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Impacts of Indian and Atlantic oceans on ENSO

3

in a comprehensive modeling framework

4

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Abstract

16 The impact of the Indian and Atlantic oceans variability on El Niño-Southern-Oscillation (ENSO)
17 phenomenon is investigated through sensitivity experiments with the SINTEX-F2 coupled model. For
18 each experiment, we suppressed the Sea Surface Temperature (SST) variability in either the Indian or
19 Atlantic oceans by applying a strong nudging of the SST toward a SST climatology computed either
20 from a control experiment or observations.

21 In the sensitivity experiments where the nudging is done toward a control SST climatology, the Pacific
22 mean state and seasonal cycle are not changed. Conversely, nudging toward an observed SST
23 climatology in the Indian or Atlantic domain significantly improves the mean state and seasonal cycle,
24 not only in the nudged domain, but also in the whole tropics.

25 These experiments also demonstrate that decoupling the Indian or Atlantic variability modifies the
26 phase-locking of ENSO to the annual cycle, influences significantly the timing and processes of
27 ENSO onset and termination stages, and, finally, shifts to lower frequencies the main ENSO
28 periodicities. Overall, these results suggest that both the Indian and Atlantic SSTs have a significant
29 damping effect on ENSO variability and promote a shorter ENSO cycle. The reduction of ENSO
30 amplitude is particularly significant when the Indian Ocean is decoupled, but the shift of ENSO to
31 lower frequencies is more pronounced in the Atlantic decoupled experiments. These changes of ENSO
32 statistical properties are related to stronger Bjerknes and thermocline feedbacks in the nudged
33 experiments.

34 During the mature phase of El Niño events, warm SST anomalies are found over the Indian and
35 Atlantic oceans in observations or the control run. Consistent with previous studies, the nudged
36 experiments demonstrate that these warm SSTs induce easterly surface wind anomalies over the far
37 western equatorial Pacific, which fasten the transition from El Niño to La Niña and promote a shorter
38 ENSO cycle in the control run. These results may be explained by modulations of the Walker
39 circulation induced directly or indirectly by the Indian and Atlantic SSTs. Another interesting result is
40 that decoupling the Atlantic or Indian oceans change the timing of ENSO onset and the relative role of
41 other ENSO atmospheric precursors such as the extra-tropical Pacific Meridional Modes or the
42 Western North Pacific SSTs.

43

44 **Keywords:** El Niño-Southern Oscillation; Indian Ocean; Atlantic Ocean; ocean-atmosphere
45 interactions; coupled climate model.

46

47 **1. Introduction**

48

49 El Niño–Southern Oscillation (ENSO) is the dominant mode of climate variability in the
50 Tropics (Clarke 2008). ENSO has a far-reaching effect leading to extensive floods or
51 anomalous droughts in many regions of the globe. Thus, advanced and accurate forecast of
52 ENSO has significant applications, but is still a challenging problem as illustrated by the
53 complete failure of the recent forecasts issued in boreal spring of 2014 by the different
54 operational predictions groups, which indicated a high chance of a major El Niño evolving
55 over the summer, autumn and winter of 2014 (Tollefson 2014). Basic properties of ENSO are
56 now rather well understood and simulated by current Coupled General Circulation Models
57 (CGCMs). However, anticipating its behavior before boreal spring and understanding its
58 relationship with other tropical and extra-tropical regions are still related and challenging
59 problems.

60

61 While the classical picture is that the Indian and Atlantic oceans only play a passive role in
62 ENSO evolution, many new studies suggest an active role of these two basins on ENSO.
63 First, there is a lot of studies focusing on the role of Indian Ocean Basin-wide (IOB) warming
64 or Indian Ocean Dipole (IOD) in ENSO onset and evolution (Kug and Kang 2006; Kug et al.
65 2006a, 2006b; Ohba and Ueda 2007; Izumo et al. 2010; Luo et al. 2010; Dayan et al. 2014).
66 They pointed out that there are significant differences in ENSO evolution with and without
67 IOB warming or IOD in observations. It has been suggested that the ENSO-forced IOB
68 warming might affect atmospheric circulation over the western Pacific to fasten the turnabout
69 of the ENSO cycle (Kug and Kang 2006). Izumo et al. (2010) claimed that skillful ENSO
70 forecasts are possible 14 months in advance with the help of the IOD. These results are
71 related to the fact that zonal wind anomalies associated with the IOB or IOD may propagate
72 from the equatorial Indian Ocean into the equatorial Pacific before the onset of El Niño or
73 during the transitions from El Niño to La Niña and are thus useful parameters to overcome the
74 spring predictability barrier of ENSO (Barnett 1983; Clarke and Van Gorder 2003; Clarke
75 2008; Kug et al. 2010). Sea Surface Temperature (SST) anomalies over the South Indian
76 Ocean may also act as a remote forcing to promote wind anomalies in the western equatorial
77 Pacific (Dominiak and Terray 2005; Terray 2011; Boschat et al. 2013). These recent studies
78 imply that the Indian Ocean may be an integral part of the ENSO dynamics. But, the relative
79 roles of the IOD, IOB or South Indian Ocean SSTs in ENSO are difficult to assess from
80 observations because these relationships involve a complex interplay of ocean and

81 atmospheric processes in both the Indian and Pacific oceans (Santoso et al. 2012; Dayan et al.
82 2014).

83
84 The relationships between the tropical Atlantic Ocean and ENSO have also been the focus of
85 many recent studies (Dommenges et al. 2006; Rodriguez-Fonseca et al. 2009; Jansen et al.
86 2009; Ding et al. 2012; Keenlyside et al. 2013; Ham et al. 2013a, b; Polo et al. 2014;
87 Kurchaski et al. 2015). Several of these studies emphasized that the so-called Atlantic El Niño
88 events during boreal summer can modulate ENSO variability through a modulation of the
89 Walker circulation (Frauen and Dommenges 2012; Ding et al. 2012; Polo et al. 2014).
90 Interestingly, some Atlantic Niñas coincide with the strongest El Niño cases (i.e., 1982/1983
91 and 1997/1998) in the twentieth century. This implies that Atlantic Niñas may help to develop
92 the strongest El Niño events in the Pacific and improve ENSO predictions (Keenlyside et al.
93 2013). The lead relationship between Atlantic variability and ENSO also exists in boreal
94 winter and early spring, involving the southern Atlantic (Terray 2011; Boschat et al. 2013) or
95 the subtropical North Atlantic (Ham et al. 2013a,b; Dayan et al. 2013). Thus, a large variety
96 of mechanisms may account for the statistical relationships between the Atlantic and Pacific
97 basins in addition to the traditional ENSO teleconnection (Chang et al. 2006).

98
99 The complex interactions between ENSO, Indian and Atlantic oceans make it difficult, and
100 probably almost impossible, to determine their cause-and-effect relationships using
101 observational analyses or forced atmospheric experiments (Dayan et al. 2014). All the
102 puzzling results discussed above need to be verified with a comprehensive CGCM in order to
103 demonstrate causal relationships between Atlantic or Indian ocean variability and ENSO.
104 CGCM experiments that can isolate the coupling processes within and external to the Pacific
105 Ocean are thus a useful alternative for testing the various hypotheses.

106
107 Basin decoupling or partially coupled experiments have already been performed to study the
108 inter-basin interactions between the Indian, Atlantic and Pacific oceans (e.g., Yu et al. 2002,
109 2009; Yu 2005; Wu and Kirtman 2004; Kug et al. 2006b; Santoso et al. 2012 for the Indian
110 Ocean; Rodriguez-Fonseca et al. 2009; Ding et al. 2012; Keenlyside et al. 2013; Polo et al.
111 2014 for the Atlantic Ocean). However, very few studies have performed decoupling or
112 partially coupled experiments for both the Indian and Atlantic oceans in exactly the same
113 modeling framework, except Dommenges et al. (2006) or Frauen and Dommenges (2012).

114

115 Frauen and Dommenges (2012) argued that knowing the initial conditions and simulating the
116 evolution in the tropical Atlantic is more important than the knowledge of Indian Ocean
117 evolution for ENSO predictability. On the other hand, Keenlyside et al. (2013), using
118 numerical experiments with observed Atlantic SSTs nudged into a coupled model, found that
119 Atlantic variability during boreal winter and spring is irrelevant for ENSO prediction, only
120 Atlantic Ocean SSTs during boreal summer do matter for ENSO. Similarly, the CGCM
121 results obtained so far on the impact of Indian Ocean SSTs on ENSO are somewhat
122 controversial. Some of these previous CGCM studies find that Indian Ocean coupling could
123 enhance ENSO variability (Yu et al. 2002, 2009; Wu and Kirtman 2004), while others found
124 the reverse (Dommenges et al. 2006; Kug et al. 2006a; Frauen and Dommenges 2012; Santoso
125 et al. 2012) or no impact of the Indian Ocean coupling on ENSO amplitude (Yeh et al. 2007).

126
127 The inconsistencies in these numerical experiments could be partly due to model biases, too
128 short simulations, the use of partly simplified coupled models or flux adjustments, or different
129 decoupling strategies. The most common approach in the different studies is to replace the
130 simulated SSTs in a basin with a monthly SST climatology. However, some studies used a
131 SST climatology estimated from observations (Yu et al. 2002, 2009), while others used a SST
132 climatology from a control simulation in the decoupled experiments (Wu and Kirtman 2004;
133 Kug et al. 2006b; Santoso et al. 2012). These different approaches may induce different
134 conclusions because SST biases may vary from one model to another. Furthermore, by using
135 SST climatology from observations, these SST biases are removed, but we cannot ascertain
136 that the changes of variability in the decoupled experiment are due to the SST biases or to the
137 absence of variability in the basin in which the ocean–atmosphere coupling has been turned
138 off.

139
140 Here, we employ a fully coupled climate model without any flux adjustments and a realistic
141 ENSO variability to address these issues. More precisely, we use a series of basin-decoupling
142 CGCM experiments using both observed and simulated SST climatologies to discriminate the
143 relative roles of Indian and Atlantic basins on ENSO variability. While our modeling results
144 are consistent with some previous studies and support the hypothesis that both the Indian and
145 Atlantic SST variability accelerates the decaying phase of El Niño, they also suggest that
146 Indian and Atlantic variability plays an important role in ENSO onset and modulates ENSO
147 feedbacks.

148

149 The article is organized as follows. In section 2, we give a brief overview to the SINTEX-F2
150 coupled model, the design of the decoupled experiments and the statistical tools used here.
151 Section 3 is devoted to the validation of ENSO variability in SINTEX-F2 and to a description
152 of the changes of ENSO properties and feedbacks directly associated with Indian and Atlantic
153 variability in the decoupled experiments. Sections 4 and 5 focus more specifically on the
154 onset and decaying stages of ENSO, which account for many of the simulated ENSO
155 changes. Conclusions and prospects for future work are given in section 6.

156

157 **2. Coupled model, experimental design and statistical tools**

158

159 2.a Model setup and sensitivity experiments

160 In order to study the impact of the tropical Atlantic and Indian oceans on ENSO, the
161 SINTEX-F2 CGCM is employed (Masson et al. 2012). The atmosphere model is ECHAM5.3
162 and is run at T106 spectral resolution (about 1.125° by 1.125°) with 31 hybrid sigma-pressure
163 levels (Roeckner et al. 2003). The oceanic component is NEMO (Madec 2008), using the
164 ORCA05 horizontal resolution (0.5°), 31 unevenly spaced vertical levels and including the
165 LIM2 ice model (Timmermann et al. 2005). Atmosphere and ocean are coupled by means of
166 the Ocean-Atmosphere-Sea Ice-Soil (OASIS3) coupler (Valcke 2006). The coupling
167 information, without any flux corrections, is exchanged every 2 h. The model does not require
168 flux adjustment to maintain a stable climate, and simulates a realistic mean state and
169 interannual variability in the tropical Pacific (Masson et al. 2012). It has also been shown to
170 perform very well in ENSO prediction studies (Luo et al. 2005a). The performance of the
171 SINTEX-F2 model in simulating the seasonal cycle in the Indian areas has been assessed in
172 Terray et al. (2012) and Prodhomme et al. (2014) and is not repeated here.

173

174 We first run a 210-years control experiment (named REF hereafter) with the standard coupled
175 configuration of SINTEX-F2. The mean SST climatology estimated from the REF
176 simulations has been compared to SST data from 1979-2012 taken from the Hadley Center
177 Sea Ice and Sea Surface Temperature dataset version 1.1 (HadISST1.1; Rayner et al. 2003).
178 The tropical areas are generally too warm and extra-tropics too cold (Figs. 1a,b). Mean
179 deviations from observed SST is less than 1°C over much of the tropical Indian and Pacific
180 oceans, with exceptions in the Humbolt and California upwelling regions where warmer

181 biases are found. The REF simulation also exhibits a slight cold tongue bias in the central
182 equatorial Pacific, but this bias is very limited compared to many other CGCMs. The SSTs in
183 the Atlantic are much less realistic, with a strong warm bias (over 3°C) in the eastern
184 equatorial and Benguela regions, a problem found in many state-of-the-art CGCMs (Richter
185 et al. 2014).

186

187 To study in a common framework, the interactions of the tropical Indian and Atlantic oceans
188 with ENSO, a partial coupled configuration of the model is used, with full coupling
189 everywhere and no flux corrections, except in a specific area (e.g. the tropical Atlantic or
190 Indian oceans), where model SSTs are nudged to a daily varying SST climatology obtained
191 from the control run or observations, following the protocol described in Luo et al. (2005a).
192 The method essentially modifies the non-solar heat flux provided by the atmosphere to the
193 ocean by adding a correction term that scales with the SST error of the model. For example, if
194 the model SST is too warm we decrease the heat flux. The damping constant used is -2400 W
195 $\text{m}^{-2} \text{K}^{-1}$ and this large feedback value is applied between 30°S and 30°N in each domain. This
196 value corresponds to the 1-day relaxation time for temperature in a 50-m ocean layer. A
197 Gaussian smoothing is finally applied in a transition zone of 5° in both longitude and latitude
198 at the limits of the SST restoring domains. This large correction, using daily climatology,
199 completely suppresses the SST variability in each corrected region.

200

201 First, we perform two 110 years sensitivity experiments (named FTIC and FTAC hereafter),
202 where the model SSTs in the tropical Indian and Atlantic oceans are nudged to the SST
203 climatology obtained from the REF simulation. These two experiments are initialized with the
204 same initial conditions as REF in order to allow a direct comparison between the different
205 simulations. Thus, in these experiments there are no change in the Indian or Atlantic SST
206 mean state and seasonal cycle compared to the REF simulation, but SST interannual
207 variability is suppressed in the corrected region. In order to assess the robustness of our
208 results with respect to the SST biases of the model (especially for the tropical Atlantic where
209 the model's SST biases are particularly significant in REF), two additional experiments,
210 named FTIC-obs and FTAC-obs, were also conducted in which the SST damping is applied
211 towards a daily climatology computed from the AVHRR only daily Optimum Interpolation
212 SST (OISST) version 2 dataset for the 1982–2010 period (Reynolds et al. 2007). These
213 simulations have a length of 50 years. In these runs, the large feedback value applied fully

214 removes the SST biases with respect to the observed SST climatology in addition of
215 suppressing the SST variability in the restoring domain. It is important to keep in mind that
216 the main focus of these two new experiments will be again to delineate, at the first order, if
217 ENSO is significantly different with or without inclusion of the tropical Indian or Atlantic
218 SST variability in the coupled simulation. In particular, we will not discuss in details the
219 changes of the Pacific mean state and variability induced specifically by the correction of the
220 SST biases in the Indian and Atlantic Oceans in the FTIC-obs and FTAC-obs runs. Table 1
221 summarizes the specifications of the different sensitivity experiments used here. Finally, in
222 the following analyses, the first 20 years of the REF, FTIC and FTAC, and the first 10 years
223 of FTIC-obs and FTAC-obs have been excluded due to the spin-up of the coupled model.

224

225 2.b Maximum Covariance Analysis

226 To trace the mechanisms governing the ENSO modifications in our set of experiments, we
227 use the Maximum Covariance Analysis (MCA) approach described in Masson et al. (2012).
228 In short, MCA describes the linear relationships between two fields by estimating the
229 covariance matrix between these two fields and computing the Singular Value Decomposition
230 (SVD) of this covariance matrix for defining some pairs of spatial patterns, which describe a
231 fraction of the total square covariance (Bretherton et al. 1992). The MCA results in spatial
232 patterns and time series. The k th Expansion Coefficient (EC) time series for each variable is
233 obtained by projecting the original monthly anomalies onto the k th singular vector of the SVD
234 of the covariance matrix. Using the ECs from the MCA, two types of regression maps can be
235 generated: the k th homogenous vector, which is the regression map between a given data field
236 and its k th EC, and the k th heterogeneous vector, which is the regression map between a given
237 data field and the k th EC of the other field. The k th heterogeneous vector indicates how well
238 the grid point anomalies of one field can be predicted from the k th EC time series of the other
239 field. The chosen geographical domain for the various MCA computations covers the Pacific
240 Ocean (defined in the area 30°S-30°N and 120-290°E. Changing the latitude boundaries of
241 this domain yields leading spatial patterns of variability that are almost similar and did not
242 change the main results presented in section 3.

243

244 The Square Covariance Fraction (SCF) is a first simple measure of the relative importance of
245 each mode (e.g. of each singular triplets of the covariance matrix between the two fields) in
246 the relationship between two fields. The correlation value (r) between the k th ECs of the two

247 fields and the Normalized root-mean-square Covariance (NC), introduced by Zhang et al.
248 (1998), indicate how strongly related the coupled patterns are. These statistics will be used in
249 section 3 to assess the strength of various atmosphere-ocean feedbacks in the different
250 experiments.

251

252 **3. ENSO statistics, feedbacks and evolution**

253

254 3.a ENSO statistics in observations and REF

255 We first assess how the ENSO variability is realistic in REF, especially how the observed
256 tropical SST evolution during ENSO events is reproduced. Such a validation is particularly
257 important, as many of the discrepancies found about the influence of the Indian and Atlantic
258 oceans on ENSO in previous modeling studies may be related to the errors in the particular
259 CGCM used in each case.

260

261 The difference between the observed and simulated spatial patterns of SST variability is
262 shown in Fig. 1d. As in observation, the maximum of SST variability in REF is found in the
263 central equatorial Pacific Fig. 1c. This maximum is slightly weaker than observed (up to -
264 0.2°C), but is not shifted westward as in many other CGCMs, thanks to the coupling method
265 described in Luo et al. (2005b). REF also shows some discrepancies with observations in the
266 eastern Pacific. The SST variability is overestimated over the southeast Indian Ocean,
267 particularly along the Java-Sumatra coast, which is one of the centers of the IOD. On the
268 other hand, the SST variability is significantly reduced over the southeast Atlantic Ocean.
269 This is consistent with the strong SST biases in the tropical Atlantic mean state (Fig. 1b).

270 In order to provide a quantitative assessment of the tropical SST variability, the SST standard-
271 deviations of the Niño-3.4, IOB, IOD indices and the equatorial Atlantic mode (defined by the
272 classical ATL3 index) in observations and REF are shown in Table 2. As noted before, REF
273 has a slightly lower Niño-3.4 SST variability than observed. In agreement with Terray et al.
274 (2012), standard-deviation of the IOD index is much stronger than in observations, but the
275 amplitude of the IOB is realistic. Finally, the simulated ATL3 SST variability is significantly
276 reduced. These important biases affecting the equatorial Indian and Atlantic variability are
277 unfortunately very common in current coupled models and are related to an erroneous
278 representation of the Bjerknes, wind-evaporation-SST and cloud-radiation-SST feedbacks,

279 which govern the evolution and amplitude of the IOD and equatorial Atlantic modes (Liu et
280 al. 2014; Li et al. 2015; Richter et al. 2014; Kucharski et al. 2015).

281

282 We now examined the seasonal dependence of ENSO SST variability (Fig. 2a). The monthly
283 standard deviation of Niño-3.4 SSTs is the highest from November to February and is about
284 1-0.9°C in the control simulation and 1-1.1°C in observations. The lowest standard deviations
285 are observed during boreal spring in both observations and REF, suggesting that the model
286 reproduces a reasonable phase-locking to the annual cycle and that El Niño onset usually
287 occurs during boreal spring as observed. The results concerning the nudged experiments will
288 be discussed below (see section 3.b).

289

290 In order to document how the ENSO frequency is simulated, power spectrum analysis is used
291 (Fig. 2b). The spectral density for the observations is estimated from the HadISST1.1 dataset
292 for the period 1900–2012 after removal of the seasonal cycle and trend. A similar
293 preprocessing was done for the different simulations. As expected, the dominant period for
294 the observed Niño-3.4 SST index is about 4 years with a broad spectrum between 2 and 6
295 years. The Niño-3.4 SST spectrum in REF is highly realistic with a similar broad spectral
296 peak. Furthermore, the simulated ENSO spectrum is always inside the 99% confidence
297 interval estimated from the observed spectrum (dashed lines in Fig. 2b). This is again a
298 distinctive feature of SINTEX-F2 compared to many CMIP5 models, which still fail to
299 reproduce the observed ENSO frequency (Jha et al. 2013). This good agreement between the
300 observed and simulated ENSO power spectra provides confidence to use SINTEX-F2 to
301 investigate the changes of ENSO frequency associated with Indian or Atlantic oceans
302 variability as well as the mechanisms connecting the variability in the three basins, in the next
303 sections.

304

305 Figure 3 shows the observed and simulated correlations of tropical SSTs with the Niño-3.4
306 SST during December-January (DJ hereafter), when ENSO is in its mature phase. The ENSO-
307 developing (decaying) phase is defined as a prior (posterior) period of the ENSO mature
308 phase (DJ). The years for the ENSO-developing and decaying phases will be denoted “year0”
309 and “year+1”, respectively. Similarly, we will use the notation “year-1” to denote the year
310 preceding the ENSO-developing year in REF. The modeled El Niño has its onset during the
311 boreal spring as observed. The simulated spatial pattern of extra-tropical SST anomalies in the

312 North and South Pacific before or at the ENSO onset period matches the observations. Cold
313 SST anomalies are also observed and simulated before El Niño onset over the tropical Indian
314 and Atlantic oceans. However, these SST signals are mainly confined in the tropics in REF,
315 while they extend significantly in the southern ocean in observations. Also in the tropical
316 Atlantic Ocean, significant cold anomalies are found north of the equator in REF rather than
317 in the southern hemisphere as in observations during boreal spring and summer of year0 (Fig.
318 3).

319

320 It is interesting to note that, conversely to many CCGMs, the ENSO events in REF are not too
321 narrowly confined to the equator and do not extend too far to the west. Furthermore, the life
322 cycle of the ENSO events in the model, as seen from the SST anomalies, agrees well with
323 observations, exhibiting realistic teleconnections with the extra-tropics, as well as over the
324 Indian and Atlantic Oceans (Fig. 3). As an illustration, SINTEX-F2 is able to reproduce
325 realistic correlations of ENSO with IOD and IOB modes during boreal fall and winter,
326 respectively. Nevertheless, the simulated IOD starts too early, extends too far west along the
327 equator, is too strong (see Table 2) and is strongly correlated with ENSO compared to
328 observations. This is a persisting bias of the SINTEX model, which is related to the
329 overestimated strength of the wind-thermocline-SST and wind-evaporation-SST positive
330 feedbacks and a too shallow thermocline in eastern equatorial Indian Ocean during boreal
331 summer and fall (Fischer et al. 2005; Terray et al. 2012). To a large extent, El Niño (La Niña)
332 events tend to be accompanied by Atlantic Ocean warming (cooling) from late boreal summer
333 onward both in observations and REF (Fig. 3). Subsequent to the El Niño mature phase, the
334 tropical Pacific SST anomalies start to decay at the beginning of year+1 and most simulated
335 El Niño events terminate before or during boreal summer of year+1 as observed (see Fig. 12
336 later).

337

338 In summary, SINTEX-F2 simulates well many of the ENSO statistical properties, such as the
339 spatial pattern of tropical Pacific SST standard deviation, the ENSO teleconnections, the
340 power spectrum of Niño-3.4 SSTs, the seasonal phase locking of ENSO variability (Masson et
341 al. 2012). This is entirely consistent with its success in climate predictability studies (Luo et
342 al. 2005a). This good agreement provides confidence to further investigate the mechanisms
343 connecting variability in the three tropical basins with this CGCM. In particular, we now
344 carefully compare the ENSO statistics between the nudged and REF simulations.

345

346 3.b ENSO statistics in the nudged experiments

347 The exclusion of the Atlantic or Indian Ocean variability has a non-significant impact on the
348 mean SST in the other tropical basins and, especially, in the tropical Pacific (Figs. 4a, c).
349 Between 40°S-40°N, the differences of the FTIC and FTAC nudged experiments with REF
350 are most often smaller than 0.05°C in amplitude inside, but also outside the nudging regions.
351 In any case, these values are much smaller than the SST differences with the observations
352 (Fig. 1b). This statement is also valid for the monthly SST climatologies computed from
353 FTIC and FTAC (not shown). These results must be kept in mind when we discuss the
354 changes of ENSO variability in these nudged experiments since these changes cannot be
355 explained by differences in the Pacific mean state or seasonal cycle in these simulations
356 compared to REF.

357

358 However, these conclusions are not valid when we consider the FTIC-obs and FTAC-obs
359 experiments (Figs. 4b, d). Interestingly, the corrections of the tropical Indian (Atlantic) warm
360 SST biases in these experiments have a significant influence on the mean state of the other
361 basins and, generally, lead to a colder mean state and a systematic reduction of the warm SST
362 bias in the tropics. This is especially true for the western-central Pacific, but also for different
363 upwelling regions such as the eastern equatorial Atlantic in FTIC-obs (Fig. 4b) or the South
364 American and East African coasts in FTAC-obs (Fig. 4d). These nudged experiments also
365 show a slight warming in the central-east equatorial Pacific compared to REF. Thus,
366 removing SST biases in the Atlantic or Indian oceans, surprisingly, improves the simulated
367 Pacific mean state by inducing a cooling of the Indo-Pacific warm pool and a slight reduction
368 of the cold tongue bias in the central-east equatorial Pacific. This is consistent with the works
369 of Kucharski et al. (2011) or Chikamoto et al. (2012) in a global warming context, which
370 suggest that the long-term Atlantic or Indian warming trends have played a role in reducing
371 the eastern Pacific warming. Similarly, the interpretation here is that reduction of the Indian
372 or Atlantic tropical heating associated with the correction of the warm SST biases in FTIC-
373 obs and FTAC-obs promotes changes in the Walker circulation and induces an El Niño-like
374 change in the equatorial Pacific (figures not shown). This modest improvement of the
375 equatorial Pacific mean state in both the FTAC-obs and FTIC-obs experiments may also have
376 a significant impact on the simulated ENSO since many aspects of ENSO variability,

377 including its seasonal phase-locking properties, are largely dependent on the tropical Pacific
378 mean state.

379

380 Figs. 4e-h first confirm that almost all the tropical Indian (Atlantic) SST variability has been
381 removed in the nudged experiments. Outside the corrected region, the SST variability changes
382 in the nudged experiments are mainly found in the tropical Pacific. A large increase of ENSO
383 variability is found when the Indian Ocean variability is excluded (Figs 4e, f), but only a
384 marginal increase when the Atlantic Ocean is decoupled (Figs. 4g, h). The increase of the
385 standard deviation is as large as 0.3-0.5°C in the central-eastern equatorial Pacific for the
386 Indian Ocean decoupled runs. This result is opposite to the earlier findings (Yu et al. 2002,
387 2005; Wu and Kirtman 2004), but agrees well with more recent studies (Kug and Kang 2006;
388 Dommenges et al. 2006; Jansen et al. 2009; Frauen and Dommenges 2012; Santoso et al.
389 2012). On the other hand, decoupling the Atlantic SST variability leads to only a very modest
390 increase of ENSO variability, which is in contrast to some recent results (Dommenges et al.
391 2006; Frauen and Dommenges 2012). Another interesting result is that the increased ENSO
392 variability is always stronger when the nudging is done toward the observed SST climatology
393 and this increase is particularly significant in FTIC-obs.

394

395 It is well known that ENSO exhibits an asymmetric behaviour between its opposing phases
396 (Clarke 2008). This non-linear ENSO component manifests itself with a significant positive
397 SST skewness in the eastern equatorial Pacific (Masson et al. 2012; Roxy et al. 2014).
398 SINTEX-F2 has difficulties in representing this positive skewness associated with ENSO as
399 many other CGCMs (see Fig. 11 of Masson et al., 2012). However, an interesting observation
400 is that the SST skewness increased in all the decoupled experiments, suggesting that the
401 ENSO damping associated with the Indian and Atlantic oceans coupling affects more the El
402 Niño than the La Niña events (figures not shown).

403

404 We are now shifting our focus to the assessment of the changes in ENSO seasonal phase
405 locking in the nudged experiments (Fig. 2a). The peak phase of the simulated ENSO occurs in
406 boreal winter and the minimum standard deviation is found during boreal spring in all
407 simulations and the observations, but there are distinctive features between the nudged
408 experiments and REF. The most striking features are (i) the increase of ENSO variability
409 during boreal summer in nearly all the nudged experiments, (ii) the recovery of a clear and

410 realistic minimum of monthly SST variability during boreal spring (e.g. April) and the so-
411 called “spring barrier” in all the nudged experiments using an observed SST climatology
412 despite of the general increase of ENSO variability in these simulations and (iii) a less
413 pronounced minimum of monthly SST variability during boreal spring in FTIC and FTAC,
414 which use a simulated SST climatology for the nudging.

415

416 These results suggest surprisingly that the seasonal phase locking of ENSO is partly
417 associated with its coupling to the Indian and Atlantic oceans. More specifically, we interpret
418 the recovery of a more realistic seasonal phase locking of ENSO in FTIC-obs and FTAC-obs
419 as a direct consequence of the improvement of the Pacific mean state and seasonal cycle in
420 these simulations, especially a tendency for a decreased east-west SST gradient in the
421 equatorial Pacific compared to REF (see Figs. 4b, d). This change is rather modest in Figs. 4b,
422 d where all the months are considered, but is much more prominent during boreal spring
423 favoring the growth of instabilities related to El Niño onset in both FTIC-obs and FTAC-obs
424 (see Figure 14 of Prodhomme et al. 2015). Conversely, there is a substantial increase of SST
425 variability during this crucial season in FTIC and FTAC simulations, pointing again to the
426 significant role of the Indian and Atlantic mean SST biases for a realistic simulation of ENSO
427 properties. All these features are consistent with the rectification of the tropical Pacific mean
428 state in FTIC-obs and FTAC-obs, which is missing in FTIC and FTAC since the nudging is
429 done toward the REF climatology in these runs. The modification of the seasonal phase-
430 locking of ENSO in FTIC and FTAC experiments also suggests that boreal spring is no more
431 the preferential season for ENSO onset or termination, pointing to a possible change of the
432 length of the ENSO events when the Indian or Atlantic oceans are decoupled.

433

434 In order to address these aspects of ENSO variability, power spectra of monthly mean Niño-
435 3.4 SST anomalies in the nudged experiments are displayed in Fig. 2b. The spectra for all the
436 nudged experiments are outside the 99% confidence interval computed from the observed
437 Niño-3.4 SST spectrum, while the spectrum estimated from REF matches the observations.
438 The ratios between the Niño-3.4 SST spectral densities in REF and the nudged experiments
439 can be used to test the hypothesis of a common spectrum for two time series from the
440 different experiments with an F-distribution (Fig. 5; see Masson et al. 2012 for further
441 details). All the nudged experiments show more power at longer periods compared to REF
442 and these changes are significant since their corresponding spectral ratios are below the lower

443 limit of the 90% point-wise tolerance interval (computed under the assumption that the two
444 underlying spectra are the same) for periods between 8 and 10 (10 and 30) years for FTIC or
445 FTIC-obs (FTAC and FTAC-obs). This shift to longer ENSO periods is particularly strong
446 when the Atlantic Ocean is decoupled (Figs. 5c, d). Interestingly, if we consider the periods
447 between 1.5 and 3 years, almost all spectral ratios also show values above the upper limit of
448 the 90% point-wise tolerance interval, suggesting that both Indian and Atlantic oceans
449 couplings introduce significant biennial variability in the simulated ENSO (Yu et al. 2009).

450

451 Furthermore, all these results are robust since they are found independently of the details of
452 the nudging (e.g. if this nudging is done toward a simulated or observed SST climatology).
453 This suggests that the turnabout of ENSO cycle is lengthened and that ENSO's recurrence
454 may shift to lower frequencies when the coupling is turned off, especially in the Atlantic
455 Ocean. Moreover, none of these changes can be explained by a change of the mean-state in
456 the tropical Pacific in the case of the FTIC and FTAC experiments. Taking into account that
457 the mature phase of ENSO still occurs during boreal winter, there are two possible, but no
458 contradictory, explanations for these differences between the nudged experiments and the
459 control simulation: the first is that ENSO onset occurs before boreal spring, the second is that
460 the transition from El Niño to La Niña is much slower in the decoupled runs. This last
461 hypothesis is consistent with the feedback mechanism proposed by Kug and Kang (2006) as
462 far the Indian Ocean is concerned. Both explanations may also explain why the phase-locking
463 of ENSO to the annual cycle is reduced in the FTIC and FTAC decoupled experiments (see
464 Fig. 2a). These two hypotheses will be further examined in the next sections (i.e. in sections 4
465 and 5).

466

467 3.c ENSO feedbacks in the nudged experiments

468 A lead-lag correlation analysis between the simulated 20°C isotherm depth (20d hereafter)
469 and the DJ Niño-3.4 SSTs further supports the finding that decoupling the Atlantic or Indian
470 oceans leads to a longer ENSO cycle with significant changes of oceanic adjustment
471 associated with the ENSO cycle (Fig. 6). We only show the results from the FTIC and FTAC
472 experiments for conciseness and because the simulated changes cannot be explained by
473 changes in the Pacific mean state in these experiments (Figs. 4a, c). However, similar results
474 are obtained from the FTIC-obs and FTAC-obs experiments.

475 The simulated 20d anomaly in the western and central (eastern) Pacific is significantly and
476 positively (negatively) correlated with the DJ Niño-3.4 SSTs when the former leads the latter
477 by about 2 years in all the simulations (Fig. 6 *1nd row*). This feature is consistent with the
478 delayed oscillator and recharge oscillator theories in which subsurface ocean preconditioning
479 is crucial for the development of El Niño (Jin 1997; Clarke 2008). Interestingly, a similar
480 correlation pattern is found when the 20d leads the Niño-3.4 SSTs by about one year in REF,
481 but not in FTIC and FTAC (Fig. 6 *2nd row*). When the Indian or Atlantic oceans are
482 decoupled, the western Pacific positive heat content anomalies have already propagated
483 eastward and are found in the central and eastern equatorial Pacific one year in advance of the
484 peak of the Niño-3.4 SST anomaly. This suggests that El Niño has already started in FTIC or
485 FTAC, but not in REF. During El Niño mature phase, the correlation maps have similar
486 patterns and amplitudes (Fig. 6 *3rd row*). However, the correlation patterns when the DJ
487 Niño-3.4 SSTs lead the 20d by one year confirm that the Indian or Atlantic oceans coupling
488 fastens the transition from El Niño to La Niña in REF. Significant negative heat content
489 anomalies are already well established in the eastern Pacific and the extra-tropical heat
490 discharge completed during the boreal fall following the peak ENSO phase in REF. On the
491 other hand, the eastern Pacific positive heat content anomaly is not collapsed and the
492 discharge is still active in FTIC or FTAC (Fig. 6 *4nd row*). This is consistent with the shift
493 to lower frequencies in the simulated ENSO spectra in the nudged experiments.

494

495 ENSO evolution results from a number of ocean-atmosphere feedbacks (Jin et al. 2006;
496 Guilyardi et al. 2009). Thus, evaluating the balance and strength of these feedbacks may lead
497 to a better understanding of the mechanisms, which are important for the significant
498 modifications of the simulated ENSO characteristics in the nudged experiments. Here, we
499 focus specifically on the positive Bjerknes and thermocline feedbacks in the different
500 experiments with the help of different MCAs (see section 2.b).

501

502 Figures 7a, b display the leading modes derived from a MCA analysis of the SST and zonal
503 wind stress (USTR hereafter) anomaly fields over the Pacific Ocean for the REF simulation.
504 In the terminology of Bretherton et al. (1992), the fields presented in Figs. 7a, b are
505 homogenous (covariance) pattern for SST and heterogeneous pattern for USTR. The leading
506 MCA modes for these variables in the nudged experiments have similar spatial structures, but
507 with spatial loadings of slightly greater amplitude (especially for FTIC and FTIC-obs), and

508 are thus not shown here. This MCA is useful to measure the strength of the Bjerknes feedback
509 and, more particularly, the intensity of the wind response to SST anomalies in the central and
510 eastern Pacific (Clarke 2008).

511

512 As expected, the SST homogenous pattern represents the peak phase of El Niño with warm
513 SST anomalies along the equator in the central and eastern Pacific and the two cold branches
514 of the traditional “horseshoe” pattern over the western Pacific (Fig. 7a). The corresponding
515 heterogeneous USTR pattern features a westerly anomaly in the western and central
516 equatorial Pacific, and a weaker easterly anomaly to the east (Fig. 7b). This wind response is
517 reminiscent of a Gill-Matsuno response to the warm SST anomaly in the central and eastern
518 Pacific in Fig. 7a.

519

520 In ENSO dynamics, another key element is the thermocline response to the USTR anomaly in
521 the equatorial Pacific (Wyrtki 1975, Clarke 2008). To explore how this ocean-atmosphere
522 interaction is modulated in the different experiments, we show the leading MCA mode from
523 the covariance matrix between USTR and 20d anomalies for REF in Figures 7c and d. Again,
524 we only present the results for REF because the leading MCA modes in the nudged
525 experiments have similar spatial patterns (figures not shown). In all the experiments, the
526 heterogeneous 20d pattern describes a zonally tilting mode between western and central-
527 eastern tropical Pacific (Fig. 7d). The associated homogeneous USTR pattern is exactly
528 similar in structure to the zonal wind stress pattern from the SST-USTR MCAs discussed
529 above. That is the USTR pattern in Figs. 7b and c corresponds to the peak phase of El Niño
530 and illustrates that the zonal tilt of the thermocline across the equatorial Pacific reacts quickly
531 to wind stress anomalies in the western-central Pacific, partly in the form of ocean Kelvin
532 waves (Clarke 2008).

533

534 Furthermore, combining this coupled 20d-USTR mode with the USTR-SST mode leads to the
535 following interpretation, which is embedded in the two diagnostic equations of the recharge
536 oscillator model (Jin 1997; Burger et al. 2005): the eastern and central equatorial Pacific
537 warming, which is observed during the peak phase of El Niño, sets up an anomalous wind
538 stress anomaly in the western-central Pacific through a Gill-type response. This wind stress
539 anomaly, in turn, influences the east-west gradient in thermocline depth as the warm water
540 flows eastward. This leads to a weaker equatorial Pacific SST gradient and, finally, results in

541 a positive wind-thermocline-SST feedback in which SST gradients trigger anomalous winds
542 and these winds amplify the initial SST gradients.

543

544 Tables 3 and 4 present summary statistics for the MCAs, including the SCFs and NCs for the
545 leading modes in the various MCA expansions and the correlation (r) between the EC time
546 series of the left and right fields. These statistics are useful to investigate the strength of this
547 positive wind-thermocline-SST feedback in the different experiments. First we note that the
548 leading modes of the MCAs between SST, USTR and 20d account for most of the SCF
549 between these variables in the different experiments and are well separated from the lower
550 MCA modes in terms of described SCF (Tables 3 and 4). Secondly, despite the strong
551 similarity of the spatial patterns of the leading SST, USTR and 20d MCA modes in each
552 experiment, the coupling strength between these modes is strikingly different depending on
553 the activation of the Indian and Atlantic oceans coupling.

554

555 The SCF/NC/ r for the first SST-USTR MCA mode of FTIC are 74/13/0.76 compared to
556 63/10/0.74 for REF (Table 3). Moreover, this first MCA mode accounts for 39% (7%) of the
557 SST (USTR) variance in FTIC, but for only 30% (6%) in REF. Much stronger results are
558 obtained for both the SCF/NC/ r coupling coefficients and (SST and USTR) explained
559 variances in FTIC-obs consistent with the highly significant increase of ENSO amplitude in
560 this experiment (e.g. see Figs. 4f and 5b). The comparison of the FTAC, FTAC-obs and REF
561 summary statistics gives similar results with a stronger SST-USTR coupling when the
562 Atlantic Ocean is decoupled (Table 3). Thus, the Bjerknes feedback explains a larger fraction
563 of the covariance between SST and the zonal wind stress in the nudged experiments.
564 Similarly, the leading USTR-20d MCA mode captures 61, 73, 68, 75% and 69% of the total
565 SCF in REF, FTIC, FTIC-obs, FTAC and FTAC-obs, respectively (see Table 4). This mode
566 explains also more than 27, 28, 25 and 27% of the total variance of 20d in FTIC, FTIC-obs,
567 FTAC, FTAC-obs, respectively, but only 17% in REF. The NC and r statistics are again
568 stronger in the decoupled experiments. Looking more carefully to the explained variances of
569 the different variables displayed in Tables 3 and 4, an interesting observation is that the
570 increase of explained variances is particularly important for 20d and SST, but less significant
571 for USTR in the decoupled experiments. In the USTR-20d MCA (see Table 4), the explained
572 variance of USTR for FTAC-obs (6.1%) is even less than the one found for REF (6.4%), but
573 the corresponding explained variance of 20d is still much higher in FTAC-obs (27.8%)

574 compared to REF (17.5%). This suggests that the thermocline feedback is particularly
575 efficient when the Indian and Atlantic oceans are decoupled, a finding consistent with the
576 results of Santoso et al. (2012).

577

578 In summary, the overall wind-thermocline-SST feedback in the equatorial Pacific is much
579 more active when the Indian and Atlantic oceans are decoupled. This finding reconfirms that
580 ENSO behavior is different in the nudged experiments.

581

582 **4. ENSO onset phase**

583

584 To further understand possible mechanisms leading to changes in ENSO properties in the
585 Indian and Atlantic oceans decoupled runs, we focus specifically on the ENSO onset phase in
586 this section.

587

588 In REF and observations, weak cold SST anomalies extending from the eastern equatorial
589 Pacific toward the dateline are found 9 months before the ENSO peak, a structure consistent
590 with the decaying phase of La Niña and the fact that ENSO is partly a standing oscillation in
591 the tropical Pacific (Fig. 3). Interestingly, during February-March, the warmest SST
592 anomalies in the Pacific occur off the equator. Simulated Sea Level Pressure (SLP),
593 precipitation and surface wind anomalies during boreal spring regressed onto the Niño-3.4
594 SST during DJ (at the end of year0) are displayed in Figure 8a. These regression patterns
595 suggest a significant connection between El Niño onset and the mid-latitude North Pacific
596 variability in REF, which is consistent with the “Seasonal Footprinting Mechanism” (SFM)
597 and the “Pacific Meridional Mode” (PMM) discussed by Vimont et al. (2003) and Chang et
598 al. (2007), respectively. More specifically, a north-south anomalous SLP dipole is found over
599 the central North Pacific during DJ, which induces a significant weakening of the
600 northeasterly trade winds and the emergence of a boomerang warm SST structure in the
601 tropical North Pacific via changes in wind-induced latent heat flux during the following
602 months (Figs. 3 and 8a, b). These warm SST anomalies intrude into the deep tropics, displace
603 the Inter-Tropical Convergence Zone (ITCZ) northward and promote southwesterly surface
604 wind anomalies in the western equatorial Pacific from late boreal winter to spring of year0
605 (Fig. 8b). These equatorial westerlies are connected to the anomalous westerlies in the central

606 North Pacific subtropics; moreover both are part of a larger cyclonic flow centered in the
607 North Pacific (Fig. 8a). This pattern resembles the anomalous zonal wind field associated
608 with positive phase of the PMM, which is a significant ENSO precursor 7–9 months prior to
609 El Niño events in both observations and SINTEX-F2 (Boschat et al. 2013). They also support
610 the idea that the North PMM is an efficient ENSO trigger via equatorially trapped wave
611 propagation in REF as in observations (Chang et al. 2007).

612

613 This suggests that the dominant non-ENSO variability influencing ENSO onset in REF is the
614 PMM, but this does not rule out the possible impact from the Atlantic or Indian oceans
615 (Boschat et al. 2013). In fact, significant cold SST and positive SLP anomalies are also
616 observed during boreal spring (and before) in both the Indian and Atlantic oceans (Figs. 3 and
617 8a). This SST pattern may result from the previous ENSO phase (Kug and Kang 2006) or,
618 alternatively be linked to subtropical variability independent from ENSO (Terry 2011;
619 Boschat et al. 2013). In order to address the relative importance of these Indian and Atlantic
620 SST signals in ENSO onset, we now focus on the decoupled experiments.

621

622 Figure 9 displays the lag correlations between the DJ Niño-3.4 SSTs (at the end of year0) and
623 the SST fields during year-1 in FTIC and FTAC. First, we note that Indian and Atlantic
624 couplings seem to be critical for the timing of the ENSO onset, since in the decoupled runs
625 this onset occurs several months before compared to REF (e.g. during year-1). Significant
626 positive SST anomalies exist in the equatorial central Pacific as soon as August-September of
627 year-1 in FTIC and FTAC. These warm SST anomalies slowly move eastward and extend
628 meridionally during the following seasons. This illustrates the occurrence of long-lasting El
629 Niño episodes in the decoupled runs, partly due to a very slow developing phase, which last
630 for several months. This is in great contrast with REF in which the El Niño onset is a very fast
631 process and the SST El Niño pattern is fully developed in June-July of the same year (Fig.
632 3a). FTIC-obs and FTAC-obs exhibit a very similar ENSO development, which further attests
633 of the robust impact of Indian and Atlantic coupling on the ENSO onset and its timing
634 (figures not shown).

635

636 We now illustrate how ENSO is initiated in the FTIC and FTAC experiments with the help of
637 Figs. 10 and 11. During February-March of year-1, the correlation patterns suggest the
638 existence of La Niña conditions in the tropical Pacific with significant cold (warm) anomalies

639 in the eastern (western) equatorial Pacific, a positive phase of the Southern Oscillation and
640 enhanced easterlies over the equatorial Pacific (Figs. 9, 10 and 11). Consistent with these
641 zonal wind anomalies, there is a strong zonal contrast in the heat content over the tropical
642 Pacific basin during February-March of year-1 and this positive heat content in the western
643 Pacific seems to be a robust and significant precursor of El Niño nearly two years in advance
644 in FTIC and FTAC (Fig. 6).

645

646 During April-May, this La Niña-like pattern fades away in both decoupled experiments. It is
647 interesting to observe that there are no clear links during boreal spring of year-1 (or year0)
648 between the cyclonic flow in the North Pacific, the westerly wind anomalies over the western
649 equatorial Pacific and the amplitude of the ongoing El Niño event, which will peak at the end
650 of year0 (Figs. 10 and 11). This is in sharp contrast with the evolution in REF where the SFM
651 over the North Pacific plays a key role in El Niño onset during year0 (Fig. 8). Closer
652 inspection reveals that westerly wind anomalies appear over the western equatorial Pacific
653 from boreal summer of year-1 in association with El Niño onset in the decoupled runs (Figs.
654 10 and 11). But these wind anomalies evolve independently of the North Pacific extra-tropical
655 forcing and are not related to the SFM in both FTIC and FTAC, contrary to what is observed
656 in REF.

657

658 We first focus on the possible origin of the persistent positive westerly wind signal from June
659 of year-1 onward in the FTIC experiment (Fig. 10). This wind variability over the far western
660 Pacific is significantly associated with SST anomalies over the Western North Pacific (WNP),
661 namely positive SST anomalies in the western and central equatorial Pacific and negative SST
662 anomalies over the northwest Pacific from June of year-1 onward in the FTIC experiment (see
663 Fig. 9a). A similar SST pattern is seen in FTIC-obs (figure not shown). This suggests that this
664 wind variability over the western equatorial Pacific is mainly driven by the local SST related
665 to the previous La Niña episode (Weisberg and Wang 1997). However, this does not exclude
666 the role of other remote factors such as the mid-latitude atmospheric variability over the South
667 Pacific or even the cold SST and positive SLP anomalies over the tropical Atlantic during
668 boreal summer of year-1 since these two other signals exhibit a highly significant statistical
669 association with the amplitude of the ongoing El Niño event (Fig. 10). We investigate these
670 aspects next.

671

672 We first examine the potential role of the South Pacific atmospheric variability on the ENSO
673 development in FTIC. As displayed in Fig. 10, an expanded trough emerges over the South
674 Pacific, extends into the deep tropics and promotes westerly wind anomalies over the western
675 equatorial Pacific during the austral winter of year-1 in FTIC. This mid-latitude South Pacific
676 SLP variability is reminiscent of the first Pacific-South American (PSA) pattern, which is
677 dominant during the developing phase of ENSO in observations (Jin and Kirtman 2009;
678 Terray 2011; Ding et al. 2014). Ding et al. (2014) suggest that this PSA pattern imparts a
679 structure of SST anomalies over the South Pacific analogous to the PMM in the Northern
680 Hemisphere through the associated wind and latent heat anomalies. This South PMM is, in
681 turn, able to force zonal wind anomalies along the equator and has an influence on ENSO
682 development consistent with the SFM hypothesis for the Southern Hemisphere. Interestingly,
683 the SST correlation patterns during boreal summer of year-1 in FTIC experiment (see Fig. 9b)
684 are reminiscent of this South PMM. This supports the idea that the South Pacific variability
685 during austral winter may also play a role in the El Niño onset in FTIC.

686

687 However, this does not exclude other important contributing factors such as the cold tropical
688 Atlantic SSTs during boreal summer of year-1 through a modulation of the Atlantic-Pacific
689 Walker circulation (Ding et al. 2012; Polo et al. 2014). From April to July of year-1, cold SST
690 anomalies cover nearly the whole Atlantic basin in the FTIC experiment (Fig. 9a). Associated
691 with these cold Atlantic SSTs there are local positive SLP anomalies (and also negative
692 rainfall anomalies) over the equatorial Atlantic (Fig. 10). The corresponding diabatic cooling
693 modulates the Walker circulation by inducing convergent (divergent) motion over the tropical
694 Atlantic (western and central Pacific) at 200 hPa (figure not shown). This is in agreement with
695 the high (low) SLP anomalies in the tropical Atlantic (western-central Pacific) during boreal
696 summer of year-1 (Fig. 10). The induced low pressure and ascending motion anomalies over
697 the western Pacific may promote convective activity and westerly wind stress anomalies there
698 (Fig. 10). This is consistent with the mechanism described by Fonseca et al. (2009), Ding et
699 al. (2012) or Polo et al. (2014).

700

701 The main point is that all these different factors may be interrelated and collectively promote
702 the generation of persistent low-level westerlies over the far western equatorial Pacific from
703 June of year-1 onward, which may be responsible for the ENSO onset in FTIC experiment
704 since these low-level wind anomalies over the western equatorial Pacific are optimally

705 situated to influence the generation of eastward propagating oceanic Kelvin waves (Kug et al.
706 2010). Forced by this westerly wind stress, the equatorial Pacific thermocline slope is
707 decreased, with shoaling and deepening in the west and east, respectively, during year-1 (see
708 Fig. 6). At the end of year-1, these wind stress, thermocline and SST anomalies are further
709 amplified by the Bjerknes positive feedback, sustaining the development of El Niño-like
710 anomalies during year0 in FTIC.

711

712 In the Atlantic decoupled experiments, the persistent westerly wind signal over the western
713 equatorial Pacific is seen from August of year-1 onward, e.g. 2 months later than in the Indian
714 decoupled runs (Figs. 10 and 11). However, again these wind changes and their timings are
715 critical for the simulated ENSO variability in FTAC (Fig. 9b). As a result, the western Pacific
716 positive heat content anomalies migrate to the east during boreal fall and cover the whole
717 equatorial Pacific at the end of year-1 in FTAC (Fig. 6). This may trigger the ENSO onset in
718 the Atlantic decoupled runs, consistent with SST correlation patterns during year-1 displayed
719 in Fig. 9b. A closer examination of Figure 9b points again to the key-role of a dipole of SST
720 anomalies in the WNP, with negative SST anomalies over the northwest Pacific and positive
721 SST anomalies in the western equatorial Pacific, for sustaining the local low-level cyclonic
722 circulation from June onward (Weisberg and Wang 1997; Wang et al. 2012). This dipole of
723 SST anomalies has a very similar pattern in the Indian and Atlantic decoupled runs, but its
724 timing is different with a lag of two months in the Atlantic decoupled runs. Since the westerly
725 wind anomalies also appear over the western equatorial Pacific with a lag of two months in
726 the Atlantic decoupled runs, this strongly supports the idea that the SST dipole over the WNP
727 is responsible for the emergence of the westerly zonal wind anomalies through coupled local
728 interactions in the different decoupled runs (Weisberg and Wang 1997; Wang et al. 2012).

729

730 The North PMM plays a vital role in ENSO onset during year0 in the REF simulation. In this
731 section, we have presented evidences that strongly support the idea that Indian and Atlantic
732 SST variability is essential for this relationship between the PMM and ENSO onset in REF
733 since ENSO onset occurs in year-1 in all the decoupled experiments and are no more related
734 to the North PMM during boreal spring.

735

736 **5. ENSO decaying phase**

737

738 We now focus on the role of Indian and Atlantic SSTs during the decaying phase of ENSO.
739 Since the results are again similar between FTIC and FTIC-obs (or FTAC and FTAC-obs), we
740 will present only the results from FTIC and FTAC experiments.

741

742 Figure 12 displays the lag-regressions of bi-monthly SST fields from April to November of
743 year+1 with the DJ Niño-3.4 SST time series during the previous boreal winter. First, we find
744 that the ENSO-related SST anomalies in the equatorial Pacific last longer in the decoupled
745 experiments. This is consistent with the time evolution of 20d shown in Fig. 6, as well as with
746 previous studies (Kug and Kang 2006; Dommenges et al. 2006; Frauen and Dommenges
747 2012; Santoso et al. 2012). In REF, negative SST anomalies appear first over the central
748 equatorial Pacific in June-July of year+1 and persist afterward (Fig. 12). By contrast, high
749 positive SST anomalies are noted in the same area and period in FTIC, while nearly
750 climatological SST values are simulated in the central equatorial Pacific from June-July of
751 year+1 onward in FTAC. Thus, the SST evolution in FTAC is intermediate between the two
752 contrasting SST evolutions in REF and FTIC simulations. This suggests that Indian Ocean
753 SSTs may have a stronger impact on ENSO during its decaying phase. The warm ENSO-
754 related SST anomalies over the tropical Atlantic (Indian) Ocean from April-May to October-
755 November of year+1 have the same amplitude or are slightly stronger in the FTIC (FTAC)
756 compared to REF, especially at the end of year+1. However, it is intriguing to observe that
757 this persisting SST variability in the Atlantic or Indian basins in FTIC and FTAC runs,
758 respectively, are not sufficient to lead to cold SST anomalies in the Pacific and a rapid demise
759 of the ENSO events as in REF (Fig. 12).

760

761 To understand better why the equatorial Pacific SST anomalies last longer after the ENSO
762 peak in the decoupled experiments, the evolution of bi-monthly mean 10-m zonal wind
763 anomalies obtained through regression onto DJ Niño-3.4 SST in the control and decoupled
764 runs are shown in Figure 13.

765

766 During April-May of year+1, the regressed surface zonal wind patterns in all the simulations
767 are very similar with enhanced easterly (westerly) zonal wind anomalies over the western
768 (central and eastern) equatorial Pacific. However, from June-July onwards, the western

769 Pacific-enhanced easterly anomalies fade away in the decoupled runs, but persist in REF,
770 which ultimately lead to the rapid demise of eastern Pacific warm heat content anomalies and
771 a strong damping of the equatorial Pacific SST anomalies in REF (Figs. 6 and 12). These
772 anomalous easterlies over the western equatorial Pacific can induce an upwelling thermocline
773 anomaly propagating eastward that fastens the transition from El Niño to La Niña in REF
774 (Kug and Kang 2006). In contrast, the transition to a normal Pacific state is further delayed by
775 several months in both the FTIC and FTAC experiments, as westerly wind anomalies are
776 stronger, last longer over most of the west and central equatorial Pacific and the thermocline
777 remains deep in the eastern Pacific (Figs. 6 and 13). This is consistent with the enhanced
778 Bjerknes feedback in the decoupled runs (Tables 3 and 4). In other words, the 20d, SST and
779 surface zonal wind regression maps clearly show the role of the western Pacific wind
780 anomalies in hastening the transition from El Niño to La Niña in REF compared to the
781 nudged experiments.

782

783 Interestingly, despite of the absence of SST variability, strong and persistent equatorial
784 easterlies are also found from April to November of year+1 over the nudged region in both
785 FTIC and FTAC (Fig. 13). All these features are in accordance with the slowdown of the
786 Walker circulation during a long lasting El Niño when the convection and the uprising branch
787 of the Pacific Walker circulation are shifted to the east during an extended period. Also, in
788 agreement with the anomalous SST equatorial gradient in the Atlantic Ocean from June to
789 November of year+1, enhanced surface easterly wind anomalies are simulated over this basin
790 in FTIC (see Figs 12 and 13).

791

792 The 200 hPa wind and velocity potential regression maps during year+1 onto the Niño-3.4
793 SST index demonstrate that the above features are related to important changes in the Walker
794 circulation over the Atlantic and Indian oceans in FTIC and FTAC, respectively (Fig. 14).
795 Negative rainfall anomalies or much weaker positive anomalies are observed over most of the
796 nudged region in FTIC and FTAC experiments compared to REF (figure not shown). Positive
797 rainfall anomalies are restricted to a rather small western equatorial area inside of the nudged
798 region consistent with the enhanced equatorial easterlies simulated near the surface (Fig. 13).
799 Thus, over the nudged region in FTIC and FTAC, the absence of SST variations induces
800 subsidence in the atmospheric column, enhanced convergence in the upper atmosphere and
801 strong westerly anomalies at 200 hPa during year+1.

802

803 In turn, the velocity potential regression maps demonstrate that the western Pacific low-level
804 wind changes are related to modulations of Pacific Walker circulation induced by the absence
805 of the Indian or Atlantic SST anomalies in the nudged runs. Strong positive (negative)
806 velocity potential anomalies are found at 200 hPa over the Indian-Maritime Continent
807 (Pacific) region from late spring to early boreal winter in FTIC and FTAC (Fig. 14). The 200
808 hPa wind anomalies are easterly over the central equatorial Pacific, especially in FTIC. In
809 REF, these upper velocity potential and wind anomalies are much weaker, quickly fade away
810 after April-May of year+1 and are replaced by negative (positive) velocity potential
811 anomalies over the Maritime Continent (central Pacific). This suggests again a fast demise of
812 the El Niño conditions and a return to normal conditions from late boreal summer or fall of
813 year+1 in REF.

814

815 In other words, the positive Indian or Atlantic SSTs induce anomalous heating in the
816 atmosphere through local surface heat fluxes and rainfall anomalies. The anomalous heating
817 leads to a modulation of the Walker circulation (Fig. 14). These circulation changes in turn
818 cause surface wind anomalies in the far western Pacific, which are responsible for a fast El
819 Niño to La Niña transition as described above. On the other hand, the low-level wind
820 anomalies over the central and eastern equatorial Pacific remain eastward from early spring to
821 fall of year+ 1 in the different simulations and cannot explain this fast demise of El Niño in
822 REF compared to the decoupled experiments (Fig. 13). This demonstrates the existence of a
823 very close relationship between the sign and amplitude of the zonal low-level wind over the
824 far western Pacific and the time required for a transition from the warm phase to the cold
825 phase of ENSO through the generation of oceanic-upwelling Kelvin waves that change the
826 oceanic thermocline in the whole equatorial Pacific. Next, the results also indicate that these
827 anomalous easterlies over the far western equatorial Pacific are intimately linked to the
828 existence of upward motion (or positive rainfall) anomalies to the west, especially over the
829 Maritime Continent and eastern Indian Ocean in the different simulations. The different
830 experiments highlight that vertical motion over the eastern Indian Ocean and Maritime
831 Continent is not only dependent on the local SST or Pacific SST, but also on the Atlantic
832 SST.

833

834 The regressed velocity potential fields also demonstrate that the net influence on ENSO of
835 either the Indian and Atlantic SST variability is strongly dependent of what happens in the
836 other tropical basin (Fig. 14). As a first illustration, in FTIC, significant warm SST anomalies
837 and negative upper level velocity potential anomalies are simulated over much of the tropical
838 Atlantic and are stronger than in REF (Figs. 12 and 14). Despite of these highly significant
839 anomalies, the transition from El Niño to La Niña is delayed by several months in FTIC
840 compared to REF due to the exclusion of the Indian Ocean SST variability. In FTIC, there is
841 no competition between the upper level divergence centers over the central Pacific and
842 Atlantic oceans, both act in concert and promote a strong upper-level convergence into the
843 Indian Ocean, inducing subsidence and inhibiting the local rainfall over this region and the
844 Maritime Continent. This probably explains the very weak easterlies over the far western
845 equatorial Pacific in FTIC experiment despite of the warm Atlantic SST anomalies.

846

847 In a similar way, in FTAC, significant warm SST anomalies are found over the tropical Indian
848 Ocean from early boreal spring to late fall of year +1 (Fig. 12). Despite of these warm SSTs,
849 positive upper level velocity potential anomalies and negative rainfall anomalies cover the
850 whole Indian domain, due to the subsidence aloft directly or indirectly induced by the absence
851 of Atlantic SST variability in FTAC (Fig. 14). This is consistent with the reduced surface
852 easterlies over the western equatorial Pacific during the decaying ENSO phase in FTAC (Fig.
853 13). The direct effect is linked to the fact that the atmospheric response due to the absence of
854 SST variability over the Atlantic may extend eastward in the Indian Ocean (Kucharski et al.
855 2007, 2008; Losada et al. 2010). The indirect effect is associated with the fact that exclusion
856 of the Atlantic SST variability may affect ENSO through the Pacific-Atlantic Walker
857 circulation (Rodriguez-Fonseca et al. 2009; Ding et al. 2012; Frauen and Dommenges 2013)
858 and the resulting long lasting El Niño event will promote increased subsidence over the
859 tropical Indian ocean through the atmospheric bridge (Alexander et al. 2002). However, it is
860 difficult without further numerical investigations to determine if this reduced rainfall signal
861 over the Indian Ocean is related to the subsidence induced directly by the atmospheric
862 response over the Atlantic basin to the west or indirectly by the persistent El Niño conditions
863 over the tropical Pacific, which are related to the modulation of the Atlantic-Pacific Walker
864 circulation to the east.

865

866 **6. Conclusions and Discussion**

867

868 There are strong evidences of interactions between Indian, Atlantic and Pacific oceans and
869 recent studies suggest an active role of the Indian and Atlantic basins on ENSO (Kug and
870 Kang 2006; Rodriguez-Fonseca et al. 2009; Izumo et al. 2010; Luo et al. 2010; Ding et al.
871 2012; Frauen and Dommenges 2013; Boschat et al. 2013). However, the mechanisms
872 underlying these relations are not fully understood and the relative impacts of these two
873 basins remain unclear.

874

875 Thus, to better understand how the Indian and Atlantic oceans impact ENSO, we have
876 performed several long sensitivity experiments using a CGCM, which realistically simulates
877 many facets of ENSO. For each experiment, we suppressed the SST variability in either the
878 Indian or Atlantic oceans by applying a strong nudging of the SST toward a SST climatology
879 computed from a control experiment or observations. This experimental framework is the
880 same as in Frauen and Dommenges (2012), but we used a fully global CGCM, while they
881 used a hybrid model. This hybrid model does not take into account the ocean dynamics
882 outside the Pacific region and used heat flux corrections to maintain a reasonable SST
883 climatology outside the tropical Pacific. Thus, our work is a logical extension of the analysis
884 of Frauen and Dommenges (2012) since our study takes fully into account the possible
885 existence of intrinsic coupled ocean-atmosphere modes of variability in the Atlantic or Indian
886 basins, such as the IOD or the Atlantic zonal mode, which interact with ENSO (Fischer et al.
887 2005; Rodriguez-Fonseca et al. 2009; Luo et al. 2010; Ding et al. 2012) and were also
888 suggested as important ENSO precursors in recent studies (Izumo et al. 2010; Keenlyside et
889 al. 2013).

890

891 The results first suggest that both the Indian and Atlantic SSTs have a significant damping
892 effect on ENSO amplitude. The increase of ENSO variability is particularly significant when
893 the Indian Ocean is decoupled. The dominant periodicities of the simulated ENSO cycle
894 increase from about 2-5 yr in the control run to about 3-12 yr in the Indian or Atlantic
895 decoupled runs. This shift to lower frequencies is more prominent in the Atlantic decoupled
896 runs. With the help of different MCAs, we also demonstrate that the Bjerknes and
897 thermocline feedbacks are more active and are partly responsible of the ENSO changes in the
898 decoupled runs. These results are consistent with recent studies using other CGCMs

899 (Dommenget et al. 2006; Santoso et al. 2012) or hybrid models (Frauen and Dommenget
900 2012).

901

902 Surprisingly, the nudged experiments toward an observed SST climatology demonstrate that
903 the simulated Pacific equatorial SST gradient, seasonal cycle and seasonal phase locking of
904 ENSO are significantly improved when SST biases are removed in the Indian or Atlantic
905 oceans. One may thus conjecture that improving model skills in simulating Indian and
906 Atlantic oceans SST climatology may also lead to improved skill in simulating and predicting
907 ENSO with coupled models, because they are important for a realistic ENSO seasonal phase-
908 locking and simulation of the low-level wind variability over the western equatorial Pacific
909 during boreal spring.

910 On the other hand, the Pacific mean state is not modified if we used the SST climatology
911 from the control run in the decoupled experiments. This demonstrates first that the correction
912 of the Pacific equatorial SST gradient discussed above is directly linked to the mean SST
913 biases in the Atlantic and Indian oceans and second that any changes in ENSO variability in
914 the decoupled runs using a simulated climatology cannot be explained by a rectification of the
915 Pacific mean state. This strongly supports the hypothesis that both the Indian and Atlantic
916 oceans interplay with the ENSO dynamics (Kug and Kang 2006; Rodriguez-Fonseca et al.
917 2009; Izumo et al. 2010; Terray 2011; Santoso et al. 2012; Ding et al. 2012; Ham et al. 2013a,
918 b).

919

920 Many of the previous studies focused on the influence of Indian and Atlantic SST variability
921 on ENSO during the peak and decaying phases of El Niño. Moreover, to a large extent, they
922 almost all agreed that the physical processes that allow the Indian or Atlantic oceans to
923 influence ENSO, involve modulations of the Walker circulation and of the surface winds in
924 the western equatorial Pacific, which trigger eastward-propagating oceanic Kelvin waves
925 responsible for the turnabout of ENSO. However, the Indian and Atlantic basins may also
926 play a key-role in triggering ENSO events (Izumo et al. 2010; Terray 2011; Boschat et al.
927 2013; Frauen and Dommenget 2012; Ham et al. 2013a, b). Our study confirms that Indian or
928 Atlantic SSTs influence significantly both the onset and decaying stages of ENSO events.

929

930 During the decaying phase of ENSO, Indian and Atlantic SST anomalies, both in observations
931 and the control simulation, are largely forced by ENSO, and the processes we have detailed
932 during this phase, may be best described as a negative feedback on ENSO (Kug and Kang
933 2006; Ohba and Ueda 2007; Santoso et al. 2012; Dayan et al. 2014). The large agreement
934 between our CGCM results and the results of Frauen and Dommenges (2012), who used a
935 hybrid model excluding ocean dynamics outside the tropical Pacific, are consistent with this
936 interpretation. An important point of our study is however that Atlantic and Indian oceans act
937 together, and SST anomalies in one basin alone may not be the only factor controlling the
938 duration of the ENSO decaying phase. As an illustration, the strong warm anomalies over the
939 tropical Atlantic Ocean simulated during year+1 in the FTIC experiment are not sufficient to
940 fasten the El Niño to La Niña transition in the absence of Indian Ocean SST variability. This
941 is further confirmed by the results of the FTAC experiment in which the absence of Atlantic
942 SST variability counteracts the expected effects of the warm Indian Ocean SSTs on ENSO
943 during year +1.

944

945 On the other hand, the changes of ENSO evolution during the onset phase in the nudged
946 experiments strongly support the hypothesis that intrinsic SST variability over the Indian and
947 Atlantic oceans does also matter for ENSO. First, the North PMM plays a vital role in the fast
948 ENSO onset during year0 in the REF simulation as in observations. This is consistent with
949 many previous studies, which have popularized SST anomalies over the North Pacific and the
950 SFM, as a key process for the development and predictability of ENSO (Vimont et al. 2003;
951 Chang et al. 2007; Boschat et al. 2013). However, our decoupled experiments demonstrate
952 that Indian and Atlantic SSTs are instrumental for this relationship between the North PMM
953 and ENSO onset in REF since the timing of ENSO onset is advanced by about 1 year and
954 ENSO has no more links with the North Pacific variability in these simulations. This supports
955 the idea that the occurrence of western Pacific wind stress anomalies and the stochastic
956 forcing of ENSO is strongly dependent on the Indian and Atlantic SSTs, a result, which is not
957 obvious from observations or a simple analysis of the control run.

958

959 All the Indian and Atlantic decoupled runs present the evidences that ENSO is triggered in
960 year-1, i.e. more than 1 year in advance than its peak phase, which occurs at the end of year0.
961 Decoupling the Indian and Atlantic oceans reinforces the anomalous SST gradient between
962 the subtropical and tropical western Pacific during boreal summer and fall of year-1 and

963 ultimately impacts the low-level circulation over the western Pacific during the second half of
964 year-1. These wind changes and their timing are critical for ENSO variability because the
965 eastern Pacific SST is closely linked to the equatorial western Pacific wind via equatorial
966 Sverdrup balance. As a result, the western Pacific positive heat content anomalies migrate
967 relatively slowly to the east and cover the whole equatorial Pacific at the end of year-1. This
968 triggers the ENSO onset in the Indian and Atlantic decoupled runs.

969

970 Overall from our results, the importance of the Indian Ocean on ENSO seems stronger.
971 However, the major systematic errors of complex models (Richter et al. 2014), including the
972 SINTEX-F2 model used here, in simulating the equatorial Atlantic climate need to be reduced
973 before any definitive conclusion can be reached about the true forcing on ENSO induced by
974 Atlantic SST variability. It should also be noted that the determination of the relative
975 contribution of the subtropical Indian or Atlantic Ocean SST variability to the simulated
976 ENSO cannot be fully addressed by the present simulations and requires additional numerical
977 experiments. This will be addressed in a forthcoming study.

978

979 This study is based on one model and the conclusions can be model dependent. However,
980 since SINTEX-F2 simulates a realistic ENSO variability and performs very well in ENSO
981 forecasting exercises (Luo et al. 2005a), the results reported here are fully consistent with the
982 hypothesis that the skill of ENSO predictions must be higher if the tropical Indian and
983 Atlantic variability were considered. The inter-basin interactions are crucial to improve our
984 understanding and predictability of ENSO because the timing and intensity of the low-level
985 wind anomalies over the far western equatorial Pacific during boreal spring is highly
986 dependent on these inter-basin interactions.

987

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995

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1242 interaction over the North Pacific. *J Clim* 11:2473-2481
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1244

1245 **Figure captions**

1246 **Figure 1:** **a)** SST mean state in REF ($^{\circ}\text{C}$) from 190-year annual mean (years 11-210 of REF),
1247 **b)** SST means difference ($^{\circ}\text{C}$) between REF (years 11-210) and HadISST1.1 dataset (years
1248 1979-2012), **c)** SST standard deviation in REF ($^{\circ}\text{C}$) and **d)** SST standard deviation difference
1249 ($^{\circ}\text{C}$) between REF (years 11-210) and HadISST1.1 dataset (years 1979-2012).

1250

1251 **Figure 2:** **a)** Monthly standard deviations of the Niño-34 SST time series from HadISST1.1
1252 dataset (for the 1950-2012 and 1979-2012 periods) and the five experiments; **b)** Power
1253 spectra of Niño-34 SST anomalies for HadISST1.1 dataset (Rayner et al. 2003) in black, REF
1254 in red, FTIC and FTIC-obs in green, FTAC and FTAC-obs in blue. The bottom axis is the
1255 period (unit: year), the left axis is variance (unit: $^{\circ}\text{C}^2$) and both axes are in logarithm scale.
1256 Dashed black curves show the point-wise 99% confidence limits for the Niño-34 SST
1257 spectrum estimated from the observations. The observed Niño-34 SST spectrum is estimated
1258 from the 1901-2012 period.

1259

1260 **Figure 3:** Lagged correlations between bi-monthly averaged SSTs and the December-January
1261 Niño-3.4 SST for **a)** REF and **b)** HadISST1.1. The correlations are calculated beginning in
1262 February-March of year0, prior to the El Niño onset, and ending in December-January at the
1263 peak season of El Niño events. For observations, the correlations are computed from the
1264 1950-2012 period. Correlations that are above the 90% significance confidence level
1265 according to a phase-scramble bootstrap test (Ebisuzaki 1997) are contoured.

1266

1267 **Figure 4:** Maps of the difference of annual mean SST climatologies between **a)** REF and
1268 FTIC, **b)** REF and FTIC-obs, **c)** REF and FTAC and **d)** REF and FTAC-obs. Units = $^{\circ}\text{C}$. Maps
1269 of the difference of SST standard deviation between **e)** REF and FTIC, **f)** REF and FTIC-obs,
1270 **g)** REF and FTAC and **h)** REF and FTAC-obs. Units = $^{\circ}\text{C}$.

1271

1272 **Figure 5:** Logarithm of the ratio of the power spectra of the Niño-3.4 SST for **a)** REF/FTIC,
1273 **b)** REF/FTIC-obs, **c)** REF/FTAC and **d)** REF/FTAC-obs on a logarithmic scale and point-
1274 wise 90% confidence intervals (blue lines) for the logarithms of the spectral ratios for a
1275 postulated common spectrum in the two experiments.

1276

1277 **Figure 6:** Lead-lag correlations between bi-monthly averaged depth of 20°C isotherm
1278 anomalies (20d) and the December-January (year0) Niño-3.4 SST for REF (left column),
1279 FTIC (middle column) and FTAC (right column). The correlations are shown only for
1280 February-March of year-1, December-January of year-1, December-January of year0 (during
1281 the El Niño peak) and October-November of year+1. Correlations that are above the 90%
1282 significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997)
1283 are underlined.

1284

1285 **Figure 7: a)** and **b)** Maps of the spatial patterns of the leading SST-USTR MCA mode for
1286 REF. **a)** SST homogeneous vectors in °C, CI=0.1°C. **b)** USTR heterogeneous vectors in
1287 N/m², CI = 0.2 N/m². The maps were obtained by regressing the SST and USTR fields upon
1288 the normalized EC time series of SST. The SST and USTR fields are, respectively,
1289 homogenous and heterogeneous covariance patterns following the terminology of Bretherton
1290 et al. (1992). Summary statistics for this mode are given in Table 3. **c)** and **d)** same as **a)** and
1291 **b)**, but for the leading USTR-20d MCA mode for REF. Units for USTR and 20d are in N/m²
1292 and m, respectively. Summary statistics for this mode are given in Table 4.

1293

1294 **Figure 8: a)** Lead regressions between bi-monthly averaged 850 hPa wind and SLP
1295 anomalies and **b)** 10-m wind and precipitation anomalies during El Niño onset (e.g. from
1296 December-January to April-May of year0) and the December-January (at the end of year0)
1297 Niño-3.4 SST for REF. Regressions that are above the 90% significance confidence level
1298 according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined and only the
1299 850 hPa wind anomalies above the 90% significance confidence level are shown.

1300

1301 **Figure 9:** Lagged correlations between bi-monthly averaged SSTs during year-1 and the
1302 December-January (year0) Niño-3.4 SST for **a)** FTIC and **b)** FTAC. The correlations are
1303 calculated beginning in February-March of year-1, prior to the El Niño onset, and ending in
1304 February-March of year0. Correlations that are above the 90% significance confidence level
1305 according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

1306

1307 **Figure 10:** Lagged regressions between bi-monthly averaged 850 hPa wind and SLP
1308 anomalies during year-1 and the December-January (year0) Niño-3.4 SST for FTIC.
1309 Regressions that are above the 90% significance confidence level according to a phase-
1310 scramble bootstrap test (Ebisuzaki 1997) are underlined and only the 850 hPa wind anomalies
1311 above the 90% significance confidence level are shown.

1312

1313 **Figure 11:** Same as Figure 10, but for the FTAC experiment.

1314

1315 **Figure 12:** Lead regressions between bi-monthly averaged SSTs during year+1 and the
1316 December-January (year0) Niño-3.4 SST for **a)** REF, **b)** FTIC and **c)** FTAC. The regressions
1317 are calculated beginning in April-May of year+1, after the El Niño peak phase, and ending in
1318 October-November of year+1. Regressions that are above the 90% significance confidence
1319 level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

1320

1321 **Figure 13:** Same as Figure 12, but for bi-monthly 10m zonal wind anomalies.

1322

1323 **Figure 14:** Same as Figure 12, but for 200 hPa wind and velocity potential anomalies.

1324

1325 **Table captions**

1326

1327 **Table 1:** Summary of the numerical experiments with their main characteristics, including
1328 length, nudging domain and SST climatology used for the nudging in the Indian or Atlantic
1329 oceans decoupled experiments.

1330

1331 **Table 2:** Standard-deviations (in °C) of Niño-3.4 (5°S–5°N, 170–120°W), ATL3 (5°S–5°N,
1332 340–360°E), IOB (20°S–20°N, 40–110°E), and IOD SST indices in observations and REF.
1333 The IOD index is computed as the differences between the SST anomalies in a western
1334 (10°S–10°N, 60–80°E) and eastern (0–10°S, 90–110°E) box in the tropical Indian Ocean. The
1335 different indices are computed as seasonal averages as defined in the first row of the Table.
1336 These seasons correspond to the peak season of the indices in the observations. The statistics
1337 for the observations are derived from the HadISST1.1 dataset. The statistical significance of
1338 the differences between the observed and simulated standard-deviations has been assessed
1339 with the help of a Fisher test. The simulated standard-deviations significant at the 95%
1340 confidence level are in bold.

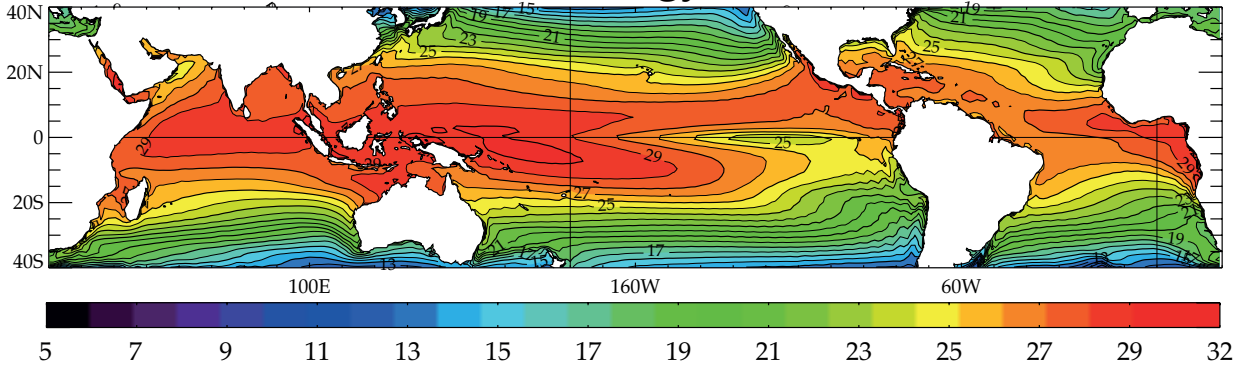
1341

1342 **Table 3:** Summary statistics for the SST-USTR MCA leading modes estimated from the
1343 REF, FTIC, FTIC-obs, FTAC and FTAC-obs experiments, including the SCFs and NCs for
1344 the leading modes in the various MCA expansions and the correlation (r) between the EC
1345 time series of the left and right fields. SCF stands for Square Covariance Fraction and NC for
1346 Normalized root-mean-square Covariance, as given in the text. As discussed by Zhang et al.
1347 (1998), the NC and r coefficients are particularly useful in comparing the strength of the
1348 coupling between the left and right fields in modes obtained from different MCAs. SSTvar
1349 and USTRvar are, respectively, SST and USTR variances accounted for by the leading mode
1350 of each MCA analysis (e.g. this is the variance of the field explained by the related Expansion
1351 Coefficient time series by linear regression).

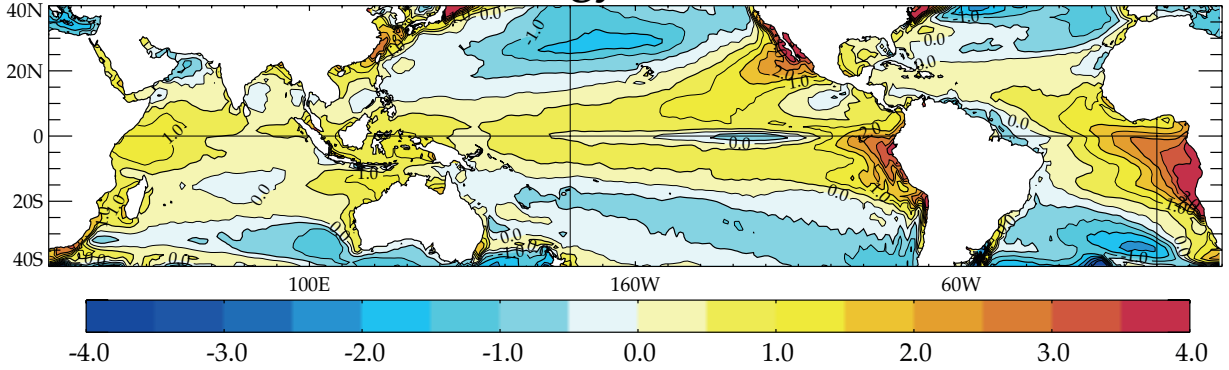
1352 **Table 4:** Same as Table 3, but for the USTR-20d MCA leading modes from the REF, FTIC,
1353 FTIC-obs, FTAC and FTAC-obs experiments.

Figure 1

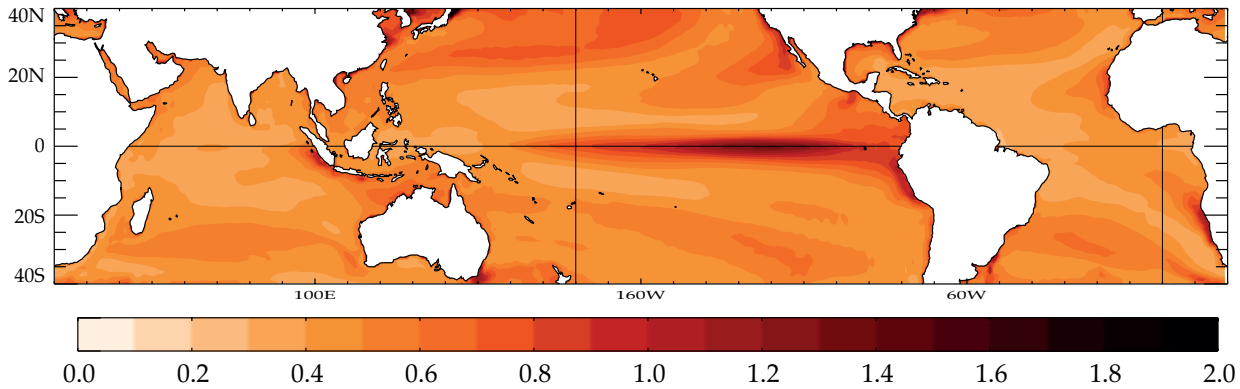
a) SST Climatology - REF (21-210)



b) SST Climatology differences - REF - HadISST



c) SST Standard-Deviation - REF (21-210)



d) SST Standard-Deviation differences REF - HadISST

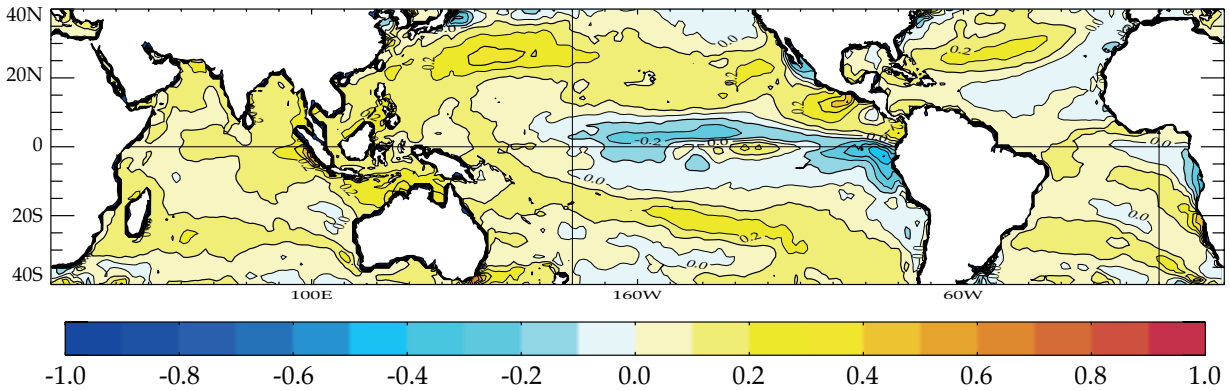


Figure 1: a) SST mean state in REF (°C) from 190-year annual mean (years 11-210 of REF), b) SST means difference (°C) between REF (years 11-210) and HadISST1.1 dataset (years 1979-2012), c) SST standard deviation in REF (°C) and d) SST standard deviation difference (°C) between REF (years 11-210) and HadISST1.1 dataset (years 1979-2012).

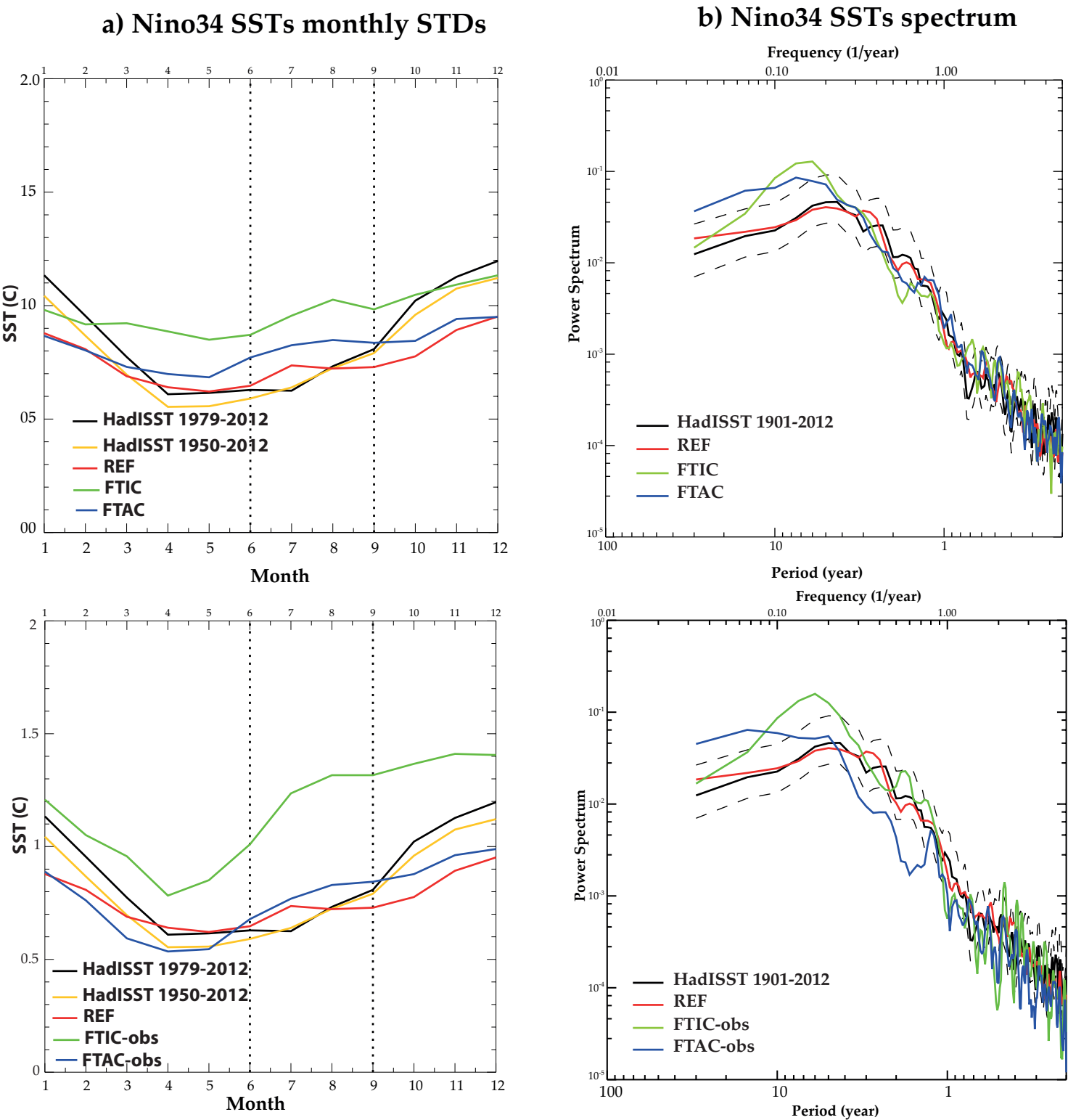
Figure 2

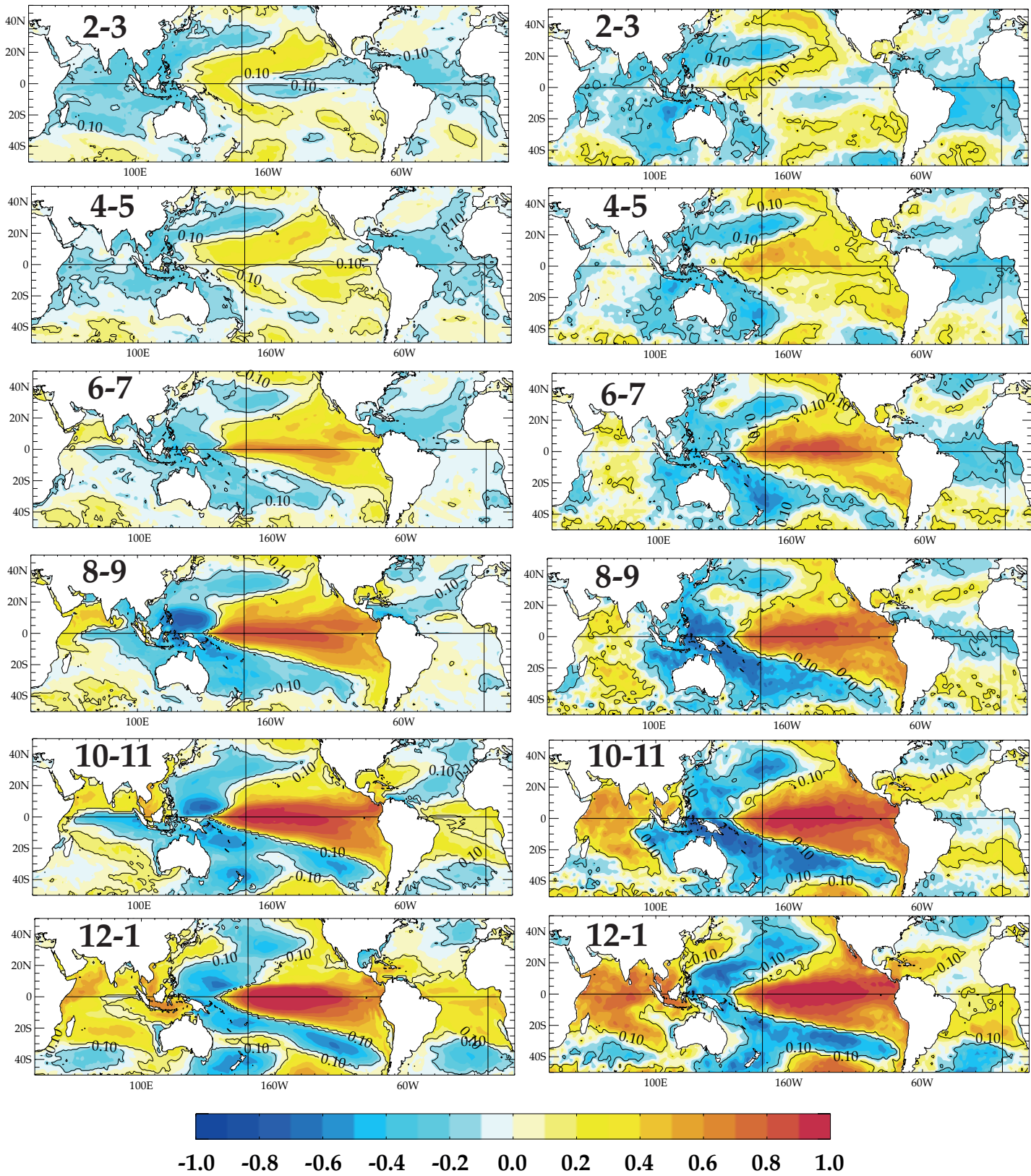
Figure 2: a) Monthly standard deviations of the Niño-34 SST time series from HadISST1.1 dataset (for the 1950-2012 and 1979-2012 periods) and the five experiments; b) Power spectra of Niño-34 SST anomalies for HadISST1.1 dataset (Rayner et al. 2003) in black, REF in red, FTIC and FTIC-obs in green, FTAC and FTAC-obs in blue. The bottom axis is the period (unit: year), the left axis is variance (unit: $^{\circ}\text{C}^2$) and both axes are in logarithm scale. Dashed black curves show the point-wise 99% confidence limits for the Niño-34 SST spectrum estimated from the observations. The observed Niño-34 SST spectrum is estimated from the 1901-2012 period.

Figure 3

Correlations Niño34 SST (12-1) SST - Year 0

a) REF (21-210)

b) HadISST (1950-2012)



-1.0 -0.8 -0.6 -0.4 -0.2 0.0 0.2 0.4 0.6 0.8 1.0

CORRELATION

Figure 3: Lagged correlations between bi-monthly averaged SSTs and the December-January Niño-3.4 SST for a) REF and b) HadISST1.1. The correlations are calculated beginning in February-March of year0, prior to the El Niño onset, and ending in December-January at the peak season of El Niño events. For observations, the correlations are computed from the 1950-2012 period. Correlations that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

Figure 4

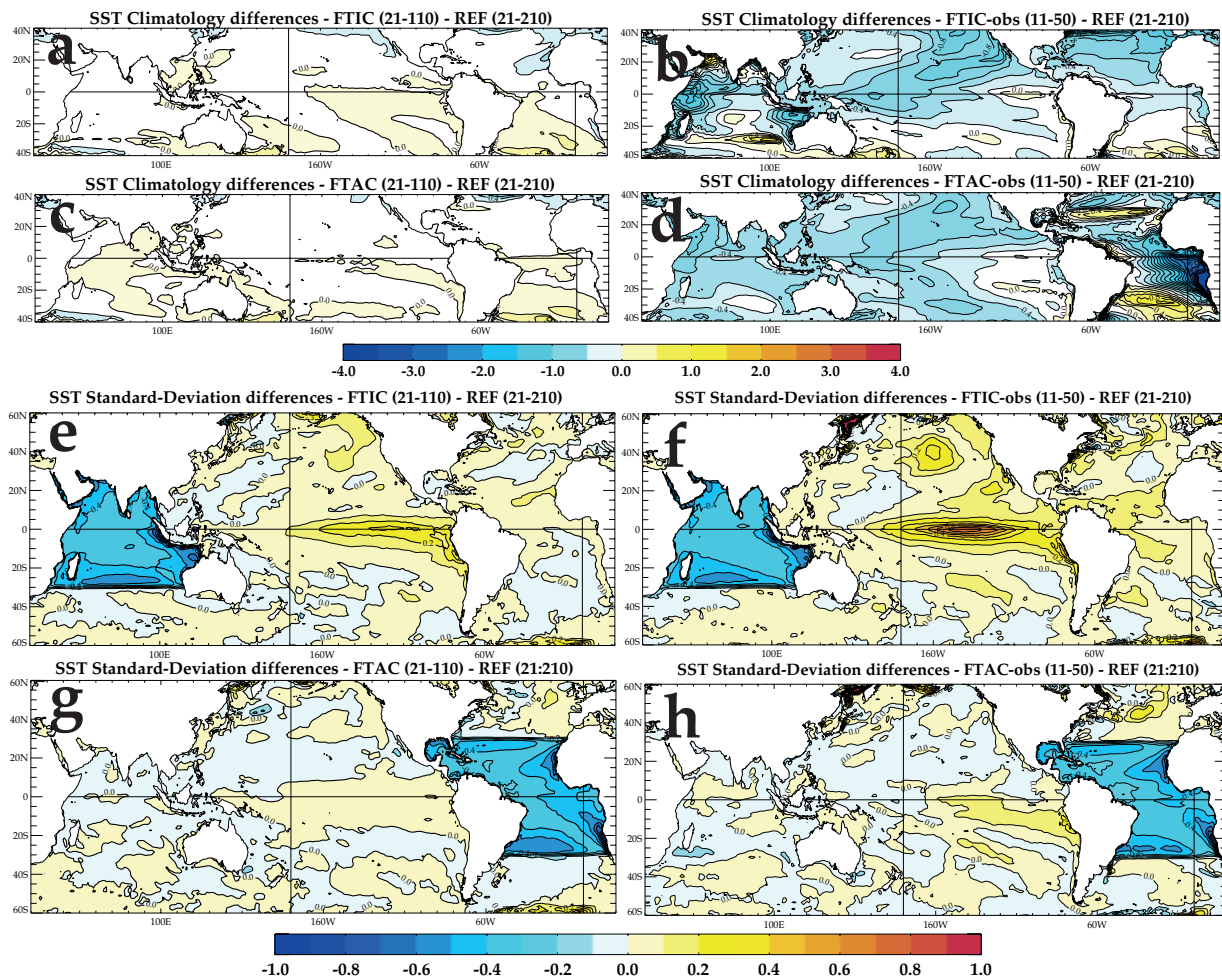


Figure 4: Maps of the difference of annual mean SST climatologies between a) REF and FTIC, b) REF and FTIC-obs, c) REF and FTAC and d) REF and FTAC-obs. Units =°C. Maps of the difference of SST standard deviation between e) REF and FTIC, f) REF and FTIC-obs, g) REF and FTAC and h) REF and FTAC-obs. Units =°C.

Figure 5

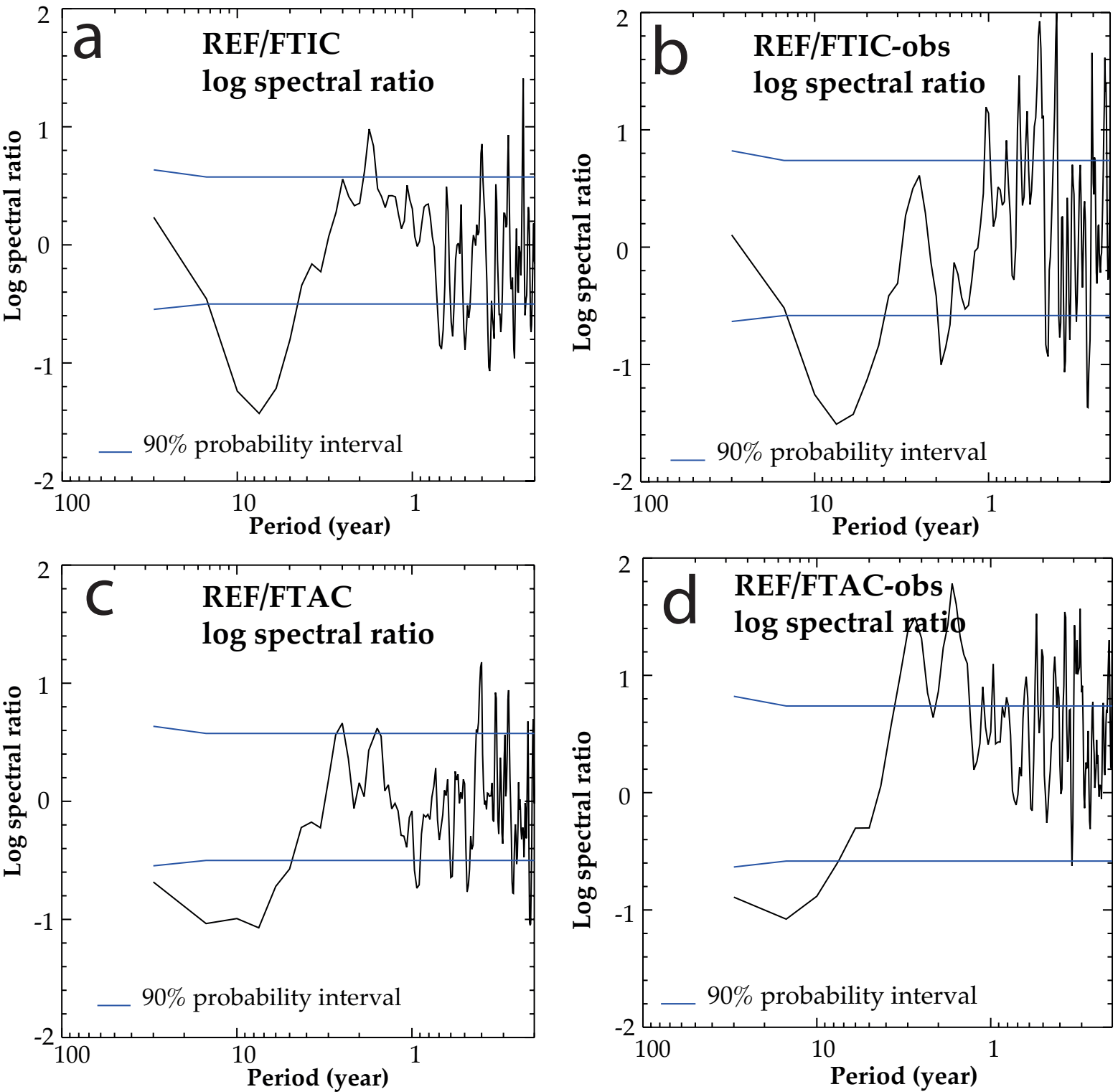


Figure 5: Logarithm of the ratio of the power spectra of the Niño-3.4 SST for a) REF/FTIC, b) REF/FTIC-obs, c) REF/FTAC and d) REF/FTAC-obs on a logarithmic scale and point-wise 90% confidence intervals (blue lines) for the logarithms of the spectral ratios for a postulated common spectrum in the two experiments.

Figure 6

Correlations Nino34 SST (12-1) 20d

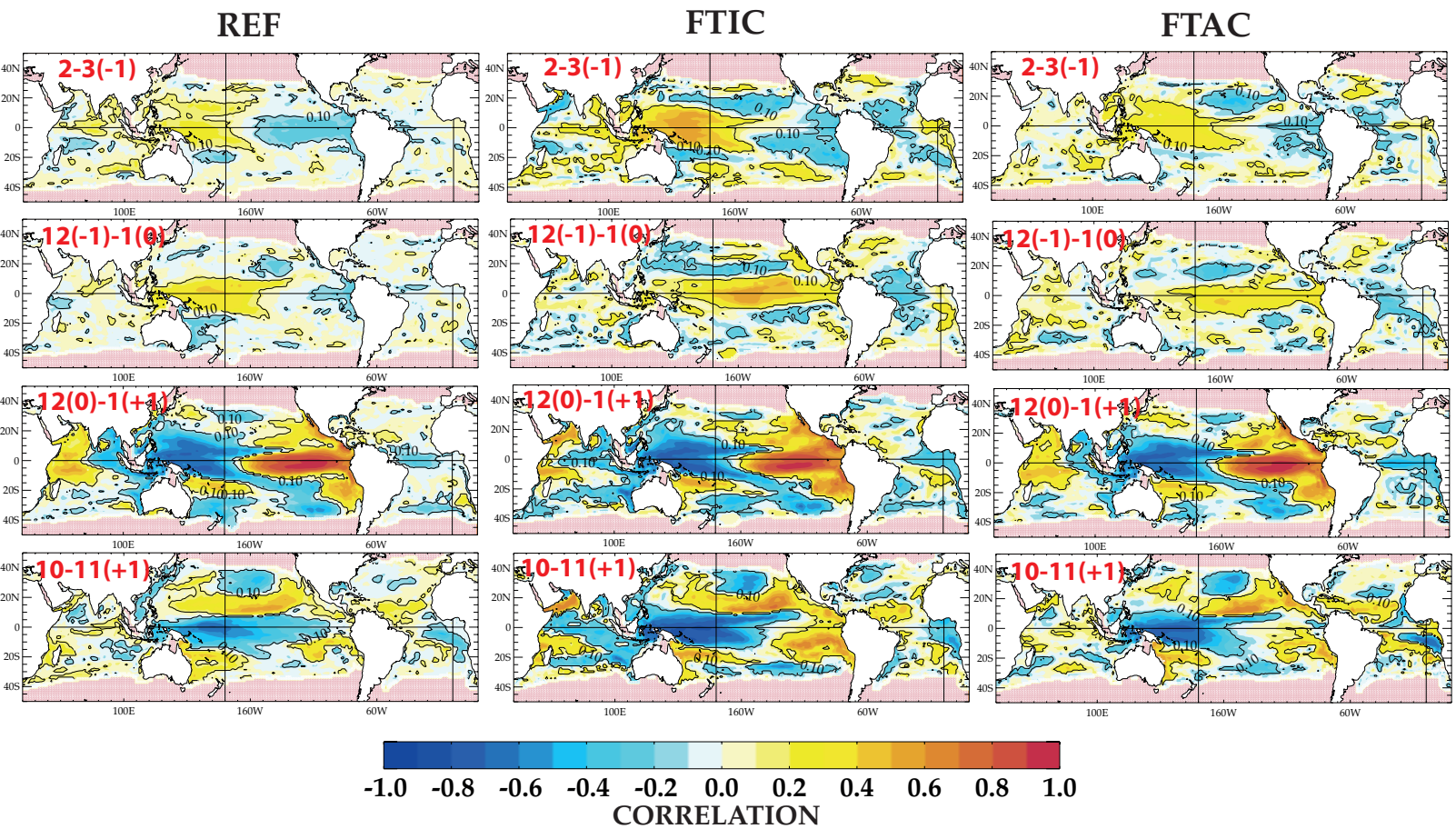


Figure 6: Lead-lag correlations between bi-monthly averaged depth of 20°C isotherm anomalies (20d) and the December-January (year0) Niño-3.4 SST for REF (left column), FTIC (middle column) and FTAC (right column). The correlations are shown only for February-March of year-1, December-January of year-1, December-January of year0 (during the El Niño peak) and October-November of year+1. Correlations that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

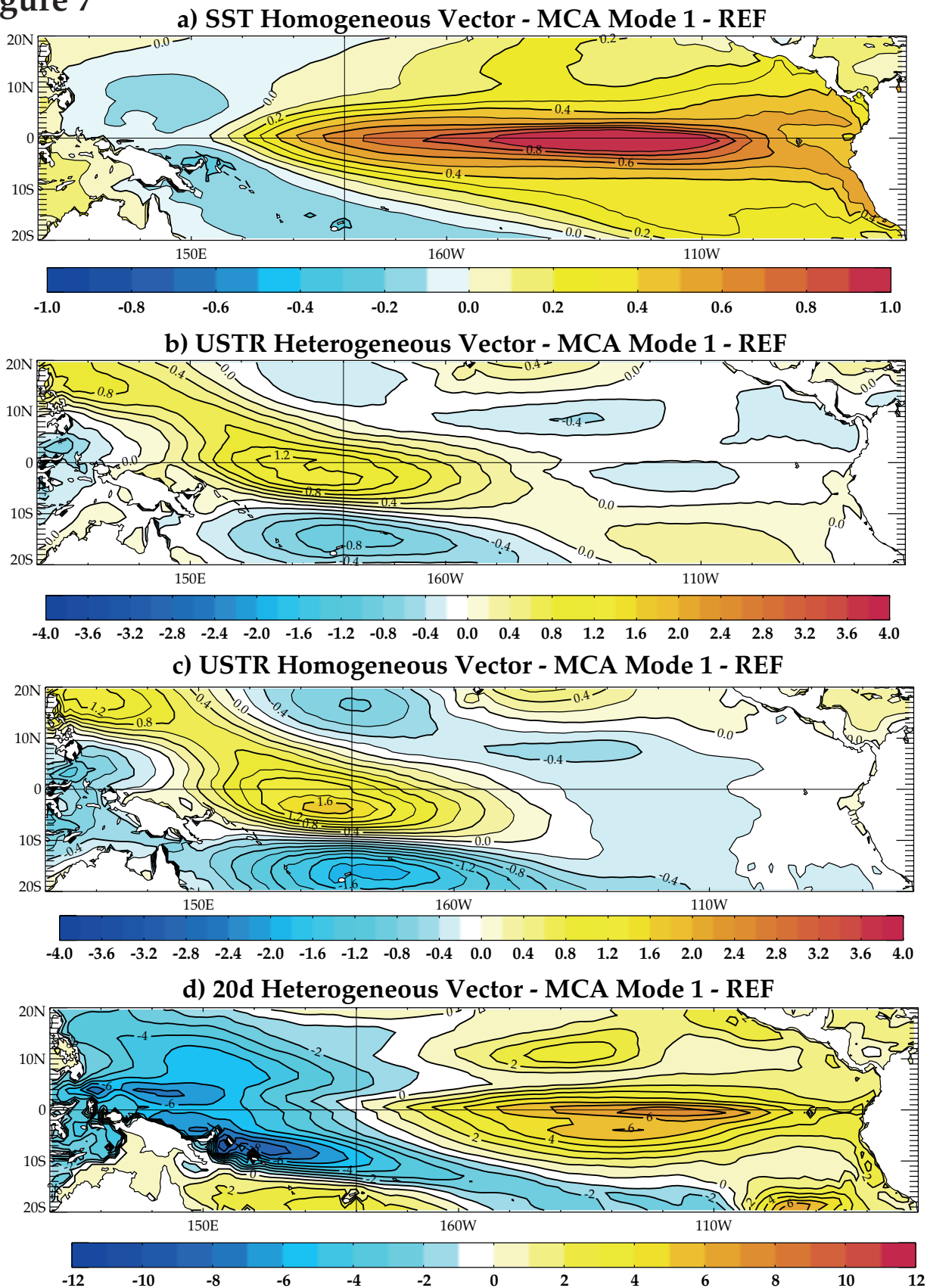
Figure 7

Figure 7: a) and b) Maps of the spatial patterns of the leading SST-USTR MCA mode for REF. a) SST homogeneous vectors in $^{\circ}\text{C}$, $\text{CI}=0.1^{\circ}\text{C}$. b) USTR heterogeneous vectors in N/m^2 , $\text{CI} = 0.2 \text{ N}/\text{m}^2$. The maps were obtained by regressing the SST and USTR fields upon the normalized EC time series of SST. The SST and USTR fields are, respectively, homogenous and heterogeneous covariance patterns following the terminology of Bretherton et al. (1992). Summary statistics for this mode are given in Table 2. c) and d) same as a) and b), but for the leading USTR-20d MCA mode for REF. Units for USTR and 20d are in N/m^2 and m , respectively. Summary statistics for this mode are given in Table 3.

Figure 8

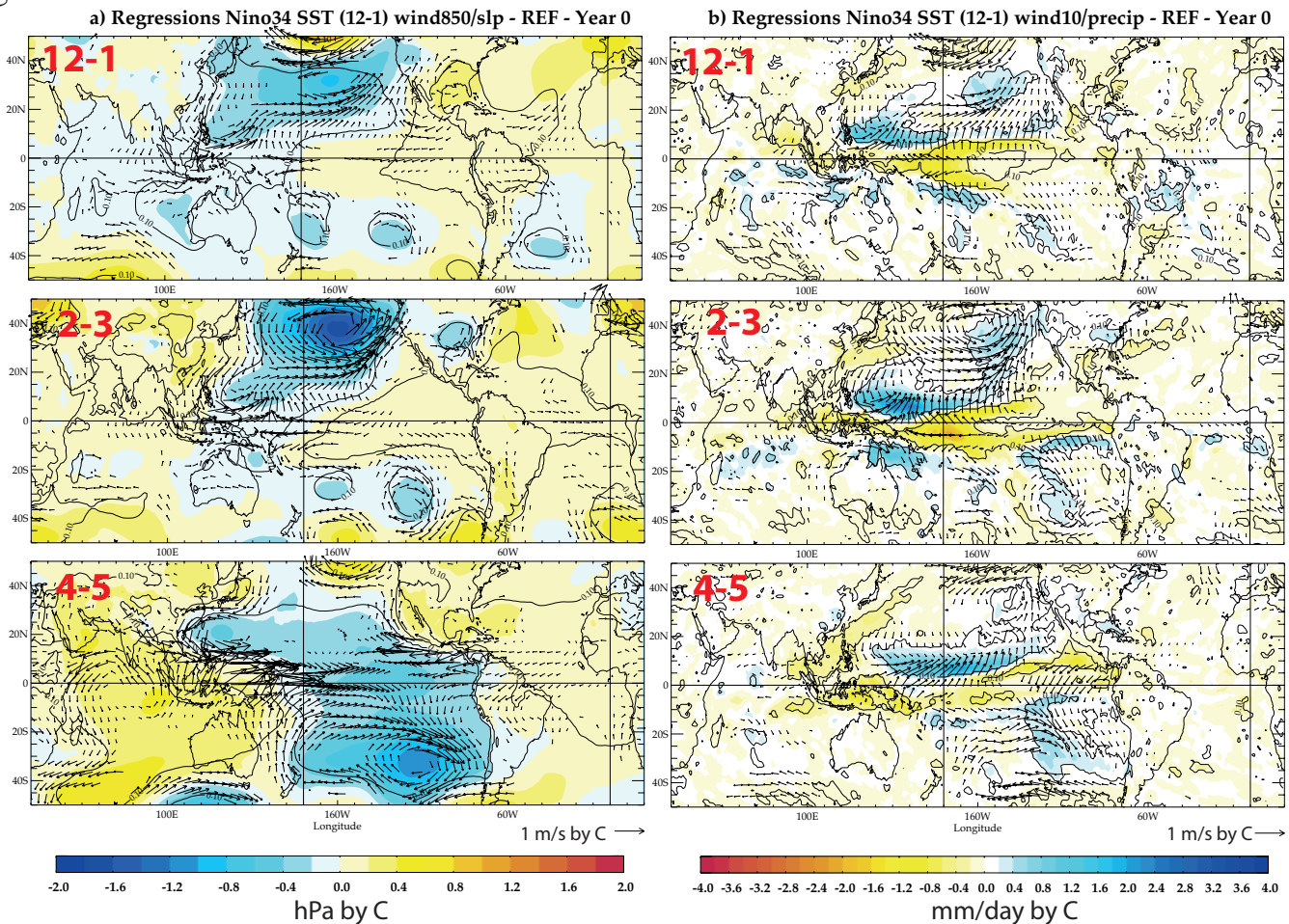


Figure 8: a) Lead regressions between bi-monthly averaged 850 hPa wind and SLP anomalies and b) 10-m wind and precipitation anomalies during El Niño onset (e.g. from December-January to April-May of year0) and the December-January (at the end of year0) Niño-3.4 SST for REF. Regressions that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined and only the 850 hPa wind anomalies above the 90% significance confidence level are shown.

Figure 9

Correlations Niño34 SST (12-1) SST

a) FTIC : YEAR -1

b) FTAC : YEAR -1

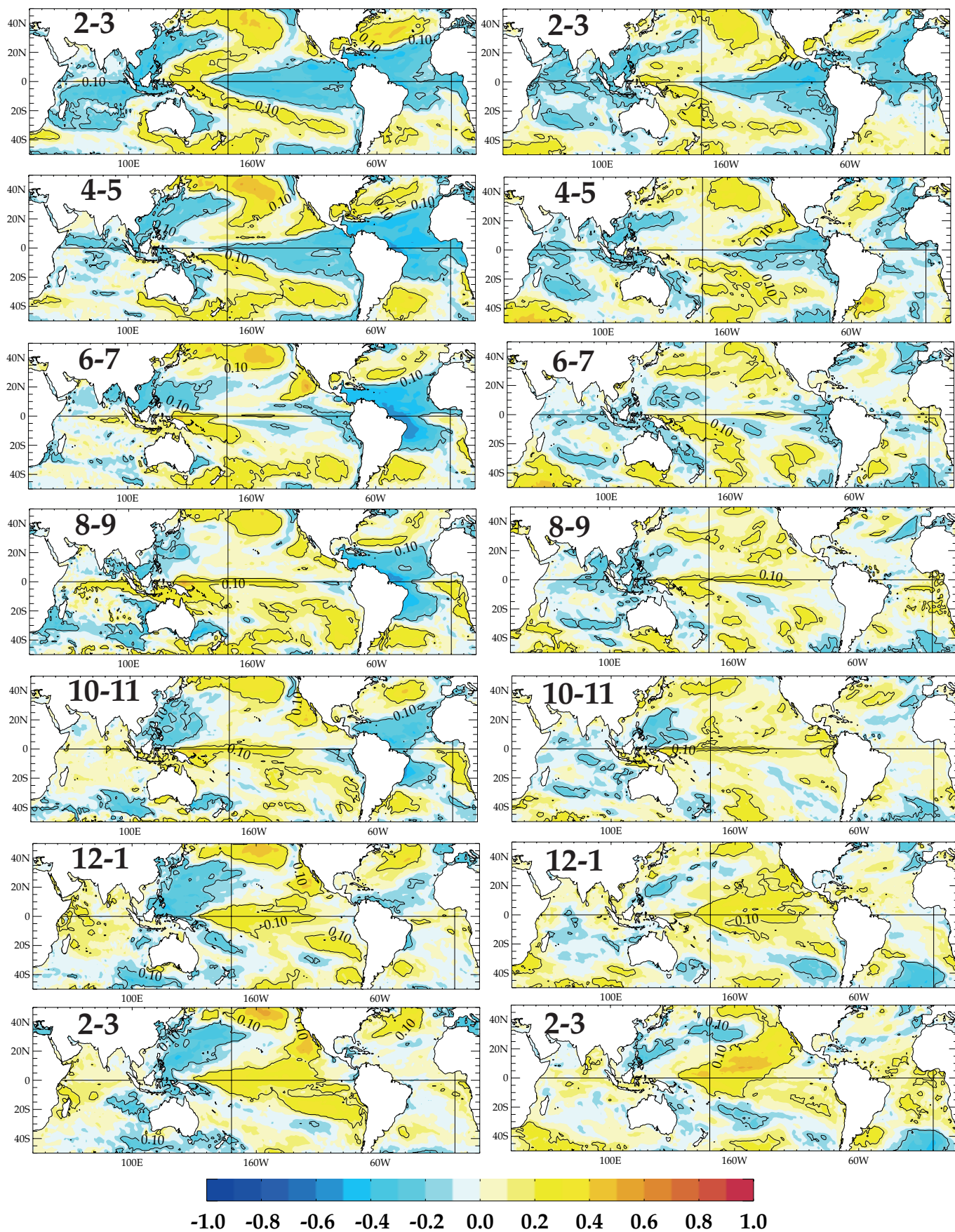


Figure 9: Lagged correlations between bi-monthly averaged SSTs during year-1 and the December-January (year0) Niño-3.4 SST for a) FTIC and b) FTAC. The correlations are calculated beginning in February-March of year-1, prior to the El Niño onset, and ending in February-March of year0. Correlations that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

figure 10 Regressions Nino34 SST (12-1) wind850/slp - FTIC - Year -1

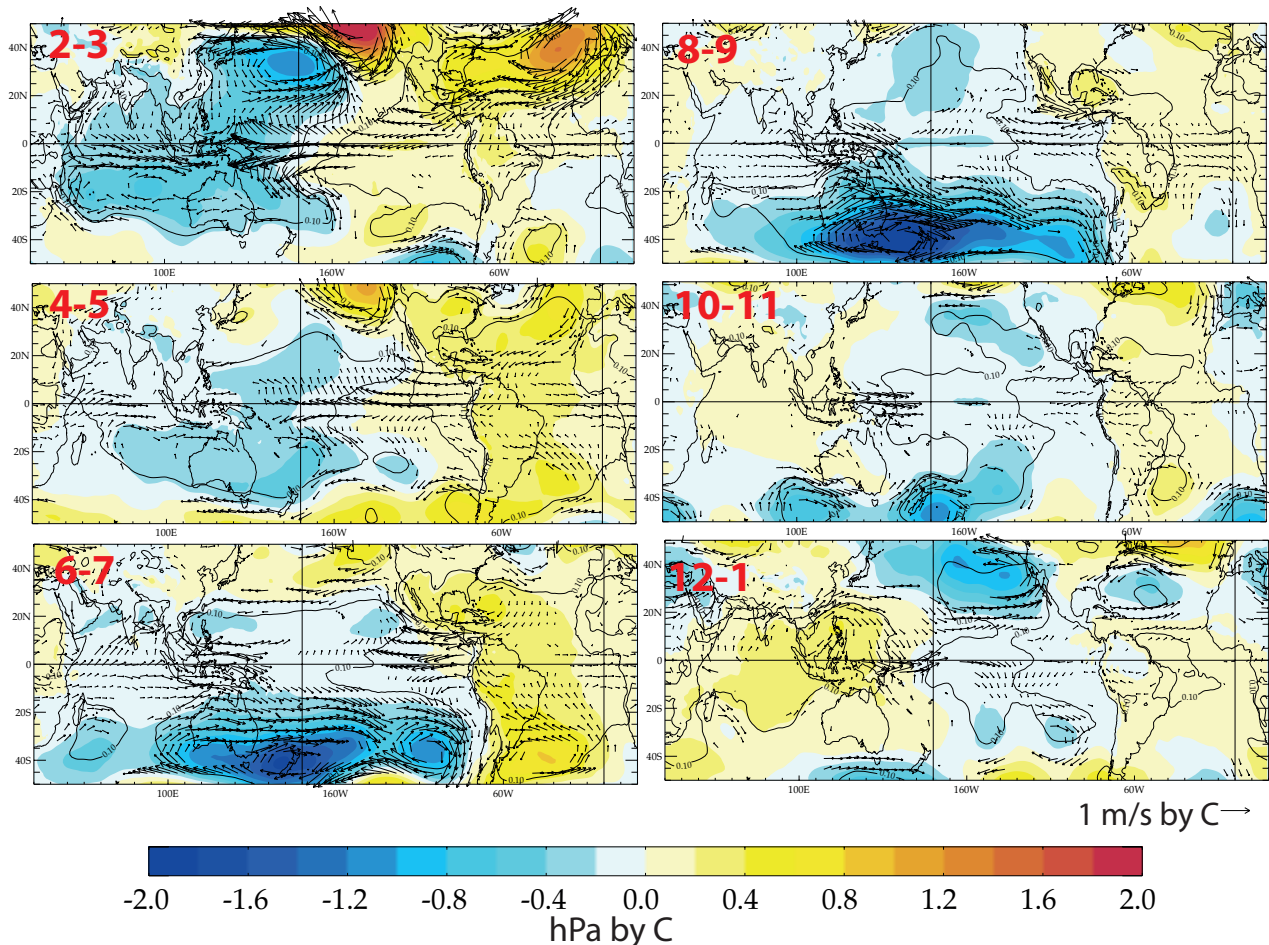


Figure 10: Lagged regressions between bi-monthly averaged 850 hPa wind and SLP anomalies during year-1 and the December-January (year0) Niño-3.4 SST for FTIC. Regressions that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined and only the 850 hPa wind anomalies above the 90% significance confidence level are shown.

figure 11 Regressions Nino34 SST (12-1) wind850/slp - FTAC - Year -1

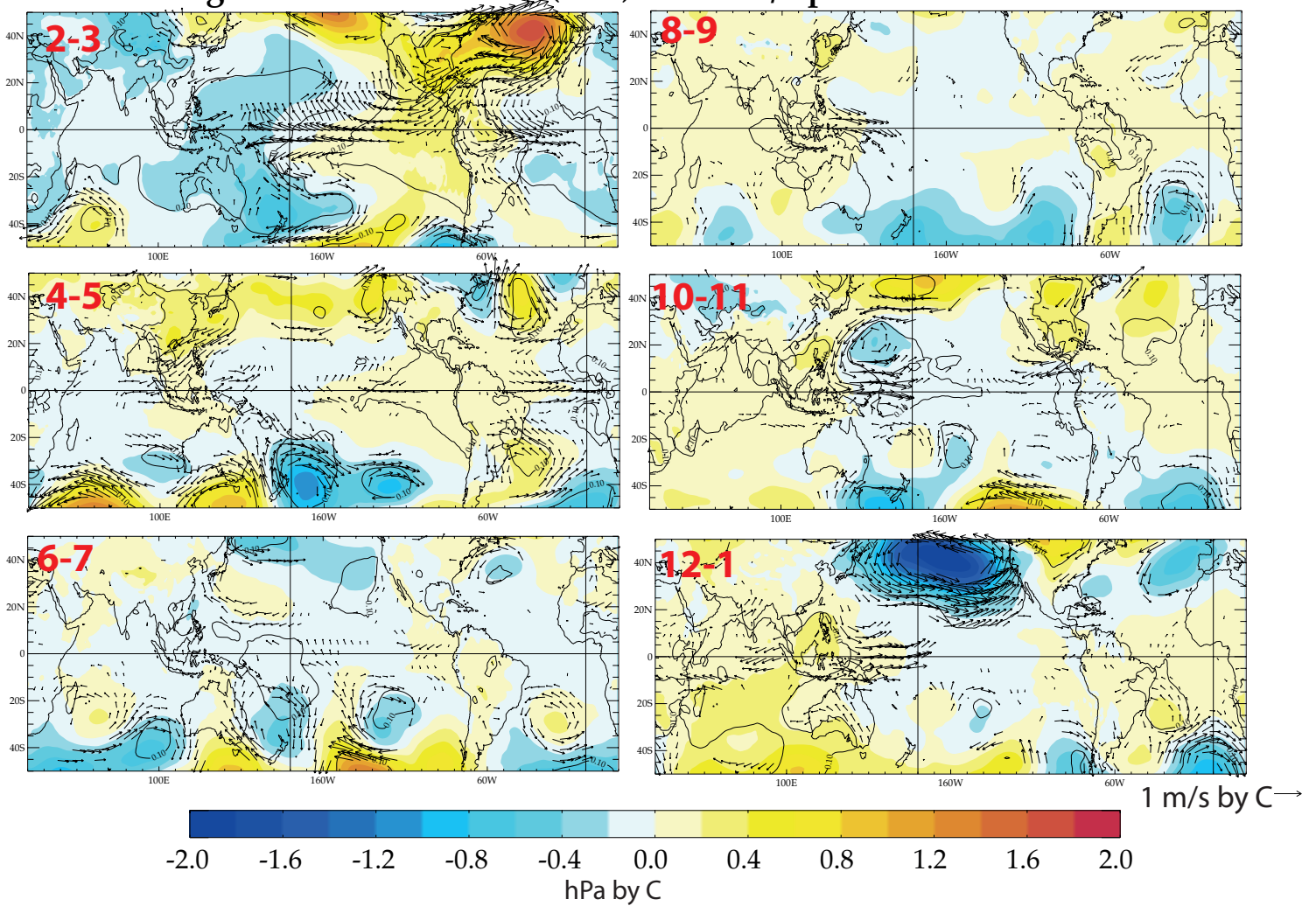


Figure 11: Same as Figure 10, but for the FTAC experiment.

figure 12 Regressions Niño34 SST (12-1) SST - Year +1

REF

FTIC

FTAC

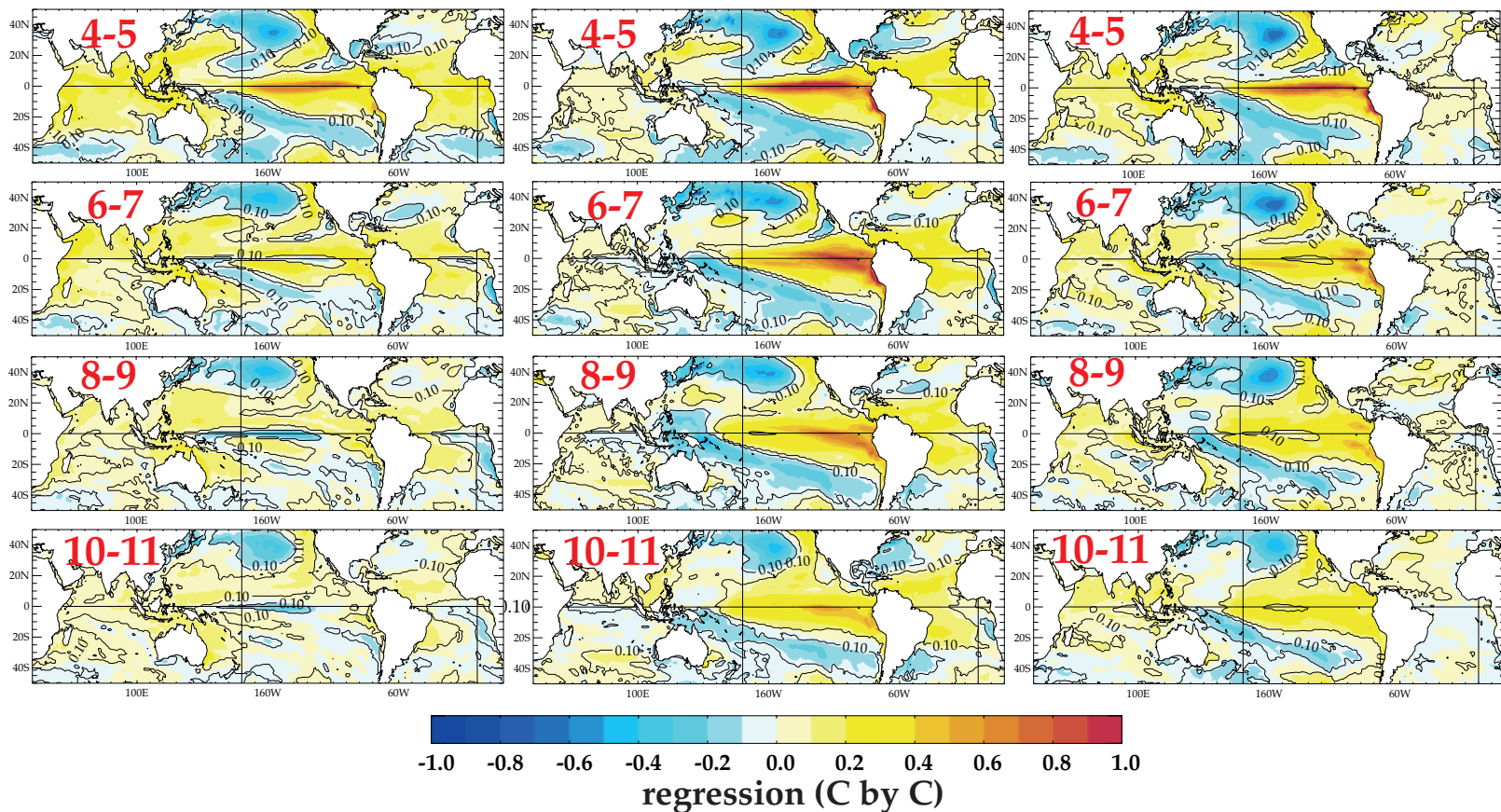


Figure 12: Lead regressions between bi-monthly averaged SSTs during year+1 and the December-January (year0) Niño-3.4 SST for a) REF, b) FTIC and c) FTAC. The regressions are calculated beginning in April-May of year+1, after the El Niño peak phase, and ending in October-November of year+1. Regressions that are above the 90% significance confidence level according to a phase-scramble bootstrap test (Ebisuzaki 1997) are underlined.

figure 13 Regressions Nino34 SST (12-1) U10 - Year +1

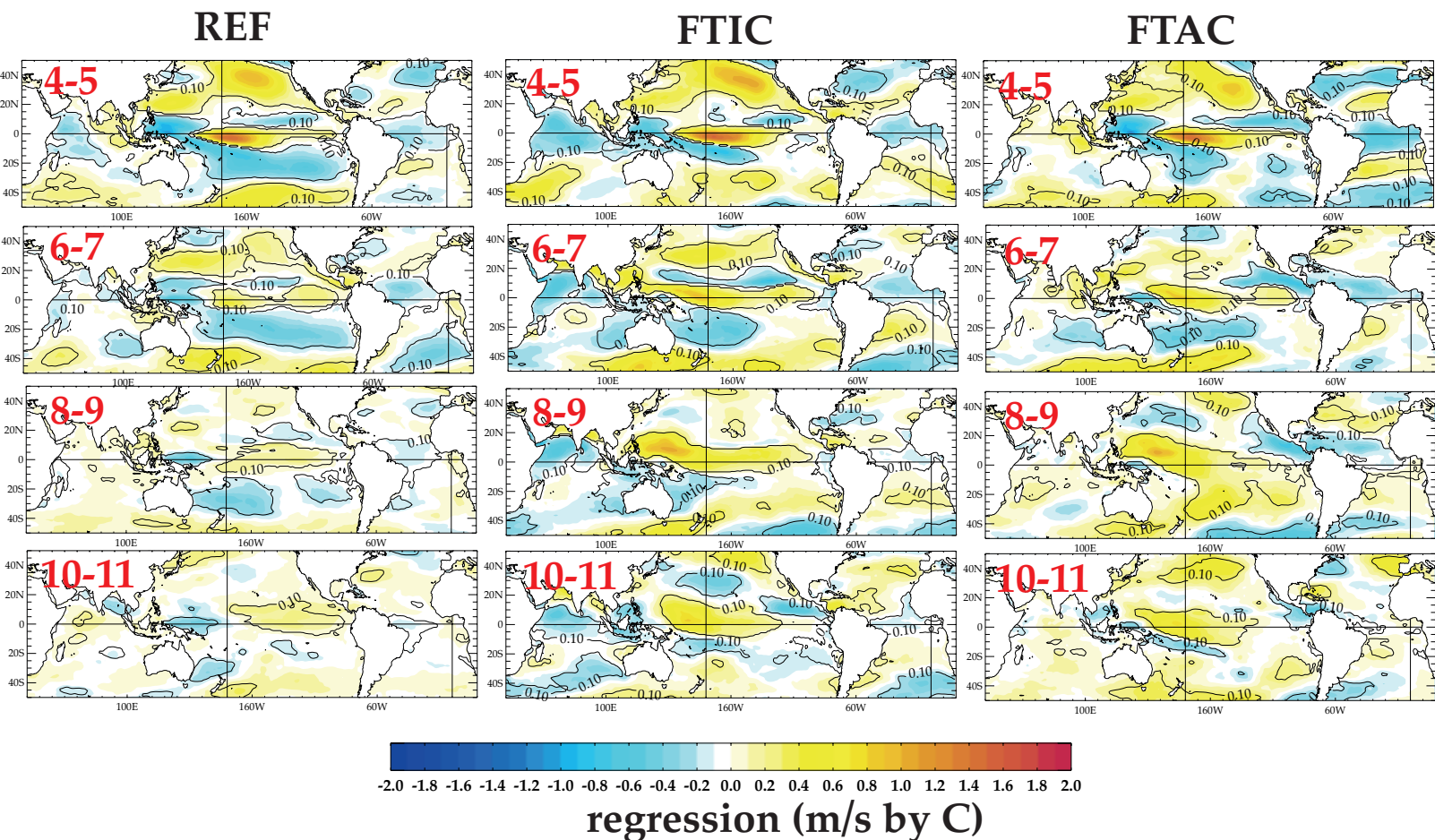


Figure 13: Same as Figure 12, but for bi-monthly 10m zonal wind anomalies.

Figure 14 Regressions Nino34 SST (12-1) 200 hPa wind and velocity potential - Year +1

REF

FTIC

FTAC

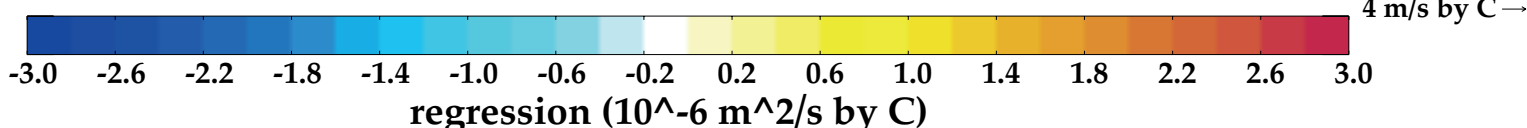
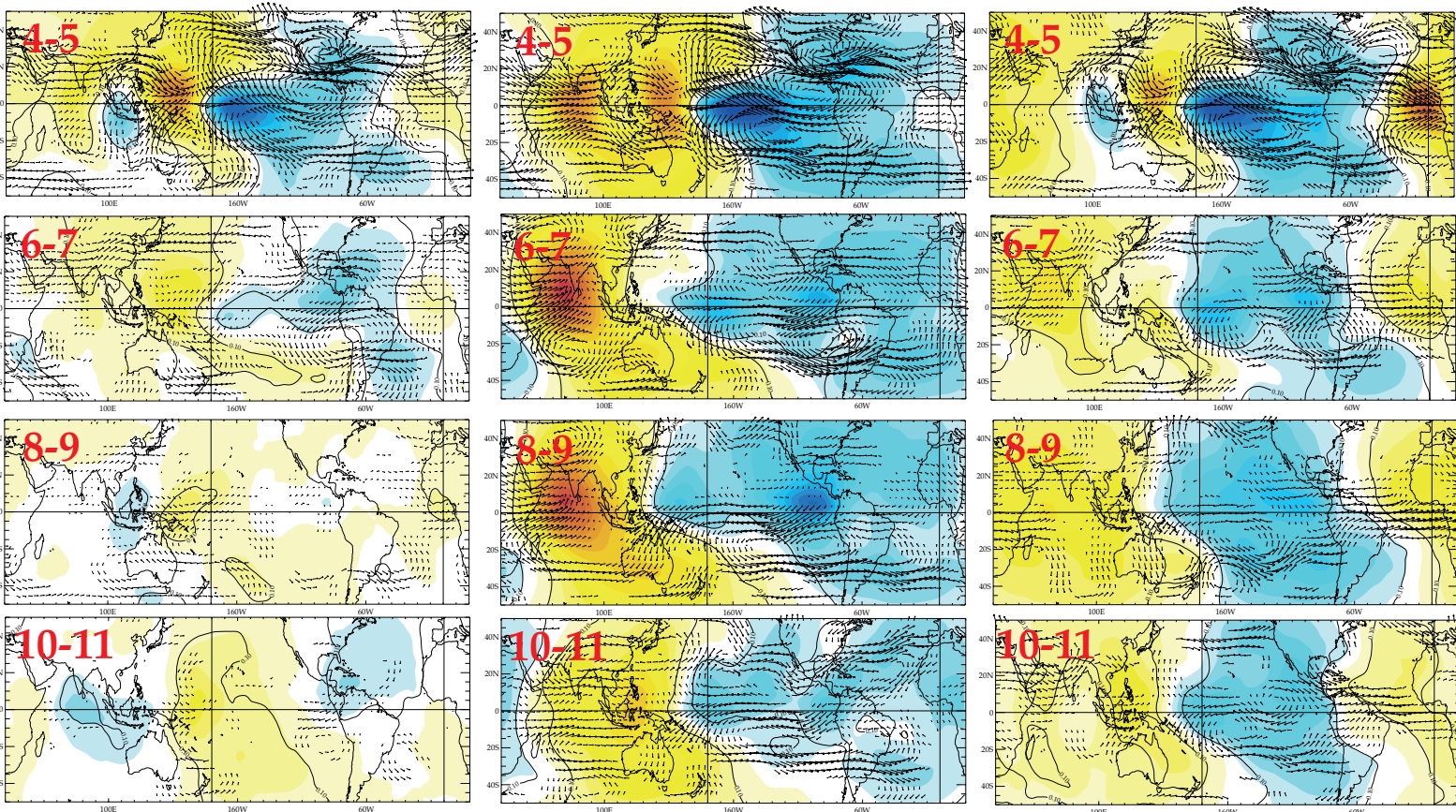


Figure 14: Same as Figure 12, but for 200 hPa wind and velocity potential anomalies.

Table 1

Name	REF	FTIC	FTIC-obs	FTAC	FTAC-obs
Correction area	None	Indian Ocean 30°E-120°E 25°S-30°N	Indian Ocean 30°E-120°E 25°S-30°N	Atlantic Ocean 100°W-20°E 25°S-25°N	Atlantic Ocean 100°W-20°E 25°S-25°N
Smoothing area	None	30°S-25°S	30°S-25°S	30°S-25°S 25°N-30°N	30°S-25°S 25°N-30°N
SST data	None	REF	AVHRR	REF	AVHRR
Time duration (Year)	210	110	50	110	50

Table 1: Summary of the numerical experiments with their main characteristics, including length, nudging domain and SST climatology used for the nudging in the Indian or Atlantic oceans decoupled experiments.

Table 2**SST standard-deviations**

Indices	Niño-3.4 DJF	ATL3 JJAS	IOB AM	IOD SON
observations	1.11	0.45	0.29	0.35
REF	0.89	0.30	0.24	0.55

Table 2: Standard-deviations (in °C) of Niño-3.4 (5°S–5°N, 170–120°W), ATL3 (5°S–5°N, 340–360°E), IOB (20°S–20°N, 40–110°E), and IOD SST indices in observations and REF. The IOD index is computed as the differences between the SST anomalies in a western (10°S–10°N, 60–80°E) and eastern (0–10°S, 90–110°E) box in the tropical Indian Ocean. The different indices are computed as seasonal averages as defined in the first row of the Table. These seasons correspond to the peak season of the indices in the observations. The statistics for the observations are derived from the HadISST1.1 dataset for the 1979-2012 period. The statistical significance of the differences between the observed and simulated standard-deviations has been assessed with the help of a Fisher test. The simulated standard-deviations significant at the 95% confidence level are in bold.

Table 3**SST – USTR (zonal wind stress) MCAs**

MCA1	SCF (%)	NC (%)	r	SSTvar (%)	USTRvar(%)
REF	63	10	0.74	30.2	6.1
FTIC	74	12.7	0.76	39.4	7.1
FTIC-obs	80	16.2	0.82	45.6	8.5
FTAC	74	12.3	0.78	34.7	7
FTAC-obs	72	11.8	0.79	37.2	6.8

Table 3 Summary statistics for the SST-USTR MCA leading modes estimated from the REF, FTIC, FTIC-obs, FTAC and FTAC-obs experiments, including the SCFs and NCs for the leading modes in the various MCA expansions and the correlation (r) between the EC time series of the left and right fields. SCF stands for Square Covariance Fraction and NC for Normalized root-mean-square Covariance, as given in the text. As discussed by Zhang et al. (1998), the NC and r coefficients are particularly useful in comparing the strength of the coupling between the left and right fields in modes obtained from different MCAs. SSTvar and USTRvar are, respectively, SST and USTR variances accounted for by the leading mode of each MCA analysis (e.g. this is the variance of the field explained by the related Expansion Coefficient time series by linear regression).

Table 4**USTR – 20d (depth of 20°C isotherm) MCAs**

MCA1	SCF (%)	NC (%)	r	USTRvar (%)	20dvar(%)
REF	61	7.8	0.73	6.4	17.5
FTIC	73	11.1	0.73	8.4	27.4
FTIC-obs	68	13.2	0.81	9.3	28.3
FTAC	75	10.2	0.76	7.2	24.9
FTAC-obs	69	9.5	0.75	6.1	27.8

Table 4: Same as Table 2, but for the USTR-20d MCA leading modes from the REF, FTIC, FTIC-obs, FTAC and FTAC-obs experiments.