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**Sea surface chlorophyll signature in the tropical Pacific during Eastern and Central  
Pacific ENSO events**

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25

26 **Abstract**

27 Recent analyses of physical measurements show the existence of a Central Pacific type of El  
28 Niño (CPEN) with a sea surface temperature warming pattern distinct from that of the  
29 “classical” Eastern Pacific El Niño (EPEN). In this study, we analyze the surface chlorophyll  
30 signature of El Niño-Southern Oscillation (ENSO), using monthly maps of satellite-derived  
31 chlorophyll anomalies between September 1997 and December 2010. We identify five typical  
32 ENSO structures. The first structure describes the lonely 1997-1998 EPEN of the period, the  
33 second and third represent La Niña, the fourth illustrates intermediate conditions, and the fifth  
34 characterizes CPEN. During the 1997-1998 EPEN, a large eastward shift of the oligotrophic  
35 warm pool and a reduction of equatorial upwelling result in negative chlorophyll anomalies  
36 east of 170°E between 10°S and 10°N. During the four CPEN events, a reduced eastward shift  
37 yields negative chlorophyll anomalies in the equatorial band, within about 160°E and 160°W  
38 only. Westward surface current in the central basin limits the expansion of the anomaly core.  
39 Negative chlorophyll anomalies that extend eastward from the equatorial anomaly core  
40 probably result from reduced upward iron fluxes linked to the deepening of the Equatorial  
41 Undercurrent. During La Niña, the westward expansion of the equatorial upwelling results in  
42 positive chlorophyll anomalies west of the date line. Away from the equatorial band,  
43 advection of oligotrophic warm pool waters by enhanced eastward countercurrents drives  
44 negative anomalies within 8-10°N and towards the Marquesas Islands during CPEN, while  
45 reduced countercurrents leads to positive chlorophyll anomaly during La Niña.

46

47 **1. Introduction**

48 It is now well established that El Niño-Southern Oscillation (ENSO) events account for an  
49 important part of the global climate variability on interannual timescales with notable impacts  
50 on environment, ecosystems, economy, and society [Glantz, 2000; McPhaden et al., 2006].  
51 Numerous studies have shown that sea surface temperatures (SST) warmer than seasonal  
52 values invade the central and, sometimes, the eastern equatorial Pacific during the warm  
53 phase (El Niño) of the ENSO cycle. However, many differences have been observed amongst  
54 the various El Niño events. Among those differences, warm SST anomalies occur both in the  
55 eastern and central Pacific during the conventional El Niño, also referred to as canonical, cold  
56 tongue, or eastern Pacific El Niño, whilst warm SST anomalies remain confined only in the  
57 central Pacific during most of the events observed in recent years [Trenberth and Stepaniak,  
58 2001; Ashok et al., 2007; Kug et al., 2009; Kao and Yu, 2009]. Interestingly, this later type of  
59 events, referred to as central Pacific El Niño, dateline El Niño, or El Niño Modoki, has been  
60 shown to be more intense in recent decades [Lee and McPhaden, 2010] and could be more  
61 frequent in a warming world [Yeh et al., 2009]. Some studies argue that this later type of  
62 events differs from conventional ENSO [Ashok et al., 2007] while other studies argue that  
63 both types belong to an overall nonlinear phenomenon [Takahashi et al., 2011]. In this study,  
64 we will refer to these two types of El Niño as Eastern Pacific (EP) and Central Pacific (CP) El  
65 Niño.

66

67 Based on atmospheric observations, the well known Southern Oscillation Index (SOI) is often  
68 used to identify the warm and cold phases of ENSO. Based on oceanic observations, several  
69 SST indices were further proposed to distinguish the EP and CP El Niño. These mainly rely  
70 on comparisons between SST anomalies, normalized or not, in the Niño3 (5°S-5°N, 150°W-  
71 90°W) and Niño4 (5°S-5°N, 160°E-150°W) regions [Kug et al., 2009; Yeh et al., 2009; Kim

72 et al., 2009]. Some studies also rely on other possible El Niño indices such as the Trans-Niño  
73 Index [TNI; Trenberth and Stepaniak, 2001], El Niño Modoki Index [EMI; Ashok et al.,  
74 2007], and other metrics [e.g. Kao and Yu, 2009; Takahashi et al., 2011; Ren and Jin, 2011].  
75 Singh et al. [2011] also proposed ENSO indices based on differences of sea surface salinity  
76 (SSS) anomaly in two equatorial regions and in the South Pacific Convergence Zone (SPCZ).  
77 Identifications of EP and CP events following several of these methods are fairly consistent.  
78 Some discrepancies however exist and reflect the complexity of the problem as events may  
79 evolve from one type to the other [Yu and Kim, 2010; Takahashi et al., 2011] or may have  
80 patterns intermediate between the EP and CP types [Kug et al., 2009]. Table 1 summarizes the  
81 classification of EP and CP ENSO events since the strong El Niño in 1997-1998 (see also  
82 Table 1 in Singh et al. [2011] for previous years).

83

84 Although physical processes responsible for SST anomaly patterns during CP El Niño are, by  
85 far, not completely understood, they seem to differ from the ones leading to EP El Niño.  
86 Basically, during EP El Niño, the weakening of the trade winds and basin wide variations of  
87 thermocline depth lead to warming in the eastern equatorial Pacific while positive SST  
88 anomalies in the central basin result from eastward advection of the warm pool [Picaut et al.,  
89 2001; Vialard et al., 2001; Kao and Yu, 2009; Kug et al., 2009]. For CP El Niño, local  
90 atmospheric forcing and zonal advection are likely mechanisms involved in the development,  
91 confinement, and decay of anomalies in the central basin [Kao and Yu, 2009; Kug et al.,  
92 2009; Singh et al., 2011]. The so-called thermocline and advection feedbacks would then be  
93 at work during EP El Niño, while the advection feedback would mostly be at work during CP  
94 El Niño.

95

96 Mechanisms that control the SST warming during EP El Niño apparently account for the  
97 decrease of biological production in the equatorial Pacific. During the 1997-1998 El Niño,  
98 nutrient- and phytoplankton-poor (surface chlorophyll lower than  $0.1 \text{ mg m}^{-3}$ ) waters of the  
99 warm pool were advected eastward to the central and eastern basins while vertical inputs of  
100 nutrients decreased in the east because of the thermocline deepening and the reduction of the  
101 upward vertical velocity [Chavez et al., 1998; Radenac et al., 2001]. The ecosystem of the  
102 equatorial upwelling region is iron-limited [Landry et al., 1997] and the thermocline  
103 deepening is associated with the depression of the Equatorial Undercurrent (EUC) that  
104 transports iron across the basin from the western Pacific [Gordon et al., 1997; Wells et al.,  
105 1999; Lacan and Jeandel, 2001]. As biologically available iron in the photic layer is mainly  
106 upwelled from the EUC [Gordon et al., 1997], strong reduction of the iron fluxes ensues from  
107 the depression of the EUC during El Niño [Barber et al., 1996; Gordon et al., 1997; Chavez et  
108 al., 1999; Friedrichs and Hofmann, 2001]. The result is a collapse of new and primary  
109 production in the equatorial Pacific during the 1997-1998 [Chavez et al., 1999; Strutton and  
110 Chavez, 2000; Radenac et al., 2001; Turk et al., 2001] and previous EP El Niño events  
111 [Barber and Kogelschatz, 1990; Barber et al., 1996]. However, to our knowledge, no study  
112 characterizes the response of the equatorial ecosystems to the subsequent CP El Niño events  
113 in 2002-2003, 2004-2005, 2006-2007, and 2009-2010, apart from the recent paper by Turk et  
114 al. [2011]. This is the goal of this study. The occurrence and strength of CP El Niño have  
115 increased since the 1990s (Lee and McPhaden, 2010) and their frequency may still increase in  
116 future as stressed by Yeh et al. (2009). Therefore, further investigations are necessary to better  
117 understand ENSO related physical-biological interactions and their impacts on biological  
118 fields and dynamics up to top predators.

119

120 The manuscript is organized as follows. Section 2 describes the satellite-derived data that  
121 allow us to describe the 1997-2010 changes in surface chlorophyll on basin scale. To set the  
122 context, section 3 compares the SST, SSS, and surface chlorophyll anomalies in the equatorial  
123 band. Then, section 4 contrasts the surface chlorophyll anomaly signatures characterizing the  
124 EP and CP El Niño, and section 5 analyzes the possible impacts of changes in surface  
125 circulation and thermocline depth on the surface chlorophyll distribution. A discussion and  
126 conclusion appear in the last section.

127

## 128 **2. Data and methods**

129 Surface chlorophyll concentrations were derived from Sea-viewing Wide Field-of-view  
130 Sensor (SeaWiFS) measurements and from the Moderate Resolution Imaging  
131 Spectroradiometer (MODIS) measurements aboard the Aqua satellite. SeaWiFS data are  
132 available between September 1997 and December 2010 and MODIS data since July 2002. We  
133 used 9 km resolution monthly composites computed by the NASA Goddard Space Flight  
134 Center (GSFC) Distributed Active Archive Center (DAAC) [McClain et al., 2004]. When a  
135 SeaWiFS monthly map was not available or had less than 60% of data available (7 maps, i.e.  
136 4.4% of the time), we used the Aqua MODIS map in order to obtain a complete chlorophyll  
137 time series from September 1997 to December 2010. For each location, chlorophyll values  
138 higher than  $3 \text{ mg m}^{-3}$  and/or that were more than five standard deviations away from the  
139 1997-2010 mean were treated as missing (Messié and Radenac, 2006). In the calculations, we  
140 interpolated the data onto a  $1^\circ \times 1^\circ$  grid following the method of Yoder and Kennelly [2003].  
141 First, we computed  $0.25^\circ \times 0.25^\circ$  maps using the maximum likelihood estimator [Campbell et  
142 al., 1995]. Then, most of the data gaps due to sparse clouds were filled by taking the median  
143 of every  $1^\circ \times 1^\circ$  cell. The remaining missing data were filled by taking the median of the first  
144 neighbors. The chlorophyll anomalies we present are relative to a mean seasonal cycle

145 calculated between 1998 and 2010. We also used monthly maps of Photosynthetically  
146 Available Radiation (PAR) estimates derived from SeaWiFS and MODIS [Frouin et al.,  
147 2003].  
148  
149 SST data was provided by the Hadley Centre for Climate Prediction and Research Sea Ice and  
150 Sea Surface Temperature HadISST1 dataset [Rayner et al., 2003]. Monthly SST maps are  
151 available since 1870 on a  $1^{\circ} \times 1^{\circ}$  grid. SSS data came from the recently-updated Delcroix et al.  
152 [2011] product for the tropical Pacific. Monthly SSS data are available on a  $1^{\circ} \times 1^{\circ}$  grid and  
153 span from 1950 to 2009. Monthly near surface currents were obtained from the Ocean Surface  
154 Current Analysis - Real time (OSCAR)  $1^{\circ} \times 1^{\circ}$  product; the geostrophic, wind-driven, and  
155 thermal-wind components of which are derived from satellite data [Bonjean and Lagerloef,  
156 2002]. Monthly anomalies of wind speed and depth of the  $20^{\circ}\text{C}$  ( $Z_{20^{\circ}\text{C}}$ ) isotherms are derived  
157 from time series recorded at the Tropical Atmosphere Ocean/Triangle Trans Ocean Buoy  
158 Network [TAO/TRITON; McPhaden et al., 1998] moorings. We also used time series of the  
159 SOI and EMI [Ashok et al., 2007].  
160  
161 Different statistical procedures have been used in the literature to discriminate ENSO features.  
162 To name a few, these include regression of anomaly onto specific ENSO indices [Trenberth  
163 and Stepaniak, 2001; Ren and Jin, 2011; Takahashi et al., 2011; Singh and Delcroix, 2011],  
164 Empirical Orthogonal Functions (EOF) analyses [Ashok et al., 2007; Park et al., 2011],  
165 combined regression-EOF analyses [Kao and Yu, 2009], neural network [Hsieh, 2001; Leloup  
166 et al., 2007], and Agglomerative Hierarchical Clustering (AHC) analysis [Kao and Yu, 2009;  
167 Singh et al., 2011]. We tested here the EOF and AHC analyses on chlorophyll in the tropical  
168 Pacific. The EOF analysis of the chlorophyll anomaly indicates that the CP El Niño signal is  
169 distributed over at least the first, third, and fourth modes (not shown). Hence, the EOF



170 technique does not distinguish properly EP and CP El Niño. A similar leakage of the signal  
171 was found for the EOF analysis of SSS in the tropical Pacific [Singh et al., 2011]. Therefore,  
172 we chose to perform an AHC analysis on the monthly surface chlorophyll anomalies to  
173 characterize ENSO related signatures. In this nonlinear composite procedure, maps are  
174 merged into clusters according to their similarity, estimated from the smallest Euclidean  
175 distance (defined as the root-mean-squared distance between each pairs of maps). The  
176 clustering procedure can be represented by a dendrogram tree that illustrates combinations  
177 made at each successive step of the analysis. This technique has been successfully performed  
178 on SST [Kao and Yu, 2009], SSS [Singh et al., 2011], and South Pacific Convergence Zone  
179 (SPCZ) location [Vincent et al., 2009] to separate ENSO signatures in the tropical Pacific. In  
180 this study, we applied the AHC method to monthly maps of surface chlorophyll anomalies for  
181 the region between 130°E-70°W and 10°S-10°N. The Indonesian Throughflow region and  
182 Central and South America coastal upwellings were excluded as regional effects may mask  
183 the basin-scale ENSO signature (Fig. 1).

184

### 185 **3. Comparing SST, SSS, and surface chlorophyll anomalies in the equatorial band**

186 The mean chlorophyll distribution and its interannual variability are presented first to set the  
187 context (Fig. 1). The mean values were computed by averaging all monthly values covering  
188 the 1997-2010 period, and the interannual variability was defined as the standard deviation of  
189 the chlorophyll anomalies (relative to the mean seasonal cycle). Figure 1a shows that the  
190 mean chlorophyll is lower than  $0.3 \text{ mg m}^{-3}$  outside the Central and South America upwelling  
191 regions, with well-marked maxima in a huge equatorial region spreading westward from the  
192 South American coast. These values are mostly due to the equatorial upwelling which brings  
193 cool, salty, and nutrient-rich waters toward the surface. Although the surface nitrate  
194 concentration is high in the upwelling, the chlorophyll content remains moderate (less than

195 0.2 mg m<sup>-3</sup> on average) because of an iron-limited and grazing-balanced ecosystem [Landry et  
196 al., 1997]. Mesotrophic waters of the equatorial upwelling are surrounded by oligotrophic  
197 waters ([chl] < 0.1 mg m<sup>-3</sup>) of the warm pool to the west and of subtropical gyres poleward  
198 (note also the moderate mesotrophic waters north of Papua New Guinea and near the  
199 Solomon Sea). The interannual variability (Fig. 1b) is high between 10°S and 10°N,  
200 especially along the equator, in the North Equatorial Countercurrent (NECC) region, around  
201 Papua New Guinea, and in the Central and South America coastal upwelling regions.

202

203 Looking at the regions of maximum interannual variability, the longitude-time distributions of  
204 the 5°N-5°S averaged SST, SSS, and surface chlorophyll anomalies are shown in figure 2. For  
205 all El Niño events, the maximum positive SST anomaly occurs during boreal winter, between  
206 September and February (Fig. 2a). The location and amplitude of the anomaly differ from one  
207 event to the other. The 1997-1998 (EP, see Table 1 and below) El Niño is the only one with  
208 very warm SST anomaly (SSTA > 3°C) that peaks east of 120°W and extends toward the  
209 central basin. During the following (CP, see Table 1 and below) El Niño events, the  
210 maximum SST anomaly is lower (sometimes less than 1°C) and mostly situated in the central-  
211 western basin, at least in 2002-2003, 2004-2005, and 2009-2010. The strongest warming (SST  
212 anomaly close to 2° C) during a CP El Niño event is reached during the mature phase of the  
213 2009-2010 event, in agreement with Lee and McPhaden [2010].

214

215 The strongest negative SSS anomalies (Fig. 2b) appear to be located west of the maximum  
216 positive SST anomalies. During the 1997-1998 EP El Niño, the SSS anomaly is negative east  
217 of 160°E, with a maximum freshening near the date line. During the following CP El Niño  
218 events, the maximum negative anomalies tend to be displaced westward by 10 to 15° of  
219 longitude, except in 2002-2003, in agreement with Singh et al (2011).

220

221 The time series of surface chlorophyll anomaly (Fig. 2c) is reminiscent of the SSS anomaly  
222 time series as strong negative chlorophyll anomalies tend to be located west of the warm El  
223 Niño-related SST anomalies. In 1997-1998, the strong anomaly core ( $< -0.12 \text{ mg m}^{-3}$ ) is near  
224  $150^\circ\text{W}$  and the surface chlorophyll concentration is below the mean seasonal value from  
225  $150^\circ\text{E}$  to the American coast. During the following events, the negative anomaly core ( $-0.08$   
226 to  $-0.10 \text{ mg m}^{-3}$ ) is located in the central-western basin, between about  $150^\circ\text{E}$  and  $180^\circ$ , while  
227 the moderate anomalies observed eastward range between less than  $-0.02 \text{ mg m}^{-3}$  in 2002-  
228 2003 and 2004-2005 to  $-0.04 \text{ mg m}^{-3}$  in 2006-2007 and 2009-2010.

229

230 Oligotrophic waters ( $[\text{chl}] < 0.1 \text{ mg m}^{-3}$ ) have been shown to be quasi-persistent in the eastern  
231 part of the equatorial warm pool while moderate mesotrophic waters (surface chlorophyll  
232 concentration slightly higher than  $0.1 \text{ mg m}^{-3}$ ) were often observed in its western part  
233 [Radenac et al., 2010]. This oligotrophic region is delimited by the black contour in Figures  
234 2a, b, c. The easternmost limit of this zone characterizes the eastern edge of the warm pool  
235 [Murtugudde et al., 1999; Stoens et al., 1999; Radenac et al., 2010] that separates waters of  
236 the eastern warm pool ( $[\text{chl}] < 0.1 \text{ mg m}^{-3}$ ) from upwelled waters ( $[\text{chl}] > 0.1 \text{ mg m}^{-3}$ ). Large  
237 zonal displacements of the eastern edge of the warm pool occur at interannual timescales and  
238 its longitudinal position is related to the phases of ENSO [Picaut and Delcroix, 1995; Le  
239 Borgne et al., 2002]. It reaches  $130^\circ\text{W}$  during the 1997-1998 EP El Niño and lies between the  
240 dateline and  $160^\circ\text{W}$  during CP events. Interestingly, the oligotrophic waters of the eastern  
241 part of the warm pool follow these movements (Fig. 2c). Thus, the maximum negative  
242 chlorophyll and SSS anomaly cores are both located west of the warmest SST anomalies.

243

244 Several La Niña years interleave between El Niño years (Table 1). When La Niña events  
245 reach their mature phase in boreal winter (e.g. in early 2008), SST in the central and eastern  
246 basins are cooler while becoming slightly warmer west of 160°E (Fig. 2a). Strong positive  
247 anomalies of SSS and surface chlorophyll are closely related west of 150°W (Figs. 2b, c).  
248 They are located at the western limit of the cool SST anomaly, in the region of zonal  
249 displacements of the eastern edge of the warm pool. Chlorophyll anomalies exceed 0.10  
250  $\text{mg m}^{-3}$  in 2010 and range between 0.05 and 0.08  $\text{mg m}^{-3}$  during the other La Niña years. A  
251 moderate increase of chlorophyll ( $< 0.03 \text{ mg m}^{-3}$ ) is observed eastward of these maxima.  
252  
253 Figure 2c shows that the chlorophyll anomalies are negative all along the equator in 1997-  
254 1998 while the core of chlorophyll anomaly remains west of 150°W during the following  
255 boreal winters (see also spatial patterns from the cluster analysis below). Therefore,  
256 comparing the chlorophyll anomalies averaged over the Niño3 and Niño4 regions during the  
257 peak period of the events (September-February; Fig. 3) mostly concurs with the classification  
258 of La Niña and CP and EP El Niño events. Chlorophyll anomalies are negative in both  
259 regions during El Niño years ( $\text{SOI} < 0$ ) and positive during La Niña years ( $\text{SOI} > 0$ ). Also,  
260 differences between the anomalies in each region are consistent with the CP El Niño as  
261 depicted by the EMI (Fig. 3). The amplitude of the chlorophyll anomaly indeed tends to be  
262 greater in Niño4 than in Niño3 during CP El Niño ( $\text{EMI} > 0$ ). Following these results, the  
263 2006-2007 El Niño is identified here as a CP event as in the SSS study [Singh et al., 2011]  
264 and in contrast to most of the SST studies (Table 1). The 2009-2010 event, acknowledged as  
265 the warmest CP El Niño during the last decades [Lee and McPhaden, 2010], yields  
266 chlorophyll anomalies in Niño4 weaker than during the 2002-2003 event. Actually, the  
267 magnitude of the chlorophyll decrease in Niño4 is linked to the eastward expansion of the  
268 oligotrophic warm pool (Fig. 2c) whereas warming in Niño4 is not (Fig. 2a). During La Niña

269 winters, the chlorophyll anomaly in Niño4 is always larger than in Niño3 (except in 1998-  
270 1999), suggesting that the strongest increase of chlorophyll occur in the region of zonal shifts  
271 of the eastern edge of the warm pool. The greatest winter increase reached in the Niño4 region  
272 is in 2010-2011.

273

#### 274 **4. Describing the spatial patterns of ENSO-related surface chlorophyll anomalies**

275 Using the AHC analysis described in section 2, we identified five clusters out of 160 monthly  
276 maps during the September 1997-December 2010 period. As detailed below, we believe they  
277 are representative of EP El Niño, equatorial recovery, La Niña, intermediate CP El Niño, and  
278 CP El Niño conditions. They occur about 6%, 6%, 34%, 28%, and 26% of the time,  
279 respectively. The associated time series and the derived composite maps are presented in  
280 figure 4. Occurrences of EP El Niño (1997-1998), CP El Niño (2002-2003, 2004-2005, 2006-  
281 2007, 2009-2010), and La Niña events (1998-1999, 1999-2000, 2000-2001, 2007-2008, 2008-  
282 2009, 2010-2011) are consistent with those found in previous studies using SST or SSS  
283 indices (Table 1) and with the comparison of chlorophyll anomalies in the Niño3 and Niño4  
284 regions (Fig. 3). Of note, the cluster analysis does not classify the end of 1998 as typical La  
285 Nina conditions although that period has been described as a La Niña year from the surface  
286 chlorophyll distribution [Murtugudde et al., 1999; Radenac et al., 2001; Ryan et al., 2002] and  
287 from other variables (Fig. 3; Table 1): the reason for this is discussed below. Intermediate CP  
288 El Niño periods occur in 2001-2002, 2003-2004, 2005-2006, and 2007 near CP El Niño years  
289 when the SOI or EMI is weak (Fig. 3).

290

291 Aside from the “all-month” AHC analysis we present, we also performed an AHC analysis  
292 using September to February months only (not shown) because CP and EP El Niño and La  
293 Niña events reach their mature phase in boreal winter for most climate variables [Kao and Yu,

294 2009] and for chlorophyll changes [Chavez et al., 1999; Strutton and Chavez, 2000; Radenac  
295 et al., 2001; see also Figure 2]. In that case, we obtained similar patterns for the CP and EP El  
296 Niño, and La Niña clusters. The strong equatorial signal (Fig. 4c) however did not show up as  
297 elevated chlorophyll concentration along the equator was observed in boreal spring and  
298 summer. Caution is thus required when selecting some months per year only in analyzing  
299 ENSO features.

300

301 Cluster 1 captures the spatial pattern of the only EP El Niño event over the study period (Fig.  
302 4b). Given the strength of this event, it is not surprising that its spatial pattern resembles that  
303 of the EOF analysis performed on the 1997-1999 chlorophyll data [Wilson and Adamec,  
304 2001]. The chlorophyll anomaly is negative over a broad region from 170°E to the American  
305 coast and maximum along the equator. The northern limit of the negative anomaly region is  
306 sharp and almost zonally oriented near 8°N while the southern limit may look smoother and  
307 extends south of 10°S (except west of the dateline). Unlike the SST anomaly pattern during  
308 EP El Niño, which is strongest near the American coast and extends westward along the  
309 equator [Rasmusson and Carpenter, 1982; Kao and Yu, 2009; Kug et al., 2009], the  
310 chlorophyll anomaly seems to have two distinct cores (lower than  $-0.075 \text{ mg m}^{-3}$ ). One core is  
311 found between the date line and 140°W and the other one east of 110°W in the equatorial  
312 region. The chlorophyll anomaly is positive in the western basin mostly between the equator  
313 and 10°latitudes.

314

315 Characteristics of surface chlorophyll during the four CP El Niño events are represented by  
316 cluster 5 (Fig. 4f). The negative anomaly pattern has an arrow shape whose head would be the  
317 core of the largest negative anomaly (less than  $-0.08 \text{ mg m}^{-3}$ ) located at the equator around  
318 170°E. Thin bands of negative chlorophyll anomaly ( $-0.02$  to  $-0.01 \text{ mg m}^{-3}$ ) stretches from the

319 anomaly core along 8-10°N to the Central American coast, while another of lower magnitude  
320 extends to the Marquesas Islands at 140°W, 10°S. The tail of the arrow would be the narrow  
321 band of moderate negative anomaly (-0.03 to -0.02 mg m<sup>-3</sup>) along the equator from the  
322 anomaly core to the American coast. The core of maximum negative surface chlorophyll  
323 anomaly matches the core of maximum negative SSS anomaly [Singh et al., 2011] during  
324 both El Niño types, and in the same way as the SST anomaly, it is shifted westwards during  
325 CP El Niño compared to EP El Niño. The three zonal bands of moderate negative chlorophyll  
326 anomaly that stretch eastward from the equatorial minimum anomaly core are specific to the  
327 chlorophyll signature. In the western Pacific, north of Papua New Guinea and Solomon  
328 Islands, chlorophyll anomalies are positive during CP El Niño events.

329

330 The timing of La Niña years, captured by cluster 3 (Fig. 4a), are consistent with the  
331 occurrence of positive SOI (Fig. 3). The La Nina patterns (Fig. 4d) somewhat mirror the CP  
332 El Nino patterns (Fig. 4f) in the western half of the basin, although the maximum anomaly is  
333 stronger and extends 5° to 10° longitude further to the west during La Niña. In agreement  
334 with figure 3, cluster 3 thus shows that the positive chlorophyll anomalies in Niño4 region are  
335 larger than in Niño3 (Fig. 4d). Narrow bands of positive anomaly stretch from the high  
336 positive anomaly core along the equator and 6°N -7°N toward the American coast and to the  
337 Marquesas Islands. Cluster 2 captures a strong enhancement of the surface chlorophyll  
338 concentration (> 0.05 mg m<sup>-3</sup>) in a narrow equatorial band by the end of 1998 and mid-2010  
339 (Fig. 4a, c) during the strong La Nina events that followed the strong 1997-1998 EP El Niño  
340 and 2009-2010 CP El Niño. Two cores with chlorophyll anomaly higher than 0.08 mg m<sup>-3</sup> are  
341 located west of the Kiribati Islands (175°E) and between 140°W and 120°W. Anomalies are  
342 negative between 10°S and 10°N in the western basin and poleward of 5° in the east.

343

344 Surface chlorophyll distribution captured by cluster 4 (Fig. 4a, e) appears when the SOI and  
345 chlorophyll anomalies in the Niño3 and Niño4 regions are weak (Fig. 2). The main feature is  
346 a negative anomaly zone in the western equatorial basin.

347

## 348 **5. Analyzing possible drivers of surface chlorophyll changes during El Niño events**

349 In this section, we investigate mechanisms that could possibly constrain the overall spatial  
350 structure of surface chlorophyll anomaly during El Niño events, relying on changes in surface  
351 zonal current, wind, and thermocline depth (assumed to be the 20°C isotherm depth,  $Z_{20^{\circ}\text{C}}$ ).  
352 We first recall the main results obtained for the 1997-1998 EP El Niño [Chavez et al., 1999;  
353 Murtugudde et al., 1999; Stoens et al., 1999; Wilson and Adamec, 2001; Strutton and Chavez,  
354 2000; Radenac et al., 2001; 2005]. Then, we choose to describe the spatial evolution of  
355 chlorophyll anomaly during CP El Niño, with a slight focus on the fairly representative 2002-  
356 2003 event, rather than on a composite, in order to preserve tiny structures whose positions  
357 are slightly different among events. Figure 5 shows the time evolution of the anomalies of  
358 zonal wind speed,  $Z_{20^{\circ}\text{C}}$ , and surface zonal current along the equator from September 1997 to  
359 December 2010. The limits of the oligotrophic region are indicated by the black contour in  
360 each panel. Anomalies of the surface current are superimposed on the surface chlorophyll  
361 anomalies during the peak period (September-February) of the 1997-1998 EP El Niño (Fig. 6)  
362 and 2002-2003 CP El Niño (Fig. 7).

363

### 364 **5.1. The 1997-1998 EP El Niño event**

365 During boreal fall of 1997, strong anomalous eastward currents dominate the equatorial  
366 Pacific (Fig. 5c; Fig. 6a-d) as a consequence of anomalous westerly winds (Fig. 5a) and  
367 forced downwelling Kelvin waves [Delcroix et al., 2000]. The South Equatorial Current  
368 (SEC) is weak or reversed and the NECC is stronger and shifted southwards [Johnson et al.,



369 2000]. As a result, nutrient-poor warm pool waters are advected eastward and reach 130°W  
370 during the mature phase in November 1997-January 1998 [Radenac et al., 2001]. The core of  
371 high negative chlorophyll anomaly is in the eastern part of the warm pool (Fig. 2c, 4b) where  
372 eastward surface currents are strong (Fig. 5c). Meanwhile, vertical supplies of nitrate and iron  
373 decrease or cease because of the reduction of the upwelling and the deepening of the  
374 nitracline and EUC [Fig. 5b; Chavez et al., 1999; Strutton and Chavez, 2000; Wilson and  
375 Adamec, 2001; Radenac et al., 2001; 2005]. A combination of these processes lead to a  
376 dramatic decrease of the surface chlorophyll and of the biological production in the central  
377 and eastern basins. The northern and southern limits of the negative chlorophyll anomaly  
378 closely match the extent of the eastward surface current anomaly until the peak period in  
379 December 1997 (Fig. 6a-d). Starting in January 1998, the surface chlorophyll anomaly  
380 weakens as strong westward surface currents begin to develop (Fig. 6d-e).

381

382 In the warm pool, the overall chlorophyll increase (Fig. 6a-f) has been ascribed to the  
383 thermocline rise (Fig. 5b) allowing nutrient input toward the lighted layer [Wilson and  
384 Adamec, 2001; Radenac et al., 2001; Turk et al., 2001]. Part of it, between 5°N and 10°N,  
385 results from enhanced chlorophyll concentration in the meanders of the nascent NECC  
386 observed seasonally between January and June and that were strong at the end of the 1997-  
387 1998 El Niño [Christian et al., 2004; Messié and Radenac, 2006]. Upwelling that develops  
388 along the coast of Papua New Guinea and Solomon Islands during favorable periods of  
389 westerly wind [Webster and Lukas, 1992; Kuroda, 2000; Ueki et al., 2003] also contributes to  
390 the chlorophyll increase in the warm pool south of the equator [Messié, 2006; Radenac et al.,  
391 2010].

392

393 **5.2. The CP El Niño events**

394 During CP El Niño events, the eastward shift of the warm pool (see the black contours in  
395 figure 5a) in response to westerly winds is of lesser magnitude than during EP El Niño. The  
396 associated equatorial eastward surface currents contribute to the development of negative  
397 chlorophyll anomaly as a consequence of the penetration of the oligotrophic waters of the  
398 warm pool toward the mesotrophic waters of the upwelling (Fig. 5c). The core of high  
399 chlorophyll anomaly remains within the zone of quasi-persistent oligotrophic waters enclosed  
400 by the  $0.1 \text{ mg m}^{-3}$  surface chlorophyll isoline in the eastern part of the warm pool (Fig. 2c),  
401 near the zone of convergence between the eastward surface current in the warm pool and the  
402 westward SEC east of it (Fig. 5c). The importance of this current convergence has been raised  
403 by Kug et al. [2009] and Singh et al. [2011] to explain how SST and SSS anomaly cores could  
404 remain confined in the central basin during CP El Niño.

405

406 During the 2002-2003 CP El Niño, the maximum negative chlorophyll anomaly is situated  
407 around  $170^{\circ}\text{E}$  in June,  $180^{\circ}\text{E}$  in August, and it stays at  $170^{\circ}\text{W}$  between September and  
408 December 2002 (Fig. 2c, 7a-d). It starts to move back to the west in January 2003 when  
409 strong westward anomalies develop. The monthly maps of surface chlorophyll and current  
410 (Fig. 7a-c) also suggest that anomalous eastward currents partly contribute to the equatorial  
411 chlorophyll decrease towards the American coast.

412

413 As discussed above for CP El Niño, negative anomalies (weaker than  $-0.015 \text{ mg m}^{-3}$ ) are  
414 constrained within  $2^{\circ}$  of the equator east of the high chlorophyll anomaly core (Fig. 4f). The  
415 zonal wind anomaly is slightly westward in the eastern basin during the CP El Niño years  
416 (Fig. 5a) and so cannot be responsible for a collapse of the upwelling in this region. Besides,  
417 variations of the vertical supply of nitrate, depending on the nutrient pool depth represented  
418 by the  $20^{\circ}\text{C}$  isotherm depth ( $Z_{20^{\circ}\text{C}}$ ), will not impact the phytoplankton growth as nitrate is

419 usually in excess in this region. The equatorial chlorophyll decrease would rather be the result  
420 of vertical displacements of the EUC that contribute modulating vertical iron fluxes and the  
421 phytoplankton growth in the iron limited ecosystem of the central and eastern equatorial  
422 Pacific. The temporal variability of the  $Z_{20^{\circ}\text{C}}$  anomaly can be used as an indicator of the EUC  
423 depth ( $Z_{\text{EUC}}$ ) anomaly (Izumo, 2005). When ADCP currents are available at the equatorial  
424 mooring sites, the bias between  $Z_{\text{EUC}}$  (calculated as the depth of the maximum eastward  
425 velocity) and  $Z_{20^{\circ}\text{C}}$  is lower than 4 m at  $170^{\circ}\text{W}$  and  $140^{\circ}\text{W}$  and the correlation coefficient is  
426 0.72 at both sites. Therefore, the variations of  $Z_{20^{\circ}\text{C}}$  reliably represent the variations of  $Z_{\text{EUC}}$   
427 that deepens by 20-40 m during CP El Niño (Fig. 5b) and could drive a significant decrease of  
428 the vertical iron flux [Chavez et al., 1999]. A strip of maximum anomaly confined to the  
429 equator is consistent with the structure of meridional velocity and a narrow band of strong  
430 divergence roughly centered along the equator [Poullain, 1993]. This further suggests that  
431 biology along the equator is sensitive to the EUC depth and iron concentration.

432

433 Off the equator, the narrow zonal band of negative chlorophyll anomaly between  $5^{\circ}\text{N}$  and  
434  $10^{\circ}\text{N}$  is observed during the four CP El Niño events. During the 2002-2003 CP El Niño event,  
435 it appears during August-September (Fig. 7a) and starts to break up during January-February  
436 (Fig. 7e). The negative chlorophyll anomaly strip follows the position of the eastward surface  
437 current anomaly (Fig. 7a-d), suggesting that nutrient- and phytoplankton-poor water from the  
438 warm pool is advected eastward by the strong NECC. Messié and Radenac [2006] showed the  
439 significant impact of the NECC variations on the surface chlorophyll seasonal variability in  
440 the western Pacific. During El Niño events, it is a basin-wide feature that participates in  
441 setting up the sharp chlorophyll front at the northern edge of the upwelling. In monthly  
442 chlorophyll maps, it is seen as a narrow band of oligotrophic waters extending from the warm  
443 pool to the Central American coast during October-December of all El Niño years (not

444 shown). South of the equator, negative chlorophyll anomalies coincide with a region of  
445 southeastward surface currents that transport nutrient- and phytoplankton-poor waters toward  
446 the Marquesas Islands (Fig. 7a-f). This feature is observed during September-February 2002-  
447 2003, 2004-2005, and 2006-2007. It develops later during February-April during the 2009-  
448 2010 CP El Niño event (not shown).

449

## 450 **6. Summary and discussion**

451 Enhanced attention has been given to global climate changes related to differences in the  
452 location of SST anomalous warming in the tropical Pacific in recent years [Trenberth and  
453 Stepaniak, 2001; Ashok et al., 2007; Kug et al., 2009; Kao and Yu, 2009; Yeh et al., 2009;  
454 Lee and McPhaden, 2010]. To our knowledge, no attempt has been made so far to contrast the  
455 biological conditions featuring the EP and CP El Niño, except for the recent study by Turk et  
456 al. [2011]. To fulfill this gap, the goal of our study was to contrast and tentatively explain the  
457 signature in chlorophyll of the EP and CP ENSO, for the 1997-2010 period.

458

459 During the SeaWiFS years, an AHC analysis of the surface chlorophyll anomaly shows that  
460 EP El Niño occurred in 1997-1998, CP El Niño in 2002-2003, 2004-2005, 2006-2007, and  
461 2009-2010, while La Niña lasted from 1998 to 2001 and from 2007 to 2009, consistent with  
462 previous studies based on SST and SSS analyses (Table 1). Both types of El Niño events are  
463 associated with an overall decrease of surface chlorophyll in the 10°S-10°N region. Yet,  
464 spatial patterns differ between events.

465

466 During the 1997-1998 EP El Niño event, negative anomalies occupy most of the equatorial  
467 basin between 10°S and 10°N, except for a chlorophyll increase in the western basin. A strong

468 negative chlorophyll anomaly ( $< -0.075 \text{ mg m}^{-3}$ ) is located at the equator between the date  
469 line and  $150^\circ\text{W}$  and the oligotrophic warm pool is displaced eastward and reaches  $130^\circ\text{W}$ .  
470

471 During the CP El Niño events, the equatorial anomaly is weaker ( $< -0.045 \text{ mg m}^{-3}$ ), shifted  
472 westwards by about  $20^\circ$  of longitude, and the eastern edge of the oligotrophic warm pool is  
473 confined west of  $160^\circ\text{W}$ . So, as already found for SST and SSS [Kug et al., 2009; Singh et al.,  
474 2011], the region of strong chlorophyll anomaly is clearly shifted westward during CP El  
475 Niño relative to EP El Niño. The location of these strong negative anomalies in the eastern  
476 part of the warm pool suggests that oligotrophic waters of the warm pool replace mesotrophic  
477 waters of the upwelling in the central basin, and that zonal advection is a major process  
478 responsible for this signature. During CP El Niño, westward surface currents in the eastern  
479 and central basins probably limit the eastward spreading of the negative chlorophyll anomaly  
480 core.

481

482 Other mechanisms affecting nutrient or light availability could also lead to a chlorophyll  
483 decrease in the central equatorial basin. One of them is the depth of the nutrient pool. The  
484 core of high chlorophyll anomaly matches the maximum sea level anomaly [Kug et al., 2009;  
485 Bosc et al., 2009] which reflects a strong depression of the thermocline [Rébert et al., 1985],  
486 closely related to the depth of the nitracline in the oligotrophic warm pool. Besides, barrier  
487 layer tends to be thick in the eastern edge of the warm pool [Maes et al., 2004; Bosc et al.,  
488 2009] and to further isolate the deep nutrient pool from the lighted surface layer [Mackey et  
489 al., 1995; Eldin et al., 1997; Murtugudde et al., 1999; Turk et al., 2001]. The nutrient pool is  
490 deep on average in the oligotrophic warm pool and the depression of the nitracline in the  
491 eastern part of the warm pool is probably a second order factor on surface chlorophyll  
492 changes compared to the impact of advection.

493

494 Another factor accounting for chlorophyll changes during EP and CP El Niño could be the  
495 reduction of incident light in the central Pacific as suggested by Park et al. [2011]. Zonal  
496 extension of oligotrophic conditions is linked to the zonal extension of the warm pool, which  
497 is the site of convective activity and moves eastward during El Niño. The average monthly  
498 value of the satellite derived Photosynthetically Available Radiation (PAR) is  $44.7 \pm$   
499  $4.4 \text{ Einstein m}^{-2} \text{ d}^{-1}$  in the eastern part of the warm pool while it is  $48.6 \pm 2.2 \text{ Einstein m}^{-2} \text{ d}^{-1}$   
500 in the upwelling region at  $140^\circ\text{W}$ . The eastward shift of the convection zone leads to PAR  
501 anomaly around  $10 \text{ Einstein m}^{-2} \text{ d}^{-1}$  in the strong chlorophyll anomaly region in 1997-1998  
502 and close to or less than  $5 \text{ Einstein m}^{-2} \text{ d}^{-1}$  during the following El Niño events. These  
503 relatively weak PAR variations, three to ten times weaker than variations at mid and high  
504 latitudes [Letelier et al., 2004; Goes et al., 2004], suggest a weak impact on the phytoplankton  
505 growth. Therefore, zonal advection and the resulting change of ecosystem in the central  
506 Pacific are probably the dominant mechanisms responsible for the strong chlorophyll  
507 anomaly, as also proposed by Messié and Chavez [submitted to J. Geophys. Res., 2012].  
508 Mesotrophic waters of the upwelling with surface chlorophyll concentration around  
509  $0.2 \text{ mg m}^{-3}$  are replaced by oligotrophic waters of the eastern part of the warm pool with  
510 surface chlorophyll concentration lower than  $0.07 \text{ mg m}^{-3}$  [Radenac et al., 2010]. However,  
511 separating the impact of the ecosystem change and of the light attenuation is not simple as  
512 oligotrophic waters move simultaneously to the convection zone, and a process study is  
513 needed to estimate each influence.

514

515 We suggest that the chlorophyll decrease along the equator east of the strong central Pacific  
516 anomaly is the consequence of reduced vertical iron fluxes linked to the deepening of the  
517 EUC during El Niño [Barber et al., 1996; Chavez et al., 1999; Wilson and Adamec, 2001;

518 Friedrichs and Hofmann, 2001]. Changes in the iron content of the EUC could also impact  
519 biology in the central and eastern Pacific. Ryan et al. [2006] hypothesized that the  
520 strengthening of the New Guinea Coastal Undercurrent (NGCUC), flowing northwestward  
521 along the northern coast of Papua New Guinea, favored transport of iron from shelf sediments  
522 that feeds the EUC during El Niño years, which in turn favored phytoplankton blooms  
523 observed in the central Pacific following El Niño events. In contrast, setting the iron  
524 concentration proportional to the NGCUC speed in the source region did not change the  
525 intensity of eastern equatorial Pacific blooms in a simulation based study [Gorgues et al.,  
526 2010]. Thus, although the impact of a continental iron source in the western equatorial Pacific  
527 on the biogeochemistry of the equatorial upwelling region has been shown in several  
528 modeling studies [Christian et al., 2002; Vichi et al., 2008; Slemmons et al., 2009; Gorgues et  
529 al., 2010], the influence of a variable iron supply into the EUC on the ecosystems of the  
530 central and eastern basins is still unclear. So is the role of tropical instability waves (TIW).  
531 Because of vigorous horizontal processes as well as upwelling and downwelling, localized  
532 maximum and minimum of phytoplankton biomass characterize TIW [Yoder et al., 1994;  
533 Strutton et al., 2001; Menkes et al., 2002]. Their net impact on phytoplankton biomass has  
534 been estimated to be positive [Barber et al., 1996; Friedrichs and Hofmann, 2001; Strutton et  
535 al., 2001] or negative [Gorgues et al., 2005]. Other studies show an enhancement or decrease  
536 of the phytoplankton biomass depending on interactions between the large scale  
537 thermocline/ferricline depth and the intensity of the local TIW dynamics [Vichi et al., 2008;  
538 Evans et al., 2009]. The reduction or absence of TIW activity during El Niño [Legeckis, 1977;  
539 Friedrichs and Hofmann, 2001; Evans et al., 2009] can also possibly contribute to the  
540 chlorophyll decrease observed along the equator.

541

542 The AHC analysis of the 1997-2010 surface chlorophyll anomaly dataset does not separate La  
543 Niña events into EP and CP events. Chlorophyll increases in the equatorial western basin and  
544 near 8-10°N and toward the Marquesas Islands are common patterns to La Niña maps.  
545 Nevertheless, the location of the equatorial core of positive anomalies and its magnitude  
546 change (much stronger in 2010 than during other events) and an equatorial increase in the  
547 central and eastern basins may not be observed. The average location of chlorophyll anomaly  
548 of the six La Niña events seen by SeaWiFS is situated west of the EP and CP El Niño  
549 anomalies. This is consistent with the La Niña signature as reported by Kug et al. [2009] and  
550 Kug and Ham [2011] who do not separate CP and EP cold events in terms of SST anomaly.  
551 Besides, the chlorophyll anomaly analysis does not evidence any symmetry between the  
552 warm and cold phases of ENSO events as suggested in Kao and Yu [2009] and Yu et al.  
553 [2010] analyses. The distinct EP and CP La Niña patterns mentioned by Singh et al. [2011]  
554 are based on a cluster analysis of SSS that captures a signal in the SPCZ region that does not  
555 show up in the chlorophyll analysis.

556

557 During La Niña years, the equatorial core of positive anomaly results from the westward  
558 expansion of the upwelling mesotrophic waters in a region where the usual surface  
559 chlorophyll concentration does not exceed  $0.1 \text{ mg m}^{-3}$  (Fig. 2c). The weak nutrient supply to  
560 the euphotic layer in the Niño4 region results from horizontal advection of nutrient-rich  
561 waters from the east and upward advection [Stoens et al., 1999]. During La Niña events,  
562 observational and modeling studies have shown that upwelling led to increased surface  
563 chlorophyll concentration in the western Pacific [Blanchot et al., 1992; Radenac and Rodier,  
564 1996; Radenac et al., 2001]. East of the strong anomaly core, the iron limitation and the  
565 grazing pressure (small phytoplankton and zooplankton species dominate the ecosystem and  
566 microzooplankton can quickly respond to changes in nano- and pico-phytoplankton biomass)



567 control the ecosystem resulting in a monotonously low phytoplankton biomass in the  
568 equatorial divergence [Chavez et al., 1991; Le Bouteiller and Blanchot, 1991; Strutton et al.,  
569 2008]. This special feature of the equatorial upwelling ecosystem could explain why no strong  
570 chlorophyll increase characterizes the overall La Niña distribution of chlorophyll anomaly  
571 along the equator east of the anomaly core (Fig. 4d). An asymmetry between El Niño and La  
572 Niña has also been observed in terms of temperature (Larkin and Harrison, 2002; An and Jin,  
573 2004). The warm phase of ENSO is often stronger than its cold phase. Nonlinear dynamical  
574 processes could impact nutrient (iron) supply and cause an asymmetry of the ENSO-related  
575 biological response as they impact the ENSO-related heat budget (An and Jin, 2004). More  
576 investigations remain to be done on this issue, especially during the CP El Niño events that  
577 dominate during the recent years. Cluster 2 in the AHC analysis (Fig. 4a, c) captured  
578 exceptions to this uniformity. 80% of the maps that compose this cluster come from the 1998  
579 La Niña year when unusual large-scale blooms [Ryan et al., 2002] were observed in the  
580 equatorial Pacific after the major 1997-1998 EP El Niño event. Chlorophyll increases  
581 subsequent to island mass effect generated by the Kiribati Islands, which behave as an  
582 obstacle to both the SEC and the EUC between February and June 1998 [Ryan et al., 2002;  
583 Messié et al., 2006] may contribute to the positive anomaly observed west of the dateline. A  
584 second bloom, the longest and more intense one with chlorophyll concentration higher than  
585  $0.8 \text{ mg m}^{-3}$ , developed between  $160^\circ\text{W}$  and  $140^\circ\text{W}$  and then spread eastward from June to  
586 September [Chavez et al., 1999; Ryan et al., 2002; Gorgues et al., 2010]. The last bloom was  
587 observed around  $130^\circ\text{W}$  during November-December 1998 [Ryan et al., 2002]. Reasons for  
588 these blooms can be both large-scale and local dynamics, such as, enhanced iron vertical  
589 fluxes because of a shallower thermocline and more active TIW during La Niña. A similar but  
590 somewhat weaker situation seemed to occur in mid-2010. Interestingly, these equatorial  
591 recovery conditions occur when strong El Niño events turn swiftly into strong La Niña: from

592 the major 1997-1998 EP El Niño to the strong 1998-1999 La Niña [Chavez et al., 1999; Ryan  
593 et al., 2002; Radenac et al., 2010] and from the strong 2009-2010 CP El Niño to the strong  
594 2010-2011 La Niña [Kim et al., 2011]. These periods coincide with phases of reduced grazing  
595 pressure as a response to El Niño related reduced phytoplankton biomass and productivity.  
596 Therefore, momentarily low grazing pressure probably combines to dynamical impacts  
597 leading to enhanced phytoplankton growth to drive high phytoplankton biomass during these  
598 recovery periods after strong El Niño events [Strutton and Chavez, 2000; Friedrichs and  
599 Hofmann, 2001; Gorgues et al., 2010].

600

601 While increased eastward advection was responsible for the chlorophyll decreases along 8-  
602 10°N and towards the Marquesas Islands during CP El Niño, the chlorophyll increases at the  
603 same locations during La Niña coincide with increased westward and poleward surface  
604 currents suggesting the influence of horizontal advection.

605

606 Situations that occurred in 2001-2002, 2003-2004, 2005-2006, and 2007 close to CP El Niño  
607 years (Figs. 4a, e) were called intermediate CP El Niño conditions. Their spatial structure  
608 (Fig. 4e) is characterized by an anomaly core of about  $-0.04 \text{ mg m}^{-3}$  located near 160°E. The  
609 persistent negative chlorophyll anomaly in the western Pacific could be specific to the period  
610 we study (recurring CP El Niño conditions) and this is why we do not refer to it as neutral  
611 conditions.

612

613 The SeaWiFS archive spans over more than 13 years during which 5 El Niño events occur.  
614 Interestingly, the AHC analysis we performed on the monthly surface chlorophyll anomaly  
615 maps separates the 1997-1998 EP episode from the other CP El Niño episodes. These results  
616 need to be refined with the help of longer observational time series, modeling outputs, and

617 theoretical work. An improve understanding of the phytoplankton distribution and its  
618 temporal variability is actually essential to anticipate biogeochemical climate-driven shifts  
619 and their consequences on ocean dynamics, carbon cycle, and marine ressources. Different  
620 phytoplankton distribution during CP and EP El Niño events could actually impact the  
621 distribution and abundance of exploited fish species such as tuna, whose fishery in the central  
622 and western tropical Pacific is one of the largest industrial fisheries of the world [Lehodey et  
623 al., 2011]. Complementing our qualitative analysis, we also need to quantify how physical  
624 and/or biological processes lead to the contrasted patterns we evidenced.

625

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639

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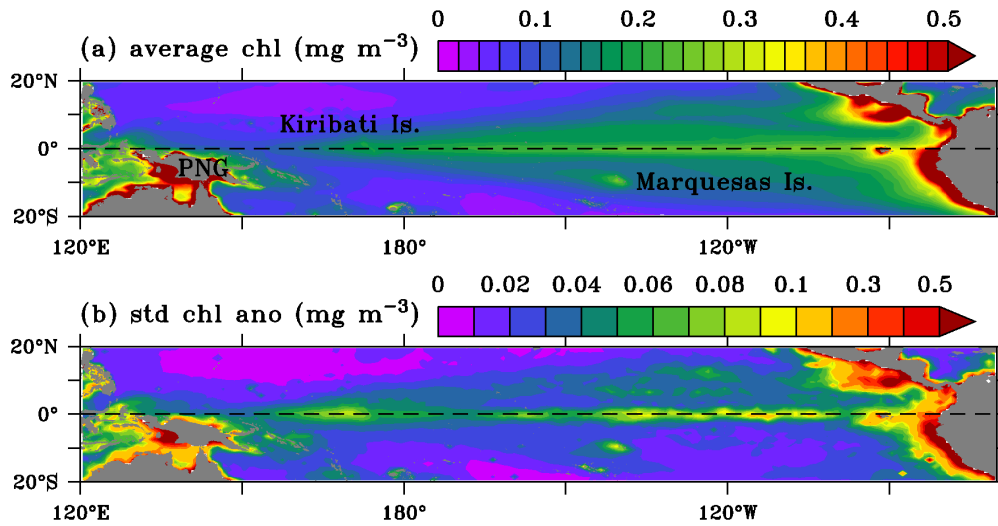
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896 **Table 1.** Classification of ENSO events as Central Pacific (CP) or Eastern Pacific (EP) El  
897 Niño (EN) or La Niña (LN). All studies used SST except (1) used SSS and (2) used  
898 chlorophyll to characterize ENSO.

|                                       | 97-<br>98 | 98-<br>99 | 99-<br>00 | 00-<br>01 | 02-<br>03 | 04-<br>05 | 05-<br>06 | 06-<br>07 | 07-<br>08 | 08-<br>09 | 09-<br>10 | 10 |
|---------------------------------------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|----|
| Ashok et al.<br>[2007]                |           |           |           |           | CP<br>EN  | CP<br>EN  |           |           |           |           |           |    |
| Kim et al. [2009]                     | EP<br>EN  | EP<br>LN  | EP<br>LN  |           | CP<br>EN  | CP<br>EN  |           |           |           |           |           |    |
| Yeh et al. [2009]                     | EP<br>EN  |           |           |           | CP<br>EN  | CP<br>EN  |           | EP<br>EN  |           |           |           |    |
| Kao and Yu<br>[2009]                  | EP<br>EN  | CP<br>LN  | CP<br>LN  |           | CP<br>EN  | CP<br>EN  | EP<br>LN  |           |           |           |           |    |
| Kug et al. [2009]                     | EP<br>EN  | LN        | LN        | LN        | CP<br>EN  | CP<br>EN  | LN        |           |           |           |           |    |
| Lee and<br>McPhaden [2010]            | EP<br>EN  | LN        |           |           | CP<br>EN  | CP<br>EN  |           | EP<br>EN  | LN        |           | CP<br>EN  |    |
| Ren and Jin<br>[2011]                 | EP<br>EN  | LN        | LN        |           | CP<br>EN  | CP<br>EN  |           | CP<br>EN  | LN        |           | CP<br>EN  |    |
| <sup>(1)</sup> Singh et al.<br>[2011] | EP<br>EN  | CP<br>LN  | EP<br>LN  | EP<br>LN  | CP<br>EN  | CP<br>EN  | EP<br>LN  | CP<br>EN  | EP<br>LN  | EP<br>LN  |           |    |
| McPhaden et al.<br>[2011]             | EP<br>EN  |           |           |           | CP<br>EN  | CP<br>EN  |           | EP<br>EN  |           |           | CP<br>EN  |    |
| <sup>(2)</sup> This study             | EP<br>EN  | LN        | LN        | LN        | CP<br>EN  | CP<br>EN  |           | CP<br>EN  | LN        | LN        | CP<br>EN  | LN |



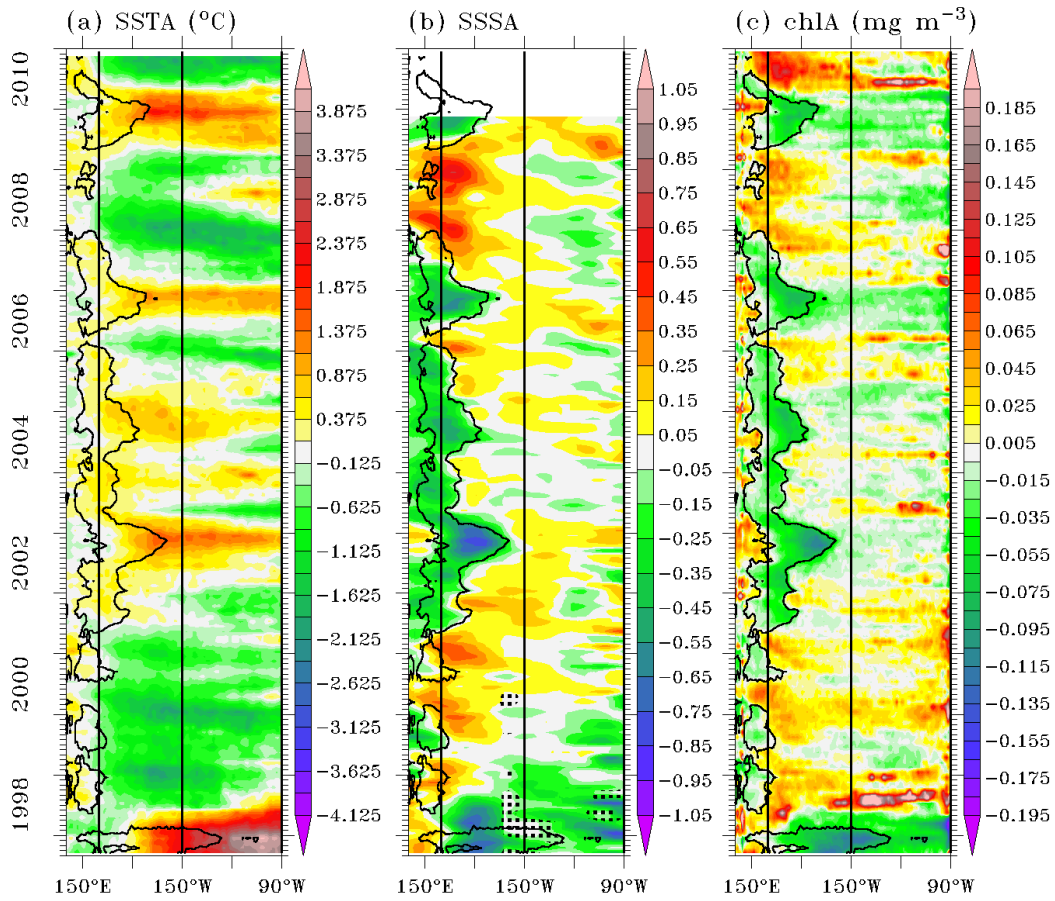




1 **Figure 1.** Maps of (a) average and (b) interannual standard deviation of the surface  
 2 chlorophyll computed over the September 1997 - December 2010 period. PNG stands for  
 3 Papua New Guinea.

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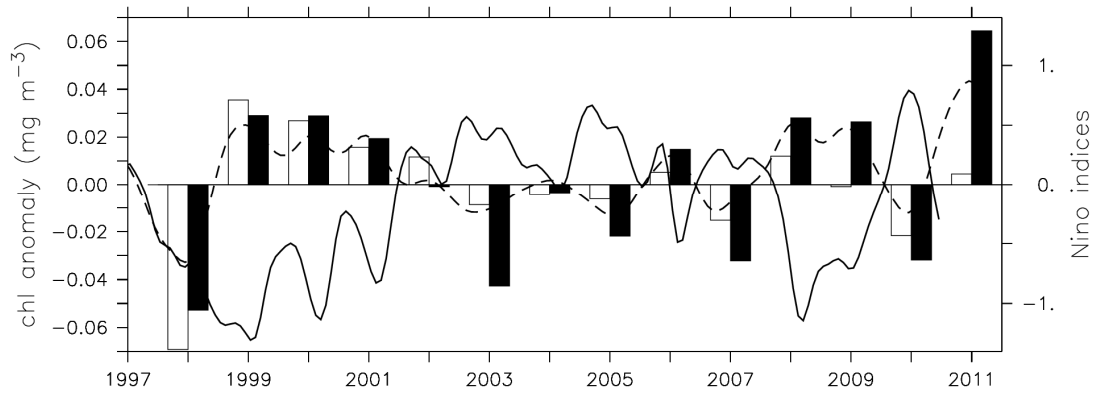
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6 **Figure 2.** Longitude-time distribution of the anomaly of (a) SST, (b) SSS, and (c) surface  
 7 chlorophyll averaged between 5°S and 5°N. The black contour encloses the region with  
 8 surface chlorophyll lower than 0.1 mg m<sup>-3</sup>. Vertical lines indicate the zonal boundaries of the  
 9 Niño3 (150°W, 90°W) and Niño4 (160°E, 150°W) regions. Regions with dots in (b) indicate  
 10 normalized SSS errors larger than 0.80.

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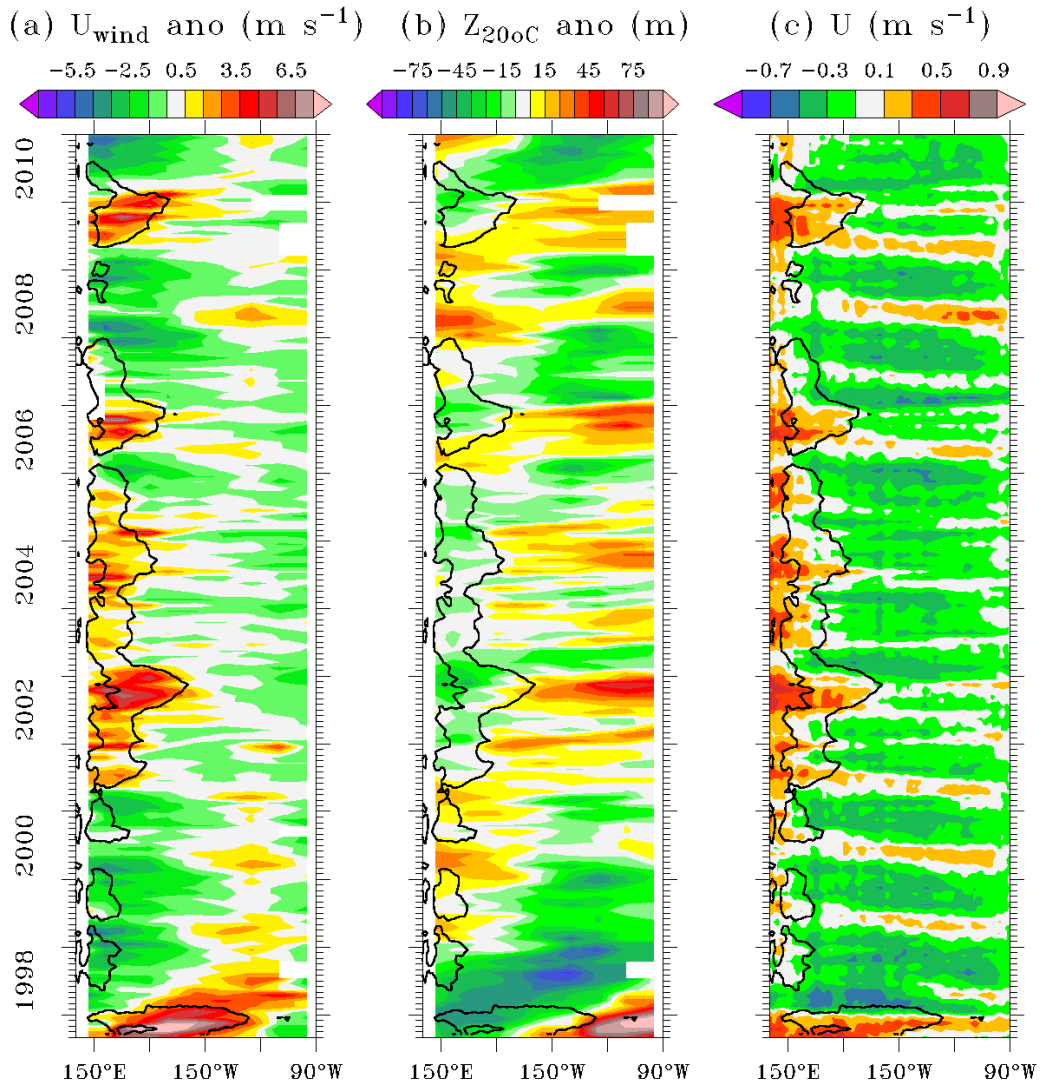


13 **Figure 3.** Surface chlorophyll anomaly averaged between September and February in the  
 14 Niño3 (150°W-90°W, 5°S-5°N; hollow bars) and Niño4 (160°E-150°W, 5°S-5°N; filled bars)  
 15 regions. EMI (full line) and 0.25×SOI (dashed line) are scaled on the right axis.

16

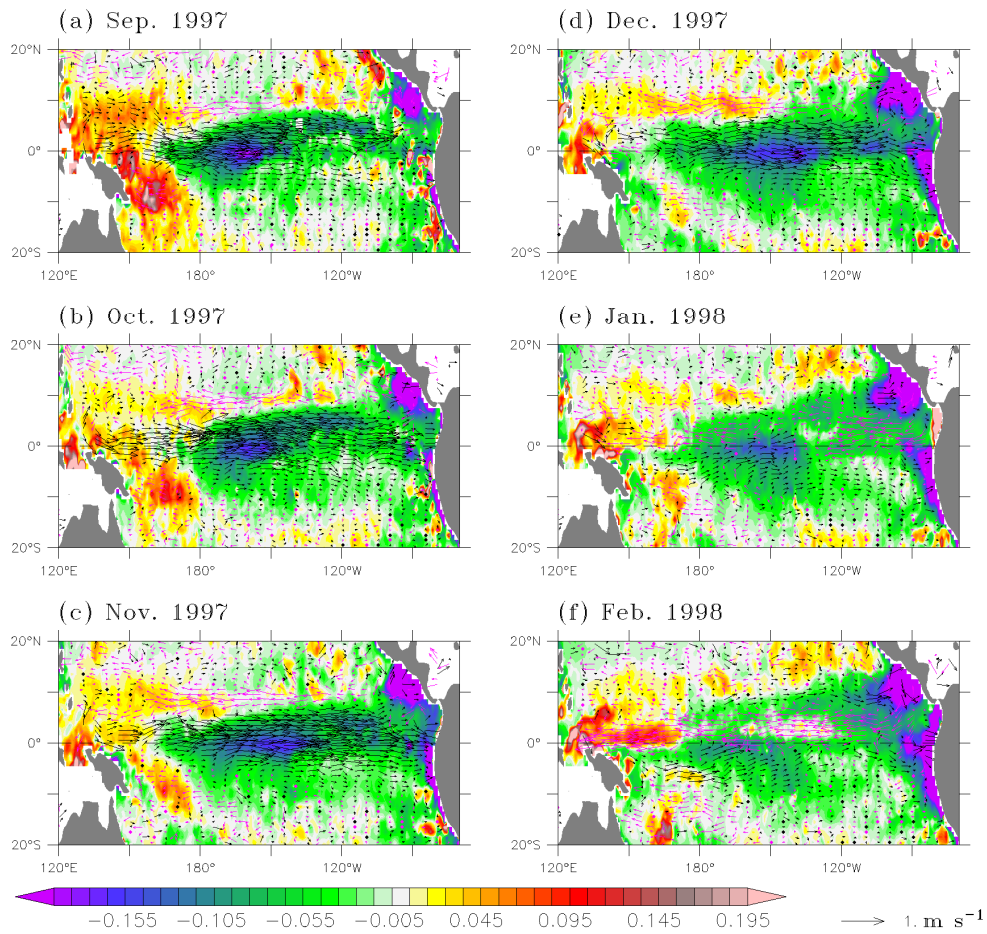
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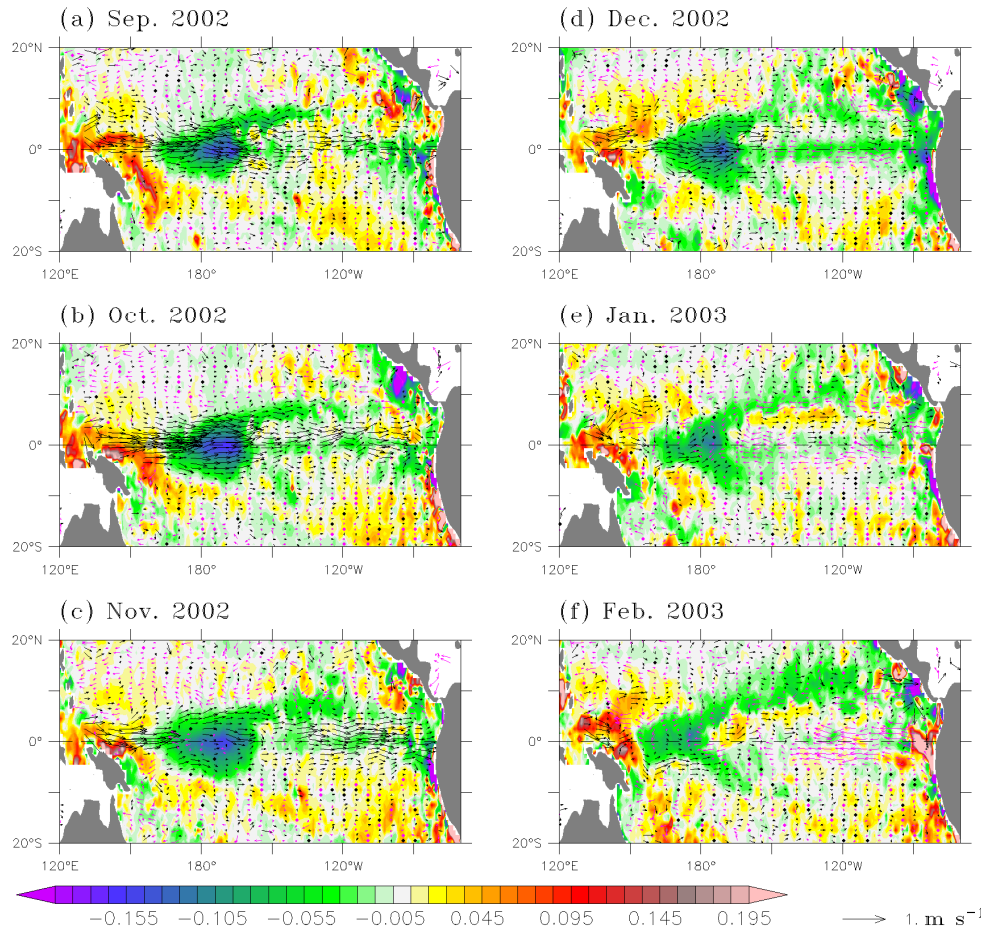
24 **Figure 5.** Longitude-time distribution along the equator of (a) zonal wind speed anomaly  
 25 (positive eastwards; colors;  $\text{m s}^{-1}$ ), (b)  $20^\circ\text{C}$  isotherm depth anomaly (positive downward; m),  
 26 and (c) zonal surface current (positive eastwards;  $\text{m s}^{-1}$ ). The black contours in (a-c) enclose  
 27 the region with surface chlorophyll lower than  $0.1 \text{ mg m}^{-3}$ .

28

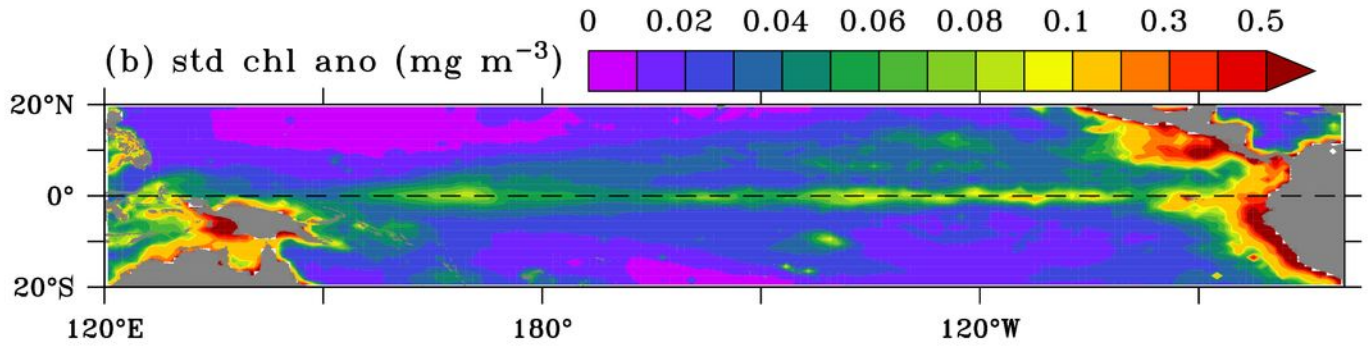
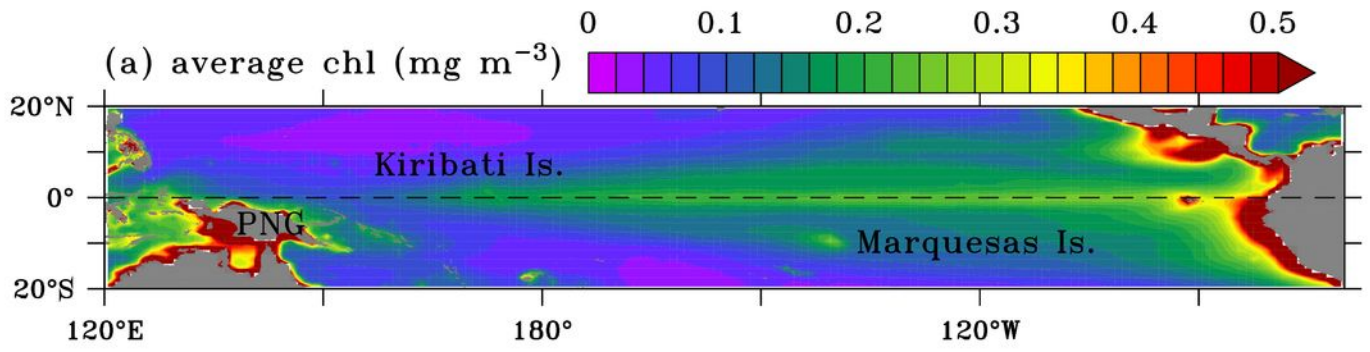


29 **Figure 6.** Maps of monthly surface chlorophyll anomaly (colors;  $\text{mg m}^{-3}$ ) and surface layer  
 30 current anomaly (vectors;  $\text{m s}^{-1}$ ) during the 1997-1998 Eastern Pacific El Niño. Current  
 31 vectors with eastward (westward) zonal component are in black (purple).

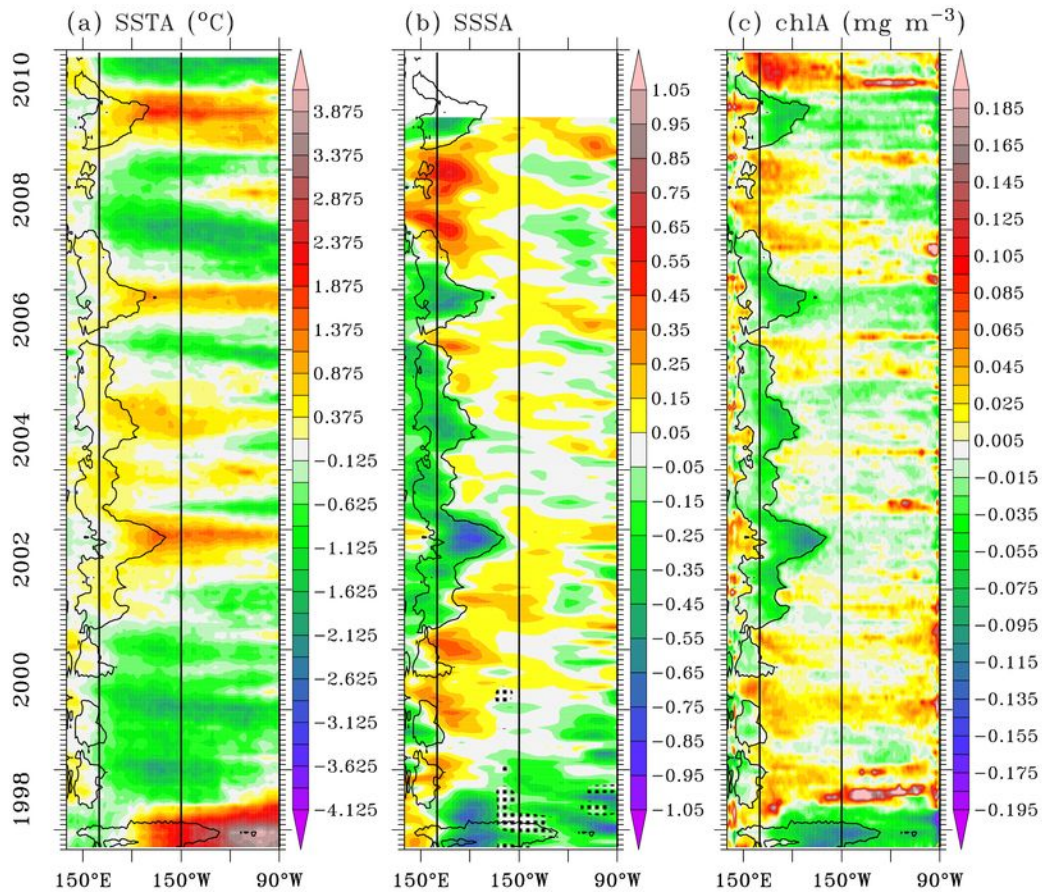
32

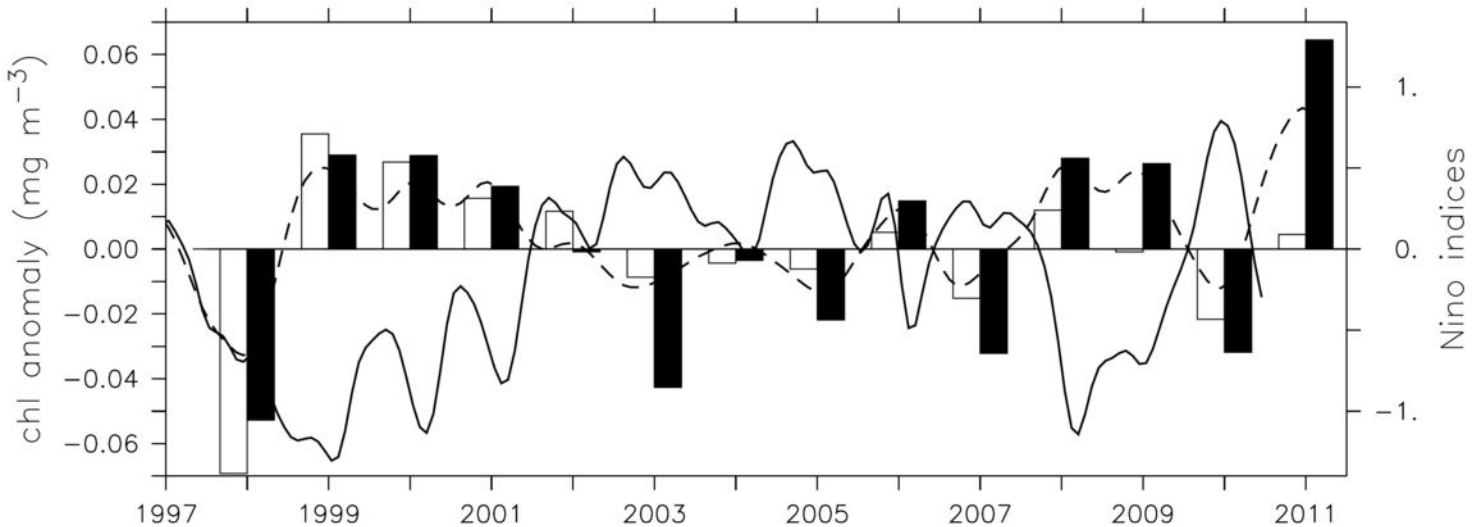


33 **Figure 7.** Maps of monthly surface chlorophyll anomaly (colors;  $\text{mg m}^{-3}$ ) and surface layer  
 34 current anomaly (vectors;  $\text{m s}^{-1}$ ) during the 2002-2003 Central Pacific El Niño. Current  
 35 vectors with eastward (westward) zonal component are in black (purple).  
 36



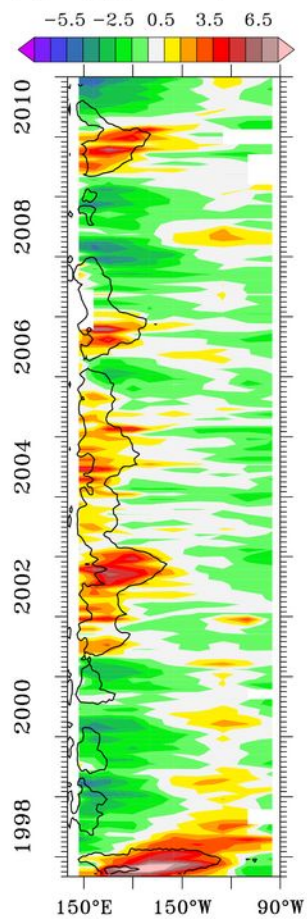




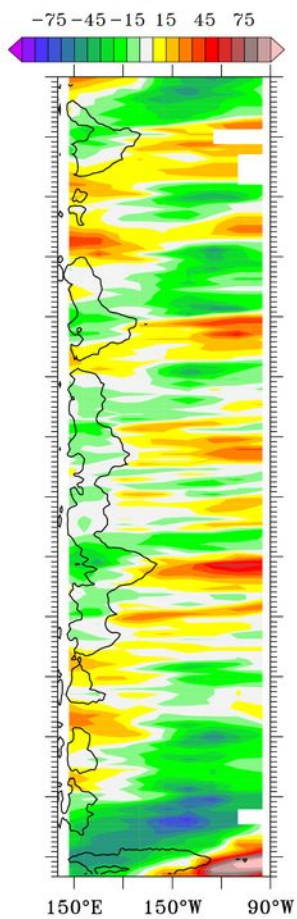




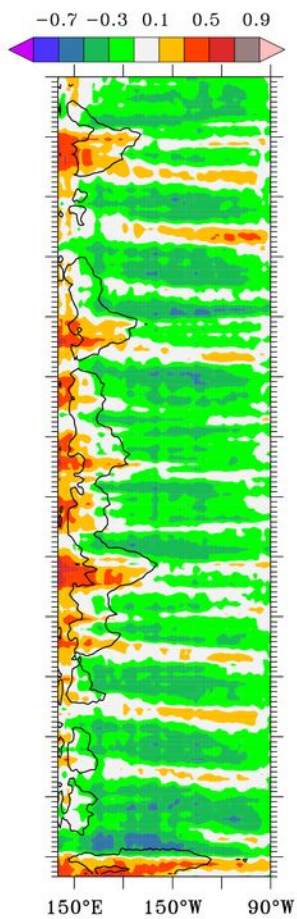
(a)  $U_{\text{wind}} \text{ ano (m s}^{-1}\text{)}$



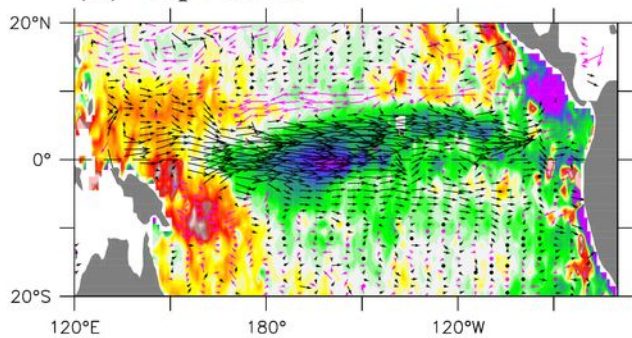
(b)  $Z_{200\text{C}} \text{ ano (m)}$



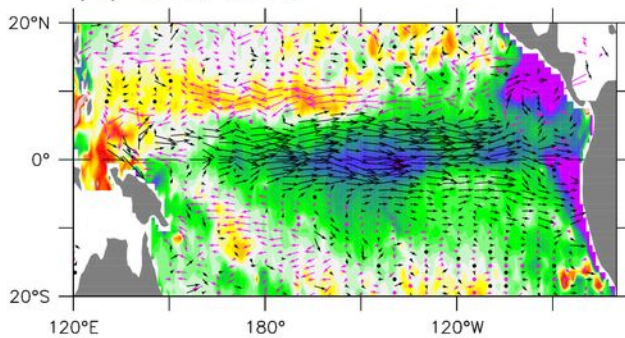
(c)  $U \text{ (m s}^{-1}\text{)}$



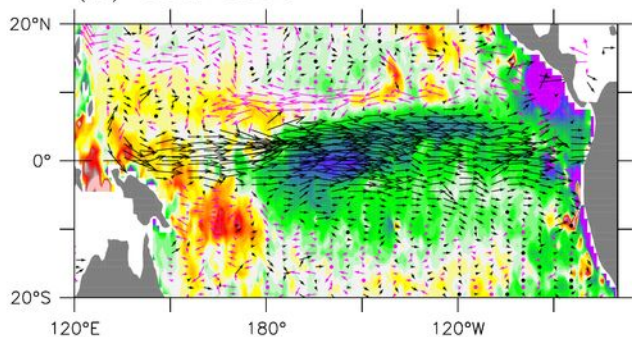
(a) Sep. 1997



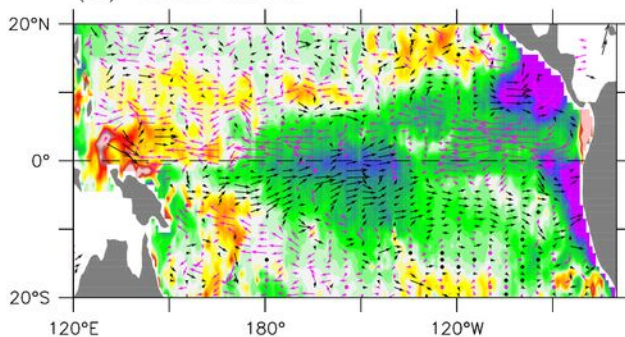
(d) Dec. 1997



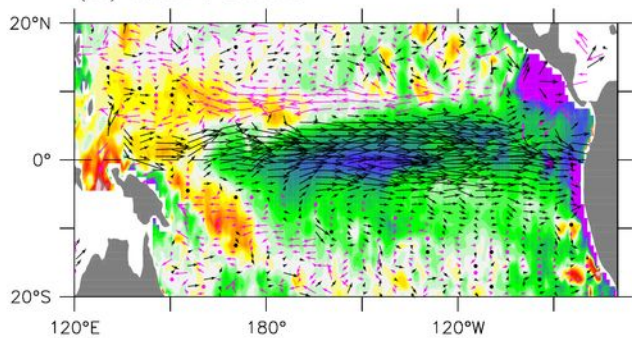
(b) Oct. 1997



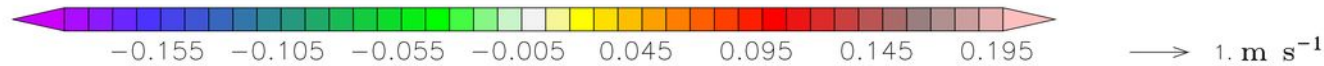
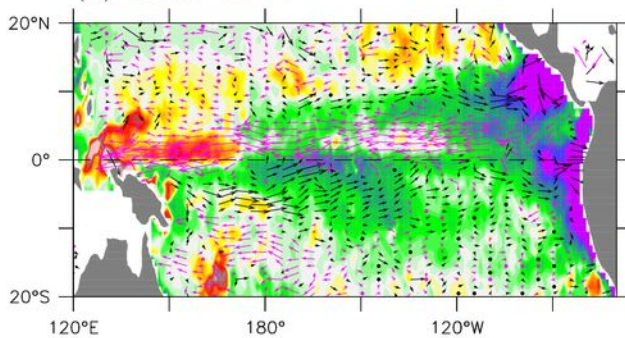
(e) Jan. 1998



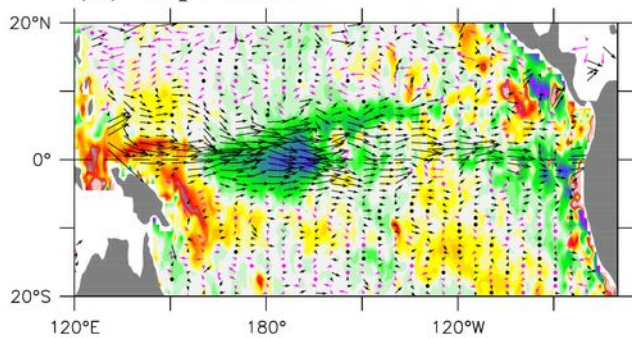
(c) Nov. 1997



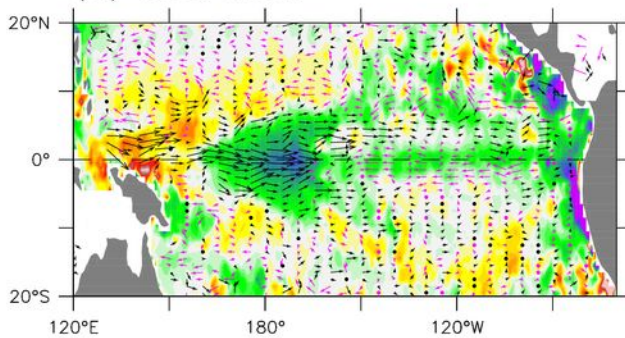
(f) Feb. 1998



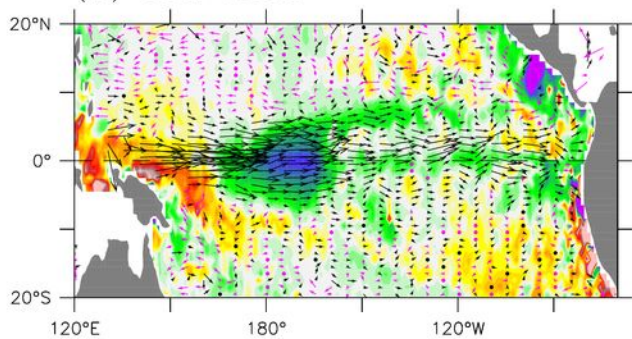
(a) Sep. 2002



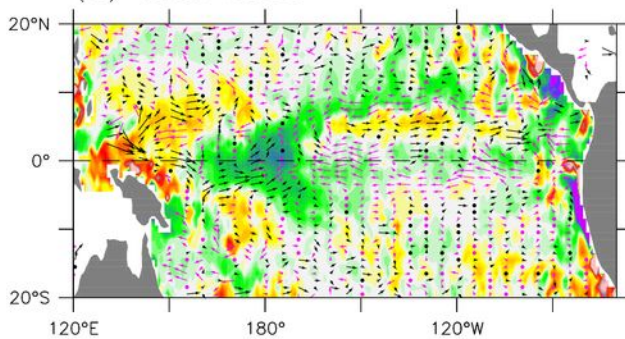
(d) Dec. 2002



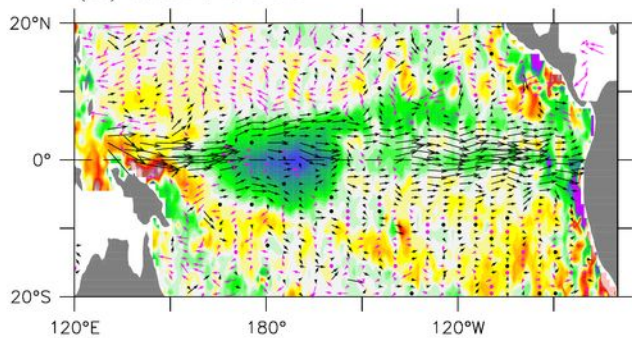
(b) Oct. 2002



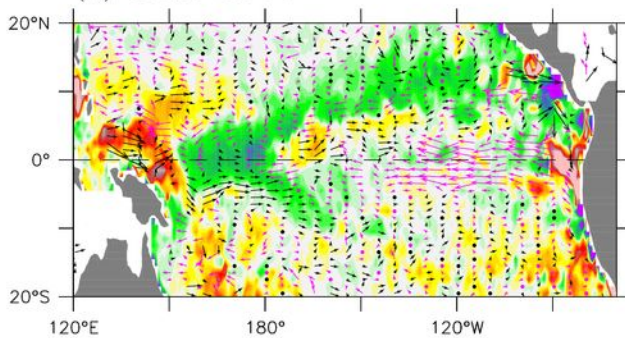
(e) Jan. 2003



(c) Nov. 2002



(f) Feb. 2003



→ 1. m s<sup>-1</sup>