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# Anatomy of the Indian Summer Monsoon and ENSO relationships in state-of-the-art CGCMs: Role of the tropical Indian Ocean

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1                                   **Anatomy of the Indian Summer Monsoon and ENSO**  
2                                   **relationships in state-of-the-art CGCMs:**

3                                   **Role of the tropical Indian Ocean**

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22

## Abstract

23 Indian Summer Monsoon (ISM) rainfall and El Niño-Southern Oscillation (ENSO) exhibit an inverse  
24 relationship during boreal summer, which is one of the roots of ISM interannual variability and its  
25 seasonal predictability. Here we document how current climate and seasonal prediction models  
26 simulate the timing and amplitude of this ISM-ENSO teleconnection. Many Coupled General  
27 Circulation Models (CGCMs) do simulate a simultaneous inverse relationship between ENSO and  
28 ISM, though with a large spread. However, most of them show significant negative correlations before  
29 ISM, which are at odd with observations. Consistent with this systematic error, simulated Niño-3.4  
30 Sea Surface Temperature (SST) variability has erroneous high amplitude during boreal spring and  
31 ISM rainfall variability is also too strong during the first part of ISM.

32 The role of the Indian Ocean (IO) in modulating the ISM-ENSO relationships is further investigated  
33 using dedicated experiments with the SINTEX-F2 CGCM. Decoupled tropical Pacific and IO  
34 experiments are conducted to assess the direct relationship between ISM and IO SSTs on one hand,  
35 and the specific role of IO feedback on ENSO on the other hand. The direct effect of IO SSTs on ISM  
36 is weak and insignificant at the interannual time scale in the Pacific decoupled experiment. On the  
37 other hand, IO decoupled experiments demonstrate that El Niño shifts rapidly to La Niña when ocean-  
38 atmosphere coupling is active in the whole IO or only in its western part. This IO negative feedback is  
39 mostly active during the decaying phase of El Niño, which is accompanied by a basin-wide warming  
40 in the IO, and significantly modulates the length of ENSO events in our simulations.

41 This IO feedback operates through a modulation of the Walker circulation over the IO, which  
42 strengthens and shifts eastward an anomalous anticyclone centered on the Philippine Sea and  
43 associated easterly wind anomalies in the equatorial western Pacific during boreal winter. In turn,  
44 these atmospheric anomalies lead to a fast ENSO turnabout via oceanic adjustment processes  
45 mediated by eastward propagating upwelling Kelvin waves. An experiment in which only the  
46 SouthEast Indian Ocean (SEIO) is decoupled, demonstrates that the equatorial SST gradient in the IO  
47 during boreal winter plays a fundamental role in the efficiency of IO feedback. In this experiment,  
48 simulated ISM-ENSO lead-lag correlations match closely the observations. This success is associated  
49 with removal of erroneous SEIO SST variability during boreal winter in the SEIO decoupled  
50 experiment. Finally, it is illustrated that most CMIP5 CGCMs exhibit similar SST errors in the SEIO  
51 during boreal winter in addition to an exaggerated SEIO SST variability during boreal fall.

52 **Keywords:** Indian Summer Monsoon; El Niño-Southern Oscillation; Indian Ocean; ocean-  
53 atmosphere interactions; coupled climate model.

54

## 55 **1. Introduction**

56 The climate of South Asia is dominated by the monsoon, which returns with remarkable  
57 regularity in each summer and provides the rainfall needed to sustain over 60% of the world's  
58 population. The Indian subcontinent is thus strongly dependent on the timing and amount of  
59 precipitation falling during the monsoon season (from June through September, JJAS  
60 hereafter) and is one of the most vulnerable regions of the world as far as water resources is  
61 concerned. Predicting and projecting (for the next century) the Indian Summer Monsoon  
62 (ISM) rainfall variability are, thus, both a scientific challenge and a key-societal need.

63 On inter-annual time scales, the ISM rainfall exhibits the multiyear (3-7 years) El-Niño-  
64 Southern-Oscillation (ENSO) frequency (Mooley and Parthasarathy 1983; Webster et al.  
65 1998; Lau and Wang 2006). ENSO is the most dominant form of ocean-atmosphere coupled  
66 variability on interannual time scales and affects climate worldwide through atmospheric and  
67 oceanic teleconnections (e.g., Tourre and White 1995; Alexander et al. 2002; Clarke 2008;  
68 Wang 2019). The effect of ENSO on the ISM has been intensively studied for decades  
69 (Walker 1924; Sikka 1980; Keshavamurthy 1982; Webster et al. 1998; Lau and Nath 2000,  
70 2012; Gadgil et al. 2004; Cherchi et al. 2007; Pillai and Anamalai 2012). During El Niños, the  
71 Walker circulation shifts eastward, inducing subsidence and dry conditions in the Indian  
72 sector and vice-versa during La Niñas (Walker 1924; Sikka 1980; Rasmusson and Carpenter  
73 1983; Webster et al. 1998; Lau and and Nath 2000, 2012). It is therefore extremely important  
74 to examine if the ENSO-ISM relationships are well simulated in state-of-the art climate and  
75 seasonal forecasting models (Annamalai et al. 2007; Terray et al. 2012; Sperber et al. 2013;  
76 Jourdain et al. 2013; Sabeerali et al. 2019; Krishna et al. 2019).

77 Though the probability of occurrence of a weak (strong) ISM during an El Niño (La Niña)  
78 is large, still ENSO can explain only about 35% the interannual variance of ISM rainfall  
79 (Krishna Kumar et al. 2006). Moreover, the ISM-ENSO relationship has considerably  
80 weakened during some periods (Torrence and Webster 1999), so that it is important to look  
81 for other sources of ISM predictability. In addition to ENSO, many studies have pointed out  
82 significant connections between ISM and the Indian Ocean (IO) Sea Surface Temperatures  
83 (SST) anomalies (Rao and Goswami 1988; Ashok et al. 2001, 2004; Gadgil et al. 2004, 2005,  
84 2007; Krishnan et al. 2003; Terray et al. 2003, 2007; Cherchi et al. 2007; Park et al. 2010;  
85 Boschhat et al. 2011; Lau and Nath 2012; Cherchi and Navarra 2013; Shukla and Huang 2016).  
86 In particular, it has been suggested that the Indian Ocean Dipole (IOD; Saji et al. 1999;  
87 Webster et al. 1999) interacts with both ENSO and ISM (Ashok et al. 2001, 2004; Drbohlav

88 et al. 2007; Loschnigg et al. 2003; Schott et al. 2009; Luo et al. 2010; Izumo et al. 2010;  
89 Wang 2019). Positive IOD events are associated with cool SST and shallow thermocline in  
90 the eastern Indian Ocean (IO), and warm SST and deep thermocline in its western part (Li et  
91 al. 2003; Gualdi et al. 2003; Spencer et al. 2005; Fischer et al. 2005; Schott et al. 2009). The  
92 way IOD can influence ISM remains controversial. Some authors suggest a direct influence of  
93 positive IOD events through moisture transport over the western IO or modifications of the  
94 local Hadley cell with increased ascendance over India, both factors enhancing ISM rainfall  
95 (Ashok et al. 2001, 2004; Gadgil et al. 2004; Behera et al. 2005; Ashok and Saji 2007). It has  
96 also been suggested that IOD counteracts the influence of ENSO on ISM (Ashok et al. 2001,  
97 2004; Ashok and Saji 2007; Ummenhofer et al. 2011; Krishnaswamy et al. 2015).

98 More recently, more complex interactions between ISM, IOD and ENSO have been  
99 evidenced, adding even more complexity to the emerging picture. It has been suggested that a  
100 strong ISM can favor a negative IOD event by producing westerly wind anomalies over the  
101 equator IO during boreal fall and that ENSO, ISM and IOD are strongly inter-related  
102 components of the Tropospheric Biennial Oscillation (TBO) in the Tropics (Terry 1995;  
103 Meehl 1997; Meehl and Arblaster 2002, 2003; Loschnigg et al. 2003; Terry et al. 2003,  
104 2005b; Drbohlav et al. 2007; Webster and Hoyos 2010; Li and Hsu 2017). IODs have also  
105 been suggested as a potential trigger of ENSO events (Luo et al. 2010; Izumo et al. 2010;  
106 Zhou et al. 2015; Jourdain et al. 2016; Wieners et al. 2017ab; Cai et al. 2019). It has been  
107 further discovered that the the Indian Ocean Basin (IOB) mode, associated with ENSO-  
108 related subsidence during the decaying phase of El Niños (e.g., Klein et al. 1999; Alexander  
109 et al. 2002; Xie et al. 2009, 2016; Schott et al. 2009; Wang 2019), provides an important  
110 forcing on ISM variability (Yang et al. 2007; 2010; Park et al. 2010; Boschhat et al. 2011,  
111 2012; Chowdary et al. 2017).

112 Because of the interactive nature of the tropical Indo-Pacific ocean-atmosphere system and  
113 the near-global patterns of ISM teleconnections summarized above, one of the best tools to  
114 simulate and predict ISM variability is a global Coupled General Circulation Model (CGCM;  
115 Wu and Kirtman 2005; Wang et al. 2005; Zhu and Shukla 2013). In order to provide reliable  
116 seasonal predictions and climate projections of ISM rainfall, it is nevertheless essential that  
117 CGCMs produce a reasonable simulation of the mean summer monsoon circulation and  
118 rainfall distribution, as well as its variability at different time scales. Unfortunately, this is still  
119 an area under rapid development, and CGCMs are still at a relatively early stage (Shukla et  
120 al., 2009). Most current CGCMS exhibit deficiencies in simulating ISM, ENSO and the IO  
121 variability (Bollasina and Ming 2013; Levine et al. 2013; Prodhomme et al. 2014, 2015; Li

122 and Xie 2012, 2014; Li et al. 2015, 2017ac; Ham and Kug 2014; Bellenger et al. 2014;  
123 Annamalai et al. 2017; Terray et al. 2018). So far, we have only a poor understanding of the  
124 relative roles of local ocean-atmosphere coupling (e.g. IO) and ENSO in the occurrence of  
125 extreme ISMs both in observations and coupled simulations (Gadgil et al. 2005; Saha et al.  
126 2016; Krishna et al. 2019). In particular, the way the IOD and IOB modes influence ISM and  
127 interact with ENSO remains unclear (Wu and Kirtman 2004ab; Achuthavarier et al. 2012; Li  
128 et al. 2015, Annamalai et al. 2017), and may limit drastically ISM seasonal predictability and  
129 the accuracy of ISM projections (Sabeerali et al. 2019; Li et al. 2017ac).

130 Taking into account the large uncertainties in the sign and amplitude of the ENSO-IO-ISM  
131 relationships, we will document in this work, (i) the ability of state-of-the-art CGCMs to  
132 simulate the ENSO-ISM relationship and (ii) the possible role of the IO (e.g. the IOD and  
133 IOB modes) in modulating this relationship. More precisely, we will document errors in the  
134 simulation of the ENSO-ISM relationship in CMIP5 and some seasonal forecasting CGCMs  
135 and analyze a set of decoupling experiments performed with a global CGCM, the SINTEX-F2  
136 model (Masson et al. 2012), to diagnose the possible origins of these errors. Such decoupling  
137 approach has been already successfully used to analyze the role of Indian and Atlantic oceans  
138 on ENSO (Ohba and Ueda 2007; Luo et al. 2010; Santoso et al. 2012; Ohba and Watanabe  
139 2012; Terray et al. 2016; Kajtar et al. 2017), the impacts of SST errors on ISM (Prodhomme  
140 et al. 2014, 2015), and the IOD evolution and its forcing mechanisms in the absence of ENSO  
141 (Fischer et al. 2005; Behera et al. 2005, 2006; Yang et al. 2015; Cretat et al. 2017, 2018;  
142 Stuecker et al. 2017; Wang et al. 2016, 2019).

143 This paper is organized as follows. The coupled models, the sensitivity experiments and  
144 validation datasets used in this study are described in Section 2. In Section 3, we document  
145 the ISM-ENSO relationships in observations and the errors in simulating these relationships  
146 by current CGCMs. In Section 4, we analyze decoupling experiments performed with the  
147 SINTEX-F2 CGCM, in which tropical Pacific or IO SST variability is removed, in order to  
148 assess the “intrinsic” role of IO SSTs on ISM, ENSO and the ISM-ENSO relationship and to  
149 understand their potential roles on the errors in the simulation of the ISM-ENSO relationship.  
150 The final section summarizes the main results of the present work and discusses if and how  
151 these results apply to CMIP5 models.

152

## 153 **2. Observed datasets, coupled models and sensitivity decoupling experiments**

### 154 **2.a Observed datasets**

155 To study ISM variability and its teleconnection with ENSO, we have used three different  
156 observed rainfall datasets. Firstly, we considered the extensively used classical All-India-  
157 Rainfall (AIR) dataset, based on quality-controlled rain-gauge data, for the period 1870-2012  
158 (Parthasarathy et al., 1994; Krishna Kumar et al. 1999; Bodai et al. 2020). The second rainfall  
159 dataset included in our analysis is a high-resolution (e.g.  $0.25^\circ \times 0.25^\circ$ ) quality-controlled  
160 gridded (daily) rainfall dataset, for the period 1901-2013, as obtained from the Indian  
161 Meteorological Department (IMD; see Pai et al. 2015 for precise details). Finally, we also  
162 used the Global Precipitation Climatology Project rainfall dataset (GPCP at  $1^\circ \times 1^\circ$  spatial  
163 resolution; Huffman et al. 2001), which combines measures of precipitation gauges and  
164 satellite data. GPCP is analysed for the 1979-2012 period. For both IMD and GPCP, the ISM  
165 Rainfall (ISMR) time series is defined as the average of rainfall anomalies for the land grid  
166 points in the region  $5^\circ\text{N}$  to  $25^\circ\text{N}$  and  $70^\circ\text{E}$  to  $95^\circ\text{E}$ .

167 The Niño-3.4 SST (monthly average of SST anomalies in the region  $5^\circ\text{S}$  to  $5^\circ\text{N}$  and  $170^\circ$   
168 to  $120^\circ\text{W}$ ) time series is chosen for the ENSO index since in observations the strongest  
169 correlations between ISMR and tropical Pacific SSTs occur over this region (not shown). This  
170 is consistent with past studies (Krishna Kumar et al. 2006; Jourdain et al. 2013). However,  
171 this practical choice implies that our study mainly focuses on canonical ENSO events rather  
172 than on El Niño Modoki events (Ashok et al. 2007). The Niño-3.4 SST time series is  
173 estimated from the Hadley Centre Sea Ice and SST dataset (HadISST1.1; Rayner et al. 2003).  
174 But similar results (not shown) are obtained with other SST datasets such the ERSST dataset  
175 (Huang et al. 2017) or the AVHRR only daily Optimum Interpolation SST version 2  
176 (OISSTv2) dataset (Reynolds et al. 2007).

## 177 2.b Climate models

178 In order to give a complete overview of the capacity of current CGCMs to simulate the  
179 ENSO-ISM relationship, we first considered monthly mean outputs from Coupled Model  
180 Intercomparison Project Phase 5 (CMIP5) coupled models available at url: [http://cmip-  
181 pcmdi.llnl.gov/cmip5/data\\_portal.html](http://cmip-pcmdi.llnl.gov/cmip5/data_portal.html) (Taylor et al. 2012). We analyzed the twentieth  
182 century simulations of 25 CMIP5 models (see Table S1 in Supplementary Materials for list of  
183 models) and for all the models we use the first ensemble member (“r1i1p1” from CMIP5  
184 database) from each model. We also used long free runs of two state-of-the-art global  
185 CGCMs, the CFSv2 (Saha et al. 2014) and the SINTEX-F2 (Masson et al. 2012). The lengths  
186 of these control simulations are 80 and 210 years, respectively, for the CFSv2 and SINTEX-  
187 F2 CGCMs.

188 The CFSv2 is the CGCM adopted for operational seasonal forecasting in the US by the  
189 National Centers for Environmental Prediction (NCEP; Saha et al. 2014) and in India, in the  
190 framework of the Monsoon Mission project (Rao et al. 2019). Its atmospheric model, the  
191 Global Forecast System (GFS) is run at T126 spectral resolution (about  $0.9^\circ$  by  $0.9^\circ$ ) with 64  
192 sigma-pressure hybrid levels. Its oceanic component has a  $0.25$ - $0.5^\circ$  horizontal resolution, 40  
193 vertical levels and includes an ice model. The atmosphere and ocean exchange quantities such  
194 as heat and momentum fluxes every half an hour, with no flux adjustment. See Saha et al.  
195 (2014) for further details on the CFSv2 model.

196 The SINTEX-F2 is used by JAMSTEC (e.g. Japan) for operational seasonal forecasting  
197 (Luo et al. 2005; Doi et al. 2016). A comprehensive description of SINTEX-F2 can be found  
198 in Masson et al. (2012). The atmospheric component is ECHAM5.3 and is run at T106  
199 spectral resolution (about  $1.125^\circ$  by  $1.125^\circ$ ) with 31 hybrid sigma-pressure levels (Roeckner  
200 et al. 2003). The oceanic component is NEMO (Madec 2008), using the ORCA05 horizontal  
201 resolution ( $0.5^\circ$ ), 31 unevenly spaced vertical levels and including the LIM2 ice model  
202 (Timmermann et al. 2005). The coupling information, without any flux corrections, is  
203 exchanged every 2 h using the Ocean-Atmosphere-Sea Ice-Soil coupler (Valcke 2006). The  
204 performance of the SINTEX-F2 model in simulating the mean state and interannual  
205 variability in the Indo-Pacific areas has been assessed many times and is not repeated here  
206 (Masson et al. 2012; Terray et al. 2012, 2016, 2018; Prodhomme et al. 2014, 2015; Cretat et  
207 al. 2017, 2018).

208 Note, finally, that the SINTEX-F2 and CFSv2 configurations used here employ fixed  $\text{CO}_2$   
209 concentrations corresponding to present day conditions. This is consistent with the use of  
210 recent observations (e.g. mostly 1979-2012) for validation of model outputs.

211 In all simulations, we characterize the ISM by the JJAS average precipitation over India  
212 ( $5^\circ$ - $25^\circ\text{N}$  and  $70^\circ$ - $95^\circ\text{E}$ , land only; ISMR index hereafter) and ENSO by the Niño-3.4 box-  
213 average SST as in observations.

## 214 2.c Partial decoupling experiments

215 In addition to the 210-year control run described above (named CTRL hereafter), we  
216 perform three partial decoupling experiments (named FTPC, FTIC and FSEIC hereafter),  
217 where the model SSTs in the tropical Pacific, Indian oceans and SEIO region are,  
218 respectively, nudged to a daily SST climatology obtained from CTRL (see Table 1 for details  
219 and definition of the acronyms used to design the nudged experiments), as described in Luo et  
220 al. (2005) and Appendix. The damping constant used is  $-2400 \text{ W m}^{-2} \text{ K}^{-1}$ , which corresponds



221 to the 1-day relaxation time for temperature in a 50-m ocean layer. In these experiments there  
222 are no significant changes in SST mean state and seasonal cycle in the nudged regions, but  
223 also in the whole Tropics compared to CTRL (see Figs. 1b-d and 2), but, SST variability is  
224 suppressed in the nudged region (not shown; see Terray et al. 2016 and Cretat et al. 2017 for  
225 illustration). The FSEIC experiment is designed to assess the role of the SST SEIO  
226 variability, and by extension of the IOD (as the SEIO region almost matches the eastern pole  
227 of the traditional IOD index in observations). Furthermore, despite the SST SEIO bias is  
228 modest in the annual mean (Fig. 1a), this region is affected by severe seasonally varying SST  
229 mean and variability biases in the SINTEX model (Fig. 2; Fischer et al. 2005; Terray et al.  
230 2012; Prodhomme et al. 2014; Cretat et al. 2017). The CGCM simulates a too shallower  
231 thermocline in the eastern Indian Ocean during boreal summer and fall. The thermocline-  
232 surface coupling is thus amplified in the annual cycle with deeper and cooler water easily  
233 upwelled at the surface, cooling drastically the SST mean and enhancing the SST variability  
234 in the SEIO during boreal summer and fall (Fig. 2). In addition, the SEIO SST is affected by a  
235 strong warm mean bias and also a too strong SST variability during late boreal winter in  
236 CTRL (Fig. 2). The FSEIC experiment will therefore be useful to understand the specific role  
237 of this amplified SST variability in the SEIO during these two critical seasons on the  
238 simulated ENSO, ISM and their relationships.

239 Finally, in order to discuss the remote or local origins of these SST errors in the SEIO in  
240 Section 5, an additional partial decoupling experiment (named FTFC-obs hereafter) will be  
241 considered. This run is similar to the FTFC experiment, excepted for the use of a daily  
242 climatology computed from the OISSTv2 dataset for the 1982-2010 period (Reynolds et al.  
243 2007) in the nudging procedure (Table 1). In this FTFC-obs run, the large feedback value  
244 applied removes the SST mean biases (in CTRL) with respect to the observed SST  
245 climatology, in addition of suppressing the SST variability in the restoring tropical Pacific  
246 domain (Fig. 1e).

247 Table 1 summarizes the specifications of the different sensitivity experiments used here  
248 and the different nudging domains are displayed in Fig.1b-d. Finally, in the analyses  
249 described in Sections 3, 4 and 5, the first 10 years of all simulations have been excluded due  
250 to the spin-up of the coupled model.

251

### 252 **3. ISM-ENSO relationships in observations and CGCMs**

253 We first document the amplitude and timing of the relationship between ENSO and ISM in

254 observations in order to provide a basis for a fair assessment of the performance of current  
255 CGCMs with respect to this metric in the next subsection.

### 256 3.a Observed ISM-ENSO relationships

257 In order to illustrate this relationship, we computed the lead-lag correlations between the  
258 different ISMR time series and the Niño-3.4 SST bimonthly time series in a three years  
259 window from the beginning of year -1 to the end of year +1, the year 0 referring to the year of  
260 the ISM season (Fig. 3a-b). Correlation coefficients have been estimated with and without  
261 detrending of the different rainfall and SST time series in order to assess the robustness of the  
262 results with respect to anthropogenic related trends and nonstationarity of the time series.  
263 Detrending has been performed with locally weighted regression, a non-parametric method  
264 for fitting a smoothed regression curve to data through local smoothing (Cleveland and  
265 Devlin 1988). As results are similar, we only show the results from detrended data here.

266 Consistent with the TBO pattern, involving ENSO and ISM (Yasunari 1990; Meehl 1997;  
267 Meehl and Arblaster 2002; Loschnigg et al. 2003; Terray et al. 2003, 2005b; Webster and  
268 Hoyos 2010), two distant correlation peaks of opposite signs are noted (Fig. 3a-b). Positive  
269 correlations are evident one year before the monsoon for most ISMR indices and time  
270 periods. These significant positive correlations preceding ISM have largely amplified during  
271 recent decades, e.g. after the 1976/77 climate shift and have been explained by a delayed  
272 ENSO effect on the ISM, mediated by the IOB warming, which follows the strong El Niño  
273 events, such as the 1982-83 and 1997-98 events (Yang et al. 2007, 2010; Park et al. 2010;  
274 Boschat et al. 2011, 2012. The correlations switch sign around April-May (AM) and become  
275 again statistically significant only in June-July (JJ) of year 0 (Webster and Yang 1992;  
276 Webster et al. 1998). These negative correlations between Niño-3.4 SSTs and ISMR peak  
277 during boreal fall and fade away progressively after, e.g. during the peaking and decaying  
278 phases of El Niño (Fig. 3a-b). The significant negative correlation between ISMR and Niño-  
279 3.4 SSTs during boreal summer of year 0 implies that warmer (cooler) SSTs over these  
280 regions will suppress (enhance) ISM rainfall consistent with the studies summarized in the  
281 introduction. This synchronous effect can be termed the “direct” ENSO effect by opposition  
282 to the “indirect” effect mediated by the IOB warming induced by ENSO (Wu et al. 2012).

283 Fig. 3a-b also demonstrates that the synchronous correlation between ISM and ENSO is  
284 decreasing for more recent (and shorter) time periods. Interestingly, the lead correlations  
285 between ENSO of year -1 and ISM follow an opposite evolution. The decreasing synchronous  
286 correlation has lead to the suggestion that the ISM-ENSO relationship has weakened during

287 recent decades for reasons which are still a matter of debate (Krishna Kumar et al. 1999,  
288 2006; Torrence and Webster 1999; Gershunov et al. 2001; Ashok et al. 2001, 2004, 2007;  
289 Kinter et al. 2002; Annamalai et al. 2007; Kucharski et al. 2007, 2008; Chen et al. 2010;  
290 Boschat et al. 2012; Li and Ting 2015; Srivastava et al. 2015; Cash et al. 2017; Yun and  
291 Timmermann 2018; Feba et al. 2019; Bodai et al. 2020).

### 292 3.b Simulated ISM-ENSO relationships

293 We first focus on the capacity of the SINTEX-F2 and CFSv2 coupled models to simulate  
294 the ISM-ENSO relationship (Fig. 4a). The two CGCMs are able to reproduce the synchronous  
295 negative correlation between ISM and ENSO, though with varying amplitude (Fig. 4.a).  
296 However, before ISM, the two CGCMs show large discrepancies from observations with  
297 negative and significant correlations during an extended period of several months before ISM.  
298 In the CFSv2 model, the lead-lag correlations suggest that ISM is linked to ENSO *before ISM*  
299 *rather than during and after ISM* since the maximum negative correlation occurring at lag 0,  
300 e.g. JJ of year 0. Moreover, after ISM, the negative correlation quickly fades away in  
301 disagreement with observations. The SINTEX-F2 performs slightly better, but shares the  
302 same deficiency during the pre-onset period.

303 We next examined the lead-lag correlations between the Niño-3.4 SST and ISMR in  
304 CMIP5 CGCMs (Fig. 5a). Here, the correlation coefficients are computed from the first  
305 member of each model's ensemble of historical runs and for the period 1950-2000. ISM and  
306 its relationships to ENSO in CMIP5 models have already been investigated in many studies  
307 (Sperber et al. 2013; Terray et al. 2012); Jourdain et al. 2013; Ramu et al. 2018), but most of  
308 them focus on the synchronous relationship (e.g. during boreal summer) between the two  
309 phenomena. Here we examine the lead-lag relationships between the two phenomena in more  
310 details following the same framework as used above for observations, the SINTEX-F2 and  
311 CFv2 models, complementing the results of Jourdain et al. (2013).

312 Many CMIP5 CGCMs are able to reproduce the synchronous negative correlation between  
313 ISM and ENSO, although in many models the amplitude of this correlation is erroneous and  
314 the diversity (e.g. spread) is large between the CGCMs (Figure 5a). However, a striking  
315 feature is that almost all of the CMIP5 CGCMs fail to capture the observed lead-lag  
316 relationships between ENSO and ISM. In particular, most CGCMs show erroneous large  
317 negative correlations before ISM, which are completely absent in observations, and the  
318 observed TBO pattern with positive correlations preceding the ISM by nearly one year is  
319 lacking in all CMIP5 CGCMs. In a similar fashion, the simulated correlations are much

320 weaker than observed after the boreal summer of year 0.

321 In other words, current CGCMs show large deficiencies in recovering the observed lead-  
322 lag relationships between ISM and ENSO, which is of paramount importance for seasonal  
323 forecasting (Gadgil et al. 2005) or climate projections of ISM rainfall (Li et al. 2017ac). It is  
324 therefore necessary to go a step further in the model validation to analyze more thoroughly  
325 the reasons of this deficiency in current CGCMs, which is the goal of the next sections.

### 326 3.c Simulated ENSO seasonal phase-locking

327 Though considerable improvements in the simulation of ENSO have been made during the  
328 past decades, current CGCMs still need to be improved with regard to realistically  
329 representing ENSO (Bellenger et al. 2014). As an illustration, the tendency to produce a  
330 double intertropical convergence zone in the Pacific basin, a poor representation of the SST  
331 annual cycle and mean pattern in the tropical Pacific, and a substantial underestimation of  
332 ENSO variability are recurrent biases shared by many past and present CGCMs (AchutaRao  
333 and Sperber 2002, 2006; Li and Xie 2012; 2014; Bellenger et al. 2014). Thus, a biased  
334 representation of ENSO in the tropical Pacific itself can be a primary plausible reason for an  
335 improper simulation of the ISM-ENSO relationship in current CGCMs (Jourdain et al. 2013).  
336 However, Indian and Atlantic SSTs can also matter as they have an overall damping effect on  
337 ENSO and modulate the length of the ENSO events (Dommenges et al. 2006; Terray et al.  
338 2016).

339 The phase locking of ENSO events to the annual cycle, with a tendency to peak at the end  
340 of the calendar year, the amplitude and length of El Niño events (e.g. ENSO periodicities) are  
341 among the most distinctive characteristics of ENSO (Rasmusson and Carpenter 1983; Clarke  
342 2008; Ham and Kug 2014; Li and Hsu 2017). Obviously, the incorrect annual phase-locking  
343 of ENSO's variability is a plausible candidate for explaining the current failure of state-of-  
344 the-art CGCMs in the simulation of the ISM-ENSO lead-lag relationships, since most coupled  
345 models do simulate a boreal summer monsoon over India (Figs. 4d and 5d). In Figures 4b and  
346 5b, we show the monthly standard deviations of the Niño-3.4 SST anomalies from  
347 observations, the SINTEX-F2, CFSv2 and CMIP5 CGCMs, respectively. Observed ENSO  
348 variability typically peaks in boreal winter and diminishes in boreal spring with relatively  
349 modest variability in boreal summer and early fall (Fig. 4b), which is related to the fact that El  
350 Niño's onset frequently occurs in boreal spring or early boreal summer (Clarke 2008).

351 It is apparent that both the SINTEX-F2 and CFSv2 models have a preference for relatively  
352 high SST variability in the Niño-3.4 region during boreal winter, as observed (Fig. 4b).

353 However, the monthly Niño-3.4 SST standard deviations in these models are higher than  
354 observed during boreal spring suggesting that ENSO onset is early than observed or that  
355 ENSO decay is slower than observed in the CGCMs. This bias is consistent with the lead-lag  
356 ISM-ENSO relationships simulated in the two models, especially the unrealistic negative  
357 correlations between ISM and the Niño-3.4 SST preceding the ISM onset (Fig. 4b). This is  
358 further confirmed by a lead-lag correlation analysis of SSTs with the ISMR index in the  
359 SINTEX-F2 model (see Fig. S1a-f or Figures 3 and 5 of Boschat et al. (2012) for similar  
360 analysis on observations). Also consistent with these erroneous negative correlations, the ISM  
361 rainfall monthly standard deviations are also exaggerated at the beginning of ISM in the  
362 simulations (Fig. 4c).

363 On the other hand, many CMIP5 models show a different seasonal Niño-3.4 SST  
364 evolution, with the peak of canonical El Niños occurring at any season (Fig. 5b; Ham and  
365 Kug 2014; Taschetto et al. 2014) and only a few models have Niño-3.4 SST variability above  
366 observations during boreal winter (Fig. 5b). This can be related to many factors, especially  
367 spatial shift of simulated SST variability over the equatorial Pacific often associated with the  
368 cold tongue bias (Collins et al. 2010; Li and Xie 2014; Swapna et al. 2015). However, most of  
369 them also produce a too large and erroneous ENSO variability during boreal spring as the  
370 SINTEX-F2 and CFSv2 models (Figs. 4b and 5b). Consistent with this deficiency and the  
371 erroneous negative correlations before ISM (Fig. 5a), the simulated ISM rainfall monthly  
372 standard deviations are again exaggerated at the beginning of ISM in most CMIP5 models  
373 (Fig. 5c). This exaggerated simulated ENSO variability during boreal spring and early  
374 summer can be due to (i) ENSO onsets occurring earlier in the CGCMs, (ii) ENSO events  
375 lasting longer with an extended maturing phase compared to observations, or both.

376 While SST mean biases, ENSO pattern or position of peak and deficient ENSO  
377 teleconnections may also modulate ISM-ENSO relationships in CGCMs (Jourdain et al.  
378 2013), we will show in the next section that length and seasonal phase-locking of ENSO  
379 events are key-factors for a realistic simulation of the lead-lag ISM-ENSO relationships.

380

#### 381 **4. Indian Ocean impacts on the ENSO-ISM relationship in the SINTEX-F2 CGCM**

382 To disentangle the role of IO SST variability in shaping the ISM-ENSO relationships, we  
383 now analyzed decoupling experiments performed with SINTEX-F2 (see Section 2c).

##### 384 4.a Intrinsic ISM teleconnections

385 We first investigate the possible “intrinsic” role of SST anomalies outside of the tropical  
386 Pacific on the ISM, as an important factor in shaping the ENSO-ISM relationship. The  
387 analysis is done with the help of the FTFC simulation, where ENSO is removed. By contrast,  
388 in the next subsection, we will investigate the possible role of IO SST anomalies on ENSO  
389 itself, as a second factor in shaping the simulated ENSO-ISM relationship with the help of the  
390 other sensitivity experiments in which the tropical IO SST variability, or parts of it, is  
391 suppressed.

392 Figure 6.a-d displays the regressions of the JJ and August-September (AS) 850-hPa wind  
393 and rainfall fields onto the ISMR index for CTRL and FTFC. As expected, Fig. 6.a-d show  
394 enhanced rainfall over India and the neighboring oceanic regions, as well as a strengthening  
395 of both the monsoon low-level winds over the Arabian Sea and the Somali jet along the  
396 African coast. Next, ISM rainfall anomalies are very sensitive to the perturbation of the  
397 Walker circulation that is characteristic of La Niña/El Niño episodes in CTRL (Fig. 6.a-b).  
398 Positive rainfall anomalies develop over the maritime continent and neighboring ocean areas  
399 and a negative IOD like rainfall pattern emerged over the equatorial IO during AS of strong  
400 ISMs and La Niña episodes, consistent with previous works (Webster et al. 1998; Lau and  
401 Nath 2000, 2003, 2012; Cretat et al. 2017).

402 By contrast, in FTFC, the atmospheric response associated with the strong/weak ISMs is  
403 mostly confined northward of 10°N and is characterized by an anomalous rainfall pattern  
404 opposing India and the northwest Pacific region (Fig. 6c-d; Terray et al. 2005a; Wang 2006;  
405 Gu et al. 2010; Kosaka et al. 2013; Cretat et al. 2017). Interestingly, ISMR rainfall standard  
406 deviations are only slightly reduced (at most of 0.2 mm/day, see Fig. 8c) in FTFC. This result  
407 may seem contradictory with the expectation that ENSO is the main driver for ISM  
408 interannual variability, but is consistent with the large internal variability associated with ISM  
409 and the fact that ISM is also an active player in tropical climate (Yasunari 1999; Kirtman and  
410 Shukla 2000). Further, boreal summer rainfall and wind anomalies are very weak outside this  
411 north Indo-Pacific region, especially over the equatorial IO.

412 We now examine the extent to which SST forcing outside the tropical Pacific may explain  
413 ISM rainfall variability in the absence of ENSO. This is done through a lead-lag correlation  
414 analysis between the ISMR index and bi-monthly SST and 200-hPa velocity potential  
415 anomalies in the CTRL and FTFC experiments (Figs. S1 and 7). Correlations between ISMR  
416 and SSTs in CTRL confirm the existence of a significant statistical association between ISM,  
417 ENSO and IOD (Fig. S1a-f), as discussed in the Introduction. Some significant correlations  
418 are also seen in the extratropical Pacific and the tropical Atlantic during boreal summer and

419 winter (Fig. S1c-f), which are also consistent with well-known ENSO teleconnections (Wang  
420 2019). On the other hand, correlations between ISMR and SSTs are strikingly different in  
421 FTFC (Fig. 7a-f). They are weak before, during and after ISM, except for a significant  
422 cooling (warming) of the Arabian Sea (the south of the bay of Bengal) during boreal summer  
423 (Fig. 7a-f). This local response is fully consistent with a “slave ocean scenario” in which the  
424 SST is mainly controlled by evaporative cooling and local upwelling associated with the  
425 strengthening of the monsoon low-level winds and the sea-saw between ISM and the  
426 northwest Pacific monsoon discussed above (see Fig. 6c-d). Moreover, simulated ISM  
427 variability is not significantly associated with SST anomalies or modulations of the Walker  
428 cell over the equatorial IO during boreal summer and fall (Fig. 7c-e and i-k) despite the fact  
429 that IODs do exist in the absence of ENSO in FTFC (Fig. 2b; Cretat et al. 2017, 2018). These  
430 results invalidate the hypothesis that the direct impact of IOD on ISM may induce a  
431 significant modulation of the ENSO-ISM relationship in SINTEX-F2.

432 On the other hand, the results in CTRL (Fig. S1) are fully consistent with a scenario in  
433 which ENSO, ISM and IOD are strongly inter-related components of the TBO in the Tropics  
434 (Meehl and Arblaster 2002, 2003; Drbohlav et al. 2007; Webster and Hoyos 2010) since a  
435 strong ISM is significantly correlated with La Niña and a negative IOD in CTRL (Fig. 6a-b  
436 and S1). However, the FTFC experiment demonstrates that this ENSO-ISM-IOD coupling is  
437 mainly due to the ENSO forcing on both ISM and the equatorial IO separately, and not to a  
438 direct association between ISM and the equatorial IO, at least in this CGCM.

439 Furthermore, the 200-hPa velocity potential anomalous patterns associated with ISMR  
440 variability confirm the hypothesis that ISM is an active player for the whole tropical  
441 circulation in the absence of ENSO. The only physically consistent anomalous 200-hPa  
442 velocity potential pattern emerging in Fig. 7g-l is the strong upper-level divergence flow  
443 associated with ISM itself and the induced significant strengthening of the subtropical  
444 anticyclones over the Pacific and South Atlantic (Fig. 7i-j). Thus, the upper-level circulation  
445 anomalies during boreal summer in FTFC are dominated by an enhancement of the zonal  
446 monsoon cyclone-subtropical anticyclone circulation driven by the east-west differential  
447 heating (Rodwell and Hoskins 2001; Chen 2003). On the other hand, in CTRL, the 200-hPa  
448 velocity potential correlation pattern (Fig. S1g-l) has a more large-scale structure and is  
449 tightly and significantly linked to ENSO evolution (Fig. S1a-f), as expected (Webster et al.  
450 1998).

451 In summary, the results of this section strongly suggest that direct (e.g. when ENSO is  
452 removed) effects of interannual SST anomalies outside of the tropical Pacific on ISM are

453 weak and do not play a direct role in shaping the ENSO-ISM relationship in the SINTEX-F2  
454 model.

#### 455 4.b Indian Ocean impacts on ENSO

456 However, the tropical IO can also influence the ENSO-ISM relationship by directly  
457 modulating ENSO in the tropical Pacific itself. Many studies have also highlighted the strong  
458 impact of the IOB in accelerating the El Niño to La Niña transition in the tropical Pacific  
459 (Kug and Kang 2006; Dommenges et al. 2006; Ohba and Ueda 2007; Jansen et al. 2009;  
460 Okumura et al. 2011; Ohba and Watanabe 2012; Kug and Ham 2012; Santoso et al. 2012;  
461 Terray et al. 2016; Xie et al. 2009, 2016; Kajtar et al. 2017). Several authors suggested that  
462 positive IOBs generate easterly wind anomalies over the western Pacific during boreal winter  
463 and spring as an atmospheric Kelvin wave response to the local diabatic warming (Annamalai  
464 et al. 2005; Terao and Kubota 2005; Kug and Kang 2006). Watanabe and Jin (2002) also  
465 argued that a positive IOB strengthens an anomalous low-level North West Pacific  
466 Anticyclone centered on the Philippine Sea (NWPA, also referred to as Philippine Sea  
467 anticyclone or western North Pacific subtropical high in the literature; see Wang et al. 2000)  
468 during the mature-decayed phase of El Niño. Collectively, these atmospheric anomalies may  
469 fasten the transition from El Niño to La Niña since the associated zonal wind stress anomalies  
470 over the western Pacific force upwelling oceanic Kelvin waves that propagates eastward  
471 along the Equator and favor the transition to the La Niña phase (Kug and Kang 2006; Ohba  
472 and Ueda 2007; Ohba and Watanabe 2012; Kug and Ham 2012; Wang 2019). Some recent  
473 studies suggest, on the other hand, that the major forcing for the NWPA during boreal winter  
474 is from the central Pacific, as rainfall anomalies and associated diabatic warming over the IO  
475 are very modest during this season and, thus, cannot trigger an atmospheric Kelvin wave  
476 intruding into the western Pacific (Chen et al. 2016; Wu et al. 2017ab; Li and Hsu 2017; Li et  
477 al. 2017b). Luo et al. (2010) and Izumo et al. (2010) also found that IOD has highly  
478 significant impacts on both ENSO variability and predictability.

479 Our decoupling experiments allow to reinvestigate the significance of the IO feedback  
480 during boreal winter and spring in a consistent modeling framework, and may challenge the  
481 view that this IO negative feedback is weak during boreal winter (Li et al. 2017b). Moreover,  
482 what may be the impact of IO variability on the ENSO-ISM relationships despite the modest  
483 role of IOD on ISM suggested above?

484 To answer this question, Figure 8a-b displays the Niño-3.4 SST-ISM lead-lag  
485 correlations and the monthly Niño-3.4 SST standard deviations in the FTIC and FSEIC



486 experiments in which IO or SEIO SST variability is removed (see Table 1 and Section 2c).  
487 The negative Niño-3.4 SST-ISM correlations before the ISM onset and ISMR standard  
488 deviations during (early) boreal summer are substantially amplified in FTIC, while the ISMR  
489 seasonal cycle remains the same (Fig. 8a, c-d). In other words, FTIC showed prominent  
490 negative correlations with the Niño-3.4 SSTs *before rather than synchronous and after* the  
491 ISM. These errors, already present in CTRL, are thus significantly amplified in FTIC. By  
492 contrast, the simulated ISM and ENSO relationships in FSEIC closely resemble the observed  
493 ones (Fig. 8a). FSEIC is able to recover the slow change of sign of the ISM-ENSO  
494 correlations from positive to negative from year -1 to year 0 as observed. The differences  
495 between the simulated and observed correlations for the different lags have been tested for  
496 statistical significance with a Fisher Z transform approach (see Morrison 1990; pp 102-106)  
497 and FSEIC is the only simulation, which does not show any statistical difference with  
498 observations at the 95% confidence level for any lags. Interestingly, this success is not related  
499 to changes in the SST mean state (Figs. 1d and 2a) or ISMR seasonal cycle (Fig. 8d).  
500 Evidences will be presented in this section to demonstrate that these large improvements are  
501 related to changes in the characteristics of the model's El Niño during its decaying phase in  
502 FSEIC.

503 Consistent with a strong damping of ENSO variability by IO variability, the Niño-3.4 SST  
504 standard deviations increase significantly in FTIC (Fig. 8b). This increase is more prominent  
505 during boreal spring and summer, which correspond to ENSO transitions. This suggests that  
506 ENSO is less phase-locked to the annual cycle and that the length of ENSO events has  
507 increased in FTIC. On the other hand, the Niño-3.4 SST variability is decreased in FSEIC  
508 compared to both CTRL and FTIC. Interestingly, the Niño-3.4 SST monthly standard  
509 deviations, as simulated by FSEIC, now match the observed Niño-3.4 SST standard  
510 deviations during ENSO transitions (e.g. boreal spring). This suggests that ENSO events start  
511 later or last shorter in FSEIC. In order to quantify more rigorously these differences between  
512 the simulations, power spectra of the simulated Niño-3.4 SST time series are shown in Figure  
513 9. Both FTIC and FSEIC spectra are strikingly different from CTRL spectrum, e.g. ENSO is  
514 predominantly biennial or quasi-biennial in FSEIC while it lasts much longer in FTIC than in  
515 CTRL and these results are statistically significant at the 95% confidence level (Fig. 9).  
516 These important differences not only confirm that IO SSTs interact significantly with ENSO  
517 in the SINTEX-F2 model, but also suggest that the spatial and temporal details of IO SST  
518 variability play also a substantial role in the damping (or enhancement) and the length of the  
519 simulated ENSO.

520 We first focus on the synchronous ISM-ENSO teleconnection during boreal summer. The  
521 rainfall and 850-hPa regressions onto the ISMR index in FTIC have spatial patterns similar to  
522 those in CTRL during JJ and AS (Fig. 6e-f). This demonstrates again that ENSO forcing is  
523 critical and that local ocean-atmosphere coupling is only a secondary factor for the generation  
524 of these anomalous patterns. However, consistent with a stronger (and partly erroneous)  
525 ENSO teleconnection during early summer (Fig. 8a,c), the rainfall and 850-hPa anomalies  
526 have greater amplitude almost everywhere in the Indo-Pacific region during JJ than AS in  
527 FTIC (Fig. 6e-f). The monsoon flux over the western IO is further increased during the whole  
528 ISM in FTIC compared to CTRL, resulting in enhanced rainfall over the Arabian Sea and  
529 western India (Fig. 6e-f). Interestingly, a negative IOD like rainfall pattern also emerges along  
530 the Equator in FTIC despite the absence of local coupling, demonstrating again the strong  
531 ENSO forcing on the regional rainfall pattern (Fig. 6e, f). In FSEIC, the 850-hPa wind and  
532 rainfall regressions onto the ISMR index show that the ISM-ENSO teleconnection gradually  
533 increases from JJ to AS in contrast to FTIC (Figs. 6g-h).

534 However, the interactions between the Indian and Pacific oceans need not to be restricted  
535 to the boreal summer, and may concern other seasons as well, and still be important for  
536 shaping the ENSO-ISM relationship. Indeed, the negative lagged correlations of ISM with the  
537 Niño-34 SSTs during boreal winter and spring preceding ISM are very different, while the  
538 lead correlations between ISM and the following Niño-34 SST anomalies are very similar  
539 among the simulations (Fig. 8a), suggesting that the IOB may play a prominent role in the  
540 ENSO damping and explain the relative success of FSEIC compared to both CTRL and FTIC.

541 In order to test the hypothesis that the IOB produces a direct forcing on ENSO itself in the  
542 simulations, Figs. 10 to 12 display lagged regressions of tropical Indo-Pacific SST, rainfall,  
543 850-hPa wind and depth of 20°C isotherm anomalies during year +1 onto the December-  
544 January (DJ) Niño-3.4 SST index in the different simulations. Here, numerals such as 0 and  
545 +1 denote the ENSO developing (0) and decay (+1) years, respectively. Lagged regressions of  
546 850-hPa stream function, 850- and 200-hPa velocity potential anomalies during February-  
547 March (FM) of year +1 onto DJ Niño-3.4 SST index are also displayed in Fig 13 in order to  
548 diagnose physical processes responsible of the differences in ENSO damping between the  
549 simulations.

550 First, the results confirm the robustness of the link between the decaying phase of El Niño  
551 events and the IOB in CTRL and FSEIC in which ocean-atmosphere coupling is (partly)  
552 active in the IO (Figs. 10a and 10c). As expected, the SST regression patterns over the  
553 tropical Pacific during FM(+1) show enhanced warming over the central and eastern Pacific

554 and a well-developed cold horseshoe signature in the extra-tropical Pacific in all experiments.  
555 However, the Pacific SST anomalies are more confined meridionally in the FSEIC  
556 experiment during FM(+1) and AM(+1) (Fig. 10c). The IO warming is basin-wide and  
557 uniform in CTRL and absent by design in FTIC (Fig. 10a-b). On the other hand, decoupling  
558 the SEIO enhances the warming in the other IO areas, especially over the western IO, and  
559 leads to a substantial anomalous SST equatorial gradient in the IO during FM(+1) (Fig. 10c).  
560 The modulation of the Walker circulation during El Niño induces local atmospheric  
561 subsidence (Fig. 13), which reduces cloud cover and forces near surface wind anomalies (Fig.  
562 11) and surface heat fluxes over the IO, and, thus, initiates and sustains the basin-wide IOB  
563 warming in CTRL and FSEIC (Klein et al. 1999; Alexander et al. 2002; Schott et al. 2009).

564 Next, Fig. 10 illustrates that the Pacific SST anomalies during boreal spring and summer of  
565 the decaying phase of ENSO are different in the three simulations. First, the El Niño related  
566 warm anomalies persist up to the boreal summer of year +1 in FTIC, but have disappeared by  
567 the end of boreal spring in FSEIC, while CTRL exhibits an intermediate evolution of tropical  
568 Pacific SST anomalies between these two extreme scenarios. These results are consistent with  
569 the different seasonal phase locking of Niño-3.4 SST variability in the simulations (Fig. 8b).  
570 Cold equatorial Pacific SST anomalies emerge during boreal summer in FSEIC, signaling the  
571 beginning of a La Niña episode. Because of the slow response of the IO to ENSO forcing, the  
572 IOB persists until the following boreal summer in both CTRL and FSEIC (Figs. 10a, c).  
573 Surprisingly, IOB warming is more intense in FSEIC than in CTRL despite of the quick  
574 demise of the El Niño SST pattern in FSEIC.

575 In summary, these results first demonstrate that the IOB has a considerable impact on the  
576 anomalous climate during the decaying phase of ENSO in SINTEX-F2 (Figs. 10 to 12),  
577 consistent with previous studies (Kug and Kang 2006; Ohba and Ueda 2007; Ohba and  
578 Watanabe 2012; Terray et al. 2016; Cai et al. 2019; Wang 2019). However, the significant  
579 differences between FSEIC and CTRL highlight that the details of the IOB warming,  
580 especially the SST differences between the western IO and SEIO during late boreal winter, do  
581 also matter a lot, which is an important result of the present study.

582 In order to provide an explanation, we focus on the atmospheric circulation during FM(+1)  
583 (Fig. 13), when the differences in IO SST anomalies between the simulations are particularly  
584 well developed (Fig. 10 first row of panels). During the peak of El Niño, all the simulations  
585 show an enhanced convective activity over the central equatorial Pacific and reduced rainfall  
586 over the Maritime Continent (not shown) and such anomalous diabatic heating pattern is still  
587 seen during FM(+1), despite that Maritime Continent rainfall starts already to recover during

588 this season (Fig. 11 first row of panels). The associated 850- and 200-hPa velocity potential  
589 anomalies reveal that the Walker circulation shifts eastward in response to the diabatic  
590 warming (cooling) due to the enhanced (reduced) convection over the central Pacific  
591 (Maritime Continent). The 850-hPa stream function anomalies reveal a quadrupole structure  
592 straddling the Equator and asymmetric about it, which can be interpreted as a direct Rossby  
593 wave response to this anomalous convective pattern in all experiments (Fig. 13a,e,i). This  
594 four-cell structure plays a fundamental role in the maintenance of the rainfall and anomalous  
595 low-level circulation (Fig. 11, first row of panels), and is largely contributed by the tropical  
596 Pacific SST anomalies, which are very similar in the different simulations.

597 Focusing now on the IO and west Pacific, important differences in both the position and  
598 strength of the NWPA and its southern hemisphere counterpart exist between the simulations  
599 (Fig. 13a,e,i), which can directly be related to the different SST anomalies in the IO (Fig. 10  
600 first row of panels). First, in both CTRL and FSEIC, in which IOB warming is present, these  
601 anticyclones are shifted eastward and the NWPA becomes stronger than its southern  
602 hemisphere counterpart. Second, these shifted anticyclones are further considerably  
603 strengthened in FSEIC compared to CTRL in response to a well-developed anomalous east-  
604 west circulation over the equatorial IO due to the enhanced SST zonal gradient in this run  
605 (Fig. 13j-k). Such an anomalous Walker cell over the IO is obviously absent in CTRL (Fig.  
606 13b-c) in which the IOB warming is basin-wide and uniform (Fig. 10a). Moreover, this zonal  
607 dipole pattern in 850- and 200-hPa velocity potential anomalies in FSEIC, with anomalous  
608 ascending (descending) motion in the western (eastern) IO, is exactly similar to the observed  
609 pattern (see Fig. 5 of Li et al. 2017b for example).

610 Such modulation of the Walker circulation over the IO seems to be the key-factor to  
611 explain the strength of the IO negative feedback on ENSO during boreal winter in FSEIC as  
612 positive rainfall anomalies are restricted to a small region in western IO and are seen in all the  
613 runs, including FTIC in which IOB is absent (Fig. 11 first row panel). It is therefore unlikely  
614 that such restricted rainfall anomaly is associated with a well-defined basin-wide diabatic  
615 heating anomaly, which may cause an atmospheric Kelvin wave response over the Western  
616 Pacific, consistent with the view of Chen et al. (2016) or Li et al. (2017b). On the other hand,  
617 the modulation of the Walker circulation over the IO induces more upper level convergence  
618 over the Maritime Continent and western Pacific in FSEIC (Fig. 13k), which strengthens  
619 considerably the NWPA and its southern counterpart (Fig. 13i), and the associated easterly  
620 wind anomalies over the western Pacific (Fig. 13l).

621 Thus, our results highlight that the IOB warming, but also the existence of an anomalous

622 SST gradient over the equatorial IO during boreal winter, are critical for the strength/position  
623 of the NWPA, the generation of the associated anomalous easterlies over the western  
624 equatorial Pacific and, finally, the turnabout of ENSO phases in our CGCM. As the equatorial  
625 easterlies over the western equatorial Pacific are weaker in FTIC compared to CTRL and  
626 much weaker than those in FSEIC (Fig. 13d,h,l), they may consequently have only a modest  
627 influence on the Pacific SSTs one or two seasons later (Fig. 10b).

628 This working hypothesis, that the IOB warming and its spatial structure play a fundamental  
629 role in hastening the ENSO transition through the generation of upwelling oceanic Kelvin  
630 waves in both CTRL and FSEIC, is further supported by the regressions of the 20°C isotherm  
631 depth (e.g. a proxy for the thermocline depth) anomalies during year +1 (Fig. 12). The slow  
632 ocean dynamic adjustment in the tropical Pacific is different among the simulations. The  
633 negative heat content anomalies (e.g. a shallower thermocline), which appear over the western  
634 Pacific during the peak of El Niño, have a faster eastward propagation along the equatorial  
635 Pacific in CTRL and FSEIC (Fig. 12a,c). This eastward propagation is very gradual in CTRL  
636 (Fig. 12a), but somewhat abrupt, from AM to JJ of year +1, in FSEIC (Fig. 12c). Moreover,  
637 FSEIC shows shallower 20°C isotherm anomalies than CTRL along the equatorial Pacific  
638 (Fig. 12a,c). The enhanced zonal wind stress associated with the stronger easterly wind  
639 anomalies over the western Pacific in FSEIC induces an enhanced vertical displacement of  
640 the equatorial thermocline via Ekman pumping and the upwelling Kelvin wave response is  
641 stronger in FSEIC. The arrival of these eastward Kelvin waves into the central equatorial  
642 Pacific erodes the local surface warming (Fig. 10c). Consequently, lagging two to four  
643 months behind the remote easterly wind anomalies over the western Pacific, cold SST  
644 anomalies emerged quickly and expand westward over the central and eastern equatorial  
645 Pacific during boreal summer of year +1 in FSEIC. Such ENSO transition is delayed in CTRL  
646 (Fig. 10a) in agreement with the more modest NWPA and associated anomalous easterlies  
647 over the western equatorial Pacific during FM(+1) seen in this run.

648 Consistent with the reversal of the equatorial anomalous SST gradient in the Pacific,  
649 anomalous easterlies gradually emerge over the central Pacific during boreal summer of year  
650 +1 in FSEIC (Fig. 11c). On the other hand, the easterly wind anomalies over the western  
651 Pacific, seen during FM(+1) in FTIC, disappear quickly during AM(+1) (Fig. 11b). As a  
652 consequence, both the El Niño-related thermocline and SST anomalies persist and we observe  
653 a revival of the El Niño-related circulation and rainfall anomalous patterns during boreal  
654 summer of year (+1) in FTIC. Finally, the evolution is again intermediate between these two  
655 extremes in CTRL (Fig. 11a).

656 In agreement with these different El Niño evolutions in the simulations, we note that ISM  
657 is again highly deficient during boreal summer of the decaying phase of El Niño in FTIC,  
658 weak during JJ only in CTRL and, finally, normal or strong in FSEIC (Fig. 11 third and  
659 fourth rows ). These assertions are further verified by computing regressions of boreal  
660 summer 200-hPa velocity potential anomalies onto the Niño-3.4 SST time series of the  
661 preceding boreal winter (Fig. 14). During El Niño decaying year, La Niña (El Niño)  
662 conditions prevail with a significantly strengthening (reduction) of the Walker circulation in  
663 the Indo-Pacific region in FSEIC (FTIC). Furthermore, strong (weak) upper level divergence  
664 is seen over the Indian region in FSEIC (FTIC) consistent with enhanced (reduced) diabatic  
665 heating associated with the ISM rainfall anomalies during boreal summer (Figs 11b-c third  
666 and fourth rows). Again, CTRL exhibits intermediate conditions between these two extremes  
667 with upper level convergence (e.g. subsidence) over India during JJ, but rather normal  
668 conditions in AS (Fig. 14a-b). These results demonstrate that the success to capture the  
669 seasonal evolution of the correlations between Niño-3.4 SSTs and ISMR in FSEIC (Fig. 8a) is  
670 plausibly related to the ability of this particular configuration to simulate correctly the space-  
671 time evolution of the tropical Pacific SSTs during the decaying phase of El Niño. In both  
672 CTRL and FSEIC, IOB is well developed during boreal summer of year +1 (Fig. 10a, c), but a  
673 weak (strong) ISM is simulated in CTRL (FSEIC) demonstrating that the remote ENSO  
674 forcing is still dominant in explaining the ISM response during the decaying year of El Niño.  
675 This may be understood by the fact that IOB SST anomalies are basin-wide, but of reduced  
676 amplitude during boreal summer of year +1, so that they are a significant factor for ISM  
677 variability if, and only if, tropical Pacific SST and circulation anomalies are rather weak at  
678 this stage (Figs. 10 and 11).

679

## 680 **5. Summary and Discussion**

### 681 5.a Summary

682 The present work examines the ability of current CGCMs in simulating the ISM-ENSO  
683 relationships. Consistent with previous studies (Terray et al. 2012; Jourdain et al. 2013;  
684 Sabeerali et al. 2019), we found that CGCMs used for climate projections or dynamical  
685 seasonal forecasting still exhibit significant errors with respect to this metric. Simulated ISM-  
686 ENSO correlations are already significant and negative several months before the ISM onset  
687 contrary to what is observed..The origins of these common errors in the simulation of the  
688 ISM-ENSO relationship are investigated based on several decoupling experiments performed

689 with the SINTEX-F2 CGCM.

690 First, a sensitivity experiment in which ENSO variability is removed (e.g. FTFC)  
691 demonstrates that relationships between ISM and SST variability outside the tropical Pacific  
692 are weak at interannual time scales in the absence of ENSO. As IODs still exist with  
693 exaggerated amplitude in such decoupling experiment in which ENSO is removed (Fig. 2b,  
694 see also Cretat et al. 2017, 2018), this suggests that IOD impact on ISM is not directly  
695 responsible of the modulated ISM-ENSO relationship in this CGCM.

696 On the other hand, two others decoupling experiments in which IO SST variability is fully  
697 removed (e.g. FTIC) or only in the SEIO (e.g. FSEIC) show a large spread in simulated  
698 ENSO length and decay pace. ENSO becomes predominantly quasi-biennial in FSEIC while  
699 ENSO lasts much longer in FTIC compared to CTRL. In FSEIC, cold SST and easterly wind  
700 anomalies emerge quickly in the central equatorial Pacific during boreal summer of El Niño  
701 decay years. By contrast, FTIC displays the persistence of a fully developed El Niño-like  
702 pattern during the same time period. Finally, CTRL exhibits an intermediate evolution  
703 between the above two extreme scenarios, despite that IOB warming is present in CTRL as in  
704 FSEIC. As a consequence, weak ISMs are simulated during both El Niño developing and  
705 decaying years in FTIC (and also in CTRL), while weak ISMs and El Niños tend to be  
706 followed systematically by normal/strong ISMs and La Niñas in FSEIC. Further analysis  
707 confirms that these differences in the decay phase of El Niño are responsible of the errors in  
708 the simulated ISM-ENSO lead-lag relationships in CTRL.

709 Overall, our results suggest that progresses or changes in the simulation of the ISM-ENSO  
710 relationships in CGCMs can be traced back to modifications of ENSO length and seasonal  
711 phase-locking related to the impact of the IOB on ENSO and details of its spatial structure  
712 (namely the equatorial SST gradient in the IO) during boreal winter. Surprisingly, the IOD or  
713 the direct relationship between IO SSTs and ISM only play a secondary role in modulating  
714 the ISM-ENSO relationship in our CGCM despite that it simulates an exaggerated IOD  
715 variability, with or without ENSO, as many current CGCMs (Cai and Cowan 2013; Li et al.  
716 2016; Annamalai et al. 2017).

717 Our results are also fully consistent with the suggestion that a fast ENSO transition in the  
718 tropical Pacific and the persistence of IOB from boreal winter to boreal summer provide  
719 favorable conditions for the occurrence of a strong ISM (Yang et al. 2007, 2010; Boschat et  
720 al. 2011, 2012; Park et al. 2010; Chowdary et al. 2016, 2017). The emerging La Niña episode  
721 will strengthen the ISM circulation and the IOB warming will tend to enhance the local

722 evaporation, moisten the atmosphere and reduce the moist stability; all factors tend to increase  
723 ISM precipitation in FSEIC (Park et al. 2010).

## 724 5.b Discussion

725 In order to understand the physical processes, which are responsible of these different  
726 evolutions, it is important to highlight first that mean-state errors are the same in the different  
727 decoupled experiments analyzed here (Fig. 1a-d). This means that westward extension of cold  
728 tongue bias and of the El Niño pattern in the tropical Pacific, which have been identified as  
729 playing a seminal role in errors of the simulated ENSO decay pace and responsible of large  
730 inter-model spread in CMIP3 or CMIP5 models (Kug and Ham 2012; Tao et al. 2016; Li et al.  
731 2019) cannot explain directly (or alone) the differences among the runs.

732 The crucial role of the IOB warming in hastening the El Niño to La Niña transition (well  
733 illustrated by the differences between the FTIC and CTRL) is consistent with previous studies  
734 (Watanabe and Jin 2002; Kug and Kang 2006; Ohba and Ueda 2007; Xie et al. 2009, 2016;  
735 Terray et al. 2016). On the other hand, the fact that the rapid decay of El Niño events and the  
736 emergence of a La Niña developing event are likely to be the effect of the spatial details of the  
737 IOB warming during boreal winter, rather than its simple existence or intensity per se, is an  
738 important result of the present study as many recent works suggest that the IOB capacitor  
739 effect is key mainly during the boreal summer of El Niño decay year, but not before (Chen et  
740 al. 2016; Li et al. 20017b).

741 By comparing further CTRL and FSEIC, it is demonstrated that the simulation skill of the  
742 ENSO decay pace and ISM-ENSO relationship is partly controlled by the (erroneous) SEIO  
743 SST variability simulated in the CGCM. As illustrated by Fig. 2b, this SST variability bias is  
744 characterized by an amplified seasonal cycle of SST variability with a double peak, one  
745 during boreal winter (e.g. February) and the other during boreal fall (e.g. September).  
746 Furthermore, Fig. 15b demonstrates that this erroneous double peak of SST variability in the  
747 SEIO is a bias shared by most CMIP5 models, despite SINTEX-F2 and CMIP5 models differ  
748 in their seasonal cycle of SST mean bias over SEIO (Figs. 2a and 15a). Most CMIP5 models  
749 exhibit a cold bias throughout the seasonal cycle in the SEIO while the SINTEX-F2 model  
750 exhibits a warm SST bias during boreal winter and a cold SST bias during boreal fall.

751 The cold SST mean bias and enhanced SST variability in the SEIO during boreal fall is a  
752 well-known and long-standing problem of both the SINTEX-F2 and CMIP5 models (Fischer  
753 et al. 2005; Terray et al. 2012; Prodhomme et al. 2014; Cretat et al. 2017; Cai and Cowan  
754 2013; Li et al. 2015, 2016; Jourdain et al. 2016; Annamalai et al. 2017). Surprisingly, this



755 exaggerated SEIO SST variability during boreal fall, linked to the IOD, cannot explain the  
756 differences in ENSO evolution as described above.

757 On the other hand, the exaggerated SST variability in the SEIO during boreal winter in  
758 both SINTEX-F2 and CMIP5 models is less explored in the literature despite our analysis  
759 demonstrates its significant role on ENSO length and ISM-ENSO relationship. A working  
760 hypothesis for the existence of this SST bias is, however, the cold tongue bias and the  
761 associated excessive westward extension of El Niño SST warming along the equatorial  
762 Pacific found in most current CGCMs (Masson et al. 2012; Wang et al. 2014; Li and Xie  
763 2012, 2014). These tropical Pacific biases have already been suggested to be very detrimental  
764 for a realistic simulation of both the ENSO-NWPA relationship and the negative IO feedback  
765 in CMIP5 models as they produce erroneous warm SST anomalies in the equatorial western  
766 Pacific during the El Niño decaying year, obstructing the formation of the anomalous NWPA  
767 (Kug and Ham 2012; Tao et al. 2016; Li et al. 2019). In order to investigate if the SEIO SST  
768 mean and variability biases are remotely forced by tropical Pacific errors in the SINTEX-F2  
769 model, the annual cycle of monthly means and standard deviations of SEIO SSTs simulated in  
770 the FTFC and FTFC-obs decoupled experiments (see Section 2c) are displayed in Fig. 2. As  
771 ENSO variability is removed in both FTFC and FTFC-obs, but the Pacific mean-state errors  
772 are corrected only in FTFC-obs (Figs. 1e and 2a), the results suggest that SEIO biases during  
773 boreal fall have primarily local origins, as they are still present in both FTFC and FTFC-obs.  
774 On the other hand, those found during boreal winter are partly due to the remote Pacific  
775 forcing, as the erroneous SEIO SST variability during boreal winter is significantly damped in  
776 both FTFC and FTFC-obs, and the corresponding SST mean-state error is eliminated in  
777 FTFC-obs.

778 Thus, our results complement the previous studies, which focused primarily on the ocean-  
779 atmosphere and atmospheric processes over the equatorial western Pacific to explain the  
780 relationships between the ENSO decay pace, the NWPA and the local easterlies wind  
781 anomalies during boreal winter (Wang et al. 2000; Wu et al. 2017ab; Li et al. 2017b) and their  
782 degraded representation in current CGCMs (Kug and Ham 2012; Tao et al. 2016; Li et al.  
783 2019). In SINTEX-F2, the negative IO feedback on ENSO is severely damped not only  
784 because of the local response (over the western equatorial Pacific) to the westward shift of the  
785 El Niño-related anomalies over the tropical Pacific, but also because such erroneous ENSO  
786 pattern produces an unrealistic remote forcing in the SEIO. The removal of the erroneous  
787 SEIO SST variability in FSEIC is sufficient to restore a realistic strength and position of the  
788 NWPA and associated equatorial easterlies and a correct ISM-ENSO relationship (Fig. 8)

789 related to the reduction of ENSO length. As most CMIP5 models exhibit exactly the same  
790 SST variability biases in the SEIO (Fig. 15a), it is not unlikely that correcting or removing the  
791 SEIO errors in these CGCMs will lead to similar improvements in their simulation of ENSO  
792 and ISM-ENSO relationships.

793 Past analysis of observations illustrate that the synchronous ISM-ENSO relationships has  
794 weakened, but that the impact of El Niño decaying phase on ISM has strengthened during  
795 recent decades (Fig 3; Yang et al. 2007, 2010; Boschat et al. 2012). This is fully consistent  
796 with the hypothesis of an enhanced negative IO feedback associated with the sustained IO  
797 warming (especially in the western IO), partly induced by anthropogenic forcing (Roxy et al.  
798 2014) and the more frequent occurrence of the ENSO-related IO warming in recent decades  
799 (e.g., Boschat et al. 2012; Xie et al. 2016). Our numerical experiments do suggest that the IO  
800 is effectively an important climate element to explain the evolution of the ISM-ENSO system  
801 in the 20<sup>th</sup> century, as we are able to restore a realistic simulation of the lead-lag relationships  
802 between ISM and ENSO in a CGCM by simply eliminating the erroneous SST variability in  
803 the SEIO (Fig. 10a). In other words, it is conceivable that the intensified warming and  
804 negative feedback of the IO are important factors in the long-term variability of ENSO and  
805 ISM. This may have significant implications for our future projections of ENSO and ISM  
806 variations in a changing climate.

807 Additional testing with other CGCMs should therefore be performed in the future to  
808 promote a better understanding of the nature, seasonal timing and long-term behavior of the  
809 IO feedback on ENSO and ISM. However, the present work demonstrates that tropical SST  
810 biases in current CGCMs must be drastically reduced in order to shed new light on the  
811 mechanisms underlying the ISM variability. It is unlikely that state-of-the-art CGCMs will  
812 provide robust projections of ENSO or the ISM-ENSO relationship before prior reductions of  
813 systematic errors in the different tropical basins, like the cold tongue bias and the westward  
814 shift of the ENSO pattern in the tropical Pacific, the spurious IOD-like pattern in the IO or the  
815 unrealistic representation of the Atlantic zonal mode. Such systematic errors already preclude  
816 the robust attribution of the observed evolution of the ISM-ENSO relationships during the  
817 20<sup>th</sup> century (illustrated in Fig. 3) to the anthropogenic forcing with the current CMIP5  
818 simulations, as illustrated by the persistent controversies in the literature on this topic (Chen  
819 et al. 2010; Li and Ting 2015; Srivastava et al. 2015; Cash et al. 2017; Yun and Timmermann  
820 2018; Feba et al. 2019; Bodai et al. 2020).

821 As illustrated in this work and summarized in two recent reviews (Cai et al. 2019; Wang  
822 2019), interactions between the Indian, Pacific and also Atlantic basins play a seminal role in

823 explaining large-scale phenomena, like ENSO or ISM. From a modeling perspective, our  
824 results further highlight that remote SST mean-state and variability errors in one basin, for  
825 example the IO, can also interfere with SST biases in other basins like the Pacific and play a  
826 key-role in the simulation of ENSO or ISM in current CGCMs (see also Terray et al. 2016).  
827 In other words, this inter-basin framework is also fundamental to diagnose and correct the  
828 severe biases still affecting current CGCMs.

829

## 830 **Appendix**

831

832 The SST nudging is performed inside the SST equation of the ocean model. We suppressed  
833 the SST variability in a specific domain by applying a strong nudging of the SST toward a  
834 SST climatology computed from a control experiment or observations (see Table 1 and  
835 Section 2c for details). More precisely, this is done through a modification of the non-solar  
836 heat flux provided by the atmosphere to the ocean by adding a correction term that is  
837 proportional with the SST difference with the target climatology at each time step:

838

$$839 \quad \mathbf{Q}_{ns} = \mathbf{Q}_{ns} + \mathbf{dq/dt} * (\mathbf{SST} - \mathbf{SST}_{clim})$$

840

841  $\mathbf{Q}_{ns}$ : the nonsolar heat flux received from atmosphere

842  $\mathbf{dq/dt}$ :  $-2400 \text{ W m}^{-2} \text{ K}^{-1}$  (corresponds to the heat flux needed to warm a 50m thick ocean layer  
843 of 1 K during 1 day)

844  $\mathbf{SST}_{clim}$ : the target SST daily climatology, after a linear interpolation at each time step of the  
845 day

846

847 The very strong  $\mathbf{dq/dt}$  constant used here implies that the SST variability is almost suppressed  
848 in the selected domain. A Gaussian smoothing is finally applied in a transition zone in both  
849 longitude and latitude at the limits of the SST restoring domains (see Table 1).

850

851 This approach is interesting because the ocean dynamics (e.g. also thermocline variations)  
852 will be consistent with the SST in the ocean model, which is not the case when the SST is  
853 changed in the coupling interface of the coupled model as it is commonly done in many  
854 past/recent studies.

855

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1477 **Figure captions**

1478 **Figure 1:** **a)** SST means difference ( $^{\circ}\text{C}$ ) between CTRL (years 11-210) and HadISST1.1  
1479 dataset (years 1979-2012); **b)** SST means difference ( $^{\circ}\text{C}$ ) between FTPC (years 11-110) and  
1480 CTRL; **c)** SST means difference ( $^{\circ}\text{C}$ ) between FTIC (years 11-110) and CTRL; **d)** SST means  
1481 difference ( $^{\circ}\text{C}$ ) between FSEIC (years 11-60) and CTRL and **e)** SST means difference ( $^{\circ}\text{C}$ )  
1482 between FTPC-obs (years 11-50) and CTRL. The nudging domains for FTPC, FTIC and  
1483 FSEIC experiments are shown, respectively, in panels b (blue), c (green) and d (purple).

1484

1485 **Figure 2:** **a)** Monthly means of the SEIO ( $90^{\circ}\text{E}$ - $115^{\circ}\text{E}$  and  $0^{\circ}\text{S}$ - $10^{\circ}\text{S}$ ; unit:  $^{\circ}\text{C}$ ) SST time  
1486 series from HadISST1.1 dataset (for the 1950-2012 and 1979-2012 periods) and the CTRL,  
1487 FSEIC, FTPC and FTPC-obs experiments performed with the SINTEX-F2 model; **b)**  
1488 Monthly standard deviations of the SEIO SST time series (unit:  $^{\circ}\text{C}$ ) from HadISST1.1 dataset  
1489 (for the 1950-2012 and 1979-2012 periods) and the CTRL, FSEIC, FTPC and FTPC-obs  
1490 experiments.

1491

1492 **Figure 3:** **a)** Lead-lag correlations between AIR time series and bi-monthly Niño-3.4 SSTs  
1493 for different time periods, starting from the beginning of the previous year (e.g. year - 1) to  
1494 the end of the following year of the monsoon (e.g. year +1) and **b)** Same as **a)** but using the  
1495 ISMR time series derived from the IMD dataset. X-axis indicates the lag (in 2 months  
1496 interval) for a 36 months period starting one year before the developing year of ISM and Y-  
1497 axis is the amplitude of the correlation. Thus, the coefficients corresponding to -1, 0, +1  
1498 month lags refer, respectively, to the correlations between AIR in year 0 (e.g. JJAS ISM  
1499 rainfall) and April-May, June-July and August-September Niño-3.4 SSTs, also during year 0,  
1500 and so on. Lead-lag correlations between ISMR time series estimated from GPCP dataset and  
1501 bi-monthly Niño-3.4 SSTs for the period 1979-2012 are also displayed in panels **a** and **b** for  
1502 comparison. All time series are detrended with a LOESS method (Cleveland and Devlin  
1503 1988). Black diamond symbols indicate correlations that are above the 95% significance  
1504 confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999  
1505 samples.

1506

1507 **Figure 4:** **a)** Lead-lag correlations between ISMR time series and bi-monthly Niño-3.4 SSTs,  
1508 starting from the beginning of the previous year (e.g. year - 1) to the end of the following year  
1509 of the monsoon (e.g. year +1) as simulated in the control experiments (CTRL) of the

1510 SINTEX-F2 and CFSv2 models and observed during the period 1979-2012. All time series  
1511 are detrended with a LOESS method (Cleveland and Devlin 1988). Black diamond symbols  
1512 indicate correlations that are above the 95% significance confidence level according to a  
1513 phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples. For observations, GPCP  
1514 and HadISST1.1 datasets are used, respectively, for estimating the ISMR and Niño-3.4 SST  
1515 time series; **b**) Monthly standard deviations of the Niño-3.4 SST time series from HadISST1.1  
1516 dataset (for the 1950-2012 and 1979-2012 periods) and the CTRL experiments of the  
1517 SINTEX-F2 and CFSv2 models; **c**) Monthly standard deviations of the ISMR time series  
1518 from GPCP dataset (for 1979-2012 period) and the CTRL experiments of the SINTEX-F2 and  
1519 CFSv2 models; **d**) Monthly means of the ISMR time series from GPCP dataset (for 1979-  
1520 2012 period) and the CTRL experiments of the SINTEX-F2 and CFSv2 models.

1521

1522 **Figure 5:** Same as Fig. 4, but for historical runs of 25 CMIP5 models. For CMIP models, the  
1523 period 1950-2000 is considered and time series are only linearly detrended. Statistical  
1524 significance is not plotted, but correlation coefficients with an absolute value greater than  
1525 0.25 are above the 95% significance confidence level according to a Student two-tailed t test.  
1526 For observations, GPCP and HadISST1.1 datasets are used, respectively, for estimating the  
1527 ISMR and Niño-3.4 SST time series. The ensembles mean and spread across the CMIP5  
1528 models are also displayed. The spread is computed as the ensemble mean plus and minus the  
1529 ensemble standard deviation (thick black circles in all panels).

1530

1531 **Figure 6:** **(a, b)** June-July and August-September mean rainfall (shading) and 850-hPa wind  
1532 (vectors) anomalies regressed against the ISMR index (e.g. JJAS ISM rainfall) in the CTRL  
1533 experiment; **(c, d)** same as **(a, b)** but for the FTFC experiment; **(e, f)** same as **(a, b)** but for the  
1534 FTIC experiment and **(g, h)** same as **(a, b)** but for the FSEIC experiment. Unit for the rainfall  
1535 and 850-hPa wind regression coefficients are, respectively, in mm/day by mm/day and m/s by  
1536 mm/day. Only rainfall and 850-hPa wind regression coefficients above the 95% significance  
1537 confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999  
1538 samples are shown.

1539

1540 **Figure 7:** **(a, b, c, d, e, f)** correlation coefficients between the ISMR index (e.g. JJAS ISM  
1541 rainfall) and bi-monthly SST anomalies from February-March to December-January of year 0  
1542 in the FTFC experiment. **(g, h, i, j, k, l)** same as **(a, b, c, d, e, f)** but for bi-monthly 200-hPa

1543 velocity potential anomalies. Correlation coefficients above the 95% significance confidence  
1544 level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are  
1545 contoured. In **(a, b, c, d, e, f)**, correlation coefficients in the Pacific nudging domain are  
1546 masked.

1547

1548 **Figure 8:** Same as Fig. 4, but for the FTIC and FSEIC experiments done with the SINTEX-  
1549 F2 model. The results for the control (CTRL) and FTIC experiments performed with the  
1550 SINTEX-F2 model are also displayed for comparison (only in panels **c** and **d** for FTIC).

1551

1552 **Figure 9:** Power spectra of Niño-34 SST anomalies for CTRL (red), FTIC (green) and FSEIC  
1553 (blue) experiments. The bottom axis is the period (unit: year), the left axis is variance (unit:  
1554 °C<sup>2</sup>) and both axes are in logarithm scale. Dashed red curves show the point-wise 95%  
1555 confidence limits for the Niño-34 SST spectrum estimated from the CTRL experiment.

1556

1557 **Figure 10: a)** bi-monthly SST anomalies regressed against the December-January Niño-3.4  
1558 SST time series for the following February-March, April-May, June-July, August-September  
1559 and October-November (e.g. in year +1) in the CTRL experiment; **b)** same as **a)** but for the  
1560 FTIC experiment; **c)** same as **a)** but for the FSEIC experiment. Unit for the SST regression  
1561 coefficients is in K by K. Regression coefficients reaching the 10% significance level  
1562 according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are  
1563 contoured.

1564

1565 **Figure 11: a)** bi-monthly mean rainfall (shading) and 850-hPa wind (vectors) anomalies  
1566 regressed against the December-January Niño-3.4 SST time series for the following February-  
1567 March, April-May, June-July, August-September and October-November (e.g. in year +1) in  
1568 the CTRL experiment; **b)** same as **a)** but for the FTIC experiment; **c)** same as **a)** but for the  
1569 FSEIC experiment. Unit for the rainfall and 850-hPa wind regression coefficients are,  
1570 respectively, in mm/day by K and m/s by K. Only rainfall and 850-hPa wind regression  
1571 coefficients reaching the 10% significance level according to a phase-scrambling bootstrap  
1572 test (Ebisuzaki 1997) with 999 samples are shown.

1573

1574 **Figure 12:** **a)** bi-monthly mean 20°C isotherm depth (a proxy for thermocline depth)  
1575 anomalies regressed against the December-January Niño-3.4 SST time series for the  
1576 following February-March, April-May, June-July, August-September and October-November  
1577 (e.g. in year +1) in the CTRL experiment; **b)** same as **a)** but for the FTIC experiment; **c)** same  
1578 as **a)** but for the FSEIC experiment. Unit for the 20°C isotherm depth regression coefficients  
1579 is in m by K. Regression coefficients reaching the 10% significance level according to a  
1580 phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.

1581

1582 **Figure 13:** **(a, b, c, d)** February-March 850-hPa stream function, 850-hPa velocity potential,  
1583 200-hPa velocity potential and 850-hPa zonal wind anomalies regressed against the preceding  
1584 December-January Niño-3.4 SST time series in the CTRL experiment; **(e, f, g, h)** same as **(a,**  
1585 **b, c, d)** but for the FTIC experiment; **(i, j, k, l)** same as **(a, b, c, d)** but for the FSEIC  
1586 experiment. Units for the stream function and velocity potential regression coefficients are in  
1587  $10^5 \text{ m}^2 \text{ s}^{-1}$  by K and for 850-hPa zonal wind in  $\text{m s}^{-1}$  by K. Regression coefficients reaching  
1588 the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997)  
1589 with 999 samples are contoured.

1590

1591 **Figure 14:** **(a, b)** June-July and August-September 200-hPa velocity potential anomalies  
1592 regressed against the preceding December-January Niño-3.4 SST time series in the CTRL  
1593 experiment; **(c, d)** same as **(a, b)** but for the FTIC experiment; **(e, f)** same as **(a, b)** but for the  
1594 FSEIC experiment. Units for the velocity potential regression coefficients are in  $10^5 \text{ m}^2 \text{ s}^{-1}$  by  
1595 K. Regression coefficients reaching the 10% significance level according to a phase-  
1596 scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.

1597

1598 **Figure 15:** **a)** Monthly means of the SEIO (90°E-115°E and 0°S-10°S; unit: °C) SST time  
1599 series from HadISST1.1 dataset and the historical runs of 25 CMIP5 models; **b)** Monthly  
1600 standard deviations of the SEIO SST time series from HadISST1.1 dataset and the historical  
1601 runs of 25 CMIP5 models. All observed and simulated statistics are estimated from the period  
1602 1950-2000. The ensemble mean and spread across the CMIP5 models are also displayed.  
1603 The spread is computed as the ensemble mean plus and minus the ensemble standard  
1604 deviation (thick black circles in panels **a** and **b**).

1605

1606 **Figure S1: (a, b, c, d, e, f)** correlation coefficients between the ISMR index (e.g. JJAS ISM  
1607 rainfall) and bi-monthly SST anomalies from February-March to December-January of year 0  
1608 in the CTRL experiment. **(g, h, i, j, k, l)** same as **(a, b, c, d, e, f)** but for bi-monthly 200 hPa  
1609 velocity potential anomalies. Correlation coefficients above the 95% significance confidence  
1610 level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are  
1611 contoured.  
1612

1613 **Table captions**

1614 **Table 1:** Summary of the numerical experiments with their main characteristics, including  
1615 length, nudging domain and SST climatology used for the nudging in the Indian or Pacific  
1616 oceans decoupled experiments. The nudged experiments are the Forced Tropical Pacific  
1617 Climatology (FTPC) run, the Forced Tropical Indian Climatology (FTIC) run, the Forced  
1618 South-East Indian Climatology (FSEIC) run and, finally, the Forced Tropical Pacific observed  
1619 Climatology (FTPC-obs) run. See text for more details. For the FTPC and FTPC-obs  
1620 experiments only ocean grid-points in the Pacific are included in the correction or smoothing  
1621 areas and, similarly, for the FTIC and FSEIC experiments for their respective domains. The  
1622 different correction domains are also displayed in Figure 1b-d. The observed SST  
1623 climatology used in the FTPC-obs experiment is derived from the AVHRR only daily  
1624 Optimum Interpolation SST version 2 (OISSTv2) dataset for the 1982-2010 period (Reynolds  
1625 et al. 2007).

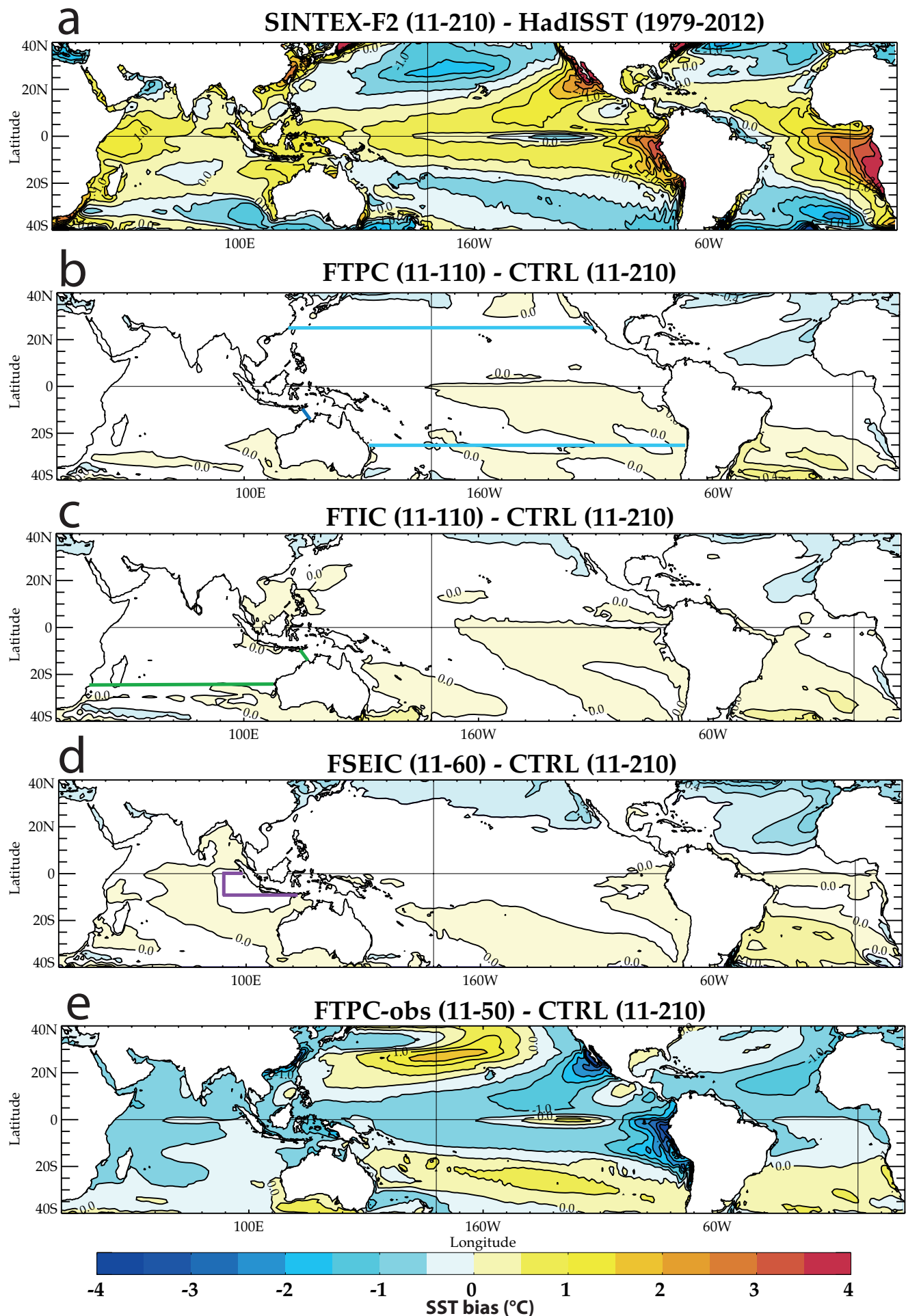
1626

1627 **Table S1:** Description of the 25 Coupled Model Inter-comparison Project phase 5 (CMIP5)  
1628 models used in our analysis. We use the historical coupled model experiments contributing to  
1629 CMIP5 (see, url: <http://pcmdi9.llnl.gov>). Here, the first member of each model's ensemble of  
1630 historical runs is used for all the analysis performed using CMIP5 models (e.g. Figures 5) and  
1631 computations are done for the period 1950-2000. Each model dataset has been re-gridded into  
1632 a  $2.5^{\circ} \times 2.5^{\circ}$  horizontal grid, using a first-order remapping procedure implemented in the  
1633 Climate Data Operators (<https://code.zmaw.de/projects/cdo>), before processing.

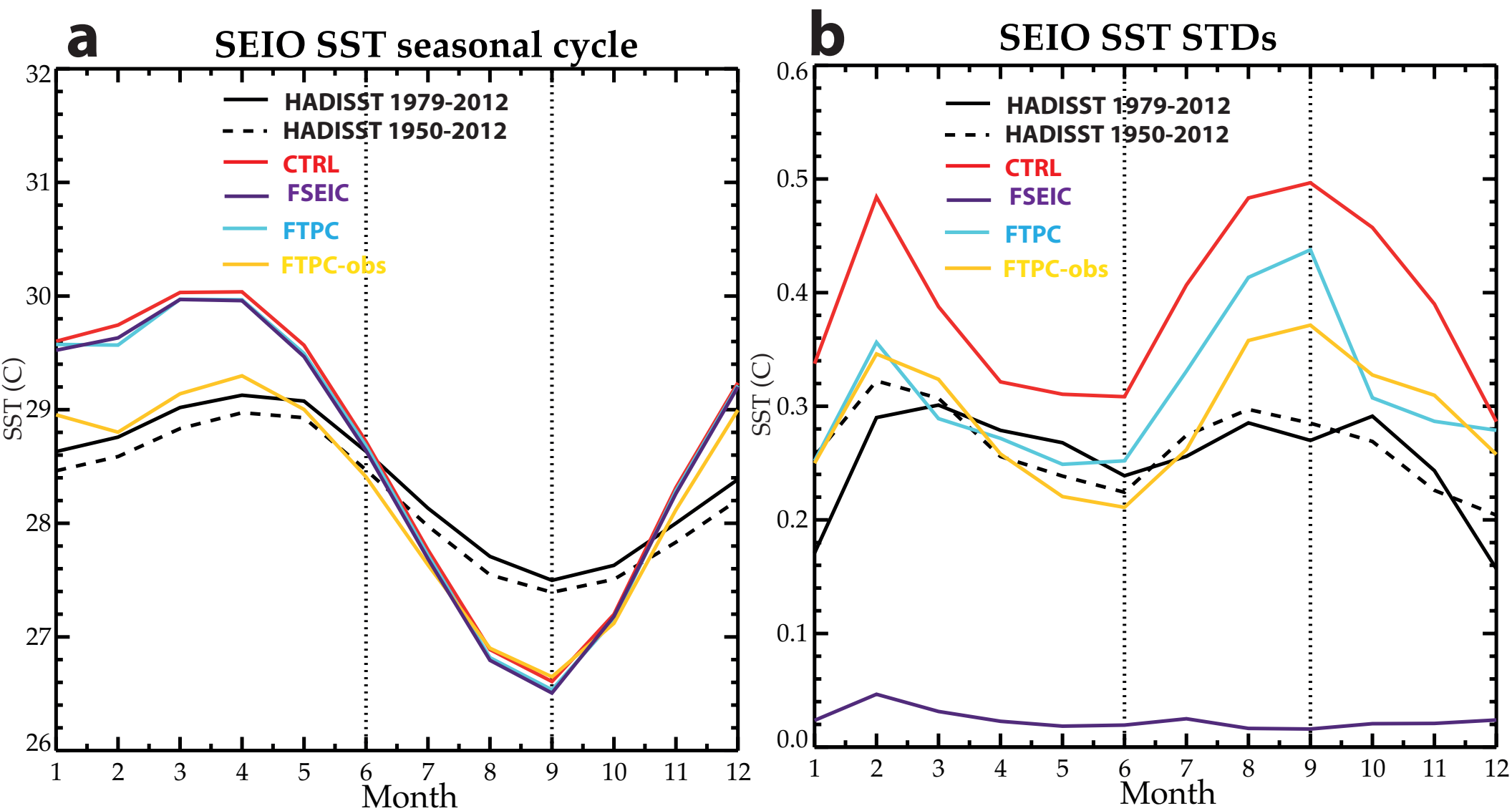
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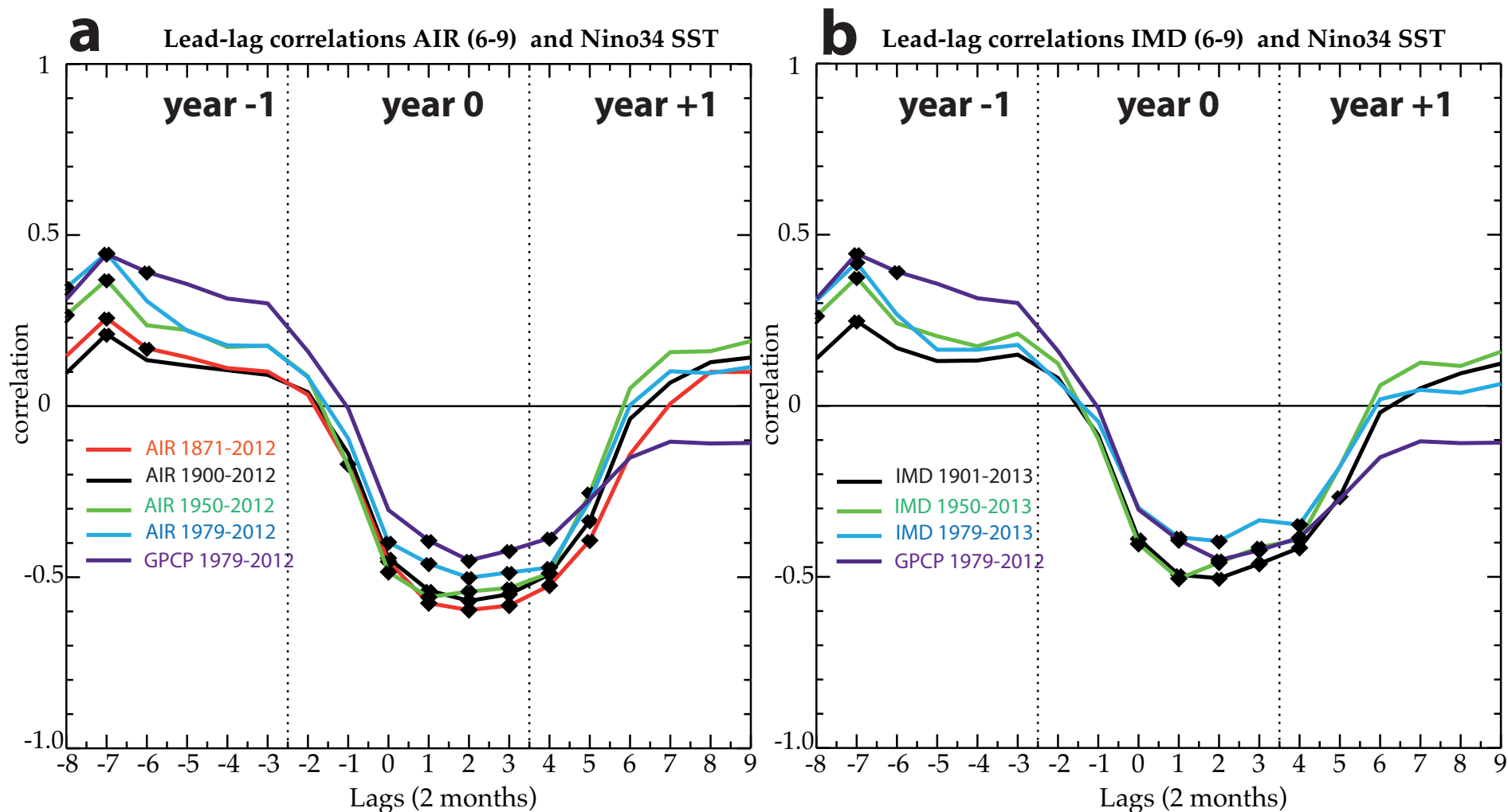




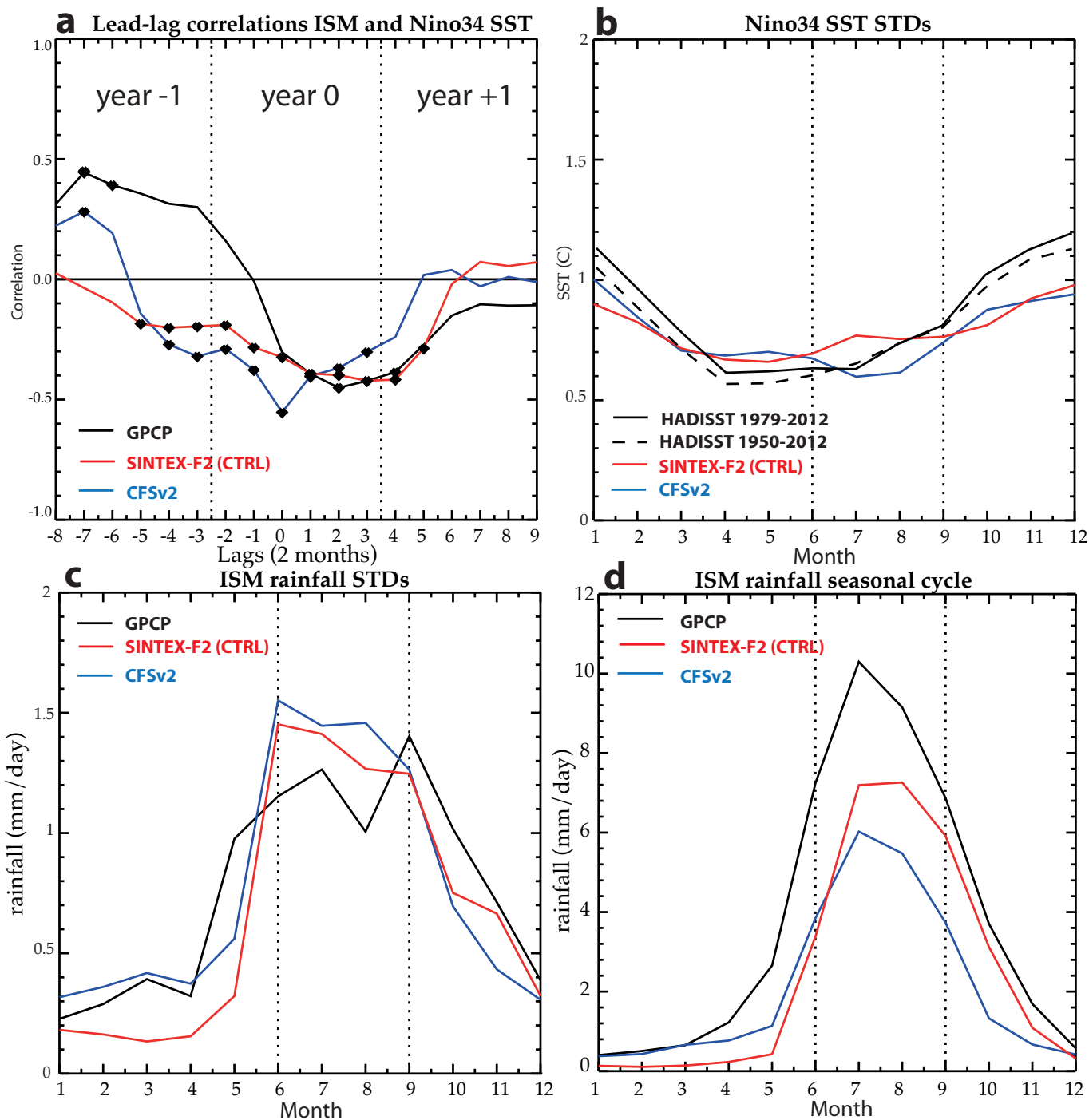
**Figure 1:** **a)** SST means difference (°C) between CTRL (years 11-210) and HadISST1.1 dataset (years 1979-2012); **b)** SST means difference (°C) between FTPC (years 11-110) and CTRL; **c)** SST means difference (°C) between FTIC (years 11-110) and CTRL; **d)** SST means difference (°C) between FSEIC (years 11-60) and CTRL and **e)** SST means difference (°C) between FTPC-obs (years 11-50) and CTRL. The nudging domains for FTPC, FTIC and FSEIC experiments are shown, respectively, in panels b (blue), c (green) and d (purple).



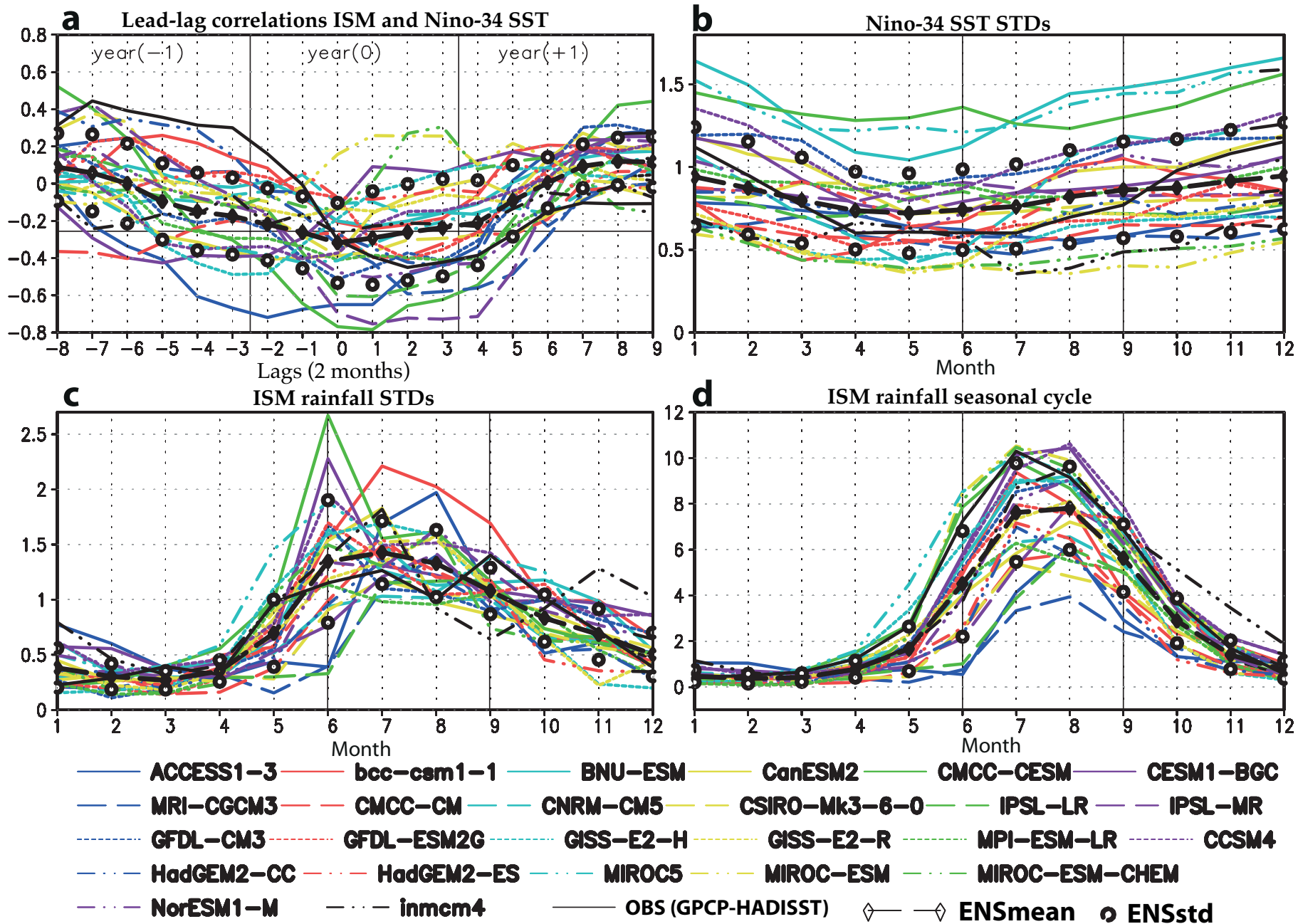
**Figure 2: a)** Monthly means of the SEIO (90°E-115°E and 0°S-10°S; unit: °C) SST time series from HadISST1.1 dataset (for the 1950-2012 and 1979-2012 periods) and the CTRL, FSEIC, FTPC and FTPC-obs experiments performed with the SINTEX-F2 model; **b)** Monthly standard deviations of the SEIO SST time series (unit: °C) from HadISST1.1 dataset (for the 1950-2012 and 1979-2012 periods) and the CTRL, FSEIC, FTPC and FTPC-obs experiments.



**Figure 3: a)** Lead-lag correlations between AIR time series and bi-monthly Niño-3.4 SSTs for different time periods, starting from the beginning of the previous year (e.g. year - 1) to the end of the following year of the monsoon (e.g. year +1) and **b)** Same as **a)** but using the ISMR time series derived from the IMD dataset. X-axis indicates the lag (in 2 months interval) for a 36 months period starting one year before the developing year of ISM and Y-axis is the amplitude of the correlation. Thus, the coefficients corresponding to -1, 0, +1 month lags refer, respectively, to the correlations between AIR in year 0 (e.g. JJAS ISM rainfall) and April-May, June-July and August-September Niño-3.4 SSTs, also during year 0, and so on. Lead-lag correlations between ISMR time series estimated from GPCP dataset and bi-monthly Niño-3.4 SSTs for the period 1979-2012 are also displayed in panels a and b for comparison. All time series are detrended with a LOESS method (Cleveland and Devlin 1988). Black diamond symbols indicate correlations that are above the 95% significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples.



**Figure 4:** **a)** Lead-lag correlations between ISMR time series and bi-monthly Niño-3.4 SSTs, starting from the beginning of the previous year (e.g. year - 1) to the end of the following year of the monsoon (e.g. year +1) as simulated in the control experiments (CTRL) of the SINTEX-F2 and CFSv2 models and observed during the period 1979-2012. All time series are detrended with a LOESS method. Black diamond symbols indicate correlations that are above the 95% significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples. For observations, GPCP and HadISST1.1 datasets are used, respectively, for estimating the ISMR and Niño-3.4 SST time series; **b)** Monthly standard deviations of the Niño-3.4 SST time series from HadISST1.1 dataset (for the 1950-2012 and 1979-2012 periods) and the CTRL experiments of the SINTEX-F2 and CFSv2 models; **c)** Monthly standard deviations of the ISMR time series from GPCP dataset (for 1979-2012 period) and the CTRL experiments of the SINTEX-F2 and CFSv2 models; **d)** Monthly means of the ISMR time series from GPCP dataset (for 1979-2012 period) and the CTRL experiments of the SINTEX-F2 and CFSv2 models.



**Figure 5:** Same as Fig. 4, but for historical runs of 25 CMIP5 models. For CMIP models, the period 1950-2000 is considered and time series are only linearly detrended. Statistical significance is not plotted, but correlation coefficients with an absolute value greater than 0.25 are above the 95% significance confidence level according to a Student two-tailed t test. For observations, GPCP and HadISST1.1 datasets are used, respectively, for estimating the ISMR and Niño-3.4 SST time series. The ensembles mean and spread across the CMIP5 models are also displayed. The spread is computed as the ensemble mean plus and minus the ensemble standard deviation (thick black circles in all panels).

### Regressions ISMR 850 hPa wind/rainfall - CTRL

Regressions - June-July

Regressions - August-September

### Regressions ISMR 850 hPa wind/rainfall - FTPC

Regressions - June-July

Regressions - August-September

### Regressions ISMR 850 hPa wind/rainfall - FTIC

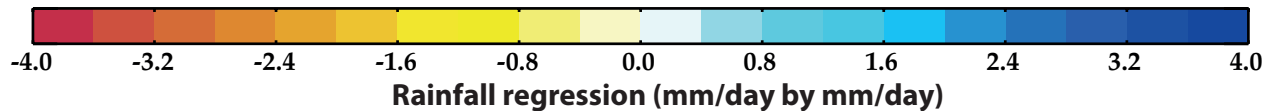
Regressions - June-July

Regressions - August-September

### Regressions ISMR 850 hPa wind/rainfall - FSEIC

Regressions - June-July

Regressions - August-September

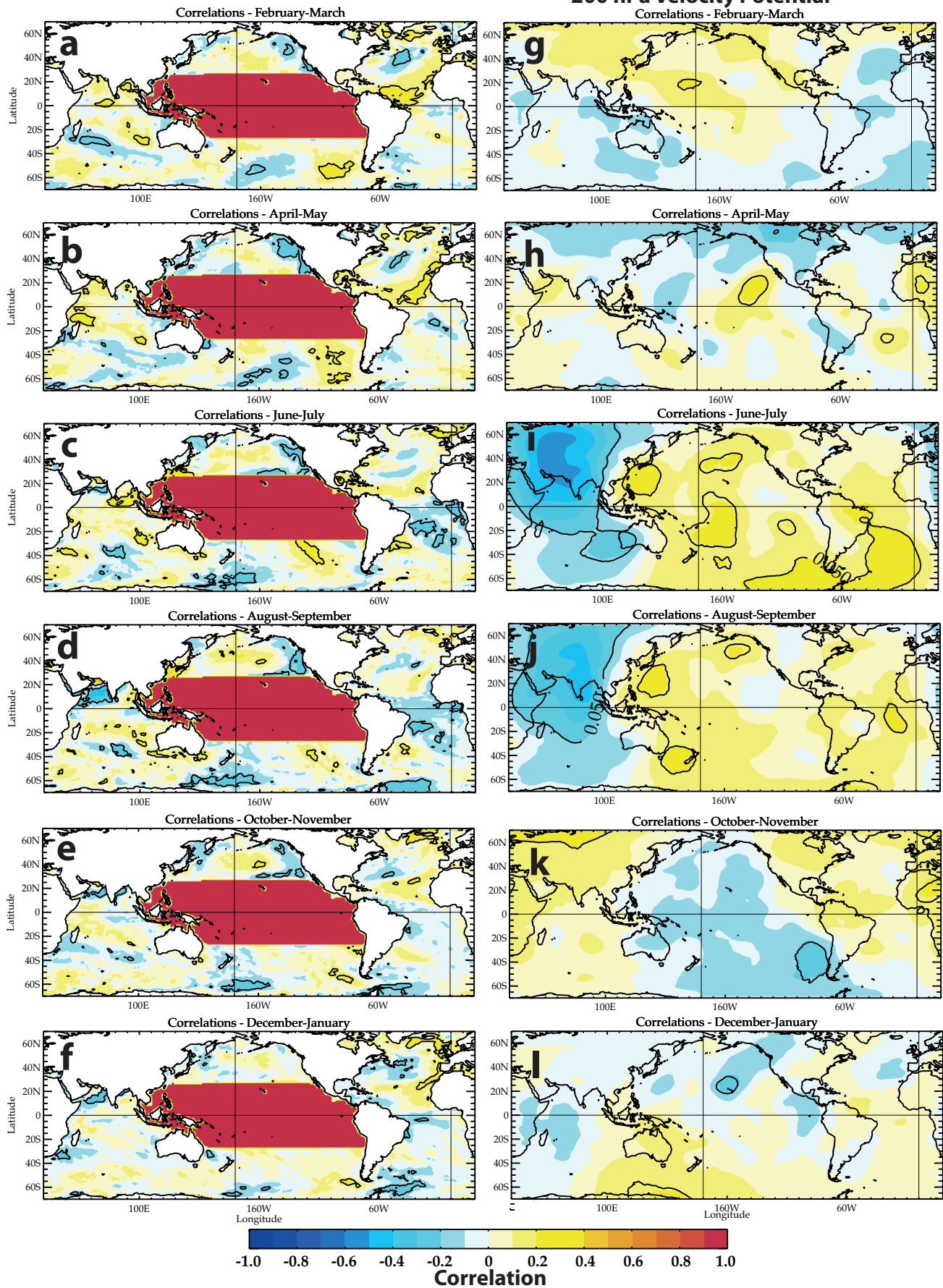


**Figure 6:** (a, b) June-July and August-September mean rainfall (shading) and 850-hPa wind (vectors) anomalies regressed against the ISMR index (e.g. JJAS ISM rainfall) in the CTRL experiment; (c, d) same as (a, b) but for the FTPC experiment; (e, f) same as (a, b) but for the FTIC experiment and (g, h) same as (a, b) but for the FSEIC experiment. Unit for the rainfall and 850-hPa wind regression coefficients are, respectively, in mm/day by mm/day and m/s by mm/day. Only rainfall and 850-hPa wind regression coefficients above the 95% significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are shown.

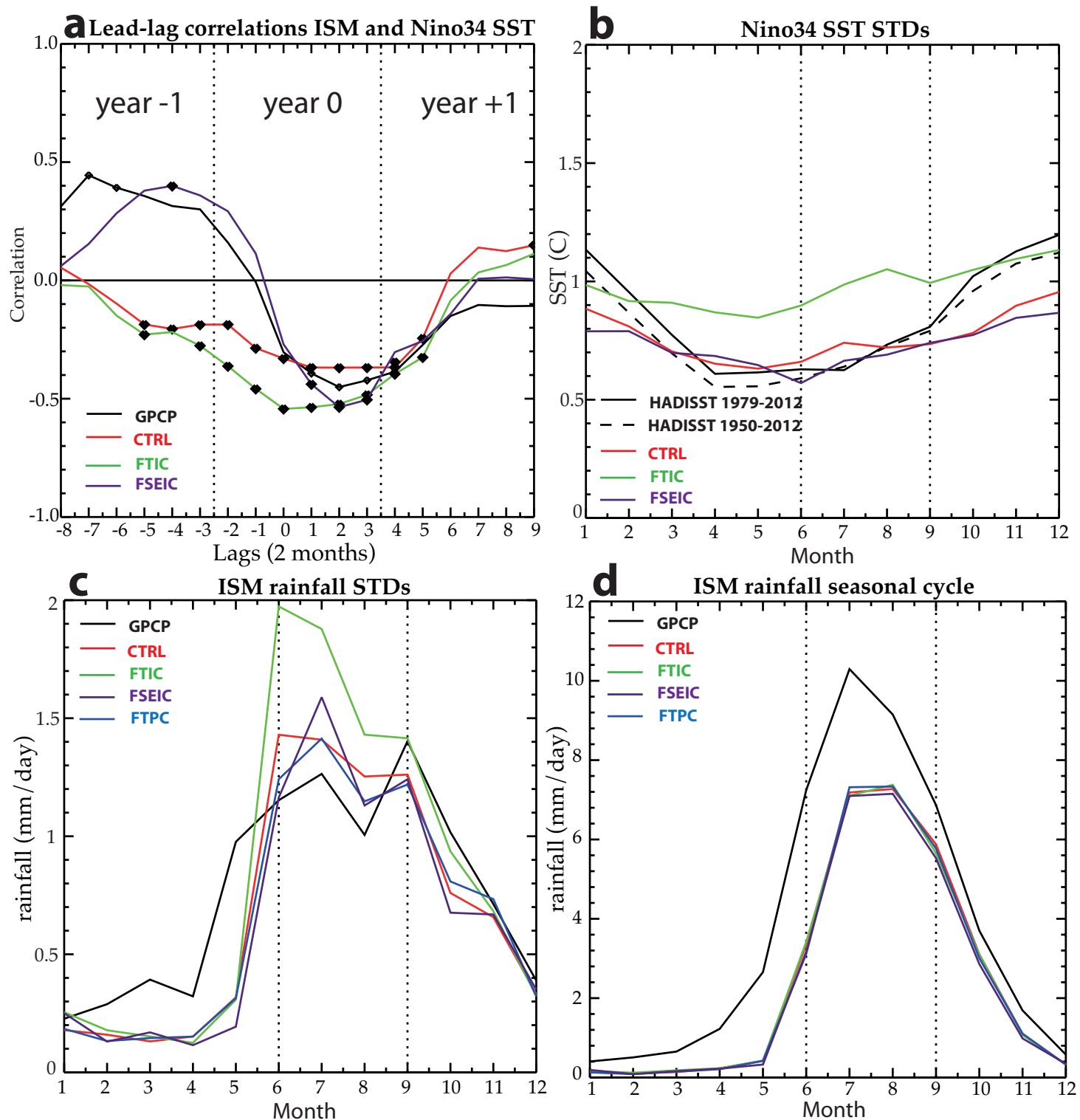
# Correlations ISM Rainfall (6-9)- FTPC - Year 0

## SST

## 200 hPa Velocity Potential

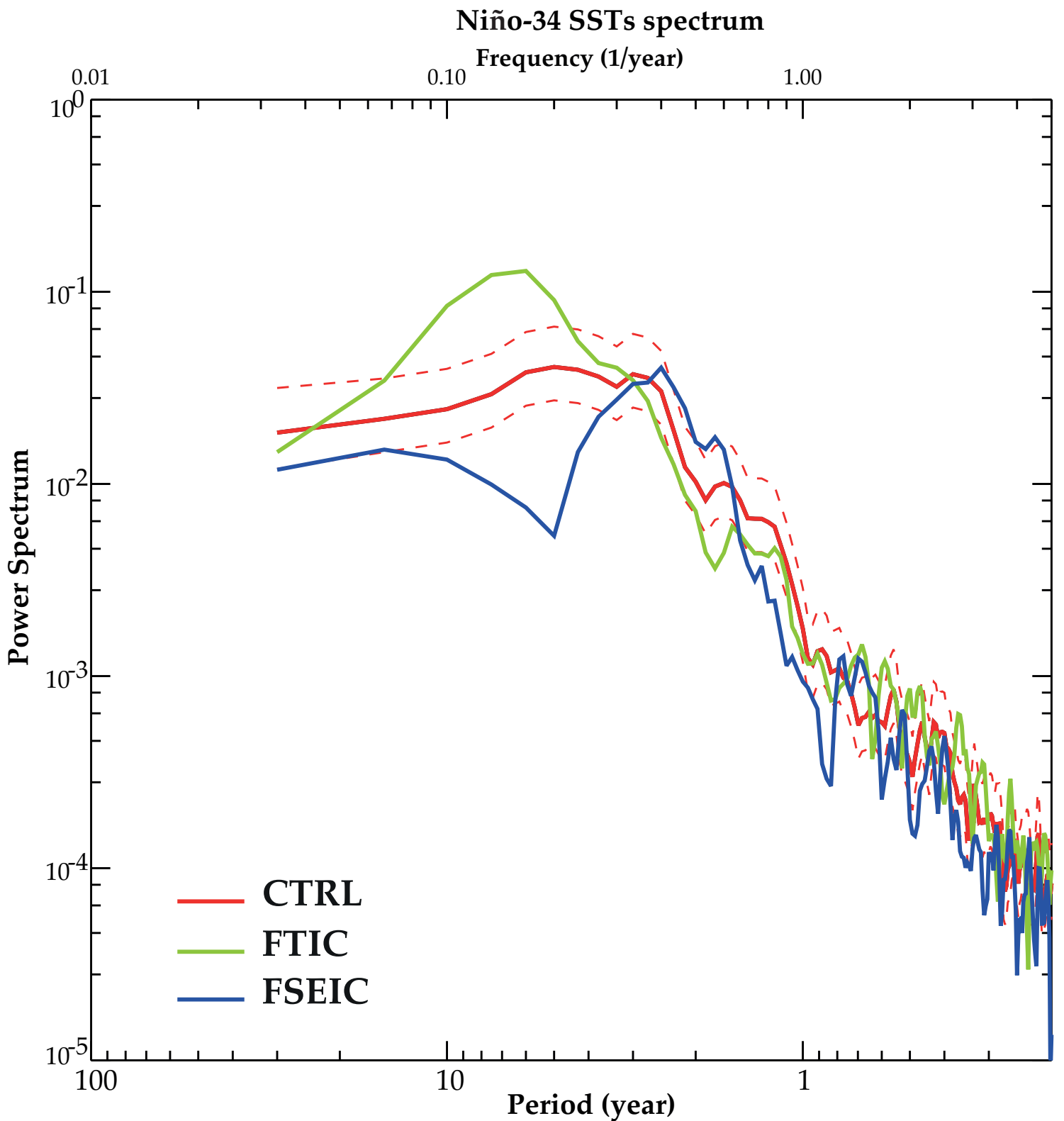


**Figure 7:** (a, b, c, d, e, f) correlation coefficients between the ISMR index (e.g. JJAS ISM rainfall) and bi-monthly SST anomalies from February-March to December-January of year 0 in the FTPC experiment. (g, h, i, j, k, l) same as (a, b, c, d, e, f) but for bi-monthly 200-hPa velocity potential anomalies. Correlation coefficients above the 95% significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured. In (a, b, c, d, e, f), correlation coefficients in the Pacific nudging domain are masked.



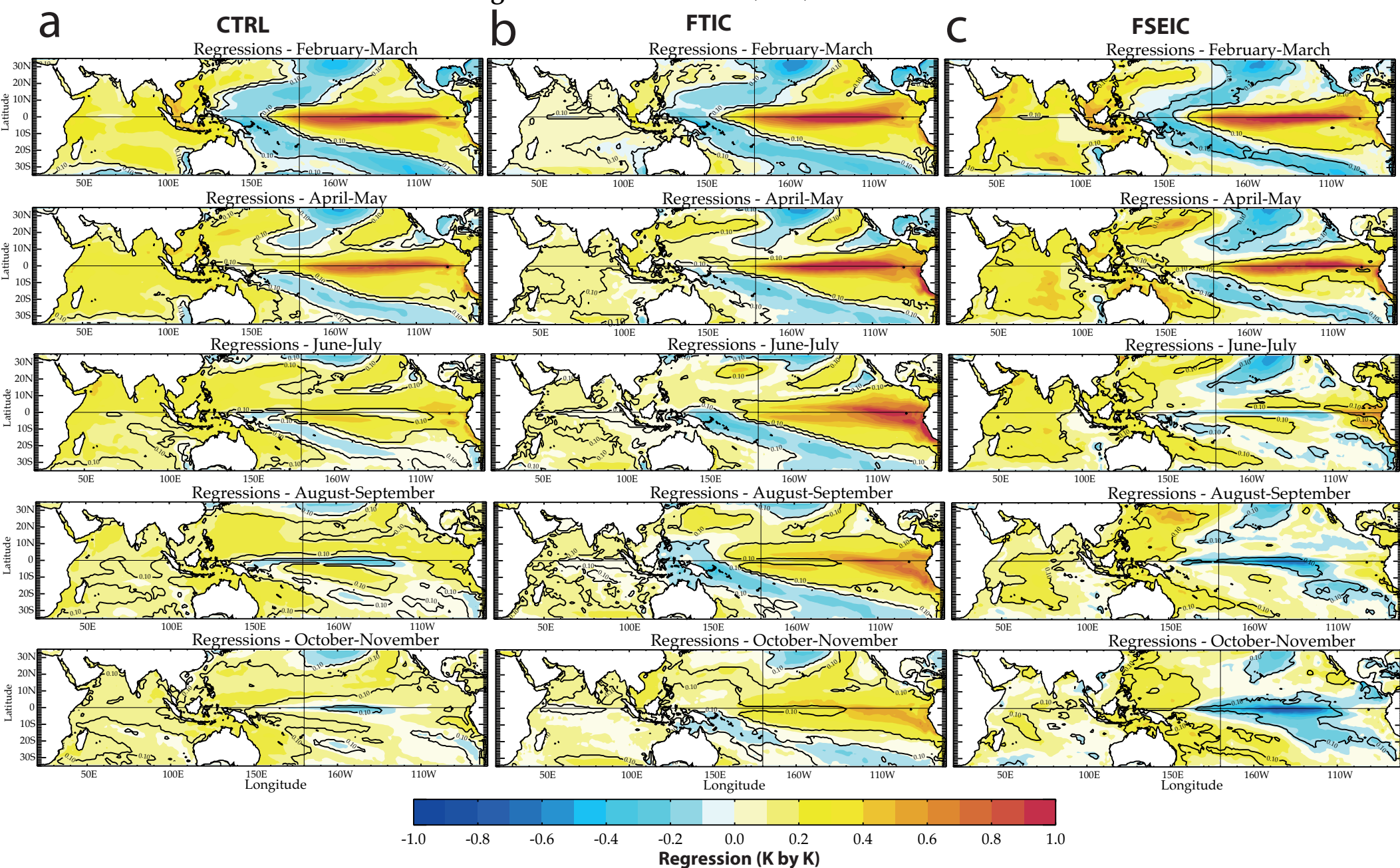
**Figure 8:** Same as Fig. 4, but for the FTIC and FSEIC experiments done with the SINTEX-F2 model. The results for the control (CTRL) and FTPC experiments performed with the SINTEX-F2 model are also displayed for comparison (only in panels **c** and **d** for FTPC).





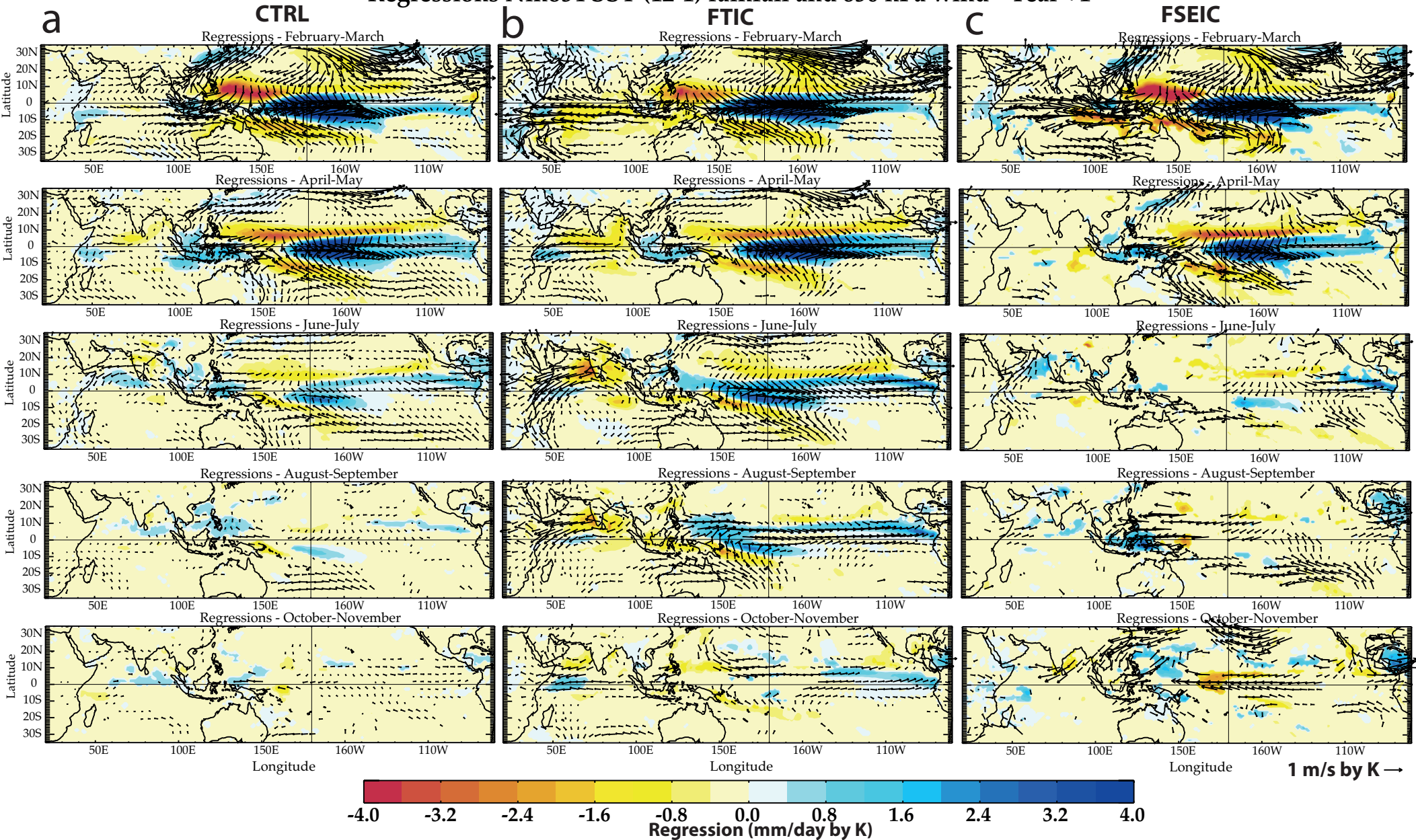
**Figure 9:** Power spectra of Niño-34 SST anomalies for CTRL (red), FTIC (green) and FSEIC (blue) experiments. The bottom axis is the period (unit: year), the left axis is variance (unit:  $^{\circ}\text{C}^2$ ) and both axes are in logarithm scale. Dashed red curves show the point-wise 95% confidence limits for the Niño-34 SST spectrum estimated from the CTRL experiment.

## Regressions Nino34 SST (12-1) SST - Year +1



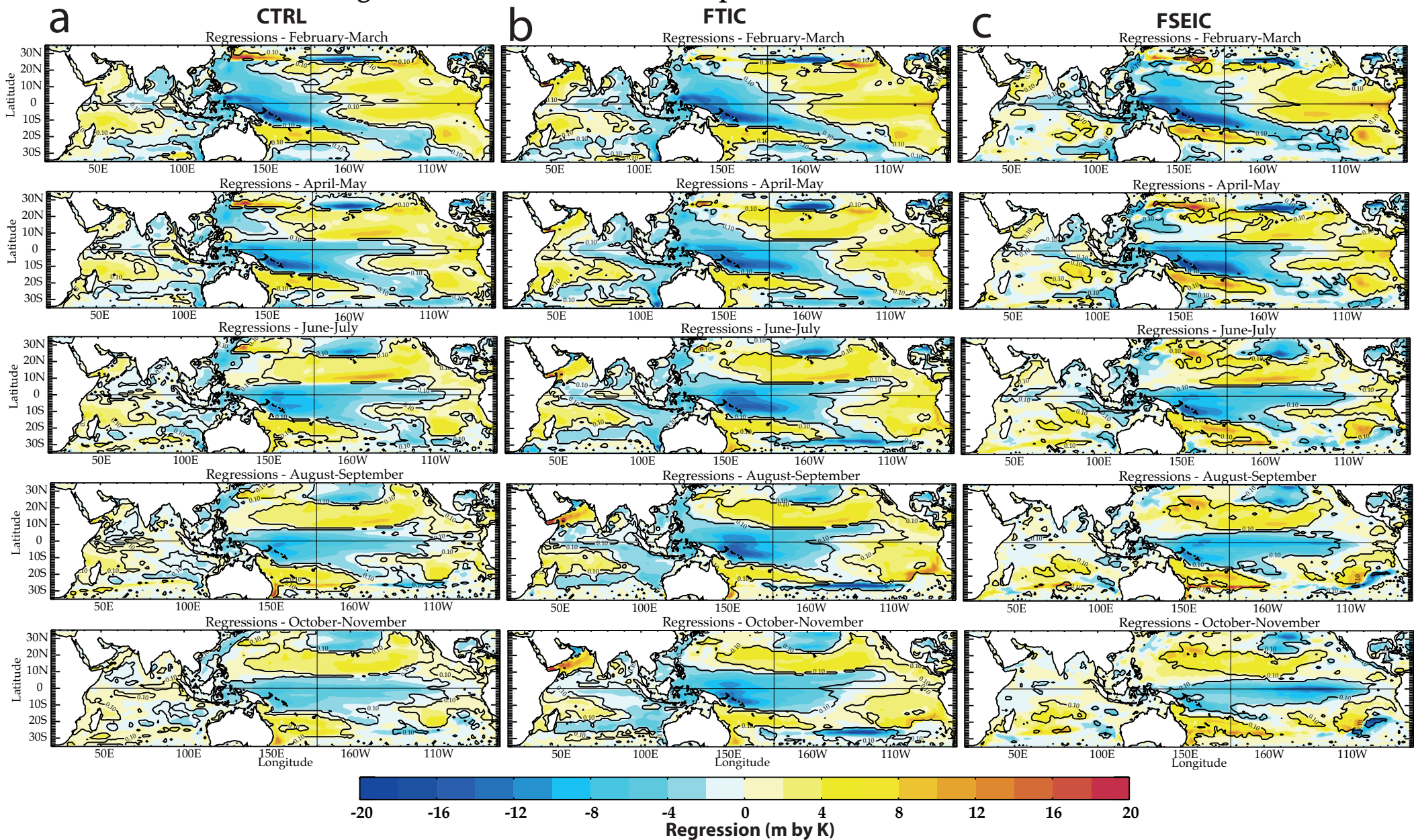
**Figure 10:** a) bi-monthly SST anomalies regressed against the December-January Niño-3.4 SST time series for the following February-March, April-May, June-July, August-September and October-November in the CTRL experiment; b) same as a) but for the FTIC experiment ; c) same as a) but for the FSEIC experiment. Unit for the SST regression coefficients is in K by K. Regression coefficients reaching the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.

## Regressions Niño34 SST (12-1) rainfall and 850 hPa wind - Year +1



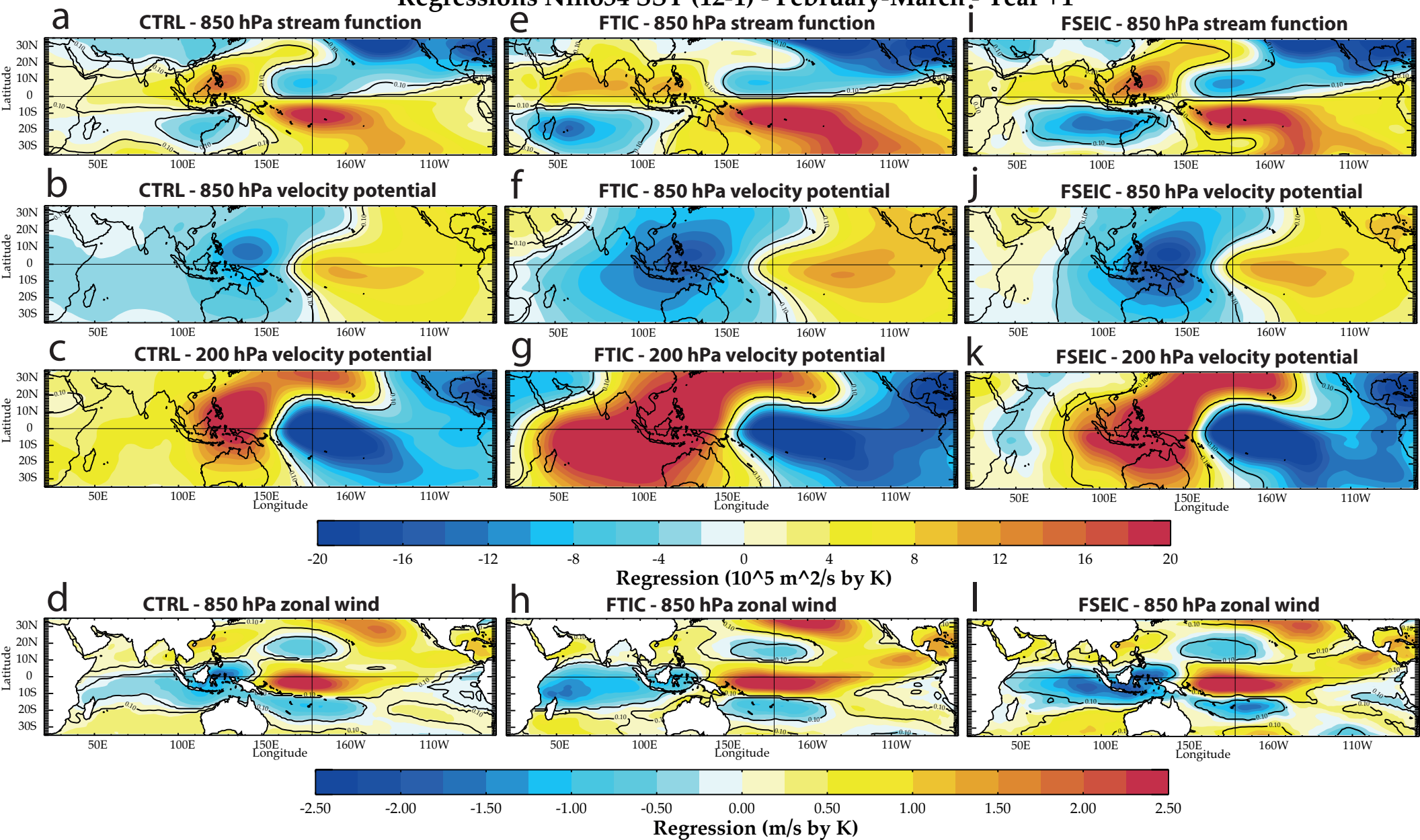
**Figure 11:** a) bi-monthly mean rainfall (shading) and 850-hPa wind (vectors) anomalies regressed against the December-January Niño-3.4 SST time series for the following February-March, April-May, June-July, August-September and October-November in the CTRL experiment; b) same as a) but for the FTIC experiment; c) same as a) but for the FSEIC experiment. Unit for the rainfall and 850-hPa wind regression coefficients are, respectively, in mm/day by K and m/s by K. Only rainfall and 850-hPa wind regression coefficients reaching the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are shown.

## Regressions Niño34 SST (12-1) depth of 20°C isotherm - Year +1



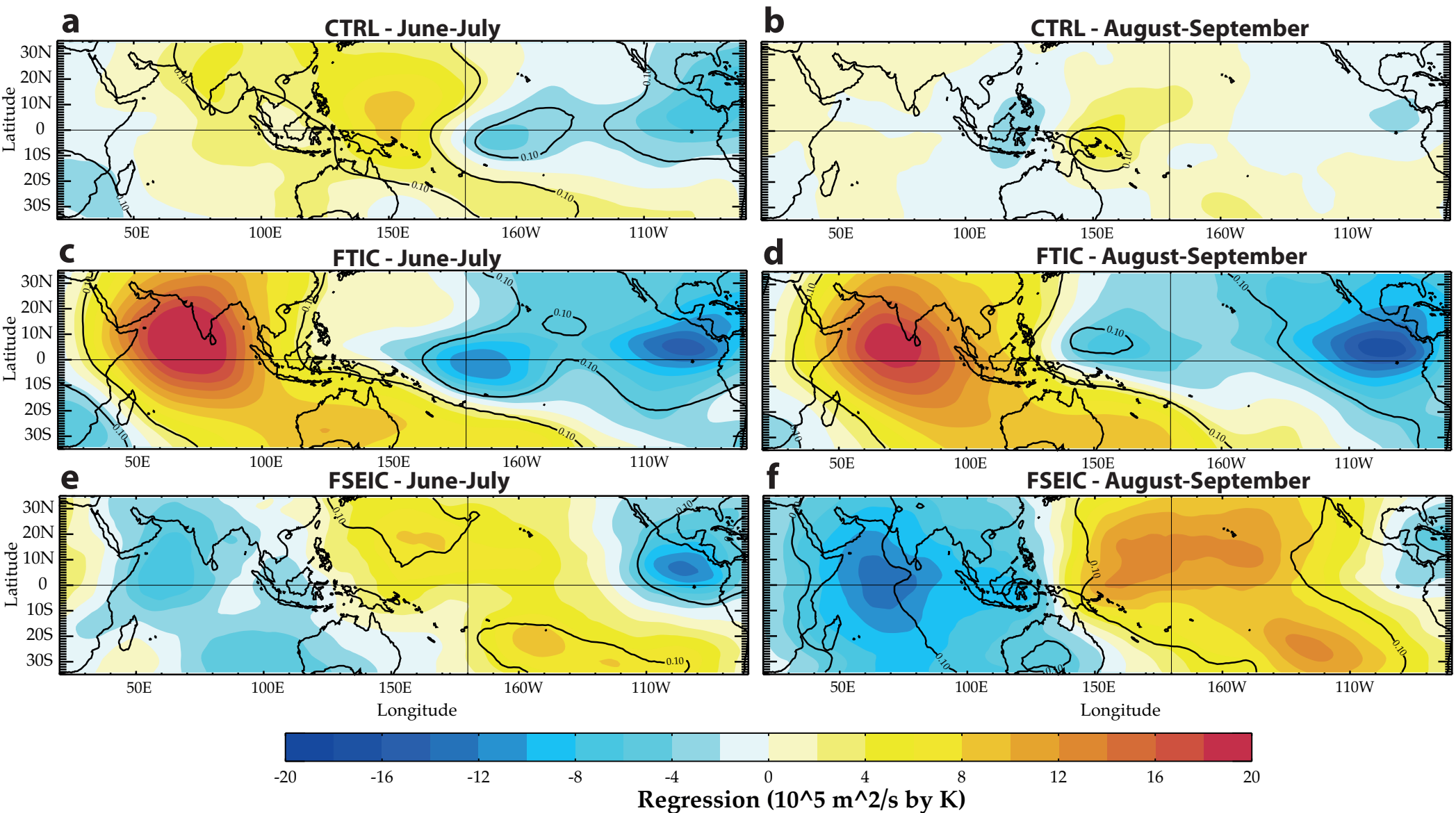
**Figure 12:** **a)** bi-monthly mean 20°C isotherm depth (a proxy for thermocline depth) anomalies regressed against the December-January Niño-3.4 SST time series for the following February-March, April-May, June-July, August-September and October-November in the CTRL experiment; **b)** same as **a)** but for the FTIC experiment ; **c)** same as **a)** but for the FSEIC experiment. Unit for the 20°C isotherm depth regression coefficients is in m by K. Regression coefficients reaching the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.

## Regressions Nino34 SST (12-1) - February-March - Year +1

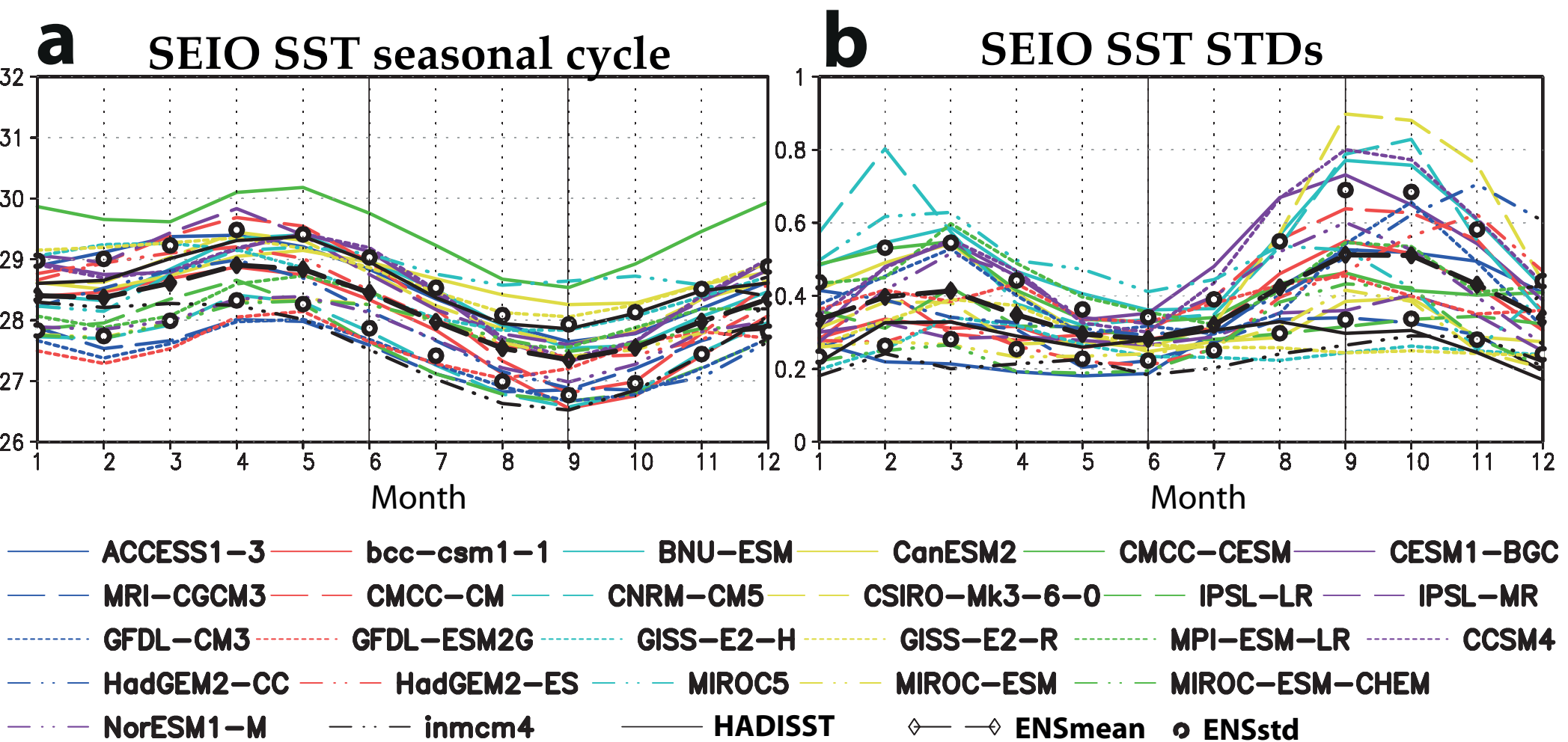


**Figure 13:** (a, b, c, d) February-March 850-hPa stream function, 850-hPa velocity potential, 200-hPa velocity potential and 850-hPa zonal wind anomalies regressed against the preceding December-January Niño-3.4 SST time series in the CTRL experiment; (e, f, g, h) same as (a, b, c, d) but for the FTIC experiment; (i, j, k, l) same as (a, b, c, d) but for the FSEIC experiment. Units for the stream function and velocity potential regression coefficients are in  $10^5 \text{ m}^2 \text{ s}^{-1} \text{ by K}$  and for 850-hPa zonal wind in  $\text{m s}^{-1} \text{ by K}$ . Regression coefficients reaching the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.

## Regressions Nino34 SST (12-1) - 200 hPa velocity potential - Year +1



**Figure 14:** (a, b) June-July and August-September 200-hPa velocity potential anomalies regressed against the preceding December-January Niño-3.4 SST time series in the CTRL experiment; (c, d) same as (a, b) but for the FTIC experiment; (e, f) same as (a, b) but for the FSEIC experiment. Units for the velocity potential regression coefficients are in  $10^5 \text{ m}^2 \text{ s}^{-1} \text{ by K}$ . Regression coefficients reaching the 10% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured.



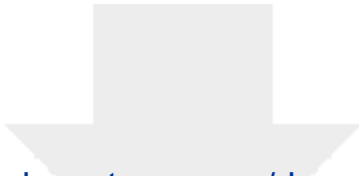
**Figure 14:** **a)** Monthly means of the SEIO ( $90^{\circ}\text{E}$ - $115^{\circ}\text{E}$  and  $0^{\circ}\text{S}$ - $10^{\circ}\text{S}$ ; unit:  $^{\circ}\text{C}$ ) SST time series from HadISST1.1 dataset and the historical runs of 25 CMIP5 models; **b)** Monthly standard deviations of the SEIO SST time series from HadISST1.1 dataset and the historical runs of 25 CMIP5 models. All observed and simulated statistics are estimated from the period 1950-2000. The ensembles mean and spread across the CMIP5 models are also displayed. The spread is computed as the ensemble mean plus and minus the ensemble standard deviation (thick black circles in panels **a** and **b**).

**Table 1**

Name	CTRL	FTPC	FTIC	FSEIC	FTPC-obs
Correction area	None	Pacific Ocean coast to coast 25°S-25°N	Indian Ocean coast to coast 25°S-30°N	Indian Ocean 90°E to coast 0°S-10°S	Pacific Ocean coast to coast 25°S-25°N
Smoothing area	None	30°S-25°S 25°N-30°N	30°S-25°S	85°E-90°E, 115°E-120°E 15°S-10°S 0°-5°N	30°S-25°S 25°N-30°N
SST data	None	CTRL	CTRL	CTRL	OISSTv2
Time duration (Year)	210	110	110	60	50

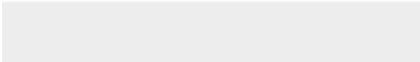
**Table 1:** Summary of the numerical experiments with their main characteristics, including length, nudging domain and SST climatology used for the nudging in the Indian or Pacific oceans decoupled experiments. The nudged experiments are the Forced Tropical Pacific Climatology (FTPC) run, the Forced Tropical Indian Climatology (FTIC) run, the Forced South-East Indian Climatology (FSEIC) run and, finally, the Forced Tropical Pacific observed Climatology (FTPC-obs) run. See text for more details. For the FTPC and FTPC-obs experiments only ocean grid-points in the Pacific are included in the correction or smoothing areas and, similarly, for the FTIC and FSEIC experiments for their respective domains. The different correction domains are also displayed in Figure 1b-d. The observed SST climatology used in the FTPC-obs experiment is derived from the AVHRR only daily Optimum Interpolation SST version 2 (OISSTv2) dataset for the 1982–2010 period (Reynolds et al. 2007).





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