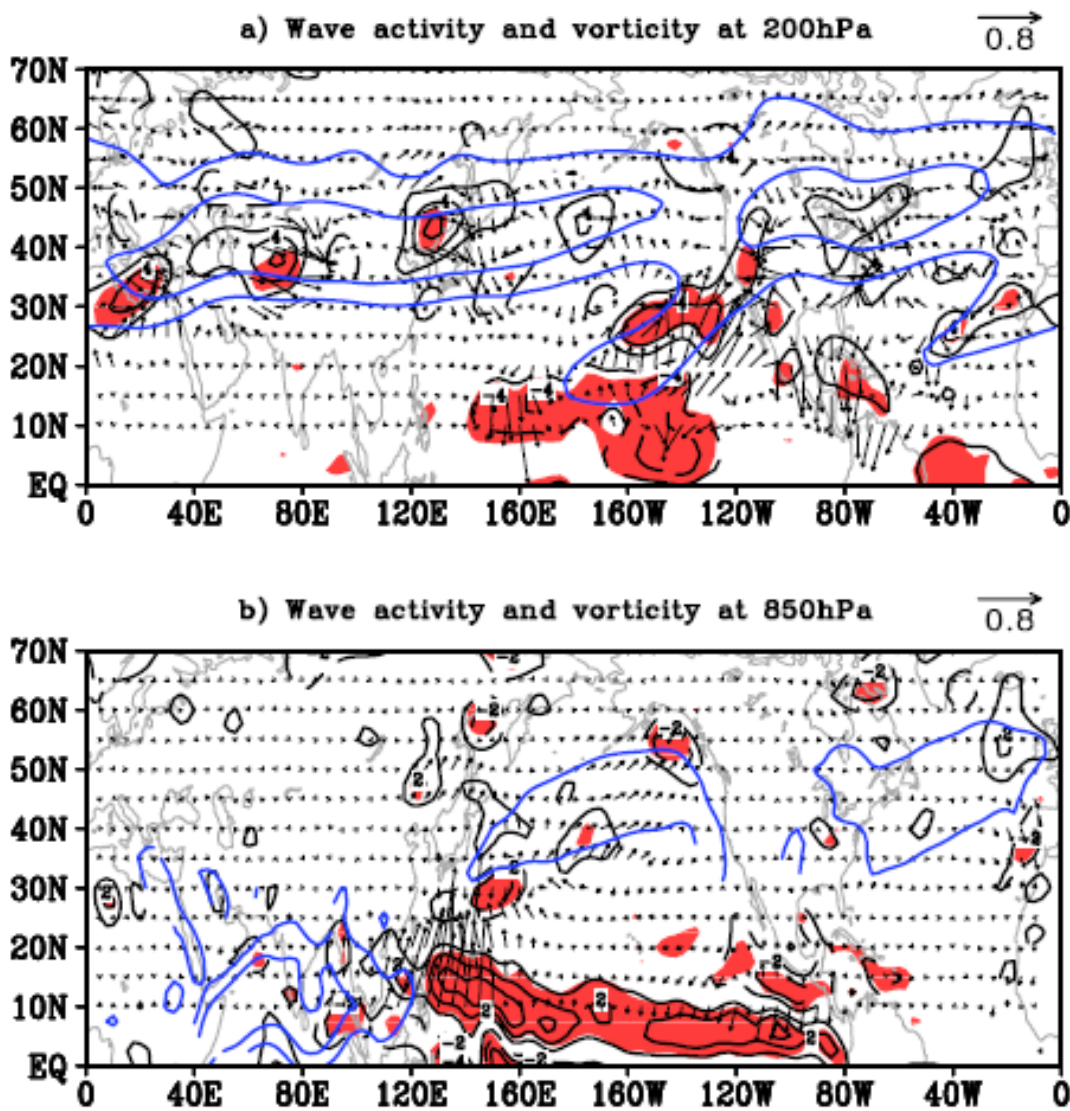




**Figure 6** Composite precipitation/OLR anomalies (shaded) from different datasets. a) PREC (CI=0.3 mm/day) and b) CRU precipitation anomalies (CI=5 mm/month) during period of 1958—2012yr. c) GPCP precipitation anomaly (CI=0.3 mm/day) and d) OLR anomaly (CI=2  $\text{w/m}^2$ ) during period of 1979—2012yr. The black contours in upper panels indicate the GPCP climatology precipitation from 1979 to 2012, with contour levels 5 and 8 mm/day respectively. White contours in the figures denote the 90% confidence level.

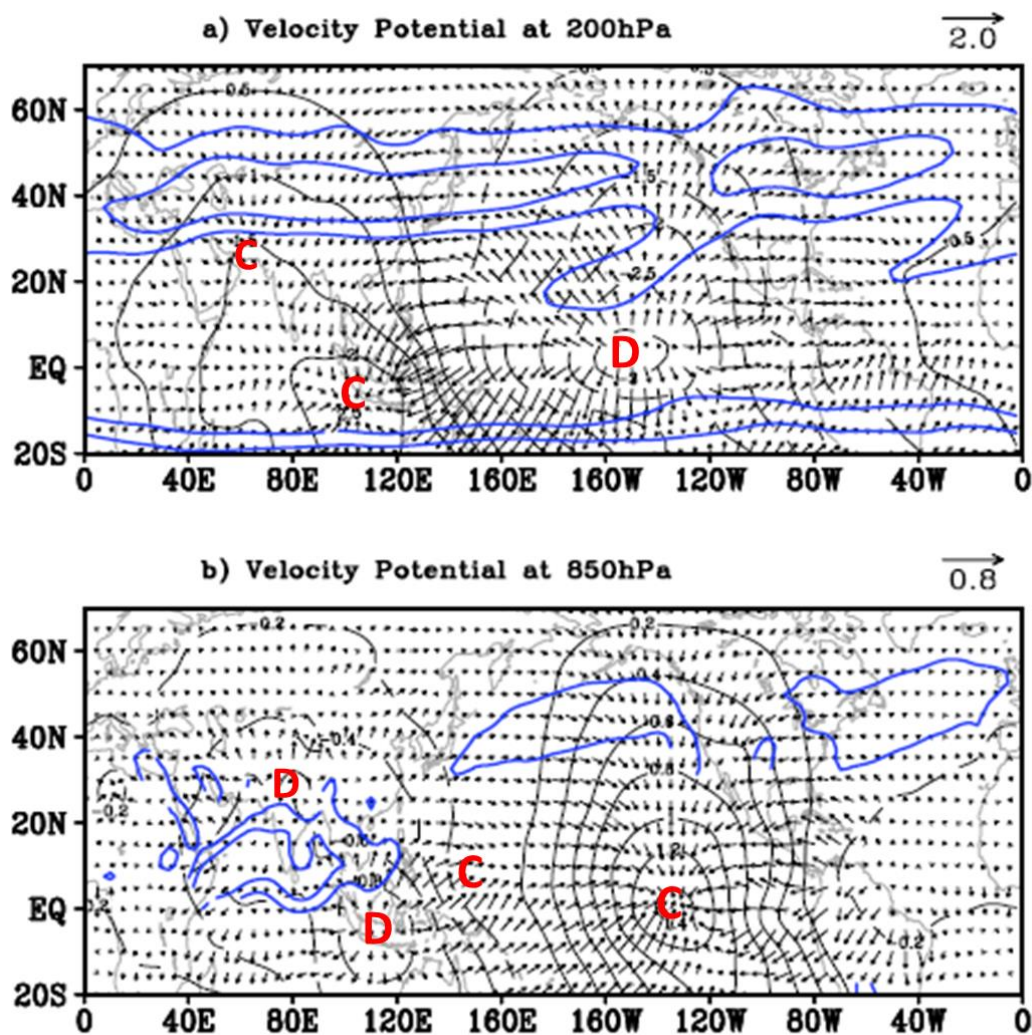


**Figure 7** Summer climatology for streamfunction (contours, unit:  $10^6 \text{m}^2 \text{s}^{-1}$ ) and horizontal wind (vector, unit:  $\text{m/s}$ ) at 200hPa (a) and 850hPa (b). In the upper / bottom panel, contour interval for streamfunction is  $10 \times 10^6 \text{m}^2 \text{s}^{-1} / 5 \times 10^6 \text{m}^2 \text{s}^{-1}$ . The blue and red shaded in the panels indicate the westerly and easterly wind with scaling at the top right corners, and the light and heavy shading denote the wind speed 10 and 20  $\text{m/s}$  at 200hPa and 5 and 10  $\text{m/s}$  at 850hPa respectively.

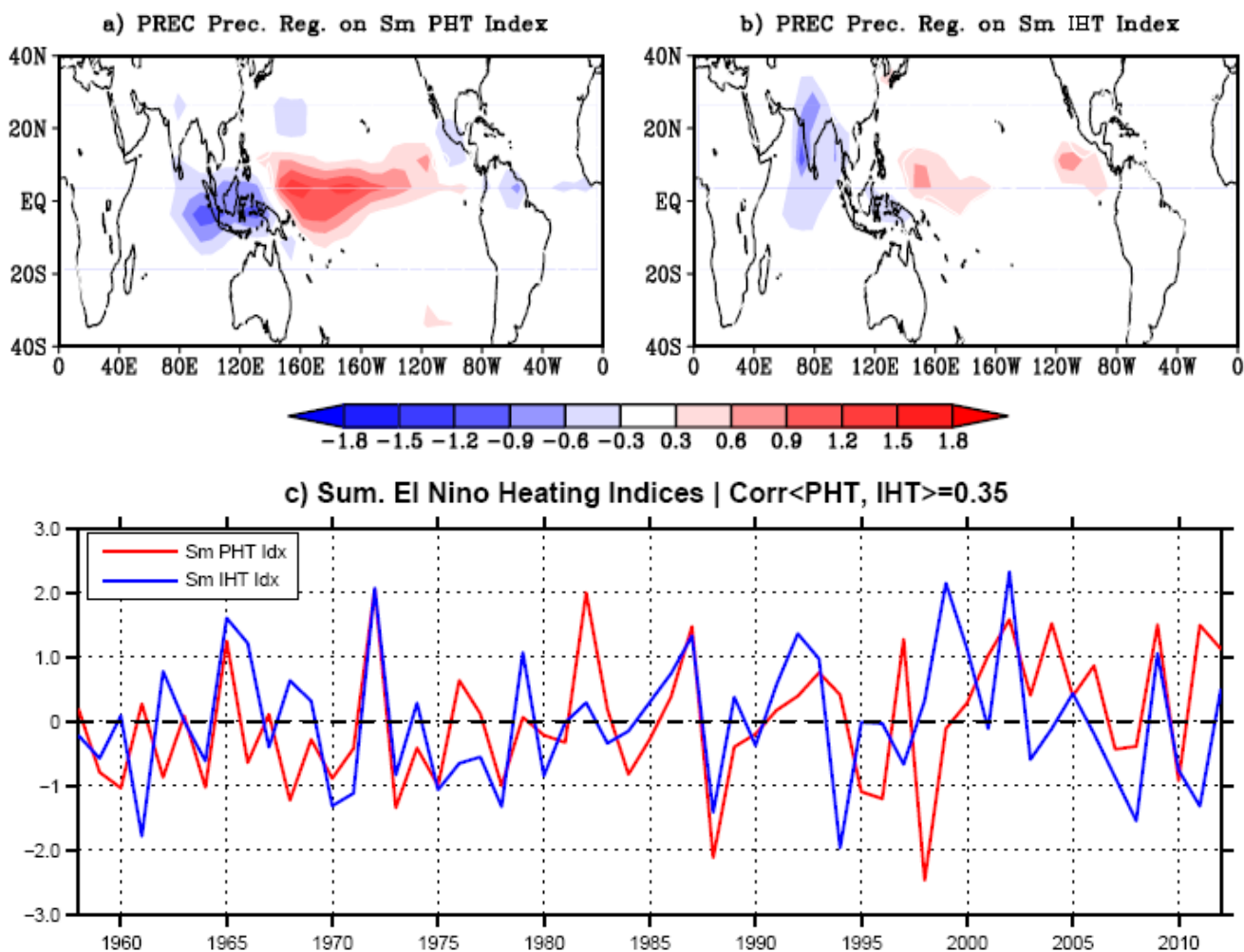


**Figure 8** Composite anomalies of vorticity (black contours, unit:  $10^{-6}\text{s}^{-1}$ ) and wave activity flux (vector, unit:  $\text{m}^2/\text{s}^2$ ) at 200hPa (a) and 850hPa (b). In upper/bottom panel, contour interval for vorticity is  $2 \times 10^{-6}\text{s}^{-1}/1 \times 10^{-6}\text{s}^{-1}$ . The superimposed blue contours in the panels denote the climatological westerly wind, with wind speed 10, 20 m/s at 200hPa and 5, 10 m/s at 850hPa respectively, as indicated by the blue shaded in Fig.7a and b. The red shaded in the figures indicates the 90% confidence level.

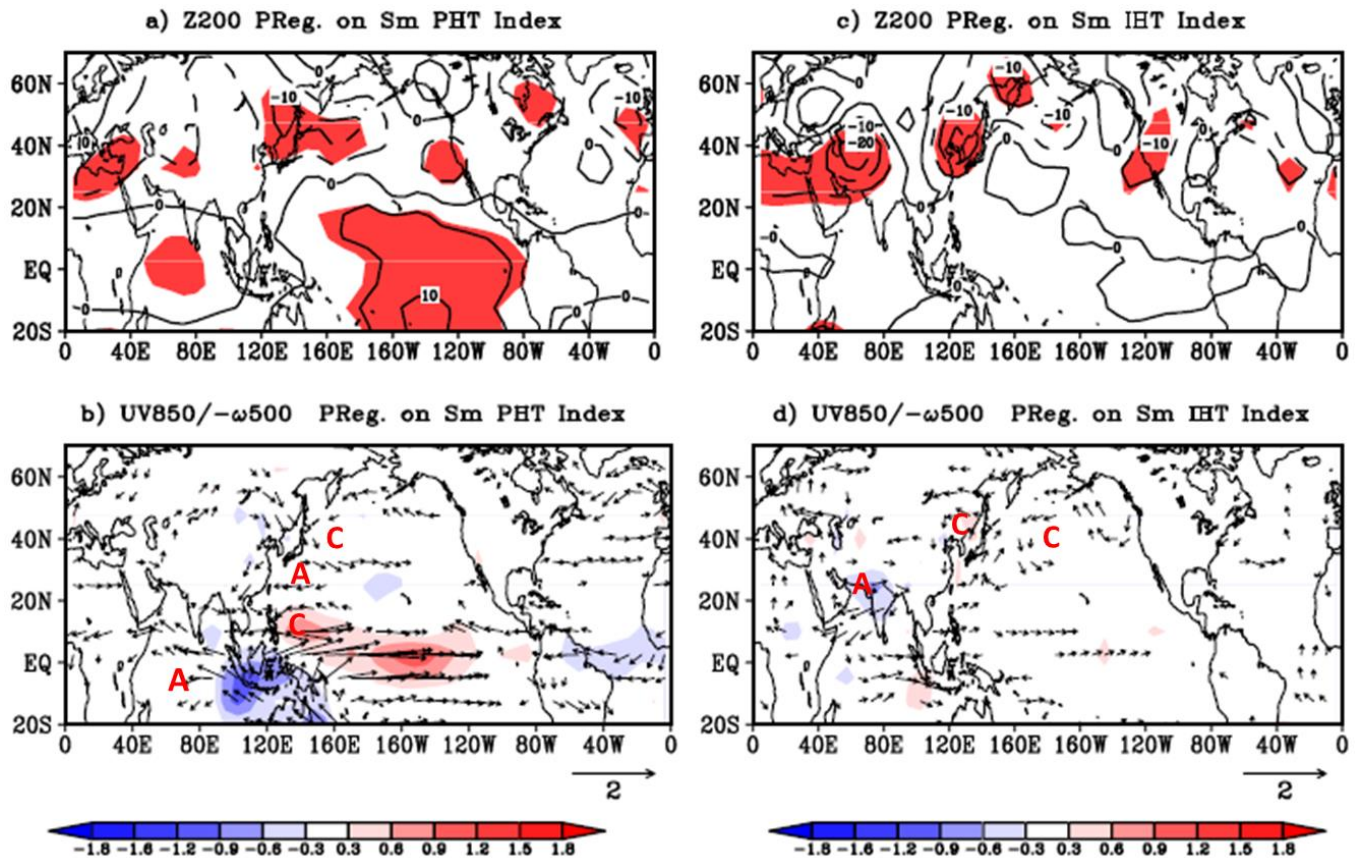




**Figure 9** Composite anomalies of velocity potential (black contours, unit:  $10^6 \text{ m}^2/\text{s}$ ) and divergence wind (vector, unit:  $\text{m/s}$ ), with climatological westerly wind (blue contours) at 200hPa (a) and 850hPa (b). In the upper/ bottom panel, contour interval for velocity potential is  $0.5 \times 10^6 \text{ m}^2/\text{s}$  /  $0.2 \times 10^6 \text{ m}^2/\text{s}$ . The superimposed blue contours in the panels denote the climatological westerly wind, with wind speed 10, 20  $\text{m/s}$  at 200hPa and 5, 10  $\text{m/s}$  at 850hPa respectively. The red letters “C” and “D” mark the convergence and divergence of the wind perturbation.

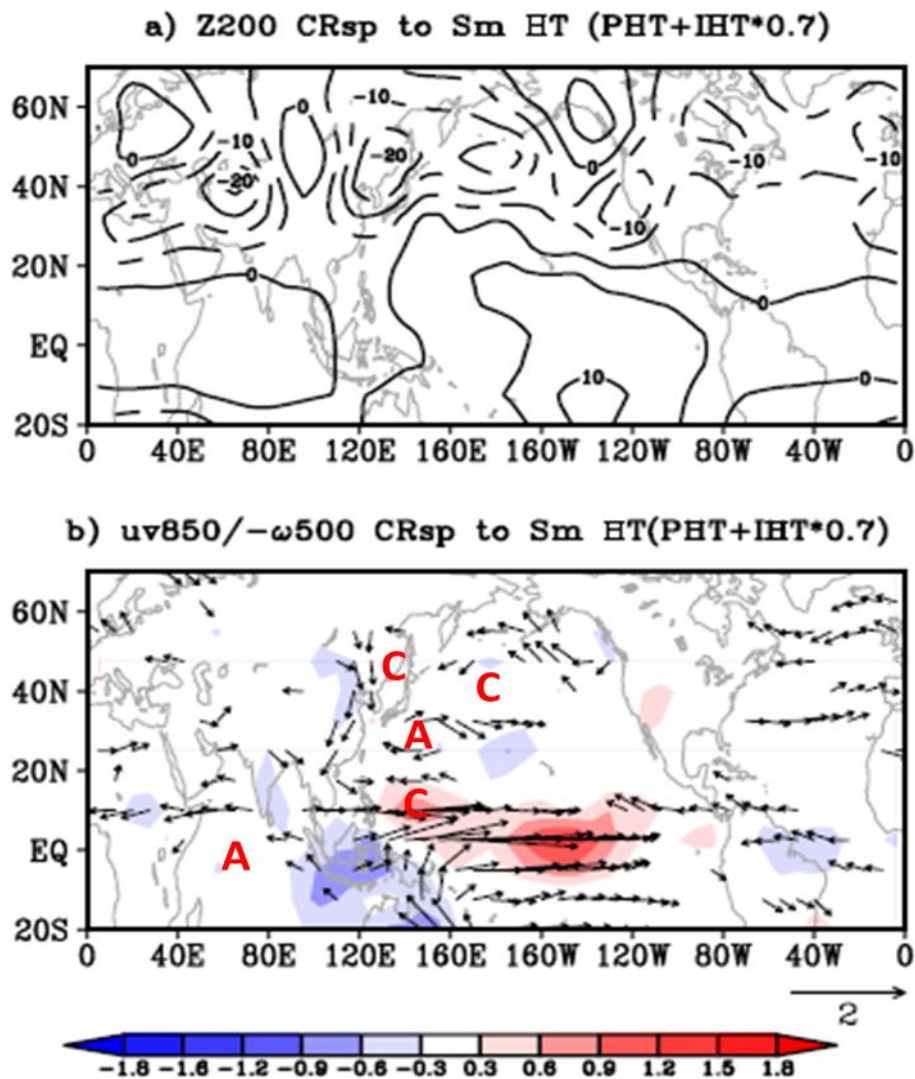


**Figure 10** Regressions of PREC precipitation on the summer El Niño-induced heating indices. a) Regression on El Niño-induced direct heating index (PHI) --- the principle component of precipitation EOF1 in the tropical Indo-Pacific ( $70^{\circ}\text{E}\sim 80^{\circ}\text{W}$ ,  $20^{\circ}\text{S}\sim 10^{\circ}\text{N}$ ). b) Regression on El Niño-induced indirect heating index (IHI) which refer to the domain average of precipitation over Northwest India ( $60^{\circ}\text{E}\sim 88^{\circ}\text{E}$ ,  $15^{\circ}\text{N}\sim 35^{\circ}\text{N}$ ). c) The corresponding indices are all normalized by their standard deviations with the red (blue) curve for El Niño-induced direct (indirect) heating index PHI (IHI).



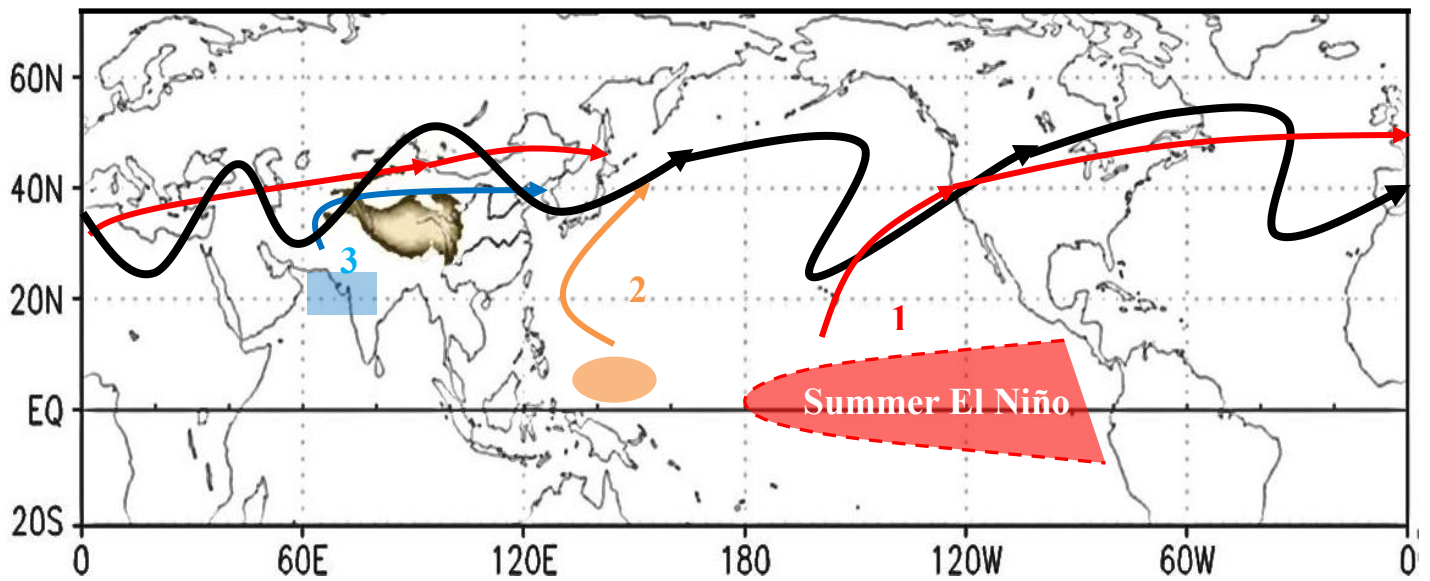
**Figure 11** Partial regressions of atmospheric variables in the tropical-northern hemisphere on summer El Niño-induced heating indices with the left panel (Fig. a and b) for the El Niño direct heating index (PHI) and the right panel (Fig. c and d) for the El Niño indirect heating index (IHI). a) and c) 200 hPa geopotential height response with solid (dashed) line for positive (negative) value (CI = 3 m). The shaded indicates the 90% confidence level. b) and d) 850 hPa wind (vector, with scaling at the bottom, unit: m/s) and 500 hPa vertical velocity (omega with opposite sign ( $-\omega$ ), shaded) responses. The small value of wind (magnitude less than 0.2 m/s) is omitted in figures. The red (blue) shading (CI =  $0.3 \times 10^{-2}$  Pa/s) indicates the air ascending (descending). And, the red letters “C” and “A” mark the cyclone and anticyclone respectively.





**Figure 12** Combined atmospheric response to summer El Niño-induced heatings, which is derived from the linear combination of the partial regressions of atmosphere to the direct heating index (PHI) (Fig.11a, b) and the indirect heating index (IHI) (Fig.11c, d), weighted by 1.0 and 0.7 respectively. a) 200 hPa geopotential height response with solid (dashed) line for positive (negative) value (CI=5 m). b) 850 hPa wind (vector, with scaling at the bottom, unit: m/s) as well as 500 hPa vertical velocity (omega with opposite sign ( $-\omega$ ), shaded) responses. The small value of the wind (magnitude less than 0.2 m/s) is omitted in the figure. The red (blue) shading (CI =  $0.3 \times 10^{-2}$  Pa/s) indicates the air ascending (descending). And, the red letters “C” and “A” mark the cyclone and anticyclone respectively.

## Mechanism of Summer El Niño influence



**Figure 13** Schematic figure showing the summer El Niño influence on atmosphere in three ways. The first path refers to the El Niño-induced vertical perturbation over the eastern-central tropical Pacific into the subtropical jet, exciting the wave propagation downstream to East Asia (as indicated by the red curve). The second is the El Niño-induced convective activity over the Maritime Continent, generating the low-level wave propagation along the East Asia coast (as indicated by the orange curve). And the third is the El Niño-induced indirect suppressed heating over Northwest India, triggering the perturbation in the subtropical jet and wave propagation to the downstream of East Asia (as indicated by the curve).