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Anatomy of the Indian Summer Monsoon and ENSO

relationship in a state-of-the-art CGCM:

Role of the tropical Atlantic Ocean

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22

Abstract

23 The main paradigm for prediction of Indian Summer Monsoon Rainfall (ISMR) is its inverse relation
24 with El Niño-Southern Oscillation (ENSO). In this study, we focus on the role of the Atlantic Ocean
25 (AO) Sea Surface Temperature (SST) variability on the ISMR. There are basically two ways by which
26 AO SSTs can impact the ISMR: a “direct pathway” in which the AO may directly force the ISMR in
27 the absence of interactions with other dominant forcings like ENSO, and an “indirect pathway” in
28 which AO forces ENSO and modulates the ENSO teleconnection to ISMR. These two pathways are
29 studied with the help of sensitivity experiments performed with a Coupled General Circulation Model
30 (CGCM). Two pairs of decoupling experiments have been done. In the first, the SST variability in the
31 tropical AO or Pacific Ocean (PO) is removed by nudging the SST in these regions from a control
32 run's SST climatology. In the second set, the SST nudging is performed from the observed SST
33 climatology, which allows us to assess the robustness of the results and the specific role of the model's
34 SST mean-state biases.

35 The direct pathway linking tropical AO SST variability onto ISMR is insignificant in the PO
36 decoupled experiments or in recent observations. Furthermore, these experiments suggest on the
37 contrary that many AO SST anomalous patterns could be forced by ISMR. On the other hand, for the
38 indirect pathway, the AO decoupled experiments demonstrate that AO SST variability modulates the
39 onset and decaying phases of ENSO events. Despite ENSO is as strong and persists longer than in the
40 control simulation, the AO SST nudging resulted in a significant weakening of the inverse relationship
41 between ENSO and ISMR. The ENSO-monsoon relationship is mainly modulated during the ENSO
42 decaying phase. The upper-level divergent wind flows mainly from the Pacific to the AO resulting in
43 rainfall suppression in the AO, but only in a weak forcing on ISMR during boreal summer of the
44 ENSO decaying year in the AO decoupled experiments. Thus, the AO rainfall variability in these
45 experiments is decoupled from the surface and mainly modulated by the upper-level convergence or
46 divergence induced by the remote ENSO forcing.

47 Finally, the rectification of the AO SST mean-state biases in the CGCM also induces an El Niño-like
48 mean pattern over the tropical Pacific during boreal spring and promotes a stronger ENSO during its
49 peak phase. This demonstrates that the prominent AO SST mean-state biases in current CGCMs
50 further complicate the dynamical prediction and simulation of ISMR and ENSO.

51 **Keywords:** Indian Summer Monsoon; El Niño-Southern Oscillation; tropical Atlantic Ocean; ocean-
52 atmosphere interactions; Walker circulation, coupled climate model.

53

54 **1. Introduction**

55 In India, the rainy season is from June to September (JJAS) and Indian Summer Monsoon
56 Rainfall (ISMR) provides 80% of India's total annual precipitation. Despite the standard
57 deviation of ISMR is only about 10% of its mean, ISMR variability has a tremendous impact
58 on water resource management, agricultural yield and India's gross domestic product (Gadgil
59 and Gadgil 2006). However, forecasting ISMR variability is still a scientific challenge (Rao et
60 al. 2019) and an active research area as it involves many factors and their complex
61 interactions (see Chowdary et al. 2021 for recent review).

62 Numerous studies have examined climatic controls on ISMR interannual variability and
63 most of them showed the role played by tropical Pacific, Indian and Atlantic oceans Sea
64 Surface Temperature (SST) anomalies (Chowdary et al. 2021). El Niño-Southern Oscillation
65 (ENSO) is the primary forcing of year-to-year ISMR variability (Webster et al. 1998).
66 However, since ENSO can only explain about 35% the interannual variance of ISMR and the
67 ISM-ENSO relationship has weakened during the latter part of the 20th century, partly in
68 response to coherent multi-decadal variability of the climate system (Kumar et al. 1999;
69 Kucharski et al. 2007; Srivastava et al. 2019; Yang and Huang 2021), it is important to look
70 for other sources of ISMR predictability.

71 First, many studies have suggested a connection between ISM and Indian Ocean (IO)
72 SSTs, especially the Indian Ocean Dipole (IOD; see reviews in Cherchi et al. 2021). The IOD
73 is an irregular interannual SST oscillation in which the eastern equatorial IO gets alternately
74 colder and then warmer than the western part during boreal fall. Positive IOD events (e.g.,
75 warm in the western IO) may enhance ISMR through moisture transport over the western IO
76 or modification of the local Hadley cell with increased ascendance over the Indian region
77 (Cherchi et al. 2021). However, the influence of IOD on both ISMR and ENSO remains a
78 controversial topic (Meehl et al. 2003; Fischer et al. 2005; Izumo et al. 2010; Cretat et al.
79 2017, 2018; Stuecker et al. 2017; Terray et al. 2021; Cherchi et al. 2021; Zhang et al. 2021a).

80 The Atlantic Ocean (AO) can also add its impact on ISMR (Kucharski et al. 2007, 2009;
81 Pottapinjara et al. 2014, 2016; Sabeerali et al. 2019; Vittal et al. 2020; Yang and Huang
82 2021). First, a basin-warming mode exists in the tropical AO and is known as Atlantic Niño
83 or the Atlantic Zonal Mode (AZM; Lübbecke et al. 2018; Cabos et al. 2019; Richter and
84 Tokinaga 2021). These Atlantic Niños peak during boreal summer and are formed due to a
85 Bjerknes feedback as for ENSO, but last only for 3-4 months due to the weaker ocean-
86 atmosphere interactions in this basin (Lübbecke et al. 2018; Cabos et al. 2019; Richter and

87 Tokinaga 2021). It is still debated whether and how ENSO affects the AZM (Tokinaga et al.
88 2019). However, Atlantic Niños give rise to important shifts in local rainfall and are
89 associated with a Matsuno-Gill atmospheric response during boreal summer (Gill 1980;
90 Kucharski et al. 2009; Li et al. 2016; Jiang and Li 2021), which may modulate the
91 tropospheric temperature gradient in the Indo-Pacific sector and ISMR (Kucharski et al. 2009,
92 Pottapinjara et al. 2014, 2016; Sabeerali et al. 2019; Jiang and Li 2021). The second leading
93 mode of tropical AO SST variability involves fluctuations of the interhemispheric SST
94 gradient in the AO and is known as the Atlantic Meridional Mode (AMM; Chiang and
95 Vimont 2004; Jiang and Li 2021). The AMM is triggered and sustained by a Wind-
96 Evaporation-SST (WES) feedback (Chiang and Vimont 2004; Cabos et al. 2019). However,
97 the Tropical North Atlantic (TNA) SST anomaly dominates AMM variability (Enfield and
98 Mayer 1997; Jiang and Li 2021) and ENSO plays a dominant role in causing the spring time
99 trade wind variability over the TNA and the generation of local SST anomalies by evaporative
100 cooling/warming through ENSO teleconnections (Enfield and Mayer 1997; Garcia-Serrano et
101 al. 2017; Jiang and Li 2019). A few studies also suggest a link between the AMM or warm
102 SST TNA anomalies and ISMR (Vittal et al. 2020; Yang and Huang 2021).

103 An important difficulty for assessing the role of AO on ISMR, is that it interacts also
104 directly with the PO and IO in a complex manner and at different time scales (Kucharski et al.
105 2009, 2011; Rodriguez-Fonseca et al. 2009; Ham et al. 2013ab, McGregor et al. 2014, 2018;
106 Li et al. 2016; Terray et al. 2016; Wang et al. 2017; Cai et al. 2019; Li et al. 2020; Jiang and
107 Li 2021; Zhang and Han 2021). Recent studies suggest that warm TNA and AZM SST
108 anomalies can force a La Niña-like SST pattern in the Pacific (Rodriguez-Fonseca et al. 2009;
109 Ding et al. 2012; Ham et al. 2013ab; Wang et al. 2017; Jiang and Li 2021). But, again the role
110 of AO in ENSO and ISMR is debated in the literature (Ding et al. 2012; Zhang et al. 2021b;
111 Richter et al. 2021). As an illustration, previous studies suggest that Atlantic Niños may
112 reduce ISMR (Kucharski et al. 2007, 2009; Pottapinjara et al. 2016), but at the same times it
113 may promote La Niña conditions over the Pacific (Rodriguez-Fonseca et al. 2009; Ding et al.
114 2012; Jiang and Li 2021), which will be associated with enhanced ISMR. Therefore, it is
115 difficult to isolate the net effect of AO SST anomalies as the players, ISM, ENSO and AO
116 modes interact with each other in multiple different ways (Kucharski et al. 2009; Ding et al.
117 2012; Ham et al. 2013b; Cai et al. 2019; Jiang and Li 2021; Yang and Huang 2021).

118 This review highlights that there are two pathways by which the AO SSTs can affect
119 ISMR, one by a “direct” forcing on the ISMR and the other, “indirect”, by the AO forcing on
120 ENSO, which, in turn, will modulate the ISMR. However, as noted above, these two

121 pathways are not really independent of each other and the complexity of these interactions
122 implies that it is very difficult to assess the distinct causal relationships between the Atlantic,
123 Indian and Pacific SSTs and ISMR, or even the net effect of AO SSTs on ISMR from
124 observations alone. Considering these difficulties, we will assess the role of the tropical AO
125 on ISMR with the help of dedicated experiments performed with a Coupled General
126 Circulation Model (CGCM).

127 This paper is organised as follows. The validation datasets, CGCM and sensitivity
128 experiments used in this study are described in Section 2. In Section 3, the observed and
129 simulated relationships between ISMR, ENSO and AO SSTs at the interannual time scale are
130 documented. In Section 4, the AO “direct” and “indirect” effects on ISMR are assessed
131 through sensitivity coupled experiments. The final section summarizes the results of the
132 present work and presents some perspectives.

133

134 **2. Datasets, coupled model and sensitivity experiments**

135 2.a Observed datasets and time series indices.

136 Multiple data sources are used for model validation. SST and atmospheric variables (e.g.,
137 850- and 200-hPa winds, velocity potentials and stream functions) are taken or computed
138 from ERA-Interim reanalysis (ERAi; Dee et al. 2011) available from 1979 onwards. The
139 depth of the 20°C isotherm (Z20) is used as a proxy of the thermocline depth and is extracted
140 from the Simple Ocean Data Assimilation reanalysis for the 1979-2010 period (Carton and
141 Giese 2008; SODA version 2.2.4). We also used the Global Precipitation Climatology Project
142 rainfall dataset (GPCP; Huffman et al. 2001), which combines measures of precipitation
143 gauges and satellite data. GPCP is analyzed for the 1979-2016 period. All these quantities are
144 interpolated onto the model resolution to foster direct comparison with the simulations.

145 To monitor ISM, ENSO and AZM variability both in observations and simulations, we
146 define three standard time series indices:

147 - The ISMR time series is defined as the average of rainfall anomalies for the land grid points
148 in the region 5°-25°N and 70°-95°E.

149 - The Niño-3.4 SST (monthly average of SST anomalies in the region 5°S-5°N and 170°-
150 120°W; Niño34 hereafter) time series is chosen for the ENSO index since in observations the
151 strongest correlations between ISMR and tropical Pacific SSTs occur over this region.

152 - The ATL3 SST (monthly average of SST anomalies in the region 3°S-3°N and 20°W-0°E)
153 time series, which is a convenient index for the AZM (Lübbecke et al. 2018).

154 Note that our analysis of observations is robust if we estimate our Nino34, ATL3 and
155 ISMR time series from the Hadley Centre Sea Ice and SST dataset (Rayner et al. 2003) and
156 the rainfall dataset obtained from the Indian Meteorological Department (Pai et al. 2015).

157 2.b Coupled model and sensitivity experiments.

158 Here we employ a CGCM, the SINTEX-F2 (Masson et al. 2012), to assess the influence of
159 AO on ISMR variability and the ISM-ENSO relationship. The different model components
160 are ECHAM5.3 atmospheric model (Roeckner et al. 2003) at T106 spectral resolution
161 ($\sim 1.125^\circ \times 1.125^\circ$) and 31 hybrid sigma-pressure levels, NEMO ocean model (Madec 2008)
162 at $0.5^\circ \times 0.5^\circ$ horizontal resolution, 31 vertical levels and the LIM2 ice model (Timmermann
163 et al. 2005). The three model components are coupled using the Ocean-Atmosphere-Sea-Ice-
164 Soil (OASIS3) coupler (Valcke 2006). The model simulates the tropical Pacific SST mean
165 state, ENSO and ISMR variability reasonably well (Masson et al. 2012, Terray et al. 2016,
166 2021; Cretat et al. 2017, 2018).

167 First, a 210-yr fully coupled ocean-atmosphere simulation is used as a control (CTRL
168 hereafter). In order to disentangle the complex interactions between ISMR, ENSO and AO
169 SST variability, two partially coupled configurations of SINTEX-F2 are used and two
170 dedicated experiments have been performed with each of these configurations (see Table 1 for
171 details). In the first partial coupled configuration, full ocean-atmosphere coupling is used
172 everywhere except in the subtropical and tropical AO (25°S-25°N band), where SST is
173 nudged toward a daily SST climatology computed from CTRL or AVHRR-V2 daily
174 Optimum Interpolation SST observations during the 1982-2010 period (Reynolds et al. 2007).
175 These two AO decoupling experiments will be called FTAC and FTAC-obs and have been
176 run for 110-yr and 50-yr, respectively. In the second partial coupled configuration, ocean-
177 atmosphere coupling is active except in the subtropical and tropical Pacific (25°S-25°N band)
178 where, again, SST is nudged toward a daily SST climatology computed from CTRL or
179 observations. These two PO decoupling experiments will be called FTAC and FTAC-obs, and
180 have been run for 110-yr and 50-yr, respectively.

181 The nudging method used in these partial decoupling experiments modifies the non-solar
182 heat fluxes in the selected domain through a correction term that completely removes the SST
183 variability in the nudging domain (Terray et al. 2021). The damping term used in this nudging
184 technique ($-2400 \text{ W m}^{-2} \text{ K}^{-1}$) corresponds to the 1-day relaxation time for temperature in a 50-

185 m ocean layer. To avoid sharp SST gradients, a buffer zone is used between the “free” ocean
186 and regions of prescribed SST forcing such that the SSTs in these buffer regions are gradually
187 merged (over 5° latitude) with the prescribed SSTs. This strong SST restoring leads to an
188 almost complete decoupling between the ocean and atmosphere in the nudging domain with
189 SSTs, which differ by no more than 0.1 K from the prescribed space-time climatology.

190 In FTAC and FTFC, there are no significant changes in SST mean-state in the nudged
191 region, but also in the whole Tropics compared to CTRL (not shown). On the other hand, in
192 FTAC-obs (and also FTFC-obs), the strong SST restoring removes the SST mean-state biases
193 present in CTRL in addition to suppressing SST variability in the selected domain. CTRL
194 exhibits a strong warm bias in the southeast AO (Fig. 1a), which is a common problem for
195 most CGCMs (Richter et al. 2014; Voltaire et al. 2019; Bi et al. 2022). This bias is
196 attributed to errors in simulating zonal trade winds during boreal spring and is related to a
197 deeper thermocline, which weakens the upwelling of cold waters in the eastern AO (Fig. 1b).
198 Consistent with this erroneous east-west SST gradient, the rainfall pattern in the tropical AO
199 is shifted southeastward in CTRL compared to observations (Fig. 1c). Also consistent with
200 these mean-state biases, CTRL simulates a weaker SST variability over the eastern equatorial
201 AO compared to observations (Fig. S1b), especially during boreal summer, which is the
202 season of maximum SST variability in observations (Fig. S1a) as this is also the season when
203 the (observed) thermocline is the shallowest. Focusing on the PO, we note that CTRL is also
204 affected by a double Inter-Tropical Convergence Zone (ITCZ) bias (Fig. 1c) and a reduced
205 ENSO amplitude during boreal winter (Fig. S1b). As we will demonstrate in the following
206 sections, the inability of CTRL to realistically reproduce the seasonal cycle in the AO is a
207 cause of concern not only for the AO region, but also for ENSO and ISMR.

208 On the other hand, the SST restoring applied in FTAC-obs is able to correct largely these
209 simulated SST, thermocline and rainfall errors in the AO and also produces some changes in
210 the two other tropical basins (Figs. 1d-f). In other words, the comparison of CTRL, FTAC and
211 FTAC-obs (or FTFC and FTFC-obs) runs is useful to isolate the specific contribution of the
212 biased SST background mean-state in the coupled model.

213 More generally, the aim of these simulations is to isolate the effects of the PO and AO SST
214 variability on the simulated ISMR and ENSO-ISM relationship. First, FTFC and FTFC-obs
215 will be used to assess the « direct » relationship between AO SSTs and ISMR in a climate
216 without any counteracting ENSO forcing. Second, FTAC and FTAC-obs are useful to assess
217 if and how AO SST variability modulates ENSO and the simulated monsoon-ENSO
218 relationship, e.g. if AO SST anomalies are able to produce an « indirect » effect on the ISMR.

219 Table 1 summarizes the specifications of the simulations used here and the different
220 nudging domains are displayed in Fig. 1a. Finally, in all the analyses described below, the
221 first 10 years of all simulations have been excluded due to the spin-up of the coupled model.

222

223 3. ISMR, ENSO and AO relationships in observations and SINTEX-F2

224 3.a AO SST variability and its relationship with ENSO in SINTEX-F2

225 Before assessing the role of AO SST variability on ISMR and its relationship with ENSO,
226 it is important to document the performance of SINTEX-F2 in simulating AO variability,
227 especially the AZM and AMM (see Introduction for details). In this way, we can first
228 appreciate if the SST, rainfall and Z20 mean-state biases discussed in Section 2 are also a
229 cause of concern for a realistic simulation of AO SST variability in CTRL. Such analysis will
230 also be useful to interpret the differences between the AO decoupled experiments and CTRL
231 in the following sections as the nudged AO region encompasses the tropical AO.

232 AZM and AMM are the two dominant modes of SST monthly anomalies in the tropical
233 AO (Lübbecke et al. 2018; Cabos et al. 2019 and references herein). Thus, Empirical
234 Orthogonal Function (EOF) analysis of observed and simulated tropical AO SST monthly
235 anomalies provides a convenient tool for describing the performance of CTRL in simulating
236 both the AZM and AMM modes and their relative importance. Note also that the different
237 datasets have been detrended before the EOF analysis. Fig. 2 displays the first two leading
238 EOFs of observed and simulated monthly SST anomalies in the tropical AO. These two EOFs
239 of ERAi and CTRL SSTs are clearly distinct from the lower EOFs as EOF3 accounts for only
240 8% of the AO SST variance in both ERAi and CTRL (not shown).

241 The first EOF of ERAi SSTs describes 25% of the AO SST variance and depicts a basin-
242 wide pattern with positive SST anomalies covering the whole tropical AO (Fig. 2a). However,
243 the spatial loadings in this EOF1 are particularly high in the coastal upwelling regions near
244 the Angola–Benguela coast and in the equatorial cold tongue region explaining why this first
245 EOF is usually associated with the AZM in the literature (Lübbecke et al. 2018; Cabos et al.
246 2019; Jiang and Li 2021). The second EOF of ERAi SSTs accounts for 23% of the AO SST
247 variance (Fig. 2b). This EOF2 depicts a cross-equatorial SST gradient in the AO, which is
248 usually interpreted as the manifestation of the AMM (Chiang and Vimont 2004; Cabos et al.
249 2019). These two leading EOFs are very similar to other published EOF analysis of AO SSTs
250 using different datasets, spatial domains or time periods, both in terms of spatial patterns and
251 variance described by these leading modes (Lübbecke et al. 2018; Cabos et al. 2019; Jiang

252 and Li 2021). Of special interest is the possible connection of these two leading EOFs with
253 ENSO, which is simply assessed here by computing the simultaneous correlation between the
254 associated amplitude monthly time series and the Nino34 index. These observed EOF modes
255 have no simultaneous relationship with ENSO ($r=-0.07$ and 0.08 for Nino34 vs EOF1 and
256 EOF2, respectively). Their lead and lag relationships with Nino34 index will be explored in
257 Section 3.c.

258 Figs. 2cd display the two leading EOFs of CTRL SST monthly anomalies in the same AO
259 domain, which explain, respectively, 28 and 16% of the SST variance. SINTEX-F2 is able to
260 simulate with a reasonable accuracy both the spatial patterns and variances described by the
261 leading EOFs of observed AO SSTs as well as their relative importance in term of explained
262 variance. Note, however, that the L-shaped structure of the anomalous SSTs linking the
263 equatorial cold tongue to the southeast AO in EOF1 of observed SSTs is not well represented
264 and shifted westward in the EOF1 of simulated SSTs. This suggests that AZM events may be
265 weaker and are partly disconnected from the upwelling region near the Angola–Benguela
266 coast in CTRL compared to observations. This error is further confirmed by the comparison
267 of observed and simulated SST monthly means and standard-deviations in the ATL3 region
268 (Fig. 3). Consistent with Fig. 1a, ATL3 region is affected by a warm mean-state bias, which is
269 particularly prominent during June–July when the observed ATL3 SST variability is
270 maximum (Fig. 3), which corresponds to the peak of AZM events in observations (Lübbecke
271 et al. 2018). By contrast, the simulated ATL3 SST variability is prominent around three
272 months earlier and is drastically reduced in amplitude. This shortcoming, which is also found
273 in many other CGCMs (Voltaire et al. 2019; Bi et al. 2022), is related to the coupled mean-
274 state biases (e.g., SST, Z20, rainfall, etc.) reducing the intensity of the equatorial cold tongue
275 during boreal summer in CTRL, especially the flatten thermocline in the equatorial AO,
276 which may reduce the thermocline feedback and, thus, weakens the local Bjerknes feedback
277 and the AZM variability (see Figs. 1 and S1). The observed ATL3 SST variability has also a
278 secondary peak in winter, but this weaker maximum is well simulated in CTRL (Fig. 3b). On
279 the other hand, the second EOF of simulated AO SSTs closely matches the second EOF
280 estimated from ERAi SSTs in terms of spatial pattern and can also be regarded as the
281 manifestation of the AMM (Figs. 2bd). Finally, the correlations between the associated two
282 amplitude time series and the Nino34 index in CTRL are, respectively, 0.36 and 0.01 , and the
283 first correlation is highly significant, even at the 99.9 % confidence level according to a
284 phase-scrambling bootstrap test (Ebisuzaki 1997). This suggests a significant association of
285 the simulated AZM with ENSO in CTRL, which is not found in observations (see above). On

286 the other hand, EOF2 of (simulated) AO SSTs, which can also be regarded as the
287 manifestation of the AMM, is not significantly associated with ENSO in CTRL; a result
288 consistent with observations.

289 In summary, the two leading modes of the tropical AO SST variability in CTRL share
290 many features with those in observations, but the simulated AZM has a much weaker
291 amplitude and a significant relationship with ENSO, suggesting a too strong ENSO
292 teleconnection to the tropical AO or vice versa.

293

294 3.b ISMR regression analysis

295 We now present the results of a lead-lag regression analysis of tropical SST, rainfall, 850-
296 hPa wind and 200-hPa velocity potential quarterly time series onto the ISMR index in order to
297 provide a clear picture of the relationships between ISMR, ENSO and AO climate variability
298 in observations and CTRL (Figures 4 and 5). The ISMR index is fixed at the JJAS season and
299 the 4-month averaged SST, rainfall, 850-hPa wind and 200-hPa velocity potential time series
300 are shifted backward and forward in time. The results are presented in a two-year window
301 from the beginning of year -1 (preceding the ISMR year) to the end of year 0, year 0 referring
302 to the year of the ISM season. Note that the results remain unchanged if the different time
303 series are detrended before the regression analysis (not shown). While it is known that the
304 ISM-ENSO relationship includes some asymmetry in observations (Terray et al. 2003, 2005;
305 Boschat et al. 2012; Chakraborty and Singhai 2021), this regression analysis is used here as a
306 first order method to assess the realism of CTRL in simulating the ISMR teleconnections.

307 The regression results from observations (Fig. 4) illustrate that ISMR is associated with
308 different phases of ENSO in a two-year window (Boschat et al. 2012; Chakraborty 2018).
309 Strong positive SST anomalies in the central and eastern PO, which are out of phase with
310 anomalies in the western part of the PO are found during year -1, consistent with the
311 occurrence of an El Niño one year before a strong ISM (Fig. 4a). The atmospheric anomalous
312 patterns are consistent with this hypothesis as they describe an eastward shift of the Pacific
313 Walker circulation with persistent westerly 850-hPa wind anomalies over the western
314 equatorial Pacific, positive rainfall anomalies and negative 200-hPa velocity potential
315 anomalies over the central Pacific during year -1 (Figs. 4bc). The regression patterns during
316 year 0 are more or less a mirror image of those during year -1 with an opposite sign (Fig. 4).
317 In other words, this analysis demonstrates that ENSO and ISMR are still highly inter-related
318 and followed a sustained biennial rhythm during recent decades (Meehl et al. 2003; Terray et

319 al. 2003, 2021).

320 Focusing now on the relationships between ISMR and IO SSTs, we note that IO SST
321 anomalies are small and insignificant during boreal winter of year -1 and the pre-monsoon
322 period of year 0. During boreal summer of year 0, the IO is also devoid of any significant SST
323 anomalies associated with ISMR variability (Fig. 4a). This suggests that the « direct » effect
324 of IO SSTs on ISMR is small (Cretat et al. 2017; Terray et al. 2021).

325 On the other hand, a significant positive correlation emerges between TNA SSTs during
326 boreal spring of year 0 and ISMR (Fig. 4a). This result is in agreement with the results of
327 Vittal et al. (2020), Yang and Huang (2021) and Ham et al. (2013ab), which suggest that
328 TNA SSTs during boreal spring are a significant precursor of ISMR and can also serve as a
329 trigger for the following La Niña event, respectively. However, this statistical relationship
330 quickly fades away during boreal summer of year 0 (Fig. 4a). Furthermore, during ISM, the
331 SST anomalies are insignificant in the tropical AO, which partly disagree with recent studies
332 highlighting the role of AZM on the ISMR (Kucharski et al. 2007, 2009; Pottapinjara et al.
333 2014), but are in agreement with the work of Ding et al. (2012).

334 In order to further elucidate the relationships between tropical SSTs, ENSO and ISMR in
335 observations, the above lead-lag regression analysis has been repeated after removing the
336 ENSO contribution to the monthly SST fields by a simple linear regression method (Fig.
337 S2a). This is equivalent to assume that ENSO has a simultaneous linear impact on tropical
338 SSTs (Kucharski et al. 2009). Despite the simplicity of this approach, the results demonstrate
339 that most of the lead-lag relationships between tropical SST anomalies and ISMR (displayed
340 in Fig. 4a) can be understood as the results of simultaneous ENSO teleconnections on both
341 ISMR and SST anomalies elsewhere, as most of the correlations are now insignificant,
342 especially those during boreal summer of year 0 (Fig. S2a). One notable exception is,
343 however, again the TNA SSTs during boreal spring of year 0, which remain significant at the
344 90% confidence level even after removing the ENSO contribution from the observed SST
345 fields. On the other hand, despite the independence of AZM and ENSO in observations (see
346 Section 3.a) and removal of the ENSO-related contribution to the SST anomalies, the tropical
347 AO SSTs are again not linearly linked to ISMR during year 0.

348 CTRL is performing relatively well in reproducing the simultaneous inverse relationship
349 between ISMR and ENSO during year 0 (Fig. 5). However, the model fails to reproduce the
350 significant positive lead correlation between TNA SSTs during boreal spring and the
351 following ISM. Also, ISMR is much more linked to ENSO during year -1 and the pre-

352 monsoon season of year 0 in CTRL (Fig. 5; see also Terray et al. 2021). This suggests an
353 overly strong control of ENSO on ISMR in CTRL as for the AZM (see Section 3.a above).
354 Removing the ENSO contribution to the CTRL monthly SST fields by a simple linear
355 regression method before the regression analysis reveals a significant contribution of the
356 subtropical North Pacific SST anomalies during year 0, but again no significant relationships
357 with TNA or tropical AO SSTs during year 0 (Fig. S2b).

358 In the context of the ENSO-ISM relationship, the physical consistency of a direct link
359 between TNA SSTs during boreal spring and ISMR independent of ENSO, as found in
360 observations, is thus questionable. This positive correlation between TNA SSTs during boreal
361 spring and ISMR may be more an artifact of the ENSO forcing on both TNA and ISMR than
362 a sign of a physical connection between TNA SSTs and ISMR (Zhang et al. 2021b). This
363 alternative scenario is also consistent with the significant La Niña SST and atmospheric
364 patterns during year 0, associated with the ENSO biennial rhythm during recent decades (Fig.
365 4a; Wang et al. 2017), and the insignificant atmospheric anomalous patterns over TNA during
366 boreal spring of year 0 in our ISMR regression analysis from observations (Fig. 4bc).

367 3.c Nino34 regression analysis

368 We now revisit the two-way interactions between AO SSTs and ENSO for a better
369 understanding of the AO SSTs and ISMR relationships. Figures S3 and S4 show the lead-lag
370 regressions between the Nino34 index during boreal winter (e.g., ONDJ season) and tropical
371 SST, rainfall and 850-hPa wind anomalies, again during a two-year window, in both
372 observations and CTRL. In the following discussion, we will refer to year 0 as the developing
373 year and year +1 as the decaying year of ENSO events, respectively. CTRL reproduces
374 reasonably well the observed lifecycle of ENSO events with El Niño onset during boreal
375 spring, a developing phase during boreal summer and fall, a peak phase during boreal winter
376 and a decaying phase during boreal spring and summer of year +1 (Fig. S3).

377 The SST and atmospheric anomalous patterns during ENSO onset (e.g., during boreal
378 spring of year 0) are very similar in observations and CTRL and are reminiscent of the
379 Seasonal Footprinting Mechanism (SFM; Vimont et al. 2003; Boschat et al. 2013). We also
380 note significant cold SST anomalies in the tropical AO and TNA during boreal spring of year
381 0 (especially in observations see Fig. S3a) consistent with past studies (Rodriguez-Fonseca et
382 al. 2009; Ham et al. 2013ab; Jiang and Li 2021). Ham et al. (2013a) argues that these TNA
383 SST anomalies may influence the tropical PO and that this forcing is mediated by the SFM
384 (Fig. S4a). Such TNA forcing on El Niño onset is also suggested in CTRL, but with much

385 weaker amplitude (Figs. S3b and S4b).

386 A first notable difference between CTRL and observations is that simulated ENSO
387 teleconnections are significantly stronger than observed, as illustrated by the enhanced IO and
388 AO basin-wide warming in CTRL at the end of year 0 (Fig. S3b). The rainfall response over
389 the central equatorial Pacific per °C of Nino34 warming during boreal winter of year 0 is
390 much stronger in CTRL; about 5 mm/day per °C against only 2.6 mm/day per °C in
391 observations (Fig. S4). The associated latent heating induces an enhanced tropospheric
392 warming, which stabilizes the atmosphere and reduces convection, cloudiness and
393 evaporation over the AO and IO more in CTRL than in observations (not shown; Chiang and
394 Sobel 2002; Chang et al. 2006).

395 The climate anomalies during year +1 depict the transition from El Niño to La Niña (or
396 vice versa since the analysis is linear) and the related changes in ENSO teleconnections (Figs.
397 S3 and S4). First, the simulated and observed SST anomalies during boreal winter of year 0
398 and spring of year +1 show that most warm TNA events can be initiated by ENSO itself (see
399 Fig. S3; Garcia-Serrano et al. 2017; Jiang and Li 2019). This TNA warming is mainly driven
400 by the weakening of the local northeasterly trade winds during FMAM of year +1 in both
401 observations and CTRL (Fig. S4). These warm TNA events may induce wind anomalies over
402 the tropical Pacific that oppose the ongoing ENSO event and accelerate its demise (Ham et al.
403 2013ab). This TNA forcing can be interpreted as a delayed negative feedback that accelerates
404 the decay of ENSO as the IO capacitor effect (Xie et al. 2009, 2016; Wang et al. 2017). This
405 is consistent with many previous CGCM studies, which found that decoupling the IO or AO
406 in CGCMs increases the length of ENSO events (Ohba and Ueda 2007; Santoso et al. 2012;
407 Terray et al. 2016, 2021; Kajtar et al. 2017).

408 A second notable difference between CTRL and observations is that, despite this stronger
409 ENSO-induced capacitor effect over the tropical IO or AO in CTRL, the simulated ENSO
410 signal in the tropical Pacific lasts longer and persists well up to boreal summer and fall of
411 year +1 in CTRL (Fig. S4b). This contradiction has already been examined in Terray et al.
412 (2021) and their results highlight that, besides the amplitude of the basin-wide IO warming,
413 the intensity of the negative IO feedback on ENSO is heavily dependent on the realism of the
414 equatorial SST gradient in the IO during the ENSO peak and decaying phases. An open
415 question, which will be examined with the help of partial decoupling experiments in the
416 following section, is to determine if the erroneous mean equatorial SST gradient in the AO
417 (Fig. 1a) is also playing a role in the longer ENSO period in CTRL.

418 In summary, this section illustrates how it is difficult to isolate the specific effect of AO
419 SST anomalies on ISMR in observations or coupled simulations as the three players, ISM,
420 ENSO and the AO SST modes, interact with each other in a complex manner.

421

422 **4. AO direct and indirect effects on ISMR in SINTEX-F2**

423 In order to assess both the possible “direct” and “indirect” effects of AO SSTs on ISMR,
424 we now focus on the analysis of the partial decoupling experiments described in Section 2b.

425 4.a AO direct effect on ISMR

426 We estimate the "direct" effect of AO SST variability on ISMR with the help of the PO
427 decoupling experiments in which ENSO has been removed.

428 In order to document how the simulated AO SST variability is modified in the absence of
429 ENSO, Fig. 6 displays the first two EOFs of tropical AO SST anomalies simulated in the two
430 PO decoupling experiments. We first note that these leading EOFs estimated from FTFC and
431 FTFC-obs are very similar both in terms of spatial pattern and described variance. EOF1
432 (EOF2) is associated with SST variability to the South (North) of the Equator and the
433 Angola–Benguela (Senegal-Mauritania) upwelling system. This suggests that, in both PO
434 experiments, these EOFs are excited by wind-driven evaporation or dynamic in each
435 hemisphere, but that they do not involve an active WES feedback, which is usually associated
436 with the AMM (Chiang and Vimont 2004; Cabos et al. 2019).

437 These EOFs can be compared to those estimated from CTRL, which are shown in Figs.
438 2cd. Consistent with the overly strong ENSO forcing on the simulated AZM found in CTRL
439 (e.g., EOF1 of CTRL SSTs in Fig. 2c), the leading EOFs in the PO experiments are no more
440 associated with SST variability in the equatorial AO and they further suggest that the two
441 poles of the AMM may vary independently of each other. This questions the physical
442 consistency of the AMM and its underlying mechanism, both in observations and CTRL, as
443 its inter-hemispheric SST dipole structure depicted by the EOF2 of (observed and CTRL) AO
444 SST anomalies may result from orthogonality constraints of the EOF analysis rather than the
445 consequence of an active WES feedback.

446 The specific role of this “ENSO-free” AO SST variability in modulating ISMR is
447 illustrated by a regression analysis of simulated SST and 200-hPa velocity potential
448 anomalies onto the ISMR index during year 0 (Figure 7). We focus on FTFC-obs in the rest
449 of this section, as the results from FTFC are similar (not shown). Overall, the direct forcing of

450 AO SST and atmospheric variability on ISMR is insignificant as the AO is devoid of any
451 significant SST or atmospheric anomalies during the pre-monsoon season (e.g., FMAM of
452 year 0) in FTAC-obs (Fig. 7). On the other hand, a well-defined atmospheric teleconnection
453 pattern exists during boreal summer with strong 200-hPa divergence and outflow anomalies
454 from the ISM region towards the south AO in FTAC-obs (Fig. 7b). This upper-level velocity
455 potential pattern is obviously triggered by the diabatic heating associated with ISMR and
456 demonstrates that ISM is a dominant feature of the boreal summer tropical circulation in the
457 absence of ENSO. Interestingly, this ISMR forcing on the AO during boreal summer also
458 promotes an AMM-like SST pattern during boreal fall with negative (positive) SST anomalies
459 in the South (North) AO (Fig. 7a). This leads to the hypothesis that it is ISMR, which is
460 forcing the AO rather than the reverse in the absence of ENSO.

461 4.b AO forcing on ENSO

462 The aim of this subsection is to isolate the effect of the AO SST mean-state bias and
463 variability on ENSO with the help of the FTAC and FTAC-obs simulations.

464 To have a first basic understanding of the ENSO behavior in these two runs, Figure 8
465 shows the Nino34 SST monthly standard deviations from observation, CTRL and the two AO
466 decoupled experiments. The observed Nino34 SST variability is most prominent during
467 boreal winter (std of about 1.2°C), and it is drastically reduced in April (about 0.6°C), which
468 marks the onset of many ENSO events (Fig. 8). CTRL underestimates the Nino34 SST
469 variability during boreal winter when ENSO events usually peak (Figs. 8 and S1). It is also
470 not able to replicate the sharp decrease of Nino34 SST variability after the ENSO peak and,
471 hence, there is an overestimation of the simulated Nino34 SST variability for a few months
472 from March till July (Fig. 8). One interesting result is how the AO decoupled experiments
473 alter this Nino34 SST variability (Frauen and Dommenges 2012; Ding et al. 2012; Kajtar et al.
474 2017). The Nino34 SST variability changes in FTAC and FTAC-obs include (i) a consistent
475 increase during boreal summer for the two runs and (ii) an enhanced variability during boreal
476 winter and a more realistic seasonal phase-locking in FTAC-obs compared to both CTRL and
477 FTAC (Fig. 8).

478 We first focus on the increase of Nino34 SST variability during boreal summer, which is
479 found in both FTAC and FTAC-obs (Fig. 8). In order to explain this feature, Figure 9 shows
480 the regressions between the Nino34 index during boreal winter of year 0 (e.g., ONDJ season)
481 and tropical SST, rainfall and 850-hPa wind anomalies during the preceding boreal winter and
482 spring (e.g., FMAM) in FTAC and FTAC-obs. As discussed in Section 3, boreal spring is the

483 season of El Niño onset and the results suggest that the SFM in the North Pacific plays a key-
484 role in ENSO onset in both observations and CTRL (Figs. S3 and S4). A robust association
485 between the SFM and El Niño onset is also found in the two AO decoupled experiments
486 (Figs. 9ab). But, this El Niño onset occurs during boreal winter of year -1, e.g., one season
487 before the El Niño onset in CTRL and observations, and one year before the El Niño peak
488 (Fig. 8), a feature, which has not been well documented in past studies (Frauen and
489 Dommenges 2012; Terray et al. 2016; Kajtar et al. 2017). Furthermore, warm SST anomalies
490 already cover the whole central and eastern equatorial PO during boreal spring of year 0 in
491 these two runs (Figs. 9cd), e.g. also one season in advance compared to observations or CTRL
492 in which this basin-wide SST anomalous pattern is seen in boreal summer (Fig. S3). The
493 associated rainfall and 850-hPa wind regression patterns during boreal spring of year 0 in
494 FTAC and FTAC-obs also describe an eastward shift of the convection center with positive
495 (negative) rainfall anomalies over the central (western) PO and westerly zonal wind
496 anomalies on the western side of the positive rainfall anomalies (Figs. 9ab). All these features
497 are fully consistent with the early El Niño onset in FTAC-obs and FTAC. Note, furthermore,
498 that the rainfall and 850-hPa regression patterns during boreal winter and spring preceding the
499 ENSO peak are very similar in the two AO decoupled experiments (Figs. 9ab). This suggests
500 that this early El Niño onset can be attributed to the common cancellation of the AO SST
501 variability in the two runs. This early El Niño onset also implies that the warm Nino34 SST
502 anomalies associated with El Niño are already well defined during boreal summer of year 0
503 consistent with the enhanced Nino34 SST variability during boreal summer found in the two
504 nudged experiments (Fig. 8). These results illustrate the important role played by AO SST
505 variability in generating spread in ENSO timing and amplitude through its influence on the
506 SFM. This is also consistent with several recent studies, which suggest that the SFM and its
507 modulation are an important source of spread in ENSO forecasts during boreal spring and
508 early summer (Ma et al. 2017; Ogata et al. 2019).

509 We now focus on the improved phase-locking and enhanced ENSO variability during
510 boreal winter found in FTAC-obs (Fig. 8). In order to illustrate the seasonal dependence in the
511 changes of ENSO variability in the different runs and isolate the role of the AO background
512 state, Figure S5 displays the seasonal differences of SST standard-deviation between FTAC-
513 obs and both CTRL and FTAC. Figs. S5a-c first confirm that almost all the tropical AO SST
514 variability has been removed in FTAC-obs. Outside the AO nudged region, the SST
515 variability changes in FTAC-obs relative to CTRL are mainly found in the tropical PO in the
516 form of an enhanced ENSO variability from boreal summer to winter (Figs. S5b-c). This is

517 consistent with the changes of Nino34 SST variability (Fig. 8). Furthermore, a large part of
518 these differences can be attributed to the corrected SST AO mean state as this seasonal pattern
519 of changes is also found in the differences of SST standard-deviation between FTAC-obs and
520 FTAC (Figs. S5d-f). In other words, decoupling the AO SST variability without restoring the
521 observed AO SST climatology leads only to a modest increase of ENSO variability,
522 especially during its peak phase (Fig. 8).

523 In order to understand why the AO background state has such an impact on ENSO, Figure
524 S6 displays the seasonal differences between FTAC-obs and CTRL climatologies of SST,
525 rainfall and Z20, and Figure 10 shows boreal spring differences of 850- and 200-hPa zonal
526 wind, velocity potential and stream function climatology between the same runs. The
527 differences between FTAC-obs and FTAC are similar as FTAC and CTRL have the same
528 mean state (not shown).

529 The correction of AO SST biases in FTAC-obs leads to drastic improvements of AO
530 rainfall and Z20 spatial distributions during all the seasons with enhanced precipitation in the
531 northwest tropical AO, reduced precipitation in the southeast tropical AO and, finally, a
532 deeper Z20 in the western AO (Figs. S6bc). These rainfall changes are consistent with the
533 imposed AO SSTs in FTAC-obs altering the regions that are above or below the threshold for
534 deep convection in FTAC-obs compared to CTRL or FTAC (Fig. S6a). An increase of
535 precipitation is also evident over the Amazon basin demonstrating that the response is not
536 purely local. These AO rainfall shifts may exert an influence on atmospheric teleconnections
537 because they alter diabatic heating (e.g., Gill 1980). However, surprisingly, while these
538 rainfall changes are significant during all seasons, the rectification of the tropical PO mean
539 state is mainly prominent during boreal spring and is characterized by a shift to an El Niño-
540 like mean state with a warmer (cooler) eastern (western) equatorial PO, a rainfall increase in
541 the central PO near the date line, a flatter thermocline (Figs. S6abc, first row) and, finally, a
542 slowdown of the mean Walker circulation across the tropical PO (Figs. 10ab). The westward
543 shift of rainfall over the AO may produce large heating anomalies at upper level centered over
544 central South America and extending in the eastern and central PO. The 850- and 200-hPa
545 velocity potential differences during boreal spring are in agreement with this hypothesis as
546 they show enhanced convergence (divergence) at 850-hPa (200-hPa) toward a large region
547 encompassing the central equatorial PO and the northwest AO and the opposite patterns
548 elsewhere in the Tropics (Figs. 10cd). The well-defined quadrupole structure of the 850- and
549 200-hPa stream functions over the eastern PO/western AO (Figs. 10ef) is further consistent
550 with the circulation expected from a Matsuno-Gill response (Gill 1980; Kucharski et al. 2009;

551 Li et al. 2016). These features directly link SST mean-state changes in the AO with the
552 rectification of the Pacific mean state during boreal spring. Such strong inter-basin
553 connectivity is not seen in the other seasons despite that AO rainfall changes are still
554 prominent (Fig. S6b). This may be related to the ITCZ and Walker cells in all oceanic basins
555 being closer to the Equator during boreal spring, thus providing ideal conditions for the
556 corrected rainfall and SST patterns in the tropical AO to influence the equatorial PO (Chang
557 et al. 2006; Tokinaga et al. 2019).

558 Not surprisingly, the rainfall and 850-hPa zonal wind variability is also significantly
559 enhanced and shifted eastward in the PO during boreal spring in FTAC-obs (Figs. 11ac),
560 while the related changes are weak in FTAC (Figs. 11bd). This is consistent with the eastward
561 shift of the mean PO SST and rainfall patterns during the same season in FTAC-obs (Figs.
562 S6ab). Furthermore, the changes of rainfall and 850-hPa zonal wind variability over the AO
563 are opposite in FTAC and FTAC-obs, with a large increase (decrease) of rainfall and 850-hPa
564 zonal wind variability over the tropical AO during boreal spring in FTAC (FTAC-obs) despite
565 the absence of AO SST variability in the two runs (Fig. 11). This again highlights strong
566 interactions between the biased AO SST mean-state and atmospheric variability in FTAC,
567 which may further perturb the ENSO onset. In other words, the El-Niño like changes of the
568 PO mean state in FTAC-obs provide more favorable conditions for El Niño to develop
569 through the Bjerknes feedback (e.g., a reduced equatorial SST gradient and a flatter
570 thermocline across the PO) and reduced atmospheric noise over the AO during boreal spring.
571 This finally leads to a much better seasonal phase-locking of the simulated ENSO and an
572 improved ENSO amplitude during its peak phase in FTAC-obs in comparison of CTRL and
573 FTAC (Fig. 8).

574 Finally, the AO decoupled experiments demonstrate that the AO SST variability
575 significantly modulates ENSO during its decaying phase. This is illustrated by the regression
576 analysis of the ONDJ Nino34 index with quarterly SST time series during the following year
577 (e.g., year +1) in FTAC-obs (Fig. 12). FTAC displays a very similar evolution (not shown).
578 The corresponding regression analyses for observations and CTRL are shown in Fig. S3. The
579 warm SST PO pattern lasts longer in FTAC-obs (and FTAC) than in CTRL and the ENSO
580 signal is still robust at the end of year +1 in this run (Fig. 12c). Overall, the results highlight
581 that the ENSO-induced AO SST anomalies in CTRL (Fig. S3b), which are on the other hand
582 very small in the FTAC-obs by design (Fig. 12), accelerate the transition from El Niño to La
583 Niña (in CTRL) during year +1.

584 4.c AO indirect effect on ISMR

585 The AO “indirect” effect on ISMR, mediated by the ISM-ENSO relationship, is now
586 explored again with the help of the AO decoupled experiments.

587 Figure 13ab shows the ISMR seasonal cycle and monthly standard deviations from
588 observation, CTRL and the two AO decoupled experiments. The ISMR seasonal cycle is not
589 altered in the nudging experiments, and they replicate the same results of CTRL (Fig. 13a).
590 Observed ISMR variability is characterized by a double peak at the onset and withdraw of
591 ISM (Fig. 13b). CTRL underestimates ISMR variability in the pre-monsoon season and
592 overestimates it during June-August. The nudging experiments reduce the simulated ISMR
593 variability during the last half of ISM, especially FTAC-obs (Fig. 13b), despite ENSO is
594 stronger during boreal summer in FTAC and FTAC-obs (Fig. 8).

595 To assess the AO “indirect” effect, Figure 13c shows the observed and simulated lead-lag
596 correlations between ISMR and Nino34 SST quarterly time series in a three-year window
597 from the beginning of year -1 (preceding the ISMR year) to the end of year +1 (following the
598 ISMR year). Consistent with Fig. 4a, there are significant positive correlations one year
599 before ISM in observations. The sign of the correlation reverses during the pre-monsoon
600 season of year 0 and the correlation gets significantly negative during boreal summer and
601 winter of year 0. These negative correlations fade away during year +1. Thus, the most
602 favorable conditions for a strong ISM are during the transitions from an El Niño during year -
603 1 to a La Niña event in year 0 in agreement with our analysis in Section 3.

604 CTRL is able to reproduce realistically the significant negative correlation between the
605 ISMR and ENSO during boreal summer of year 0 and the decrease of amplitude of this
606 negative correlation during year +1. However, the model shows large discrepancies from
607 observations with a negative correlation during several months before ISM (Fig. 13c). This
608 bias is again consistent with the results of Section 3. However, the relative success of the
609 model in reproducing the observed simultaneous relationships between ISMR and ENSO is
610 important as the analysis of the nudged experiments can then provide more insights on the
611 precise role of the AO in this system.

612 In this respect, the lead-lag correlations between ENSO and ISMR are significantly
613 different between the two nudging experiments and CTRL, with a consistent weakening of
614 the simultaneous inverse relationship between ISM and ENSO (Fig. 13c), despite Nino34
615 SST variability is higher during boreal summer in FTAC and FTAC-obs compared to CTRL
616 (Fig. 8). In FTAC-obs, the simultaneous negative correlation between ISMR and Nino34 SST
617 is not even statistically significant at the 90% confidence level, despite FTAC-obs simulates a

618 stronger ENSO than FTAC. As the warm SST mean-state bias affecting the tropical AO in
619 both CTRL and FTAC is removed in FTAC-obs, these results are consistent with the AGCM
620 experiments performed in Kucharski et al. (2007, 2009) in which a weakening of the
621 monsoon-ENSO relationship is simulated in response to a cooling trend of the tropical AO
622 and allow us to isolate the specific role of the SST AO mean-state biases on the simulated
623 monsoon-ENSO relationship. However, as the simultaneous correlation between ISMR and
624 Nino34 SST is also reduced in FTAC, the reduced AO SST variability in the two runs may
625 also play an important role in the weakening of the inverse relationship between ISM and
626 ENSO in the nudged experiments.

627 Furthermore, ISMR evolution has different flavors during the ENSO decaying year in each
628 dataset (Figure S7). First, we find more rainfall over the IO during the pre-monsoon season
629 (not shown) and over India and the Arabian Sea during the monsoon of year +1 in
630 observations (Fig. S7a). This is consistent with the warm local SSTs (Fig. S3a) and the fast
631 demise of ENSO induced upper-level subsidence during year +1 in observations (Figs. S7a
632 and 14a). During boreal summer of the ENSO decaying year, the significant 200-hPa velocity
633 potential anomalies are restricted to a regional dipole opposing upper-level divergence over
634 the Arabian Sea to upper-level convergence over the western PO in observations (Fig. 14a).
635 These features are physically consistent with the positive (negative) correlation between
636 precipitation in the Indian (northwest PO) region during ISM of year +1 and the Nino34 index
637 in observations (Fig. S7a). This confirms that with the increase (decrease) of the SST over the
638 Nino34 region during the preceding boreal winter, the ISM of the next year is enhanced
639 (suppressed) or vice versa in observations (Fig 13c; Yang et al. 2007).

640 CTRL is not able to replicate these precipitation and 200-hPa velocity potential anomalies
641 during year +1 (Figs. S7b and 14b). In CTRL, there is a significant negative correlation
642 between precipitation over the Indian region during the monsoon season of year +1 and
643 Nino34 SST (Fig. S7b). This negative ISMR anomaly is consistent with both the persistent
644 ENSO signal (Fig. S3b) and the (significant) positive 200-hPa velocity potential anomalies
645 over India during boreal summer of year +1 in CTRL (Figs. S7b and 14b).

646 The effect of removing the AO SST variability on these correlations can now be analyzed
647 (Figs. 14c and S7c), keeping in mind that the ENSO-related SST signal persists even longer
648 and is stronger in FTAC-obs than in CTRL (Fig. 12). Despite this enhanced ENSO forcing,
649 the correlation of Nino34 SST with ISMR during year +1 has vanished and is not significant
650 in FTAC-obs (Fig. S7c). This is consistent with the weakening of the simultaneous ISMR-
651 Nino34 correlation in FTAC-obs (Fig. 13c). The origin of this paradox can be seen from Fig.

652 S7c, which shows that there is a stronger negative correlation between precipitations over the
653 tropical AO during year +1 with Nino34 SST (during the preceding boreal winter) in FTAC-
654 obs compared to CTRL. Moreover, these negative correlations are also shifted westward in
655 FTAC-obs consistent with the corrected AO mean state in this simulation. This stronger
656 relationship between ENSO and the AO in FTAC-obs during year +1 is further confirmed by
657 the associated 200-hPa velocity potential signal (Fig. 14c). In the 200-hPa velocity potential
658 anomalous pattern during boreal summer of year +1 in FTAC-obs, there is a significant
659 positive correlation over the AO, while the correlation over the Indian region is near zero.
660 This implies that the upper-level divergent flow is mainly from the central PO to the TNA
661 instead toward the Indian region, which results in suppression of rainfall in the AO and a near
662 normal ISM in FTAC-obs. FTAC shows a similar evolution (not shown), but the ENSO-
663 induced subsidence over the AO during boreal summer of the ENSO decaying year is weaker,
664 presumably due to the biased AO SST mean-state in FTAC.

665 In a nutshell, these last results illustrate that, by artificially removing the SST variability
666 over the tropical AO in the nudged experiments, the ENSO signal is stronger and persists
667 longer during the ENSO decaying year, but the associated upper-level divergent winds will
668 flow mainly from the central PO to the tropical AO, resulting in rainfall suppression in the
669 AO, but only in a weak ENSO forcing on ISM during the ENSO decaying year.

670

671 **5. Conclusions and discussion**

672 In this study, we use dedicated coupled experiments to isolate both the “direct” and
673 “indirect” effects of AO SSTs on ISMR. The “direct” effect refers to the AO forcing on ISMR
674 in the absence of others dominant forcings like ENSO (Kucharski et al. 2009). The “indirect”
675 effect refers to the AO forcing on ENSO (Rodriguez-Fonseca et al. 2009; Ding et al. 2012;
676 Ham et al. 2013ab; Jiang and Li 2021), which may subsequently affect ISMR through ENSO
677 teleconnections. Furthermore, with the help of these experiments, we also identify the role of
678 AO SST mean-state biases on the simulated ISMR-ENSO relationship.

679 First, we found that the “direct” effect of AO SST variability on ISMR is insignificant.
680 Overall, the results highlight that ISMR is a major player in the tropical atmospheric
681 circulation in the absence of ENSO, even forcing an AMM SST pattern during boreal fall
682 after ISM, rather than the reverse. Several studies have suggested that the AZM may provide
683 a direct remote forcing on ISMR through a Gill-Matsuno mechanism with a Kelvin wave
684 transporting the signal to the IO (Kucharski et al. 2009; Pottapinjara et al. 2014), but we were

685 not able to isolate this “direct” effect on ISMR in our simulations or observations during
686 recent decades, even when ENSO is removed (Figs. S2 and 7), which is consistent with the
687 modeling results of Ding et al. (2012). However, it is known that current CGCMs struggle to
688 represent this relationship, partly due to their common strong AO SST mean-state biases
689 (Barimalala et al. 2012; Voldoire et al. 2019). Taking into account the warm SST mean-state
690 bias in the southeastern AO found in SINTEX-F2 (see Fig. 1a), the realism and amplitude of
691 the AZM simulated by this CGCM are severely biased (Fig. 3b) and this may also deteriorate
692 the simulated ISMR-AZM relationship in our model.

693 The global-scale effects of the corrections of tropical AO SST biases are readily apparent
694 when comparing our two AO decoupled experiments with each other and with a long free
695 simulation performed with the same model. When the AO SST mean-state bias (present in
696 CTRL and FTAC) is corrected in FTAC-obs, the rainfall pattern in the AO is shifted
697 northwestward (see Fig. 1f). The associated changes in diabatic heating produce a Matsuno-
698 Gill atmospheric response centered over the eastern equatorial Pacific and generate a trans-
699 basin (e.g., Pacific/Atlantic) atmospheric see-saw with upward motion over a large region
700 encompassing the central and eastern equatorial Pacific and the western tropical AO and
701 descending motion elsewhere in the Tropics. The overall effect is the emergence of an El
702 Niño-like mean-state pattern in FTAC-obs when these AO SST biases are corrected, which is
703 mostly significant during boreal spring (see Fig. S6). This also corresponds to the onset
704 period of El Niño events in observations and the simulations analyzed here.

705 The comparison of FTAC and FTAC-obs with CTRL suggests that the “indirect” influence
706 of the tropical AO SST variability on ISMR is significant. The main effect of AO SST
707 variability is to modulate the amplitude and length of ENSO events, especially during their
708 onset and decaying phases. First, AO SST variability plays a key-role in ENSO developing
709 phase in agreement with the results of Ham et al. (2013ab). Without AO SST variability,
710 ENSO onset, while still seasonally phase-locked and linked to the SFM over the North Pacific
711 (Vimont et al. 2003; Boschat et al. 2013; Terray et al. 2016), occurs one season before during
712 boreal winter. This finding supports the idea that AO SST variability is also a possible source
713 of ENSO spread, especially during its developing year. This may reduce ENSO predictability
714 and contributes to the spring ENSO predictability barrier (Ma et al. 2017; Ogata et al. 2019).
715 Furthermore, when AO SST mean-state biases are also corrected (e.g., in FTAC-obs), this
716 leads to increased ENSO amplitude during its peak phase (e.g. boreal winter) as well,
717 demonstrating nonlinear interactions between the mean state during boreal spring and ENSO
718 amplitude.

719 Our results also confirm that AO SST variability modulates the length of the ENSO
720 decaying phase, as ENSO is still active up to the end of the ENSO decaying year in the FTAC
721 and FTAC-obs experiments (Terray et al. 2016; Kajtar et al. 2017). During boreal winter,
722 ENSO influences the subtropical and tropical AO through the atmospheric bridge (Jiang and
723 Li 2019). These ENSO-induced SST AO anomalies may then feedback negatively on ENSO
724 and fasten the transition from El Niño to La Niña. This discharging capacitor effect of the AO
725 (Wang et al. 2017) serves as a phase-reversal mechanism for the ENSO cycle as for the IO
726 (Xie et al. 2009; Terray et al. 2016, 2021).

727 All these findings are consistent with recent modelling studies that have identified a tight
728 physical linkage between AO and PO variability on decadal timescales (Kucharski et al.
729 2011) or in a global warming context (McGregor et al 2014) and also demonstrate that the
730 prominent AO SST biases play a significant role in modulating the simulated Pacific Walker
731 circulation at both the seasonal and longer time scales in current CGCMs (Kajtar et al 2018;
732 McGregor et al. 2018; Li et al. 2020).

733 Finally, the absence of AO SST variability weakens the simultaneous inverse relationship
734 between ISM and ENSO despite ENSO is stronger during boreal summer and persists longer.
735 This result is opposite to the one found in similar IO decoupling experiments in which ENSO
736 is also stronger and more persistent (Terray et al. 2021). The origin of this paradox lies
737 mainly in the modulation of the Walker circulation when SST variability is removed in one of
738 the two oceanic basins, especially during boreal summer of the ENSO decaying year (see Fig.
739 14). The upper-level divergent wind flows mainly from the PO to the AO, resulting in rainfall
740 suppression in the AO, but in a weaker forcing on ISMR in the AO decoupled experiments
741 (Fig. 14c). On the other hand, both the stronger ENSO amplitude and the enhanced upper-
742 level convergence towards the Indian domain act in concert and result in a much stronger
743 inverse ENSO-ISMR relationship in similar IO decoupled experiments (Terray et al. 2021).

744 In conclusion, while the AO “direct” effect on ISMR is insignificant in our coupled model,
745 we highlight that the AO “direct“ effect on ENSO is significant in many aspects including the
746 ENSO triggering mechanism and AO capacitor effect, which have also a significant “indirect”
747 impact on ISMR mediated by the ENSO teleconnections. We hope that these encouraging
748 results will promote the interest of performing similar IO, PO and AO decoupled experiments
749 with other coupled models in order to verify if the insignificant “direct” effect of AO SSTs on
750 ISMR found here is model dependent or not.

751

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980 **Figure captions**

981 **Figure 1:** **a)** SST annual means difference (unit: °C) between CTRL and ERAi; **b)** Z20
982 annual means difference (unit: m) between CTRL and SODA; **c)** Rainfall annual means
983 difference (unit: mm/day) between CTRL and GPCP; **d)** SST annual means difference (unit:
984 °C) between FTAC-obs and CTRL; **e)** Z20 annual means difference (unit: m) between FTAC-
985 obs and CTRL; **f)** Rainfall annual means difference (unit: mm/day) FTAC-obs and CTRL.
986 The nudging domain for FTAC and FTAC-obs (FTAC and FTAC-obs) is shown in blue
987 (purple) in panel a. See Section 2 and Table 1 for more details.

988 **Figure 2:** Empirical Orthogonal Function (EOF) 1st and 2nd modes for detrended SST
989 anomalies with monthly means removed obtained from ERAi and CTRL. **a)** and **b)** EOF1 and
990 EOF2 from ERAi SST detrended anomalies (1979-2015 period), respectively. **c)** and **d)** EOF1
991 and EOF2 from CTRL SST detrended anomalies (11-210 period), respectively. The number
992 in parentheses for each panel gives the % of SST variance described by the EOF mode.

993 **Figure 3:** **a)** Monthly means of ATL3 SST (unit: °C) index from ERAi (blue) and CTRL
994 (orange); **b)** Monthly standard deviations of ATL3 SST (unit: °C) from ERAi (blue) and
995 CTRL (orange).

996 **Figure 4:** **a)** Quarterly SST time series (from ERAi) during the preceding (e.g. year -1) and
997 simultaneous (e.g. year 0) ISM years, regressed against the ISMR index (e.g. JJAS ISM
998 rainfall from GPCP). Unit for the SST regression coefficient is °C by mm/day. **b)** Same as **a)**,
999 but for regression using quarterly rainfall and 850-hPa wind time series. Units for the rainfall
1000 and 850-hPa wind regression coefficients are mm/day by mm/day and $m s^{-1}$ by mm/day,
1001 respectively. **c)**, Same as **a)**, but for quarterly 200-hPa velocity potential time series. Unit for
1002 the 200-hPa velocity potential regression coefficient is $10^5 m^2 s^{-1}$ by mm/day. Regression
1003 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1004 test (Ebisuzaki 1997) with 999 samples are contoured (SST or 200-hPa velocity potential) or
1005 shown (rainfall and 850-hPa wind). Quarterly time series refer to the seasons February-
1006 May, June-September, October-January and so on.

1007 **Figure 5:** **a)** Quarterly SST time series during year 0 regressed against the ISMR index (e.g.
1008 JJAS ISM rainfall) in CTRL. Unit for the SST regression coefficient is °C by mm/day. **b)**
1009 Same as **a)**, but for regression using quarterly rainfall and 850-hPa wind time series in CTRL.
1010 Units for the rainfall and 850-hPa wind regression coefficients are mm/day by mm/day and m
1011 s^{-1} by mm/day, respectively. **c)**, Same as **a)**, but for quarterly 200-hPa velocity potential time
1012 series in CTRL. Unit for the 200-hPa velocity potential regression coefficient is $10^5 m^2 s^{-1}$ by

1013 mm/day. Regression coefficients reaching the 90% significance level according to a phase-
1014 scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity potential)
1015 or shown (rainfall and 850-hPa wind).

1016 **Figure 6:** Empirical Orthogonal Function (EOF) 1st and 2nd modes for 12-monthly anomaly
1017 of detrended SST obtained from FTPC and FTPC-obs experiments. **a)** and **b)** EOF1 and
1018 EOF2 from FTPC SST detrended anomalies (11-110 period), respectively. **c)** and **d)** EOF1
1019 and EOF2 from FTPC-obs SST detrended anomalies (11-50 period), respectively. The
1020 number in parentheses for each panel gives the % of SST variance described by the EOF
1021 mode.

1022 **Figure 7:** **a)** Quarterly SST time series during year 0 regressed against the ISMR index (e.g.
1023 JJAS ISM rainfall) in FTPC-obs. Unit for the SST regression coefficient is °C by mm/day. **b)**
1024 Same as **a)**, but for quarterly 200-hPa velocity potential time series in FTPC-obs. Unit for the
1025 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression
1026 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1027 test (Ebisuzaki 1997) are contoured.

1028 **Figure 8:** Monthly standard deviations of Nino34 SST (unit:°C) from ERAi (blue), CTRL
1029 (orange), FTAC (green) and FTAC-obs (red).

1030 **Figure 9:** **a)** Quarterly rainfall and 850-hPa wind time series during boreal winter and spring
1031 regressed against the Nino34 index during the following boreal winter (e.g. ONDJ Nino34
1032 SST) in FTAC-obs. **b)** Same as **a)**, but in FTAC. **c)** Quarterly SST time series during boreal
1033 winter and spring regressed against the Nino34 index during the following boreal winter in
1034 FTAC-obs. **d)** Same as **c)**, but in FTAC. Unit for the SST regression coefficient is °C by °C.
1035 Units for the rainfall and 850-hPa wind regression coefficients are mm/day by °C and m s^{-1} by
1036 °C, respectively. Regression coefficients reaching the 90% significance level according to a
1037 phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity
1038 potential) or shown (rainfall and 850-hPa wind).

1039 **Figure 10:** **a)** 850-hPa zonal wind boreal spring means differences (unit: m s^{-1}) between
1040 FTAC-obs and CTRL; **b)** 200-hPa zonal wind boreal spring means differences (unit: m s^{-1})
1041 between FTAC-obs and CTRL; **c)** 850-hPa velocity potential boreal spring means differences
1042 (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **d)** 200-hPa velocity potential boreal spring
1043 means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **e)** 850-hPa stream
1044 function boreal spring means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **f)**

1045 200-hPa stream function boreal spring means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-
1046 obs and CTRL.

1047 **Figure 11: a)** Rainfall boreal spring standard deviation differences (unit: mm/day) between
1048 FTAC-obs and CTRL. **b)** Same as **a)**, but for FTAC and CTRL. **c)** 850-hPa zonal wind boreal
1049 spring standard deviation differences (unit: m s^{-1}) between FTAC-obs and CTRL. **d)** Same as
1050 **c)**, but for FTAC and CTRL.

1051 **Figure 12: a)** Boreal spring SST time series regressed against the Nino34 index during the
1052 preceding boreal winter (e.g. ONDJ Nino34 SST) in FTAC-obs; **b)** Boreal summer SST time
1053 series regressed against the Nino34 index during the preceding boreal winter in FTAC-obs; **c)**
1054 Boreal winter SST time series regressed against the Nino34 index during the preceding boreal
1055 winter in FTAC-obs. Unit for the SST regression coefficient is $^{\circ}\text{C}$ by $^{\circ}\text{C}$. Regression
1056 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1057 test (Ebisuzaki 1997) are contoured.

1058 **Figure 13: a)** Monthly means of ISMR (unit: mm/day) from GPCP (blue), CTRL (orange),
1059 FTAC (green) and FTAC-obs (red); **b)** Monthly standard deviations of ISMR (unit: mm/day)
1060 from GPCP (blue), CTRL (orange), FTAC (green) and FTAC-obs (red); **c)** Lead-lag
1061 correlations between ISMR and quarterly Nino3.4 SSTs starting from the beginning of the
1062 previous year (e.g. year - 1) to the end of the following year of the monsoon (e.g. year +1) in
1063 observations (blue), CTRL (orange), FTAC (green) and FTAC-obs (red). X-axis indicates the
1064 lag (in 4 months interval) for a 36 months period starting one year before the developing year
1065 of ISMR (e.g. year 0) and Y-axis is the amplitude of the correlation. Thus, the coefficients
1066 corresponding to -1, 0, +1 lags refer, respectively, to the correlations between ISMR in year 0
1067 (e.g. JJAS ISM rainfall) and February-May, June-September and October-January Niño-3.4
1068 SSTs, also during year 0, and so on. Circles indicate correlations that are above the 90%
1069 significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki
1070 1997).

1071 **Figure 14: a)** Boreal spring and summer 200-hPa velocity potential time series regressed
1072 against the preceding boreal winter Nino34 SST in ERAi. **b)** Same as **a)**, but for CTRL. **c)**
1073 Same as **a)**, but for FTAC-obs. Unit for the 200-hPa velocity potential regression coefficient
1074 is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression coefficients reaching the 90% significance level
1075 according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured.

1076 **Figure S1: a)** SST seasonal standard deviations (unit: $^{\circ}\text{C}$) computed from ERAi. **b)** SST
1077 seasonal standard deviation differences (unit: $^{\circ}\text{C}$) computed between CTRL and ERAi.

1078 **Figure S2: a)** Quarterly (residual) SST time series during the preceding (e.g. year -1) and
1079 simultaneous (e.g. year 0) ISM years regressed against the ISMR index (e.g. JJAS ISM
1080 rainfall) in ERAi and GPCP after the simultaneous linear effect of the Nino34 index has been
1081 removed from the SST time series by a linear regression method. Unit for the SST regression
1082 coefficient is °C by mm/day. **b)** Same as **a)**, but for CTRL.

1083 **Figure S3: a)** Quarterly SST time series during years 0 and +1 regressed against the Nino34
1084 index during boreal winter (e.g. ONDJ Nino34 SST) in ERAi. **b)** Same as **a)**, but in CTRL.
1085 Unit for the SST regression coefficient is °C by °C. Regression coefficients reaching the 90%
1086 significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are
1087 contoured.

1088 **Figure S4: a)** Quarterly rainfall and 850-hPa wind time series during years 0 and +1
1089 regressed against the Nino34 index during boreal winter (e.g. ONDJ Nino34 SST) in GPCP
1090 and ERAi. **b)** Same as **a)**, but in CTRL. Units for the rainfall and 850-hPa wind regression
1091 coefficients are mm/day by °C and $m s^{-1}$ by °C, respectively. Regression coefficients reaching
1092 the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997)
1093 are shown.

1094 **Figure S5: a)** SST standard-deviation differences (unit: °C) between FTAC-obs and CTRL
1095 during boreal spring (e.g. FMAM); **b)** SST standard-deviation differences (unit: °C) between
1096 FTAC-obs and CTRL during boreal summer (e.g. JJAS); **c)** SST standard-deviation
1097 differences (unit: °C) between FTAC-obs and CTRL during boreal winter (e.g. ONDJ). **d)**, **e)**
1098 and **f)** Same as **a)**, **b)** and **c)**, but for SST standard-deviation differences between FTAC-obs
1099 and FTAC.

1100 **Figure S6: a)** SST seasonal mean climatological differences (unit: °C) between FTAC-obs
1101 and CTRL; **b)** Rainfall seasonal means differences (unit: mm/day) between FTAC-obs and
1102 CTRL; **c)** Z20 seasonal means differences (unit: m) between FTAC-obs and CTRL.

1103 **Figure S7: a)** Correlation coefficients between the Nino34 index during boreal winter (e.g.
1104 ONDJ) and boreal summer (e.g. JJAS) rainfall during the following year in ERAi and GPCP.
1105 **b)** Same as **a)**, but in CTRL. **c)** Same as **a)**, but in FTAC-obs. Correlation coefficients above
1106 the 90% significance confidence level according to a phase-scrambling bootstrap test
1107 (Ebisuzaki 1997) are contoured.

1108

1109 **Table captions**

1110 **Table 1:** Summary of the numerical experiments with their main characteristics, including
1111 length, nudging domain and SST climatology used for the nudging in the AO or PO
1112 decoupled experiments. The nudged experiments are the Forced Tropical Pacific Climatology
1113 (FTPC), the Forced Tropical Pacific observed Climatology (FTPC-obs), the Forced Tropical
1114 Atlantic Climatology (FTAC) and, finally, the Forced Tropical Atlantic observed Climatology
1115 (FTAC-obs) runs. See text for more details. For the FTPC and FTPC-obs experiments only
1116 ocean grid-points in the PO are included in the correction or smoothing areas and, similarly,
1117 for the FTAC and FTAC-obs experiments. The different correction domains are also
1118 displayed in Figure 1a. The observed daily SST climatology used in the FTPC-obs and
1119 FTAC-obs experiments is derived from the AVHRR only daily Optimum Interpolation SST
1120 version 2 (OISSTv2) dataset for the 1982-2010 period (Reynolds et al. 2007).

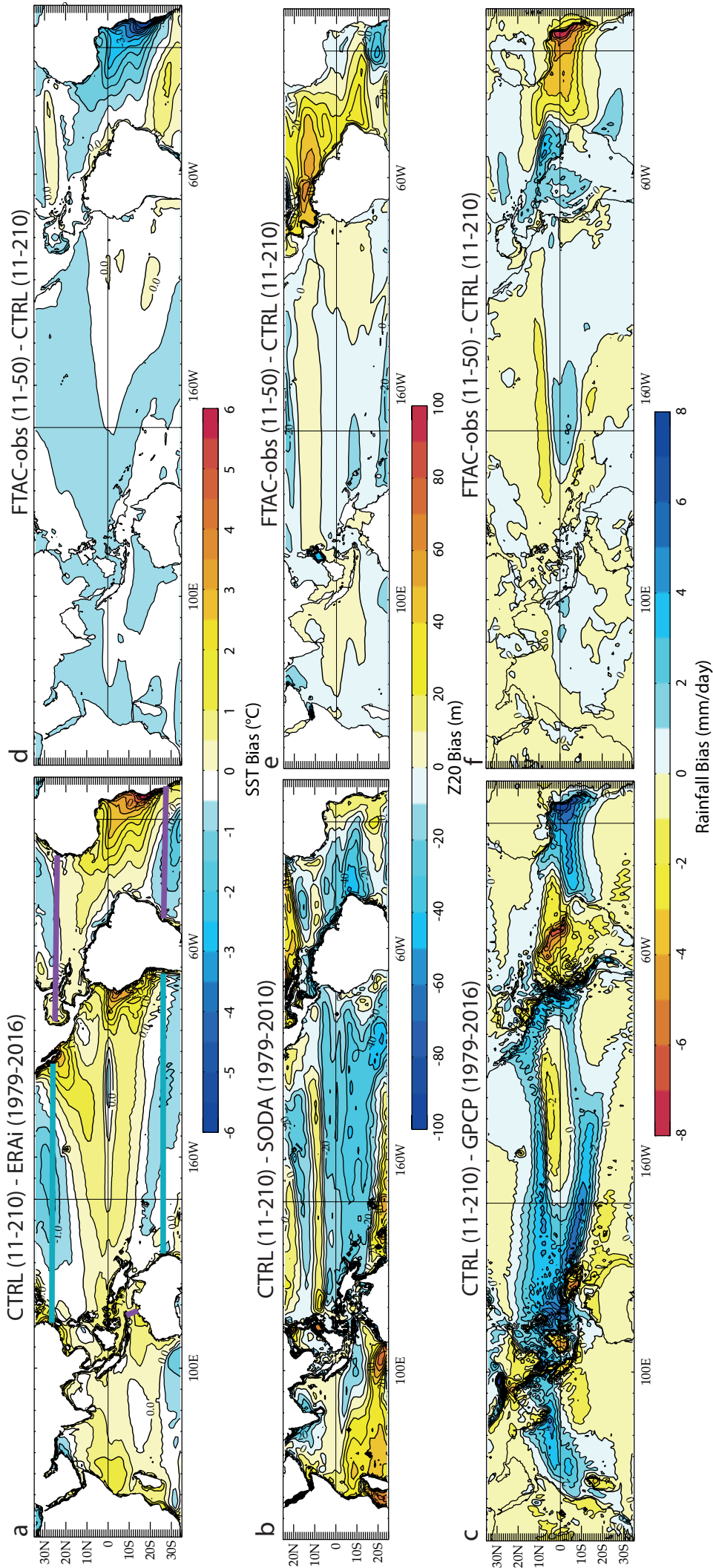


Figure 1: a) SST annual means difference (unit: °C) between CTRL and ERAI; b) Z20 annual means difference (unit: m) between CTRL and SODA; c) Rainfall annual means difference (unit: mm/day) between CTRL and GPCP; d) SST annual means difference (unit: °C) between FTAC-obs and CTRL; e) Z20 annual means difference (unit: m) between FTAC-obs and CTRL; f) Rainfall annual means difference (unit: mm/day) FTAC-obs and CTRL. The nudging domain for FTAC and FTAC-obs (FTAC and FTAC-obs) is shown in blue (purple) in panel a. See Section 2 and Table 1 for more details.

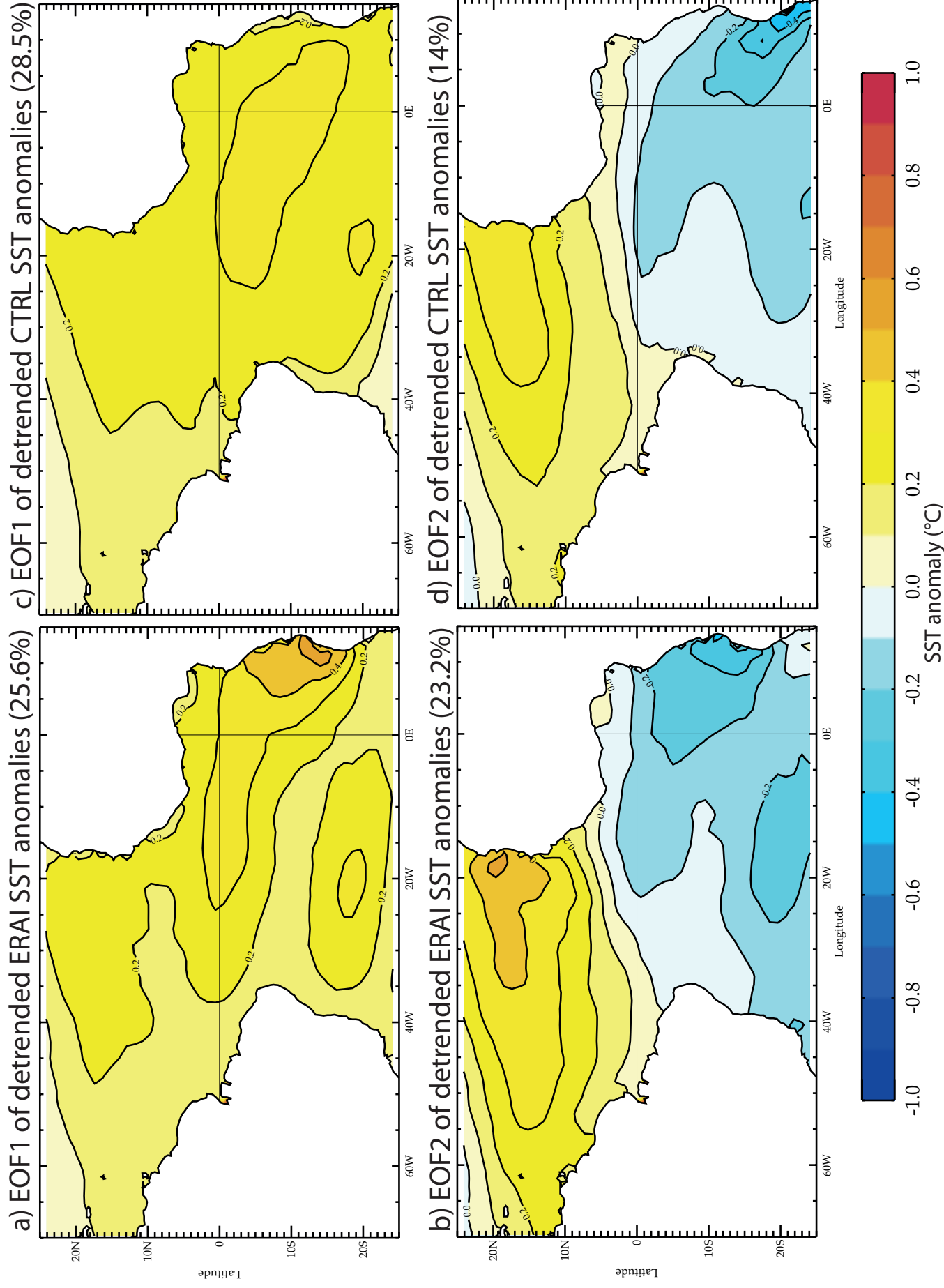
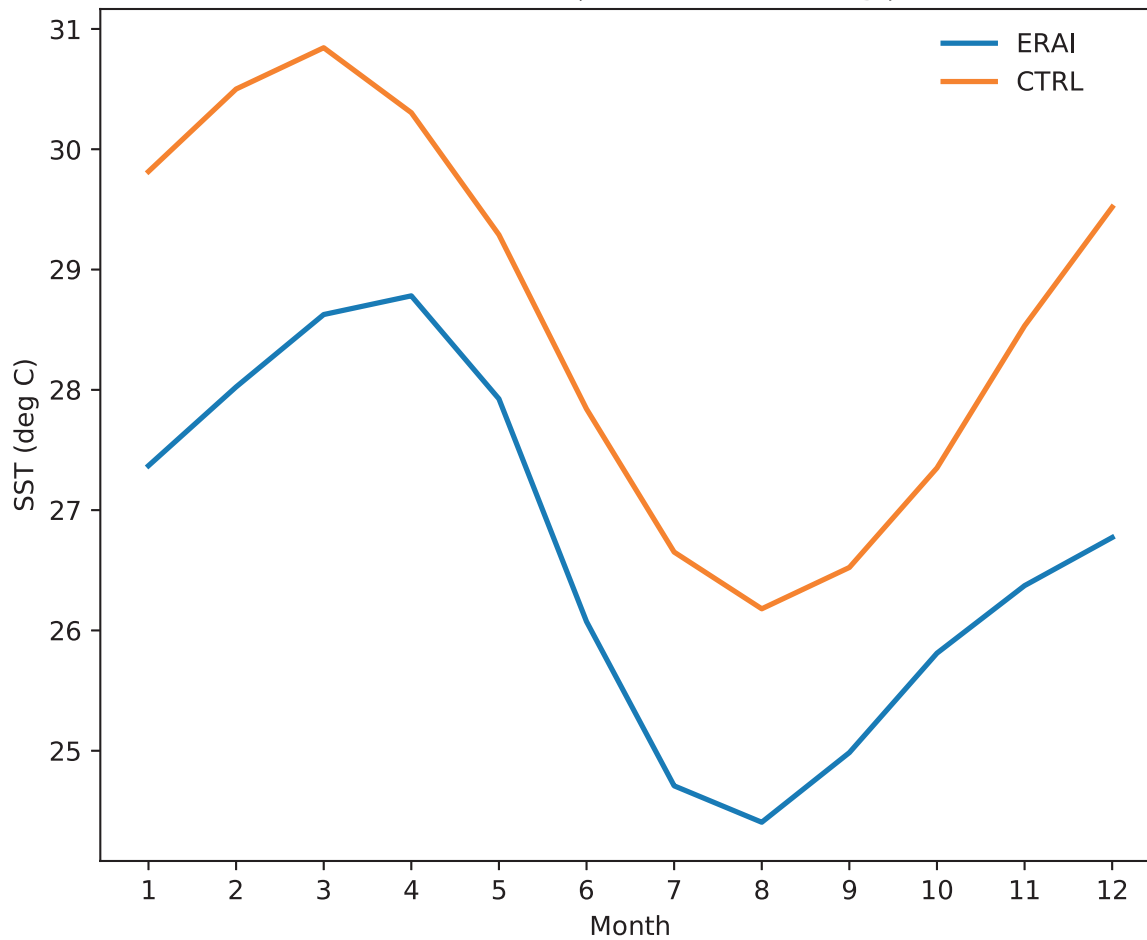


Figure 2: Empirical Orthogonal Function (EOF) 1st and 2nd modes for 12-monthly anomaly of detrended SST obtained from ERAI and CTRL. a) and b) EOF1 and EOF2 from ERAI SST detrended anomalies (1979-2015 period), respectively. c) and d) EOF1 and EOF2 from CTRL SST detrended anomalies (11-210 period), respectively. The number in parentheses for each panel gives the % of SST variance described by the EOF mode.

a) ATL3 SST monthly climatology



b) ATL3 SST monthly standard deviations

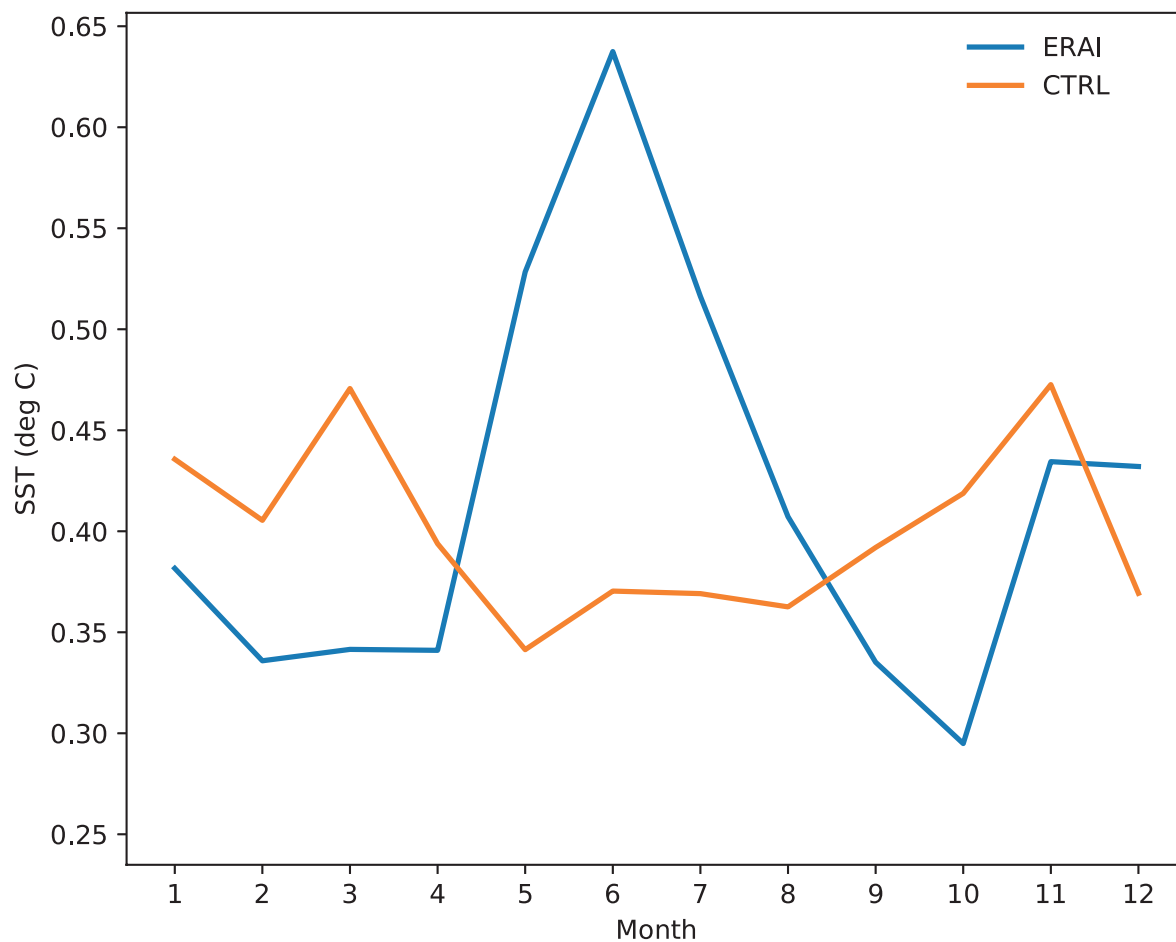


Figure 3: a) Monthly means of ATL3 SST (unit: °C) index from ERAi (blue) and CTRL (orange); b) Monthly standard deviations of ATL3 SST (unit: °C) from ERAi (blue) and CTRL (orange).

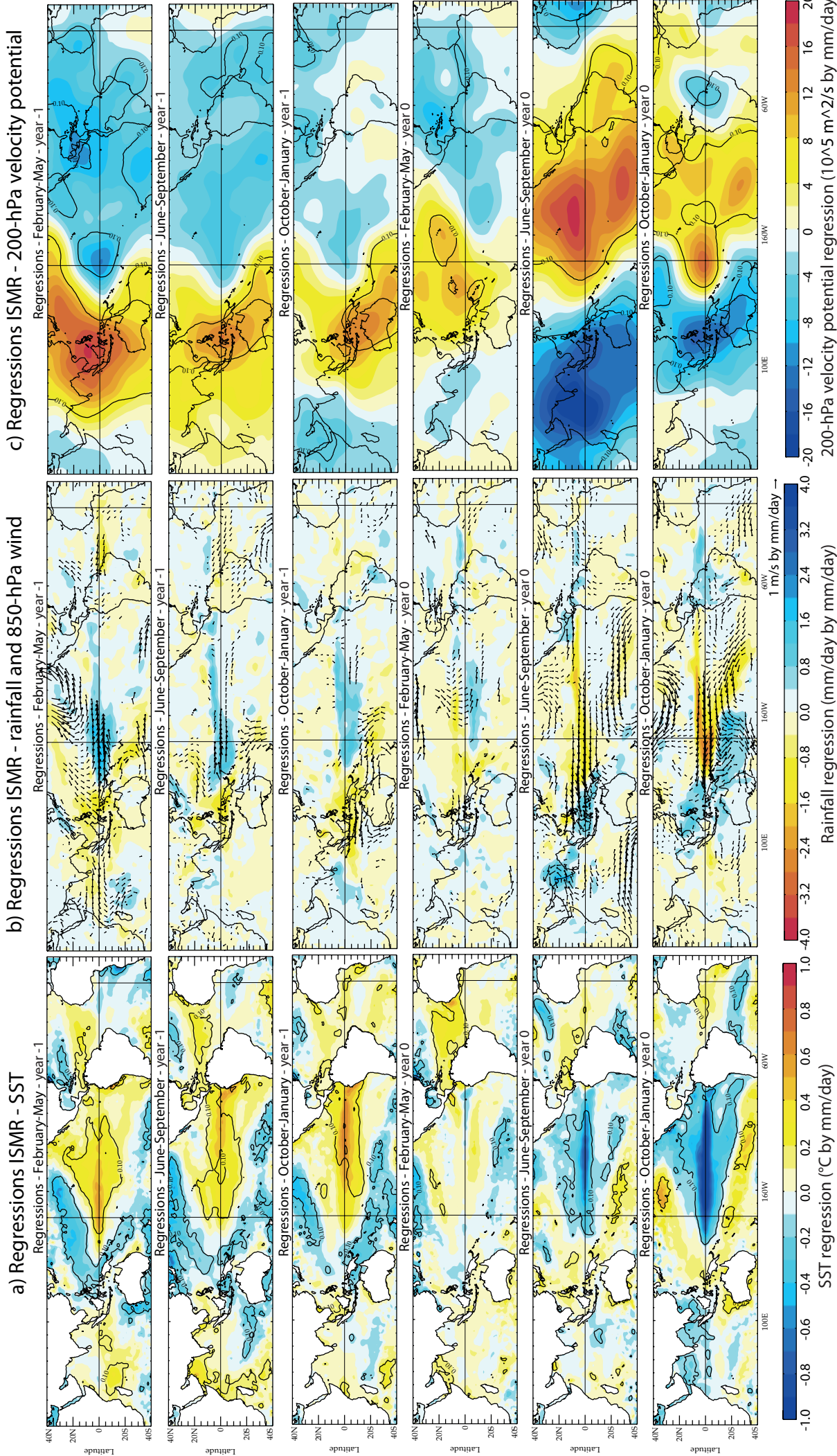
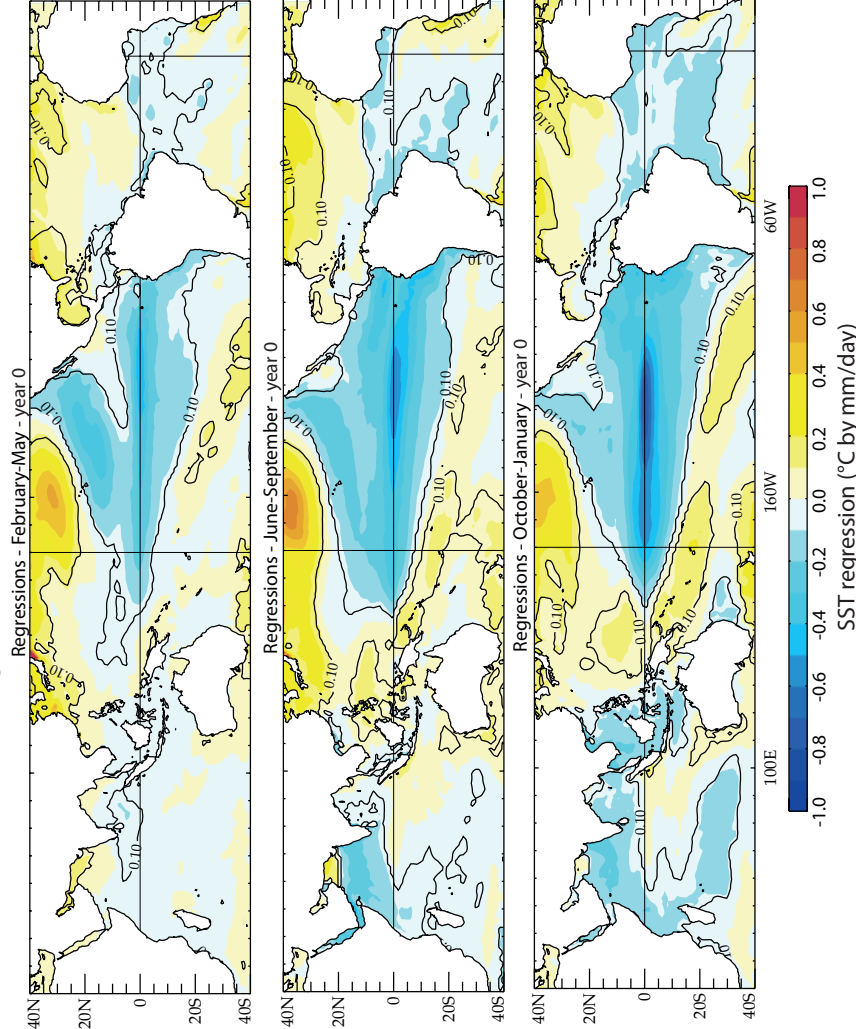
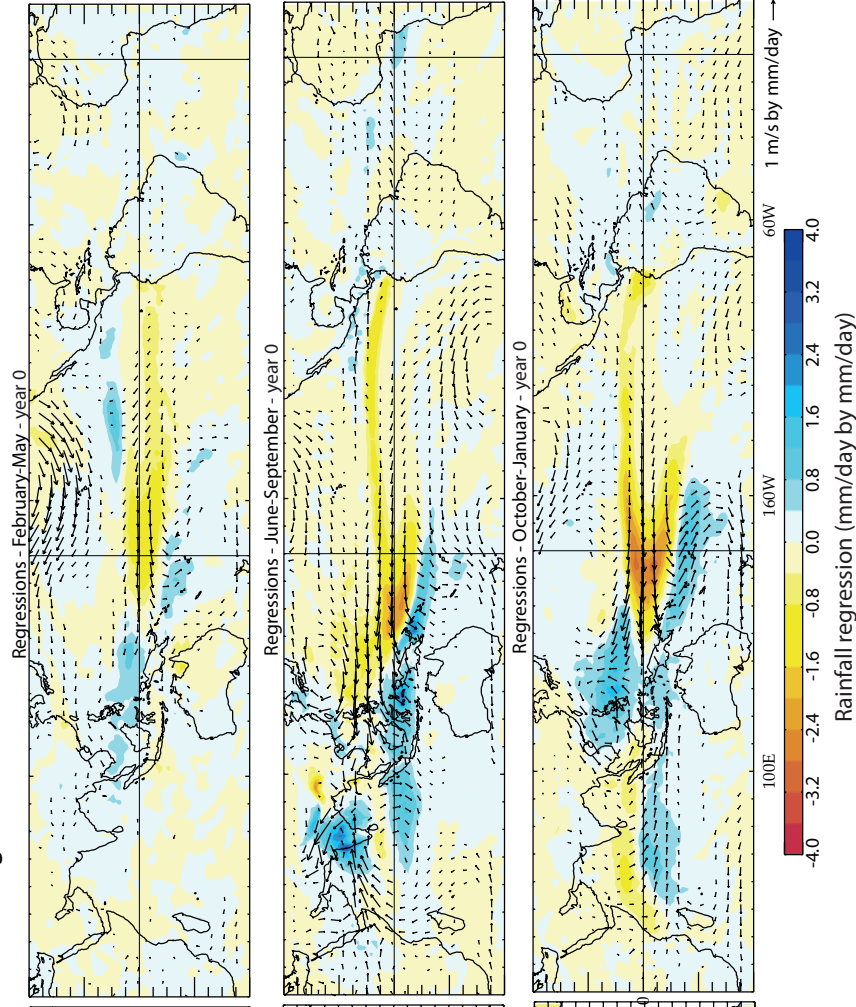


Figure 4: a) Quarterly SST time series (from ERAi) during the preceding (e.g. year -1) and simultaneous (e.g. year 0) ISM years, regressed against the ISMR index (e.g. JJAS ISM rainfall from GPCP). Unit for the SST regression coefficient is °C by mm/day. b) Same as a), but for regression using quarterly rainfall and 850-hPa wind time series. Units for the rainfall and 850-hPa wind regression coefficients are mm/day by mm/day and m s⁻¹ by mm/day, respectively. c), Same as a), but for quarterly 200-hPa velocity potential time series. Unit for the 200-hPa velocity potential regression coefficient is 10⁵ m² s⁻¹ by mm/day. Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) with 999 samples are contoured (SST or 200-hPa velocity potential) or shown (rainfall and 850-hPa wind). Quarterly time series refer to the seasons February-May, June-September, October-January and so on.

a) Regressions ISMR - SST CTRL



b) Regressions ISMR - rainfall and 850-hPa wind CTRL



c) Regressions ISMR - 200-hPa velocity potential CTRL

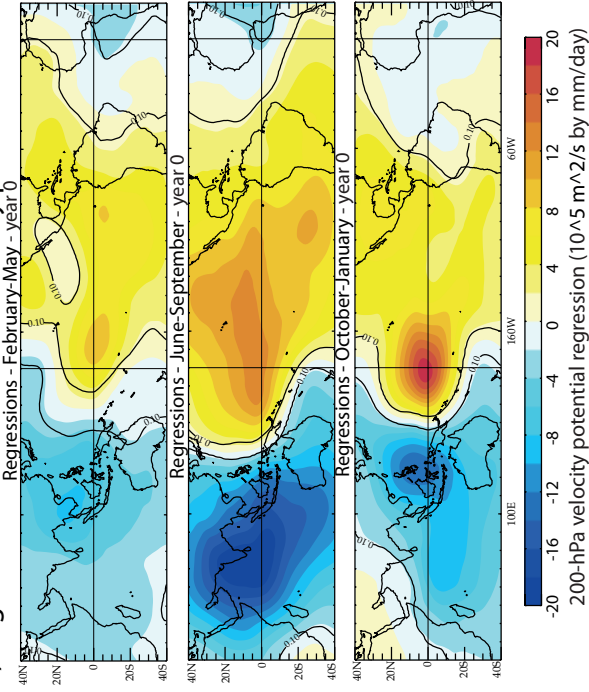


Figure 5: a) Quarterly SST time series during year 0 regressed against the ISMR index (e.g. JJAS ISM rainfall) in CTRL. Unit for the SST regression coefficient is °C by mm/day. b) Same as a), but for regression using quarterly rainfall and 850-hPa wind time series in CTRL. Units for the rainfall and 850-hPa wind regression coefficients are mm/day by mm/day and m s⁻¹ by mm/day, respectively. c), Same as a), but for quarterly 200-hPa velocity potential time series in CTRL. Unit for the 200-hPa velocity potential regression coefficient is 10⁵ m² s⁻¹ by mm/day. Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity potential) or shown (rainfall and 850-hPa wind).

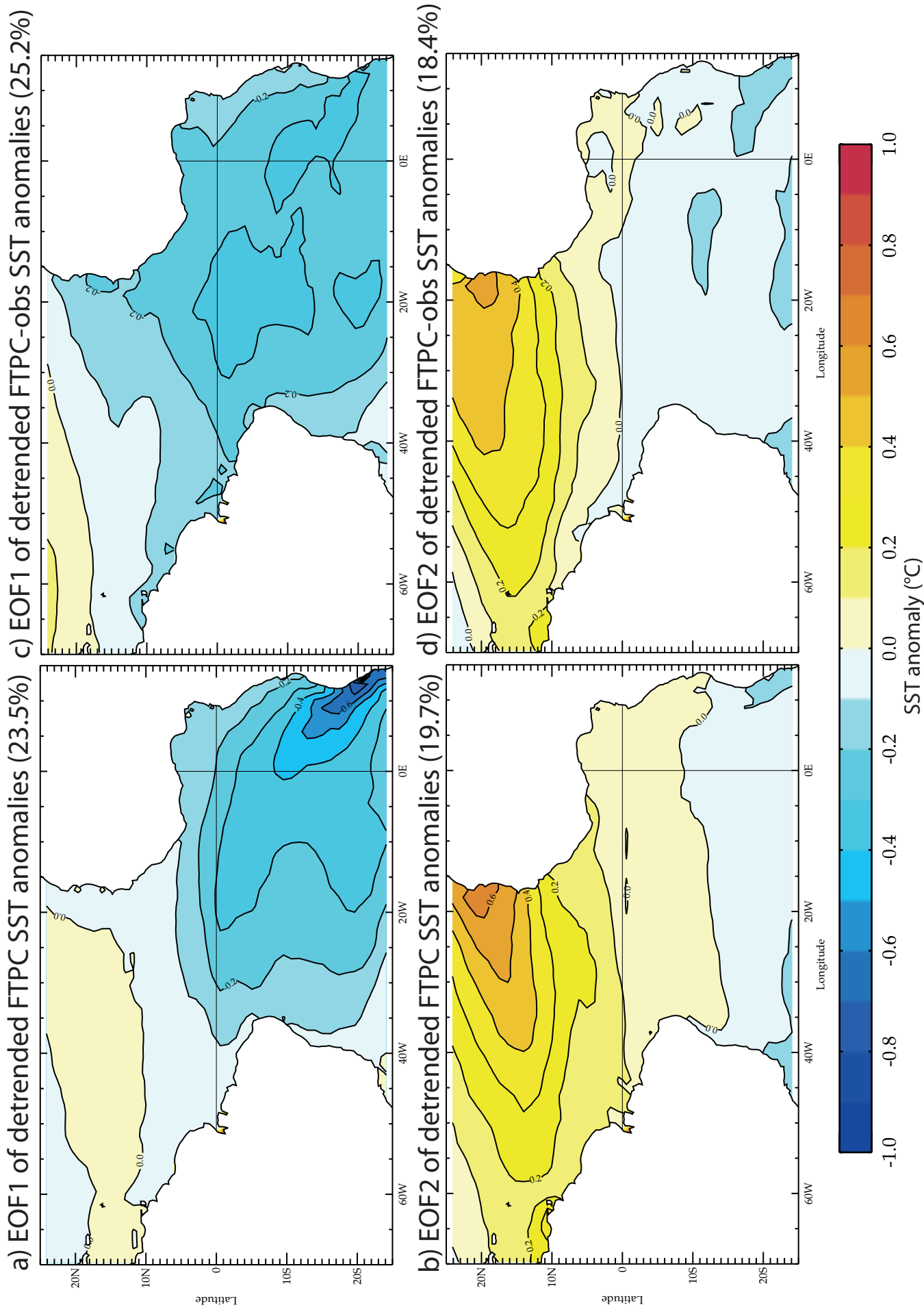
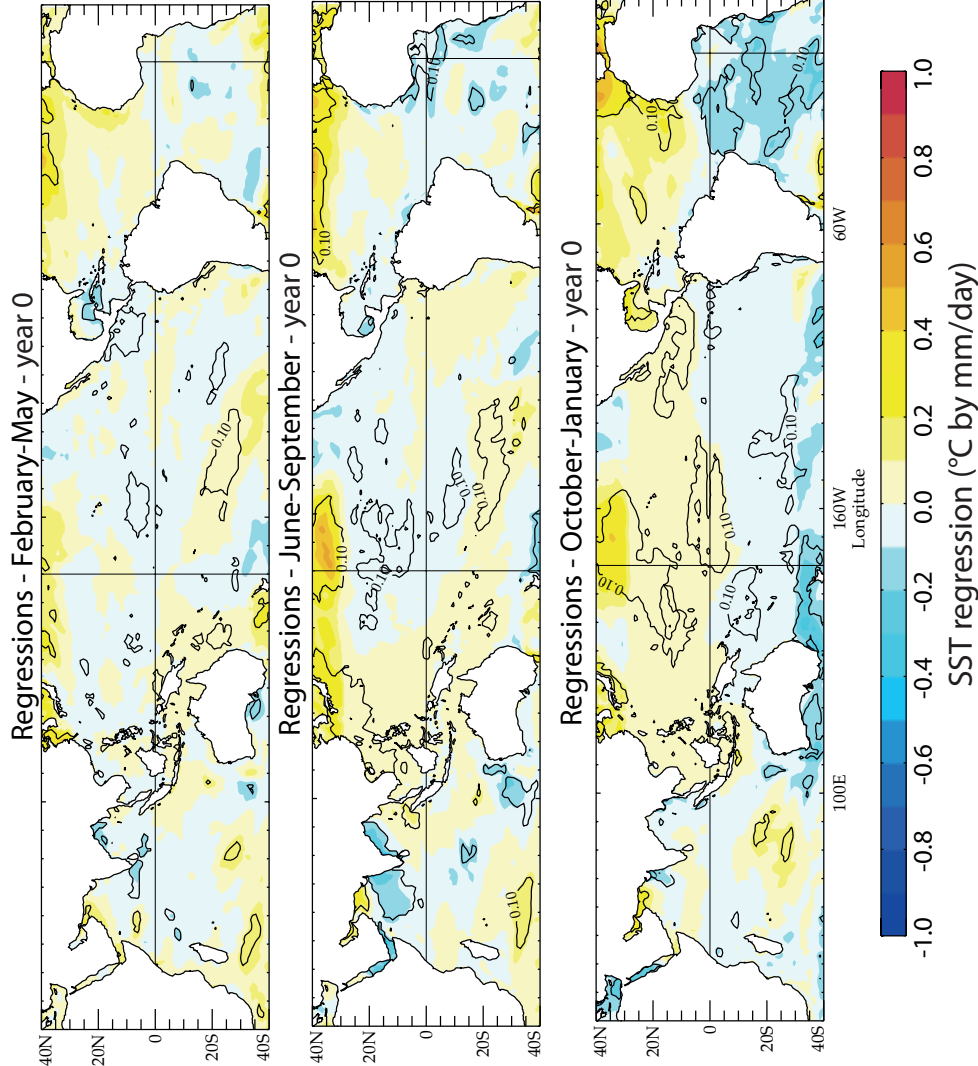


Figure 6: Empirical Orthogonal Function (EOF) 1st and 2nd modes for 12-monthly anomaly of detrended SST obtained from FTPC and FTPC-obs experiments. a) and b) EOF1 and EOF2 from FTPC SST detrended anomalies (11-10 period), respectively. c) and d) EOF1 and EOF2 from FTPC-obs SST detrended anomalies (11-50 period), respectively. The number in parentheses for each panel gives the % of SST variance described by the EOF mode.

a) Regressions ISMR - SST - FTPC-obs



b) Regressions ISMR - 200-hPa velocity potential - FTPC-obs

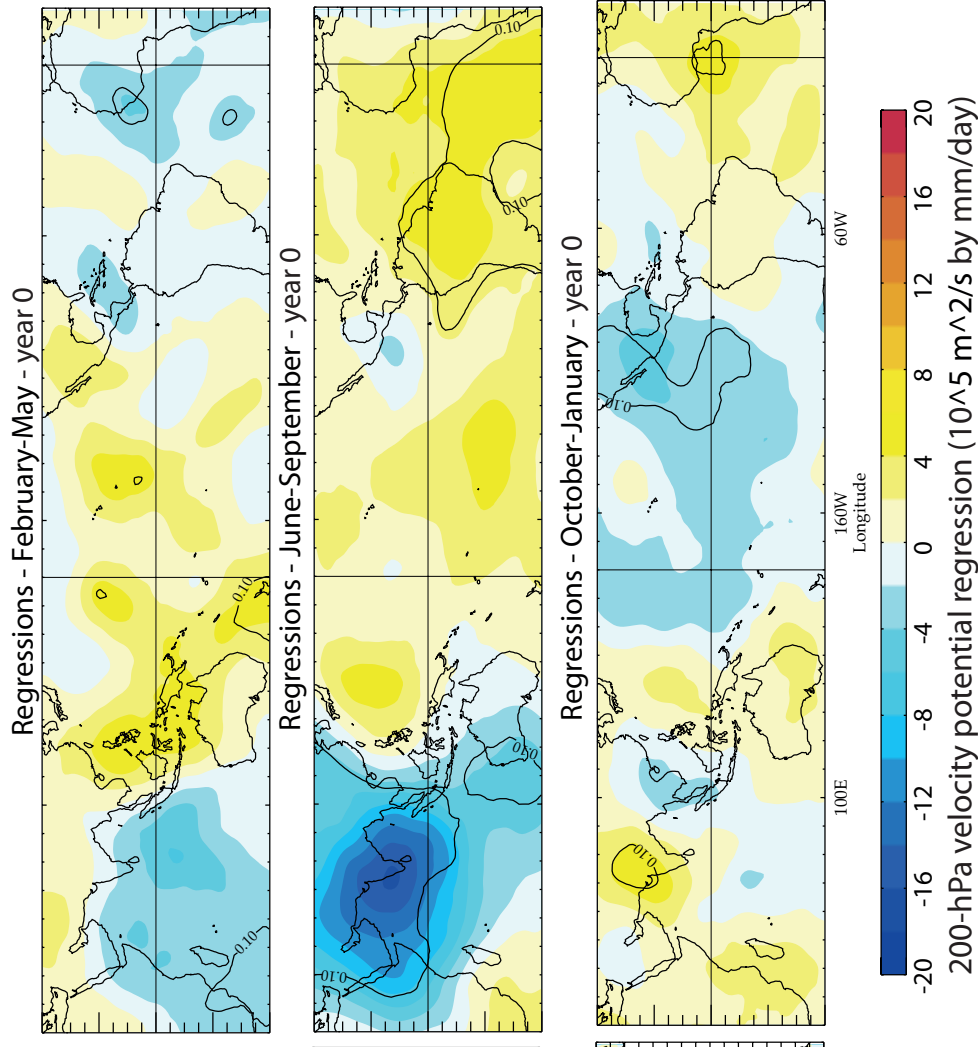


Figure 7: a) Quarterly SST time series during year 0 regressed against the ISMR index (e.g. JJAS ISM rainfall) in FTPC-obs. Unit for the SST regression coefficient is °C by mm/day. b) Same as a), but for quarterly 200-hPa velocity potential time series in FTPC-obs. Unit for the 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured.

Nino34 SST monthly standard deviations

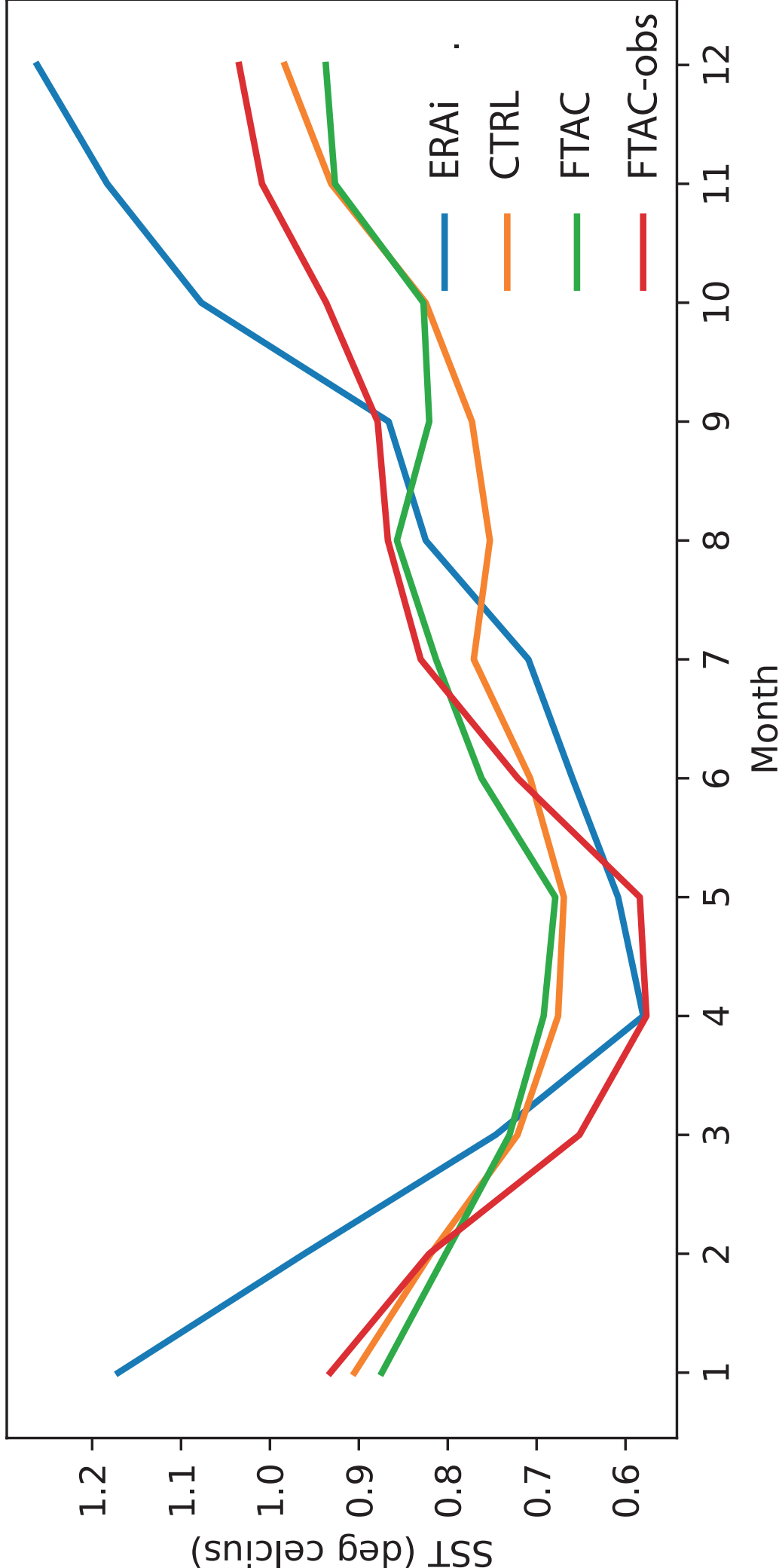
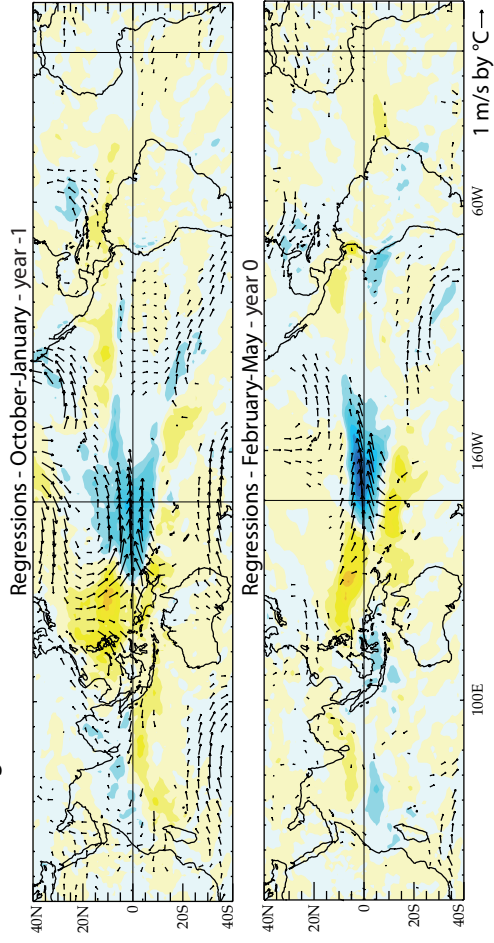
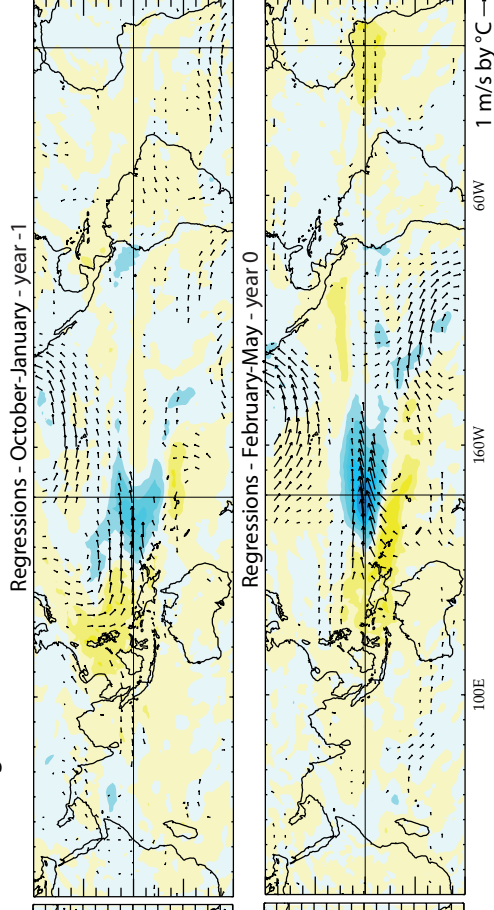


Figure 8: Monthly standard deviations of Nino34 SST (unit:°C) from ERAI (blue), CTRL (orange), FTAC (green) and FTAC-obs (red).

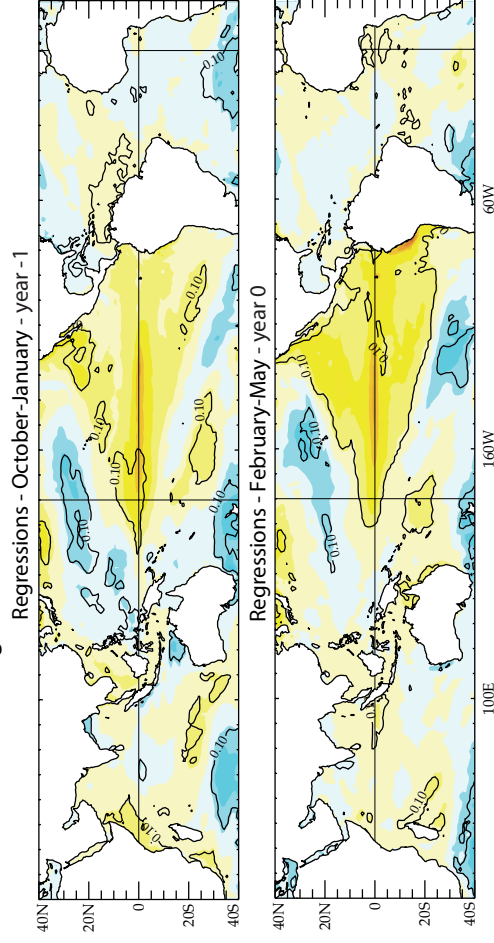
a) Regressions Nino34 (10-1) - rainfall and 850-hPa wind - FTAC-obs



b) Regressions Nino34 (10-1) - rainfall and 850-hPa wind - FTAC



c) Regressions Nino34 (10-1) - SST - FTAC-obs



d) Regressions Nino34 (10-1) - SST - FTAC

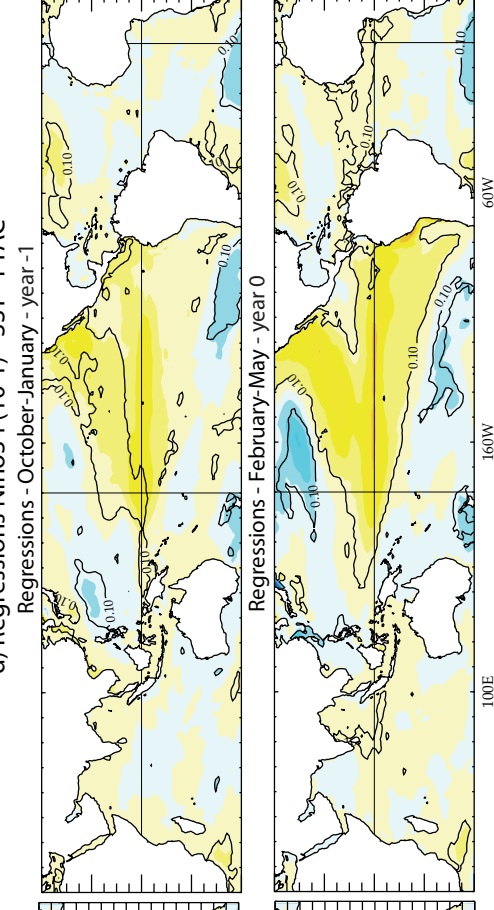
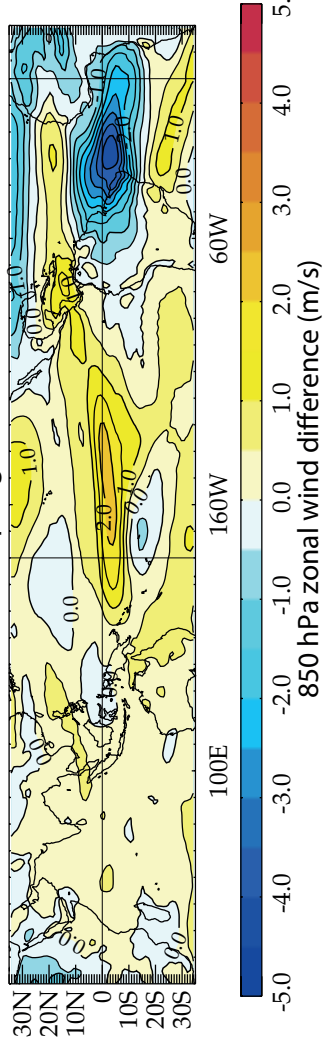
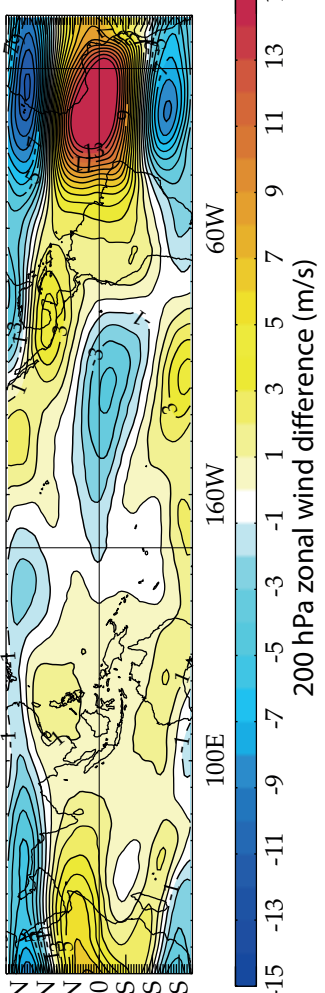


Figure 9: a) Quarterly rainfall and 850-hPa wind time series during boreal winter and spring regressed against the Nino34 index during the following boreal winter (e.g. ONDJ Nino34 SST) in FTAC-obs. b) Same as a), but in FTAC. c) Quarterly SST time series during boreal winter and spring regressed against the Nino34 index during the following boreal winter in FTAC-obs. d) Same as c), but in FTAC. Unit for the SST regression coefficient is °C by °C. Units for the rainfall and 850-hPa wind regression coefficients are mm/day by °C and m s⁻¹ by °C, respectively. Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity potential) or shown (rainfall and 850-hPa wind).

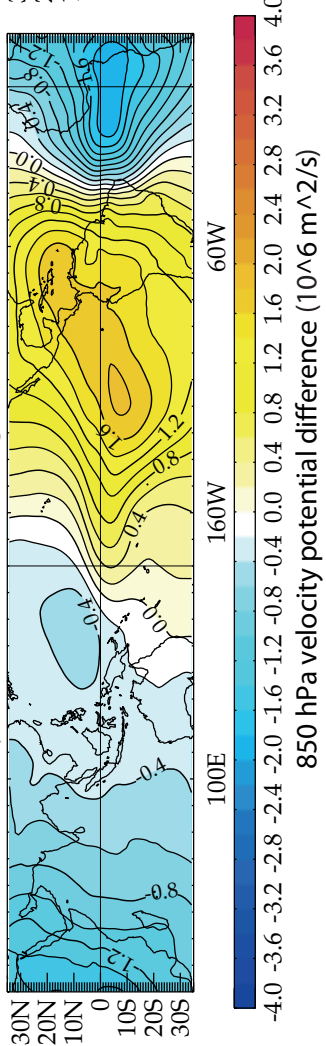
a) 850 hPa zonal wind boreal spring differences - FTAC-obs - CTRL



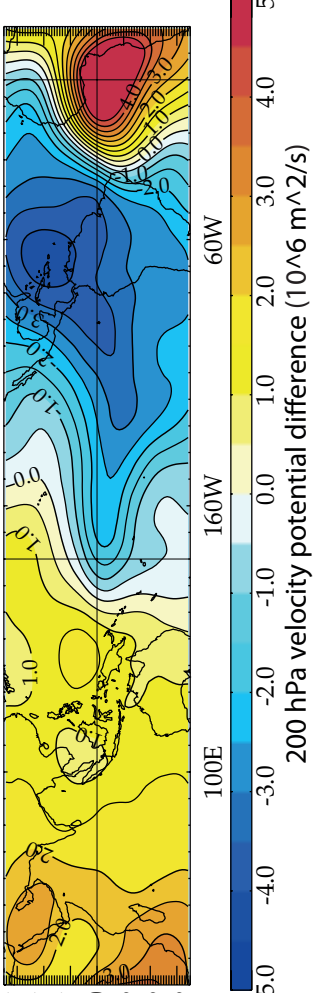
b) 200 hPa zonal wind boreal spring differences - FTAC-obs - CTRL



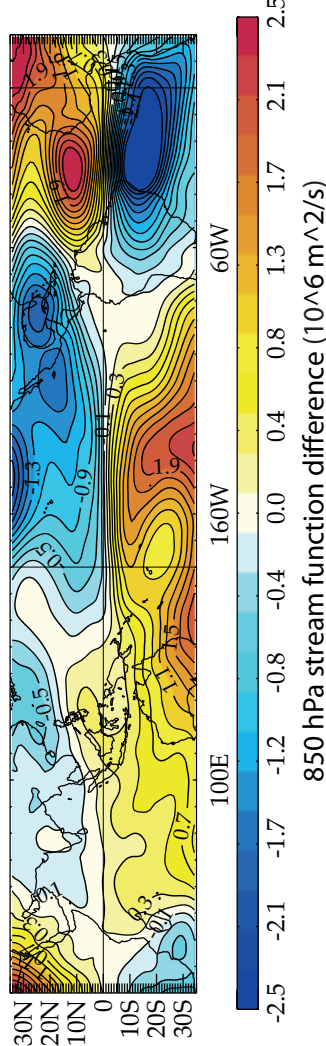
c) 850 hPa velocity potential boreal spring differences - FTAC-obs - CTRL



d) 200 hPa velocity potential boreal spring differences - FTAC-obs - CTRL



e) 850 hPa stream function boreal spring differences - FTAC-obs - CTRL



f) 200 hPa stream function boreal spring differences - FTAC-obs - CTRL

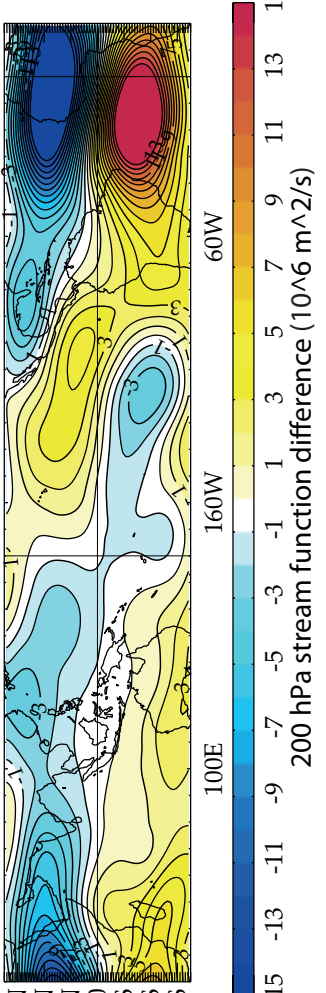


Figure 10: a) 850-hPa zonal wind boreal spring means differences (unit: $m\ s^{-1}$) between FTAC-obs and CTRL; b) 200-hPa zonal wind boreal spring means differences (unit: $m\ s^{-1}$) between FTAC-obs and CTRL; c) 850-hPa velocity potential boreal spring means differences (unit: $10^6\ m^2\ s^{-1}$) between FTAC-obs and CTRL; d) 200-hPa velocity potential boreal spring means differences (unit: $10^6\ m^2\ s^{-1}$) between FTAC-obs and CTRL; e) 850-hPa stream function boreal spring means differences (unit: $10^6\ m^2\ s^{-1}$) between FTAC-obs and CTRL; f) 200-hPa stream function boreal spring means differences (unit: $10^6\ m^2\ s^{-1}$) between FTAC-obs and CTRL.

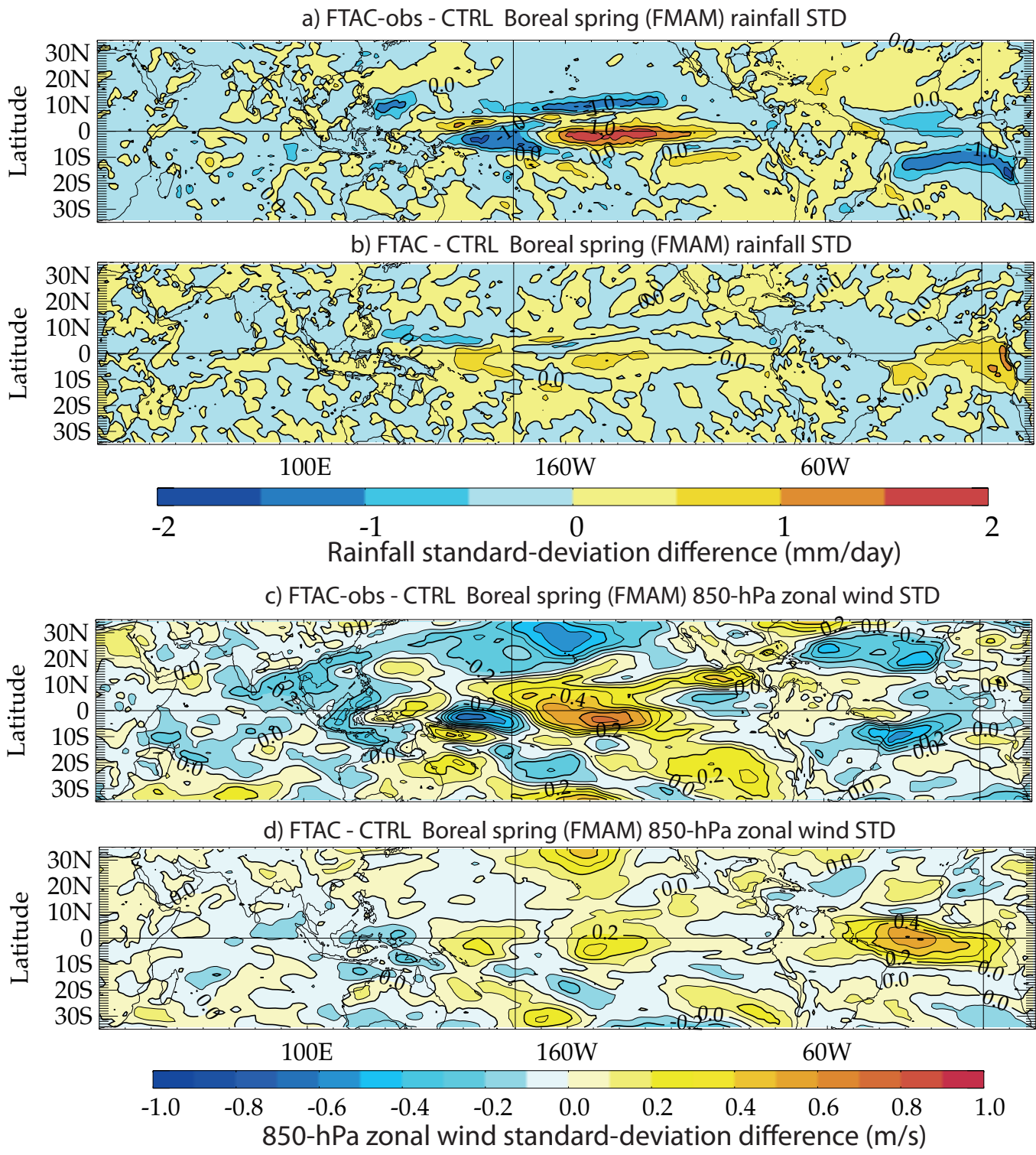


Figure 11: a) Rainfall boreal spring standard deviation differences (unit: mm/day) between FTAC-obs and CTRL. b) Same as a), but for FTAC and CTRL. c) 850-hPa zonal wind boreal spring standard deviation differences (unit: m s⁻¹) between FTAC-obs and CTRL. d) Same as c), but for FTAC and CTRL.

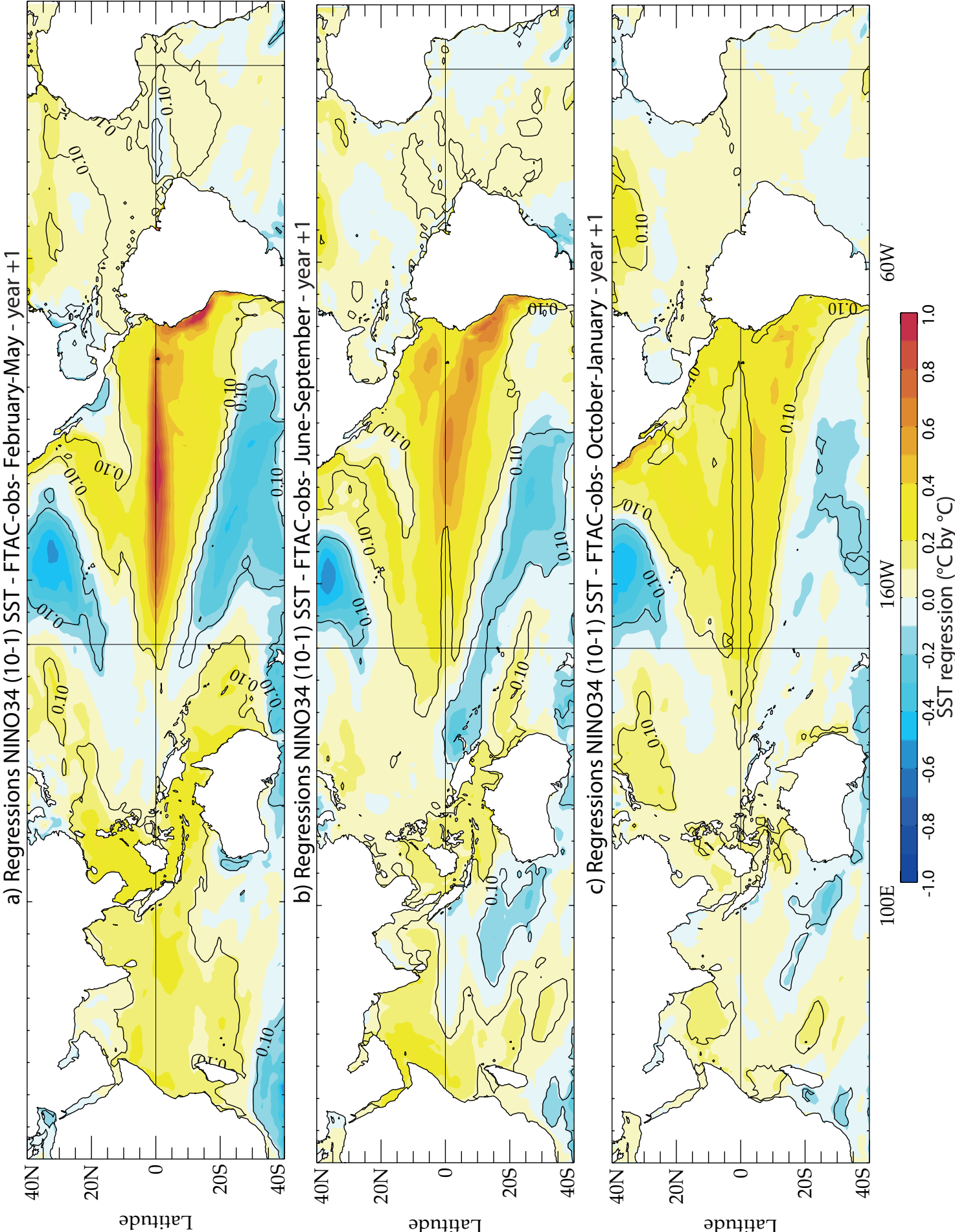


Figure 12: a) Boreal spring SST time series regressed against the Niño34 index during the preceding boreal winter (e.g. ONDJ Niño34 SST) in FTAC-obs; b) Boreal summer SST time series regressed against the Niño34 index during the preceding boreal winter in FTAC-obs; c) Boreal winter SST time series regressed against the Niño34 index during the preceding boreal winter in FTAC-obs. Unit for the SST regression coefficient is $^{\circ}\text{C by }^{\circ}\text{C}$. Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured.

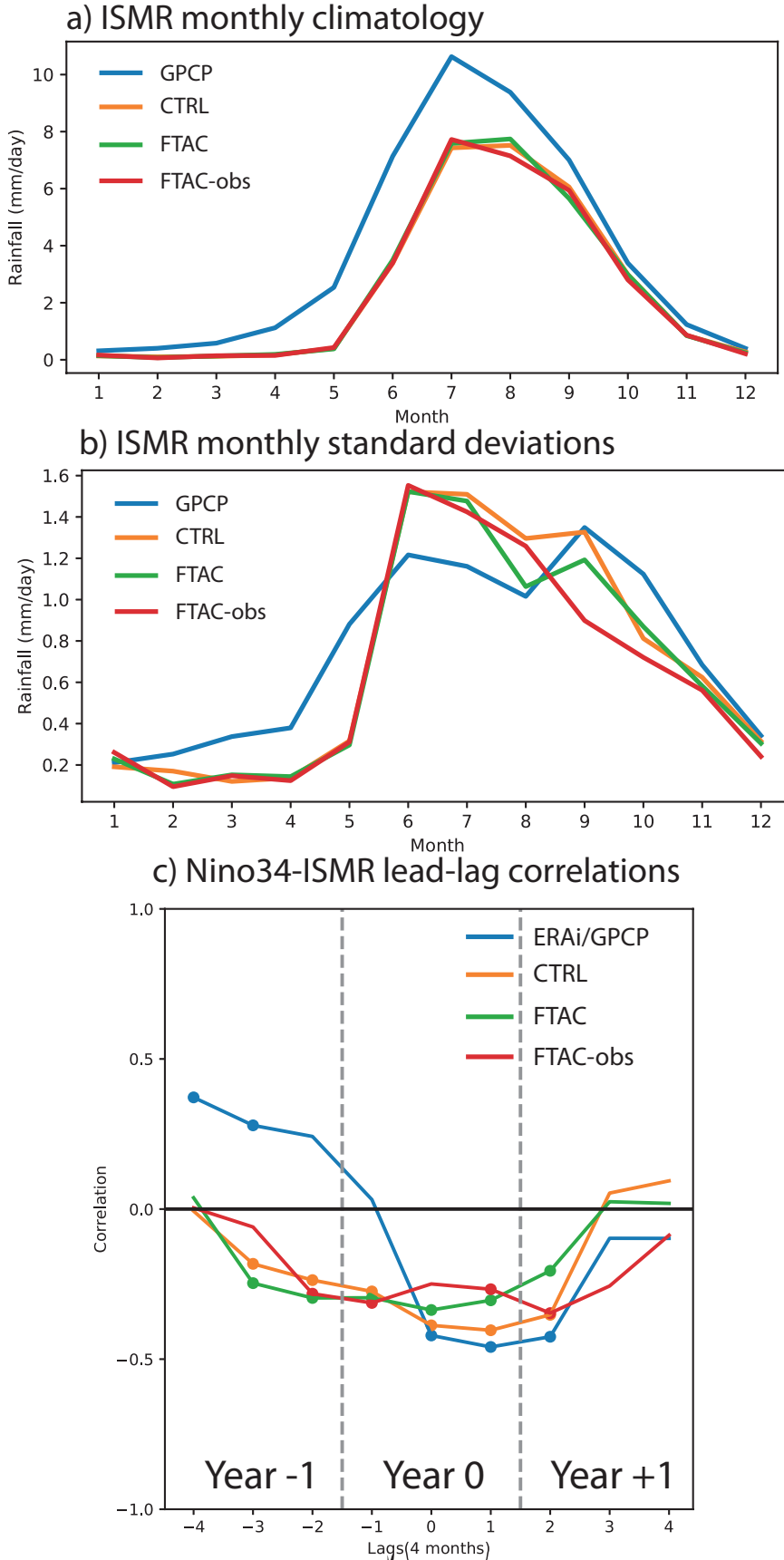


Figure 13: a) Monthly means of ISMR (unit: mm/day) from GPCP (blue), CTRL (orange), FTAC (green) and FTAC-obs (red); b) Monthly standard deviations of ISMR (unit: mm/day) from GPCP (blue), CTRL (orange), FTAC (green) and FTAC-obs (red); c) Lead-lag correlations between ISMR and quarterly Niño-3.4 SSTs 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) in observations (blue), CTRL (orange), FTAC (green) and FTAC-obs (red). X-axis indicates the lag (in 4 months interval) for a 36 months period starting one year before the developing year of ISMR (e.g. year 0) and Y-axis is the amplitude of the correlation. Thus, the coefficients corresponding to -1, 0, +1 lags refer, respectively, to the correlations between ISMR in year 0 (e.g. JJAS ISM rainfall) and February-May, June-September and October-January Niño-3.4 SSTs, also during year 0, and so on. Circles indicate correlations that are above the 90% significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki 1997).

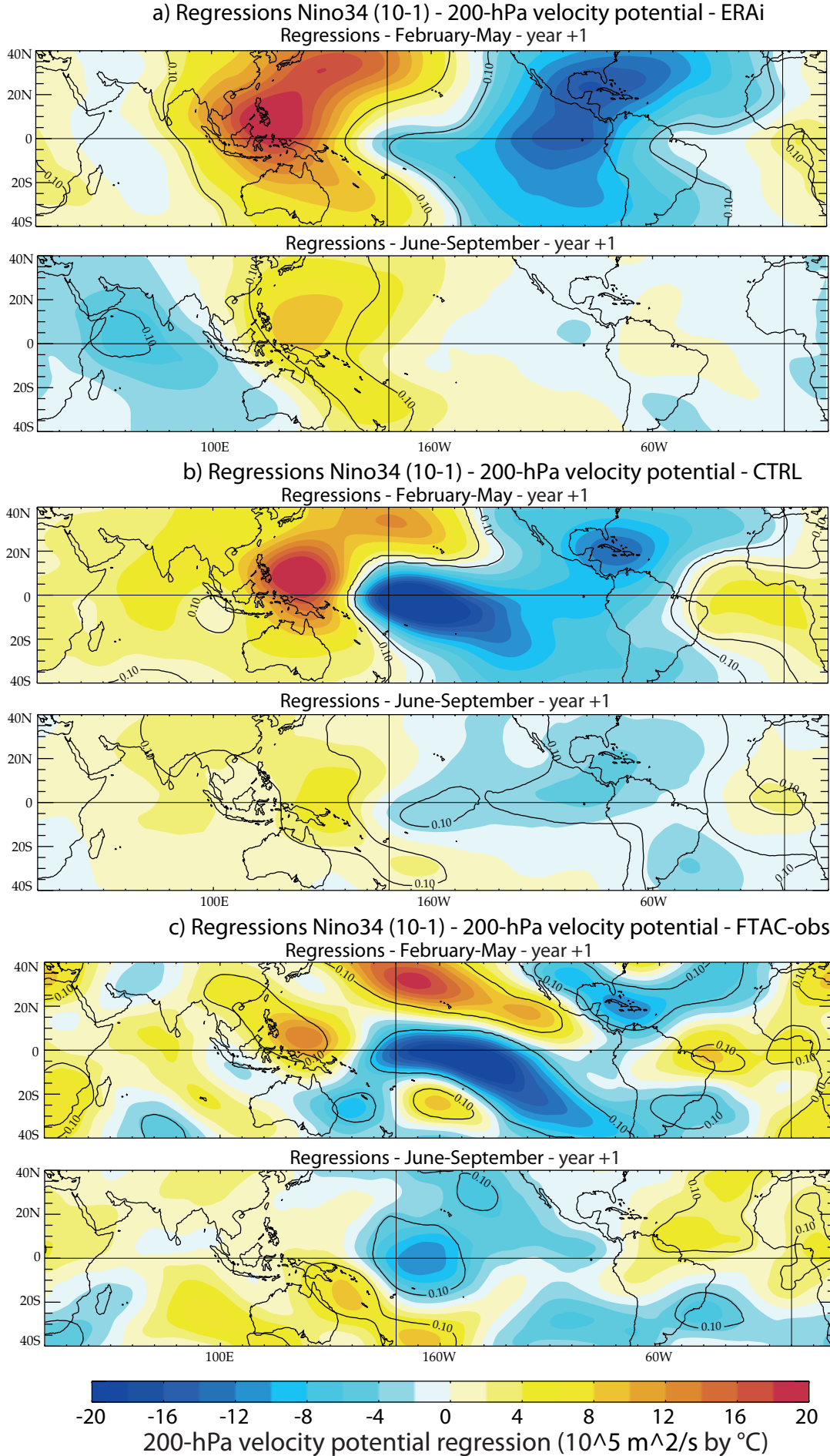
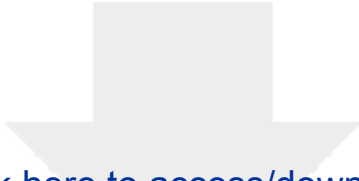


Figure 14: a) Boreal spring and summer 200-hPa velocity potential time series regressed against the preceding boreal winter Nino34 SST in ERAi. b) Same as a), but for CTRL. c) Same as a), but for FTAC-obs. Unit for the 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day . Regression coefficients reaching the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured.

Table 1

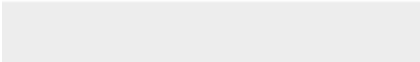
Name	CTRL	FTPC	FTPC-obs	FTAC	FTAC-obs
Correction area	None	Pacific Ocean coast to coast 25°S-25°N	Pacific Ocean coast to coast 25°S-25°N	Atlantic Ocean coast to coast 25°S-25°N	Atlantic Ocean coast to coast 25°S-25°N
Smoothing area	None	30°S-25°S 25°N-30°N	30°S-25°S 25°N-30°N	30°S-25°S 25°N-30°N	30°S-25°S 25°N-30°N
SST data	None	CTRL	OISSTv2	CTRL	OISSTv2
Time duration (Year)	210	110	50	110	50

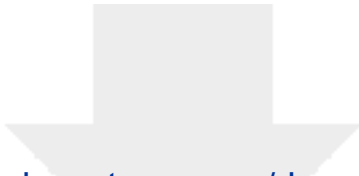
Table 1: Summary of the numerical experiments with their main characteristics, including length, nudging domain and SST climatology used for the nudging in the AO or PO decoupled experiments. The nudged experiments are the Forced Tropical Pacific Climatology (FTPC), the Forced Tropical Pacific observed Climatology (FTPC-obs), the Forced Tropical Atlantic Climatology (FTAC) and, finally, the Forced Tropical Atlantic observed Climatology (FTAC-obs) runs. See text for more details. For the FTPC and FTPC-obs experiments only ocean grid-points in the PO are included in the correction or smoothing areas and, similarly, for the FTAC and FTAC-obs experiments. The different correction domains are also displayed in Figure 1a. The observed SST daily climatology used in the FTPC-obs and FTAC-obs experiments 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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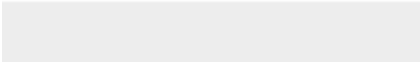
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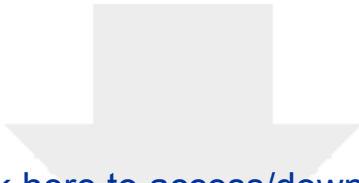




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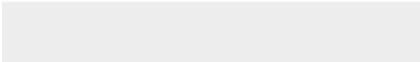
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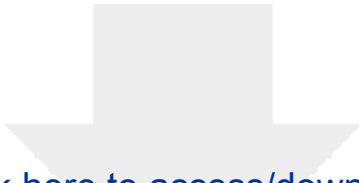




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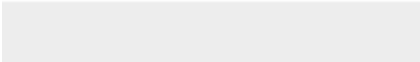
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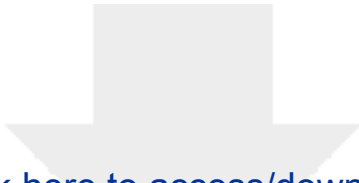




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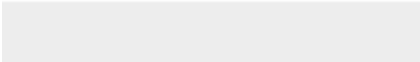
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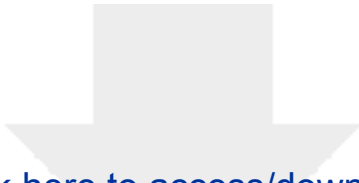




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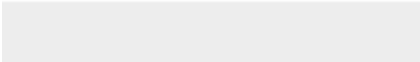
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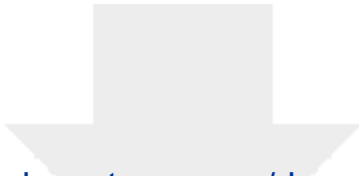




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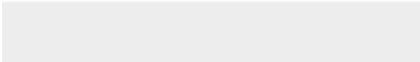
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1 **Anatomy of the Indian Summer Monsoon and ENSO**

2 **relationship in a state-of-the-art CGCM:**

3 **Role of the tropical Atlantic Ocean**

4

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6

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22

Abstract

23 The main paradigm for prediction of Indian Summer Monsoon Rainfall (ISMR) is its inverse relation
24 with El Niño-Southern Oscillation (ENSO). In this study, we focus on the role of the Atlantic Ocean
25 (AO) Sea Surface Temperature (SST) variability on the ISMR. There are basically two ways by which
26 AO SSTs can impact the ISMR: a “direct pathway” in which the AO may directly force the ISMR in
27 the absence of interactions with other dominant forcings like ENSO, and an “indirect pathway” in
28 which AO forces ENSO and modulates the ENSO teleconnection to ISMR. These two pathways are
29 studied with the help of sensitivity experiments performed with a Coupled General Circulation Model
30 (CGCM). Two pairs of decoupling experiments have been done. In the first, the SST variability in the
31 tropical AO or Pacific Ocean (PO) is removed by nudging the SST in these regions from a control
32 run's SST climatology. In the second set, the SST nudging is performed from the observed SST
33 climatology, which allows us to assess the robustness of the results and the specific role of the model's
34 SST mean-state biases.

35 The direct pathway linking tropical AO SST variability onto ISMR is insignificant in the PO
36 decoupled experiments or in recent observations. Furthermore, these experiments suggest on the
37 contrary that many AO SST anomalous patterns could be forced by ISMR. On the other hand, for the
38 indirect pathway, the AO decoupled experiments demonstrate that AO SST variability modulates the
39 onset and decaying phases of ENSO events. Despite ENSO is as strong and persists longer than in the
40 control simulation, the AO SST nudging resulted in a significant weakening of the inverse relationship
41 between ENSO and ISMR. The ENSO-monsoon relationship is mainly modulated during the ENSO
42 decaying phase. The upper-level divergent wind flows mainly from the Pacific to the AO resulting in
43 rainfall suppression in the AO, but only in a weak forcing on ISMR during boreal summer of the
44 ENSO decaying year in the AO decoupled experiments. Thus, the AO rainfall variability in these
45 experiments is decoupled from the surface and mainly modulated by the upper-level convergence or
46 divergence induced by the remote ENSO forcing.

47 Finally, the rectification of the AO SST mean-state biases in the CGCM also induces an El Niño-like
48 mean pattern over the tropical Pacific during boreal spring and promotes a stronger ENSO during its
49 peak phase. This demonstrates that the prominent AO SST mean-state biases in current CGCMs
50 further complicate the dynamical prediction and simulation of ISMR and ENSO.

51 **Keywords:** Indian Summer Monsoon; El Niño-Southern Oscillation; tropical Atlantic Ocean; ocean-
52 atmosphere interactions; Walker circulation, coupled climate model.

53

54 **1. Introduction**

55 In India, the rainy season is from June to September (JJAS) and Indian Summer Monsoon
56 Rainfall (ISMR) provides 80% of India's total annual precipitation. Despite the standard
57 deviation of ISMR is only about 10% of its mean, ISMR variability has a tremendous impact
58 on water resource management, agricultural yield and India's gross domestic product (Gadgil
59 and Gadgil 2006). However, forecasting ISMR variability is still a scientific challenge (Rao et
60 al. 2019) and an active research area as it involves many factors and their complex
61 interactions (see Chowdary et al. 2021 for recent review).

62 Numerous studies have examined climatic controls on ISMR interannual variability and
63 most of them showed the role played by tropical Pacific, Indian and Atlantic oceans Sea
64 Surface Temperature (SST) anomalies (Chowdary et al. 2021). El Niño-Southern Oscillation
65 (ENSO) is the primary forcing of year-to-year ISMR variability (Webster et al. 1998).
66 However, since ENSO can only explain about 35% the interannual variance of ISMR and the
67 ISM-ENSO relationship has weakened during the latter part of the 20th century, partly in
68 response to coherent multi-decadal variability of the climate system (Kumar et al. 1999;
69 Kucharski et al. 2007; Srivastava et al. 2019; Yang and Huang 2021), it is important to look
70 for other sources of ISMR predictability.

71 First, many studies have suggested a connection between ISM and Indian Ocean (IO)
72 SSTs, especially the Indian Ocean Dipole (IOD; see reviews in Cherchi et al. 2021). The IOD
73 is an irregular interannual SST oscillation in which the eastern equatorial IO gets alternately
74 colder and then warmer than the western part during boreal fall. Positive IOD events (e.g.,
75 warm in the western IO) may enhance ISMR through moisture transport over the western IO
76 or modification of the local Hadley cell with increased ascendance over the Indian region
77 (Cherchi et al. 2021). However, the influence of IOD on both ISMR and ENSO remains a
78 controversial topic (Meehl et al. 2003; Fischer et al. 2005; Izumo et al. 2010; Cretat et al.
79 2017, 2018; Stuecker et al. 2017; Terray et al. 2021; Cherchi et al. 2021; Zhang et al. 2021a).

80 The Atlantic Ocean (AO) can also add its impact on ISMR (Kucharski et al. 2007, 2009;
81 Pottapinjara et al. 2014, 2016; Sabeerali et al. 2019; Vittal et al. 2020; Yang and Huang
82 2021). First, a basin-warming mode exists in the tropical AO and is known as Atlantic Niño
83 or the Atlantic Zonal Mode (AZM; Lübbecke et al. 2018; Cabos et al. 2019; Richter and
84 Tokinaga 2021). These Atlantic Niños peak during boreal summer and are formed due to a
85 Bjerknes feedback as for ENSO, but last only for 3-4 months due to the weaker ocean-
86 atmosphere interactions in this basin (Lübbecke et al. 2018; Cabos et al. 2019; Richter and

87 Tokinaga 2021). It is still debated whether and how ENSO affects the AZM (Tokinaga et al.
88 2019). However, Atlantic Niños give rise to important shifts in local rainfall and are
89 associated with a Matsuno-Gill atmospheric response during boreal summer (Gill 1980;
90 Kucharski et al. 2009; Li et al. 2016; Jiang and Li 2021), which may modulate the
91 tropospheric temperature gradient in the Indo-Pacific sector and ISMR (Kucharski et al. 2009,
92 Pottapinjara et al. 2014, 2016; Sabeerali et al. 2019; Jiang and Li 2021). The second leading
93 mode of tropical AO SST variability involves fluctuations of the interhemispheric SST
94 gradient in the AO and is known as the Atlantic Meridional Mode (AMM; Chiang and
95 Vimont 2004; Jiang and Li 2021). The AMM is triggered and sustained by a Wind-
96 Evaporation-SST (WES) feedback (Chiang and Vimont 2004; Cabos et al. 2019). However,
97 the Tropical North Atlantic (TNA) SST anomaly dominates AMM variability (Enfield and
98 Mayer 1997; Jiang and Li 2021) and ENSO plays a dominant role in causing the spring time
99 trade wind variability over the TNA and the generation of local SST anomalies by evaporative
100 cooling/warming through ENSO teleconnections (Enfield and Mayer 1997; Garcia-Serrano et
101 al. 2017; Jiang and Li 2019). A few studies also suggest a link between the AMM or warm
102 SST TNA anomalies and ISMR (Vittal et al. 2020; Yang and Huang 2021).

103 An important difficulty for assessing the role of AO on ISMR, is that it interacts also
104 directly with the PO and IO in a complex manner and at different time scales (Kucharski et al.
105 2009, 2011; Rodriguez-Fonseca et al. 2009; Ham et al. 2013ab, McGregor et al. 2014, 2018;
106 Li et al. 2016; Terray et al. 2016; Wang et al. 2017; Cai et al. 2019; Li et al. 2020; Jiang and
107 Li 2021; Zhang and Han 2021). Recent studies suggest that warm TNA and AZM SST
108 anomalies can force a La Niña-like SST pattern in the Pacific (Rodriguez-Fonseca et al. 2009;
109 Ding et al. 2012; Ham et al. 2013ab; Wang et al. 2017; Jiang and Li 2021). But, again the role
110 of AO in ENSO and ISMR is debated in the literature (Ding et al. 2012; Zhang et al. 2021b;
111 Richter et al. 2021). As an illustration, previous studies suggest that Atlantic Niños may
112 reduce ISMR (Kucharski et al. 2007, 2009; Pottapinjara et al. 2016), but at the same times it
113 may promote La Niña conditions over the Pacific (Rodriguez-Fonseca et al. 2009; Ding et al.
114 2012; Jiang and Li 2021), which will be associated with enhanced ISMR. Therefore, it is
115 difficult to isolate the net effect of AO SST anomalies as the players, ISM, ENSO and AO
116 modes interact with each other in multiple different ways (Kucharski et al. 2009; Ding et al.
117 2012; Ham et al. 2013b; Cai et al. 2019; Jiang and Li 2021; Yang and Huang 2021).

118 This review highlights that there are two pathways by which the AO SSTs can affect
119 ISMR, one by a “direct” forcing on the ISMR and the other, “indirect”, by the AO forcing on
120 ENSO, which, in turn, will modulate the ISMR. However, as noted above, these two

121 pathways are not really independent of each other and the complexity of these interactions
122 implies that it is very difficult to assess the distinct causal relationships between the Atlantic,
123 Indian and Pacific SSTs and ISMR, or even the net effect of AO SSTs on ISMR from
124 observations alone. Considering these difficulties, we will assess the role of the tropical AO
125 on ISMR with the help of dedicated experiments performed with a Coupled General
126 Circulation Model (CGCM).

127 This paper is organised as follows. The validation datasets, CGCM and sensitivity
128 experiments used in this study are described in Section 2. In Section 3, the observed and
129 simulated relationships between ISMR, ENSO and AO SSTs at the interannual time scale are
130 documented. In Section 4, the AO “direct” and “indirect” effects on ISMR are assessed
131 through sensitivity coupled experiments. The final section summarizes the results of the
132 present work and presents some perspectives.

133

134 **2. Datasets, coupled model and sensitivity experiments**

135 2.a Observed datasets and time series indices.

136 Multiple data sources are used for model validation. SST and atmospheric variables (e.g.,
137 850- and 200-hPa winds, velocity potentials and stream functions) are taken or computed
138 from ERA-Interim reanalysis (ERAi; Dee et al. 2011) available from 1979 onwards. The
139 depth of the 20°C isotherm (Z20) is used as a proxy of the thermocline depth and is extracted
140 from the Simple Ocean Data Assimilation reanalysis for the 1979-2010 period (Carton and
141 Giese 2008; SODA version 2.2.4). We also used the Global Precipitation Climatology Project
142 rainfall dataset (GPCP; Huffman et al. 2001), which combines measures of precipitation
143 gauges and satellite data. GPCP is analyzed for the 1979-2016 period. All these quantities are
144 interpolated onto the model resolution to foster direct comparison with the simulations.

145 To monitor ISM, ENSO and AZM variability both in observations and simulations, we
146 define three standard time series indices:

147 - The ISMR time series is defined as the average of rainfall anomalies for the land grid points
148 in the region 5°-25°N and 70°-95°E.

149 - The Niño-3.4 SST (monthly average of SST anomalies in the region 5°S-5°N and 170°-
150 120°W; Niño34 hereafter) time series is chosen for the ENSO index since in observations the
151 strongest correlations between ISMR and tropical Pacific SSTs occur over this region.

152 - The ATL3 SST (monthly average of SST anomalies in the region 3°S-3°N and 20°W-0°E)
153 time series, which is a convenient index for the AZM (Lübbecke et al. 2018).

154 Note that our analysis of observations is robust if we estimate our Nino34, ATL3 and
155 ISMR time series from the Hadley Centre Sea Ice and SST dataset (Rayner et al. 2003) and
156 the rainfall dataset obtained from the Indian Meteorological Department (Pai et al. 2015).

157 2.b Coupled model and sensitivity experiments.

158 Here we employ a CGCM, the SINTEX-F2 (Masson et al. 2012), to assess the influence of
159 AO on ISMR variability and the ISM-ENSO relationship. The different model components
160 are ECHAM5.3 atmospheric model (Roeckner et al. 2003) at T106 spectral resolution
161 ($\sim 1.125^\circ \times 1.125^\circ$) and 31 hybrid sigma-pressure levels, NEMO ocean model (Madec 2008)
162 at $0.5^\circ \times 0.5^\circ$ horizontal resolution, 31 vertical levels and the LIM2 ice model (Timmermann
163 et al. 2005). The three model components are coupled using the Ocean-Atmosphere-Sea-Ice-
164 Soil (OASIS3) coupler (Valcke 2006). The model simulates the tropical Pacific SST mean
165 state, ENSO and ISMR variability reasonably well (Masson et al. 2012, Terray et al. 2016,
166 2021; Cretat et al. 2017, 2018).

167 First, a 210-yr fully coupled ocean-atmosphere simulation is used as a control (CTRL
168 hereafter). In order to disentangle the complex interactions between ISMR, ENSO and AO
169 SST variability, two partially coupled configurations of SINTEX-F2 are used and two
170 dedicated experiments have been performed with each of these configurations (see Table 1 for
171 details). In the first partial coupled configuration, full ocean-atmosphere coupling is used
172 everywhere except in the subtropical and tropical AO (25°S-25°N band), where SST is
173 nudged toward a daily SST climatology computed from CTRL or AVHRR-V2 daily
174 Optimum Interpolation SST observations during the 1982-2010 period (Reynolds et al. 2007).
175 These two AO decoupling experiments will be called FTAC and FTAC-obs and have been
176 run for 110-yr and 50-yr, respectively. In the second partial coupled configuration, ocean-
177 atmosphere coupling is active except in the subtropical and tropical Pacific (25°S-25°N band)
178 where, again, SST is nudged toward a daily SST climatology computed from CTRL or
179 observations. These two PO decoupling experiments will be called FTAC and FTAC-obs, and
180 have been run for 110-yr and 50-yr, respectively.

181 The nudging method used in these partial decoupling experiments modifies the non-solar
182 heat fluxes in the selected domain through a correction term that completely removes the SST
183 variability in the nudging domain (Terray et al. 2021). The damping term used in this nudging
184 technique ($-2400 \text{ W m}^{-2} \text{ K}^{-1}$) corresponds to the 1-day relaxation time for temperature in a 50-

185 m ocean layer. To avoid sharp SST gradients, a buffer zone is used between the “free” ocean
186 and regions of prescribed SST forcing such that the SSTs in these buffer regions are gradually
187 merged (over 5° latitude) with the prescribed SSTs. This strong SST restoring leads to an
188 almost complete decoupling between the ocean and atmosphere in the nudging domain with
189 SSTs, which ~~will~~ differ by no more than 0.1 K from the prescribed space-time climatology.

190 In FTAC and FTFC, there are no significant changes in SST mean-state in the nudged
191 region, but also in the whole Tropics compared to CTRL (not shown). On the other hand, in
192 FTAC-obs (and also FTFC-obs), the strong SST restoring removes the SST mean-state biases
193 present in CTRL in addition to suppressing SST variability in the selected domain. CTRL
194 exhibits a strong warm bias in the southeast AO (Fig. 1a), which is a common problem for
195 most CGCMs (Richter et al. 2014; [Voltaire et al. 2019](#); [Bi et al. 2022](#)). This bias is
196 attributed to errors in simulating zonal trade winds during boreal spring and is related to a
197 deeper thermocline, which weakens the upwelling of cold waters in the eastern AO (Fig. 1b).
198 Consistent with this erroneous east-west SST gradient, the rainfall pattern in the tropical AO
199 is shifted southeastward in CTRL compared to observations (Fig. 1c). Also consistent with
200 these mean-state biases, CTRL simulates a weaker SST variability over the eastern equatorial
201 AO compared to observations (Fig. S1b), especially during boreal summer, which is the
202 season of maximum SST variability in observations (Fig. S1a) as this is also the season when
203 the (observed) thermocline is the shallowest. Focusing on the PO, we note that CTRL is also
204 affected by a double Inter-Tropical Convergence Zone (ITCZ) bias (Fig. 1c) and a reduced
205 ENSO amplitude during boreal winter (Fig. S1b). As we will demonstrate in the following
206 sections, the inability of CTRL to realistically reproduce the seasonal cycle in the AO is a
207 cause of concern not only for the AO region, but also for ENSO and ISMR.

208 On the other hand, the SST restoring applied in FTAC-obs is able to correct largely these
209 simulated SST, thermocline and rainfall errors in the AO and also produces some changes in
210 the two other tropical basins (Figs. 1d-f). In other words, the comparison of CTRL, FTAC and
211 FTAC-obs (or FTFC and FTFC-obs) runs ~~is can be~~ useful to isolate the specific contribution
212 of the biased SST background mean-state in the coupled model.

213 More generally, the aim of these simulations is to isolate the effects of the PO and AO SST
214 variability on the simulated ISMR and ENSO-ISM relationship. First, FTFC and FTFC-obs
215 will be used to assess the « direct » relationship between AO SSTs and ISMR in a climate
216 without any counteracting ENSO forcing. Second, FTAC and FTAC-obs are useful to assess
217 if and how AO SST variability modulates ENSO and the simulated monsoon-ENSO
218 relationship, e.g. if AO SST anomalies are able to produce an « indirect » effect on the ISMR.

219 Table 1 summarizes the specifications of the simulations used here and the different
220 nudging domains are displayed in Fig. 1a. Finally, in all the analyses described below, the
221 first 10 years of all simulations have been excluded due to the spin-up of the coupled model.

222

223 3. ISMR, ENSO and AO relationships in observations and SINTEX-F2

224 3.a AO SST variability and its relationship with ENSO in SINTEX-F2

225 Before assessing the role of AO SST variability on ISMR and its relationship with ENSO,
226 it is important to document the performance of SINTEX-F2 in simulating AO variability,
227 especially the AZM and AMM (see Introduction for details). In this way, we can first
228 appreciate if the SST, rainfall and Z20 mean-state biases discussed in Section 2 are also a
229 cause of concern for a realistic simulation of AO SST variability in CTRL. Such analysis will
230 also be useful to interpret the differences between the AO decoupled experiments and CTRL
231 in the following sections as the nudged AO region encompasses the tropical AO.

232 AZM and AMM are the two dominant modes of SST monthly anomalies in the tropical
233 AO (Lübbecke et al. 2018; Cabos et al. 2019 and references herein). Thus, Empirical
234 Orthogonal Function (EOF) analysis of observed and simulated tropical AO SST monthly
235 anomalies provides a convenient tool for describing the performance of CTRL in simulating
236 both the AZM and AMM modes and their relative importance. Note also that the different
237 datasets have been detrended before the EOF analysis. Fig. 2 displays the first two leading
238 EOFs of observed and simulated monthly SST anomalies in the tropical AO. These two EOFs
239 of ERAi and CTRL SSTs are clearly distinct from the lower EOFs as EOF3 accounts for only
240 8% of the AO SST variance in both ERAi and CTRL (not shown).

241 The first EOF of ERAi SSTs describes 25% of the AO SST variance and depicts a basin-
242 wide pattern with positive SST anomalies covering the whole tropical AO (Fig. 2a). However,
243 the spatial loadings in this EOF1 are particularly high in the coastal upwelling regions near
244 the Angola–Benguela coast and in the equatorial cold tongue region explaining why this first
245 EOF is usually associated with the AZM in the literature (Lübbecke et al. 2018; Cabos et al.
246 2019; Jiang and Li 2021). The second EOF of ERAi SSTs accounts for 23% of the AO SST
247 variance (Fig. 2b). This EOF2 depicts a cross-equatorial SST gradient in the AO, which is
248 usually interpreted as the manifestation of the AMM (Chiang and Vimont 2004; Cabos et al.
249 2019). These two leading EOFs are very similar to other published EOF analysis of AO SSTs
250 using different datasets, spatial domains or time periods, both in terms of spatial patterns and
251 variance ~~described explained~~ by these leading modes (Lübbecke et al. 2018; Cabos et al.

252 2019; Jiang and Li 2021). Of special interest is the possible connection of these two leading
253 EOFs with ENSO, which is simply assessed here by computing the simultaneous correlation
254 between the associated amplitude monthly time series and the Nino34 index. These observed
255 EOF modes have no simultaneous relationship with ENSO ($r=-0.07$ and 0.08 for Nino34 vs
256 EOF1 and EOF2, respectively). Their lead and lag relationships with Nino34 index will be
257 explored in Section 3.c.

258 ~~The results (-0.07 for EOF1 and 0.08 for EOF2) suggest that these EOFs are not linearly~~
259 ~~linked to ENSO (at least when all months and no lags are taken into account) as both~~
260 ~~correlations are statistically insignificant even at the 80% confidence level according to a~~
261 ~~phase-scrambling bootstrap test (Ebisuzaki 1997).~~

262 Figs. 2cd display the two leading EOFs of CTRL SST monthly anomalies in the same AO
263 domain, which explain, respectively, 28 and 16% of the SST variance. SINTEX-F2 is able to
264 simulate with a reasonable accuracy both the spatial patterns and variances described by the
265 leading EOFs of observed AO SSTs as well as their relative importance in term of explained
266 variance. Note, however, that the L-shaped structure of the anomalous SSTs linking the
267 equatorial cold tongue to the southeast AO in EOF1 of observed SSTs is not well represented
268 and shifted westward in the EOF1 of simulated SSTs. This suggests that AZM events may be
269 weaker and are partly disconnected from the upwelling region near the Angola–Benguela
270 coast in CTRL compared to observations. This error is further confirmed by the comparison
271 of observed and simulated SST monthly means and standard-deviations in the ATL3 region
272 (Fig. 3). Consistent with Fig. 1a, ATL3 region is affected by a warm mean-state bias, which is
273 particularly prominent during June–July when the observed ATL3 SST variability is
274 maximum (Fig. 3), which corresponds to the peak of AZM events in observations (Lübbecke
275 et al. 2018). By contrast, the simulated ATL3 SST variability is prominent around three
276 months earlier and is drastically reduced in amplitude. This shortcoming, which is also found
277 in many other CGCMs (Voldoire et al. 2019; Bi et al. 2022), is may be related to the coupled
278 mean-state biases (e.g., SST, Z20, rainfall, etc.) reducing the intensity of the equatorial cold
279 tongue during boreal summer in CTRL, especially the flatten thermocline in the equatorial
280 AO, which may reduce the thermocline feedback and, thus, weakens the local Bjerknes
281 feedback and the AZM variability (see Figs. 1 and S1). The observed ATL3 SST variability
282 has also a secondary peak in winter, but this weaker maximum is well simulated in CTRL
283 (Fig. 3b). On the other hand, the second EOF of simulated AO SSTs closely matches the
284 second EOF estimated from ERAi SSTs in terms of spatial pattern and can also be regarded
285 as the manifestation of the AMM (Figs. 2bd). Finally, the correlations between the associated

286 two amplitude time series and the Nino34 index in CTRL are, respectively, 0.36 and 0.01, and
287 the first correlation is highly significant, even at the 99.9 % confidence level according to a
288 phase-scrambling bootstrap test (Ebisuzaki 1997). This suggests a significant association of
289 the simulated AZM with ENSO in CTRL, which is not found in observations (see above). On
290 the other hand, EOF2 of (simulated) AO SSTs, which can also be regarded as the
291 manifestation of the AMM, is not significantly associated with ENSO in CTRL; a result
292 consistent with ~~the one found from~~ observations.

293 In summary, the two leading modes of the tropical AO SST variability in CTRL share
294 many features with those in observations, but the simulated AZM has a much weaker
295 amplitude and a significant relationship with ENSO, suggesting a too strong ENSO
296 teleconnection to the tropical AO or vice versa.

297

298 3.b ISMR regression analysis

299 We now present the results of a lead-lag regression analysis of tropical SST, rainfall, 850-
300 hPa wind and 200-hPa velocity potential quarterly time series onto the ISMR index in order to
301 provide a clear picture of the relationships between ISMR, ENSO and AO climate variability
302 in observations and CTRL (Figures 4 and 5). The ISMR index is fixed at the JJAS season and
303 the 4-month averaged SST, rainfall, 850-hPa wind and 200-hPa velocity potential time series
304 are shifted backward and forward in time. The results are presented in a two-year window
305 from the beginning of year -1 (preceding the ISMR year) to the end of year 0, year 0 referring
306 to the year of the ISM season. Note that the results remain unchanged if the different time
307 series are detrended before the regression analysis (not shown). While it is known that the
308 ISM-ENSO relationship includes some asymmetry in observations (Terray et al. 2003, 2005;
309 Boschat et al. 2012; Chakraborty and Singhai 2021), this regression analysis is used here as a
310 first order method to assess the realism of CTRL in simulating the ISMR teleconnections.

311 The regression results from observations (Fig. 4) illustrate that ISMR is associated with
312 different phases of ENSO in a two-year window (Boschat et al. 2012; Chakraborty 2018).
313 Strong positive SST anomalies in the central and eastern PO, which are out of phase with
314 anomalies in the western part of the PO are found during year -1, consistent with the
315 occurrence of an El Niño one year before a strong ISM (Fig. 4a). The atmospheric anomalous
316 patterns are consistent with this hypothesis as they describe an eastward shift of the Pacific
317 Walker circulation with persistent westerly 850-hPa wind anomalies over the western
318 equatorial Pacific, positive rainfall anomalies and negative 200-hPa velocity potential

319 anomalies over the central Pacific during year -1 (Figs. 4bc). The regression patterns during
320 year 0 are more or less a mirror image of those during year -1 with an opposite sign (Fig. 4).
321 In other words, this analysis demonstrates that ENSO and ISMR are still highly inter-related
322 and followed a sustained biennial rhythm during recent decades (Meehl et al. 2003; Terray et
323 al. 2003, 2021).

324 Focusing now on the relationships between ISMR and IO SSTs, we note that IO SST
325 anomalies are small and insignificant during boreal winter of year -1 and the pre-monsoon
326 period of year 0. During boreal summer of year 0, the IO is also devoid of any significant SST
327 anomalies associated with ISMR variability (Fig. 4a). This suggests that the « direct » effect
328 of IO SSTs on ISMR is small (Cretat et al. 2017; Terray et al. 2021).

329 On the other hand, a significant positive correlation emerges between TNA SSTs during
330 boreal spring of year 0 and ISMR (Fig. 4a). This result is in agreement with the results of
331 Vittal et al. (2020), Yang and Huang (2021) and Ham et al. (2013ab), which suggest that
332 TNA SSTs during boreal spring are a significant precursor of ISMR and can also serve as a
333 trigger for the following La Niña event, respectively. However, this statistical relationship
334 quickly fades away during boreal summer of year 0 (Fig. 4a). Furthermore, during ISM, the
335 SST anomalies are insignificant in the tropical AO, which partly disagree with recent studies
336 highlighting the role of AZM on the ISMR (Kucharski et al. 2007, 2009; Pottapinjara et al.
337 2014), but are in agreement with the work of Ding et al. (2012).

338 In order to further elucidate the relationships between tropical SSTs, ENSO and ISMR in
339 observations, the above lead-lag regression analysis has been repeated after removing the
340 ENSO contribution to the monthly SST fields by a simple linear regression method (Fig.
341 S2a). This is equivalent to assume that ENSO has a simultaneous linear impact on tropical
342 SSTs (Kucharski et al. 2009). Despite the simplicity of this approach, the results demonstrate
343 that most of the lead-lag relationships between tropical SST anomalies and ISMR (displayed
344 in Fig. 4a) can be understood as the results of simultaneous ENSO teleconnections on both
345 ISMR and SST anomalies elsewhere, as most of the correlations are now insignificant,
346 especially those during boreal summer of year 0 (Fig. S2a). One notable exception is,
347 however, again the TNA SSTs during boreal spring of year 0, which remain significant at the
348 90% confidence level even after removing the ENSO contribution from the observed SST
349 fields. On the other hand, despite the independence of AZM and ENSO in observations (see
350 Section 3.a) and removal of the ENSO-related contribution to the SST anomalies, the tropical
351 AO SSTs are again not linearly linked to ISMR during year 0.

352 CTRL is performing relatively well in reproducing the simultaneous inverse relationship
353 between ISMR and ENSO during year 0 (Fig. 5). However, the model fails to reproduce the
354 significant positive lead correlation between TNA SSTs during boreal spring and the
355 following ISM. Also, ISMR is much more linked to ENSO during year -1 and the pre-
356 monsoon season of year 0 in CTRL (Fig. 5; see also Terray et al. 2021). This suggests an
357 overly strong control of ENSO on ISMR in CTRL as for the AZM (see Section 3.a above).
358 Removing the ENSO contribution to the CTRL monthly SST fields by a simple linear
359 regression method before the regression analysis reveals a significant contribution of the
360 subtropical North Pacific SST anomalies during year 0, but again no significant relationships
361 with TNA or tropical AO SSTs during year 0 (Fig. S2b).

362 In the context of the ENSO-ISM relationship, the physical consistency of a direct link
363 between TNA SSTs during boreal spring and ISMR independent of ENSO, as found in
364 observations, is thus questionable. This positive correlation between TNA SSTs during boreal
365 spring and ISMR may be more an artifact of the ENSO forcing on both TNA and ISMR than
366 a sign of a physical connection between TNA SSTs and ISMR (Zhang et al. 2021b). This
367 alternative scenario is also consistent with the significant La Niña SST and atmospheric
368 patterns during year 0, associated with the ENSO biennial rhythm during recent decades (Fig.
369 4a; Wang et al. 2017), and the insignificant atmospheric anomalous patterns over TNA during
370 boreal spring of year 0 in our ISMR regression analysis from observations (Fig. 4bc).

371 3.c Nino34 regression analysis

372 We now revisit the two-way interactions between AO SSTs and ENSO for a better
373 understanding of the AO SSTs and ISMR relationships. Figures S3 and S4 show the lead-lag
374 regressions between the Nino34 index during boreal winter (e.g., ONDJ season) and tropical
375 SST, rainfall and 850-hPa wind anomalies, again during a two-year window, in both
376 observations and CTRL. In the following discussion, we will refer to year 0 as the developing
377 year and year +1 as the decaying year of ENSO events, respectively. CTRL reproduces
378 reasonably well the observed lifecycle of ENSO events with El Niño onset during boreal
379 spring, a developing phase during boreal summer and fall, a peak phase during boreal winter
380 and a decaying phase during boreal spring and summer of year +1 (Fig. S3).

381 The SST and atmospheric anomalous patterns during ENSO onset (e.g., during boreal
382 spring of year 0) are very similar in observations and CTRL and are reminiscent of the
383 Seasonal Footprinting Mechanism (SFM; Vimont et al. 2003; Boschat et al. 2013). We also
384 note significant cold SST anomalies in the tropical AO and TNA during boreal spring of year

385 0, (—especially in observations ~~see~~ (Fig. S3a); consistent with past studies (Rodríguez-Fonseca
386 et al. 2009; Ham et al. 2013ab; Jiang and Li 2021). Ham et al. (2013a) argues that these TNA
387 SST anomalies may influence the tropical PO and that this forcing is mediated by the SFM
388 (Fig. S4a). Such TNA forcing on El Niño onset is also suggested in CTRL, but with much
389 weaker amplitude (Figs. S3b and S4b).

390 A first notable difference between CTRL and observations is that simulated ENSO
391 teleconnections are significantly stronger than observed, as illustrated by the enhanced IO and
392 AO basin-wide warming in CTRL at the end of year 0 (Fig. S3b). The rainfall response over
393 the central equatorial Pacific per °C of Niño3.4 warming during boreal winter of year 0 is
394 much stronger in CTRL; about 5 mm/day per °C against only 2.6 mm/day per °C in
395 observations (Fig. S4). The associated latent heating induces an enhanced tropospheric
396 warming, which stabilizes the atmosphere and reduces convection, cloudiness and
397 evaporation over the AO and IO more in CTRL than in observations (not shown; Chiang and
398 Sobel 2002; Chang et al. 2006).

399 The climate anomalies during year +1 depict the transition from El Niño to La Niña (or
400 vice versa since the analysis is linear) and the related changes in ENSO teleconnections (Figs.
401 S3 and S4). First, the simulated and observed SST anomalies during boreal winter of year 0
402 and spring of year +1 show that most warm TNA events can be initiated by ENSO itself (see
403 Fig. S3; Garcia-Serrano et al. 2017; Jiang and Li 2019). This TNA warming is mainly driven
404 by the weakening of the local northeasterly trade winds during FMAM of year +1 in both
405 observations and CTRL (Fig. S4). These warm TNA events may induce wind anomalies over
406 the tropical Pacific that oppose the ongoing ENSO event and accelerate its demise (Ham et al.
407 2013ab). This TNA forcing can be interpreted as a delayed negative feedback that accelerates
408 the decay of ENSO as the IO capacitor effect (Xie et al. 2009, 2016; Wang et al. 2017). This
409 is consistent with many previous CGCM studies, which found that decoupling the IO or AO
410 in CGCMs increases the length of ENSO events (Ohba and Ueda 2007; Santoso et al. 2012;
411 Terray et al. 2016, 2021; Kajtar et al. 2017).

412 ~~In this respect,~~ a second notable difference between CTRL and observations is that,
413 despite this stronger ENSO-induced capacitor effect over the tropical IO or AO in CTRL, the
414 simulated ENSO signal in the tropical Pacific lasts longer and persists well up to boreal
415 summer and fall of year +1 in CTRL (Fig. S4b). This contradiction has already been
416 examined in Terray et al. (2021) and their results highlight that, besides the amplitude of the
417 basin-wide IO warming, the intensity of the negative IO feedback on ENSO is heavily
418 dependent on the realism of the equatorial SST gradient in the IO during the ENSO peak and

419 decaying phases. An open question, which will be examined with the help of partial
420 decoupling experiments in the following section, is to determine if the erroneous mean
421 equatorial SST gradient in the AO (Fig. 1a) is also playing a role in the longer ENSO period
422 in CTRL.

423 In summary, this section illustrates how it is difficult to isolate the specific effect of AO
424 SST anomalies on ISMR in observations or coupled simulations as the three players, ISM,
425 ENSO and the AO SST modes, interact with each other in a complex manner.

426

427 **4. AO direct and indirect effects on ISMR in SINTEX-F2**

428 In order to assess both the possible “direct” and “indirect” effects of AO SSTs on ISMR,
429 we now focus on the analysis of the partial decoupling experiments described in Section 2b.

430 4.a AO direct effect on ISMR

431 We estimate the "direct" effect of AO SST variability on ISMR with the help of the PO
432 decoupling experiments in which ENSO has been removed.

433 In order to document how the simulated AO SST variability is modified in the absence of
434 ENSO ~~in a first step~~, Fig. 6 displays the first two EOFs of tropical AO SST anomalies
435 simulated in the two PO decoupling experiments. We first note that these leading EOFs
436 estimated from FTFC and FTFC-obs are very similar both in terms of spatial pattern and
437 described variance. EOF1 (EOF2) is associated with SST variability to the South (North) of
438 the Equator and the Angola–Benguela (Senegal-Mauritania) upwelling system. This suggests
439 that, in both PO experiments, these EOFs are excited by wind-driven evaporation or dynamic
440 in each hemisphere, but that they do not involve an active WES feedback, which is usually
441 associated with the AMM (Chiang and Vimont 2004; Cabos et al. 2019).

442 These EOFs can be compared to those estimated from CTRL, which are shown in Figs.
443 2cd. Consistent with the overly strong ENSO forcing on the simulated AZM found in CTRL
444 (e.g., EOF1 of CTRL SSTs in Fig. 2c), the leading EOFs in the PO experiments are no more
445 associated with SST variability in the equatorial AO and they further suggest that the two
446 poles of the AMM may vary independently of each other. This questions the physical
447 consistency of the AMM and its underlying mechanism, both in observations and CTRL, as
448 its inter-hemispheric SST dipole structure depicted by the EOF2 of (observed and CTRL) AO
449 SST anomalies may result from orthogonality constraints of the EOF analysis rather than the
450 consequence of an active WES feedback.

451 The specific role of this “ENSO-free” AO SST variability in modulating ISMR is
452 illustrated by a regression analysis of simulated SST and 200-hPa velocity potential
453 anomalies onto the ISMR index during year 0 (Figure 7). We ~~mainly~~ focus on FTPC-obs in
454 the rest of this section, as the results from FTPC are similar (not shown). Overall, the direct
455 forcing of AO SST and atmospheric variability on ISMR is insignificant as the AO is devoid
456 of any significant SST or atmospheric anomalies during the pre-monsoon season (e.g.,
457 FMAM of year 0) in FTPC-obs (Fig. 7). On the other hand, a well-defined atmospheric
458 teleconnection pattern exists during boreal summer with strong 200-hPa divergence and
459 outflow anomalies from the ISM region towards the south AO in FTPC-obs (Fig. 7b). This
460 upper-level velocity potential pattern is obviously triggered by the diabatic heating associated
461 with ISMR and demonstrates that ISM is a dominant feature of the boreal summer tropical
462 circulation in the absence of ENSO. Interestingly, this ISMR forcing on the AO during boreal
463 summer also promotes an AMM-like SST pattern during boreal fall with negative (positive)
464 SST anomalies in the South (North) AO (Fig. 7a). This leads to the hypothesis that it is
465 ISMR, which is forcing the AO rather than the reverse in the absence of ENSO.

466 4.b AO forcing on ENSO

467 The aim of this subsection is to isolate the effect of the AO SST mean-state bias and
468 variability on ENSO with the help of the FTAC and FTAC-obs simulations.

469 To have a first basic understanding of the ENSO behavior in these two runs, Figure 8
470 shows the Nino34 SST monthly standard deviations from observations, CTRL and the two
471 AO decoupled experiments. The observed Nino34 SST variability is most prominent during
472 boreal winter (std of about 1.2°C), and it is drastically reduced in April (about 0.6°C), which
473 marks the onset of many ENSO events (Fig. 8). CTRL underestimates the Nino34 SST
474 variability during boreal winter when ENSO events usually peak (Figs. 8 and S1). It is also
475 not able to replicate the sharp decrease of Nino34 SST variability after the ENSO peak and,
476 hence, there is an overestimation of the simulated Nino34 SST variability for a few months
477 from March till July (Fig. 8). One interesting result is how the AO decoupled experiments
478 alter this Nino34 SST variability (Frauen and Dommenges 2012; Ding et al. 2012; Kajtar et al.
479 2017). The Nino34 SST variability changes in FTAC and FTAC-obs include (i) a consistent
480 increase during boreal summer for the two runs and (ii) an enhanced variability during boreal
481 winter and a more realistic seasonal phase-locking in FTAC-obs compared to both CTRL and
482 FTAC (Fig. 8).

483 We first focus on the increase of Nino34 SST variability during boreal summer, which is
484 found in both FTAC and FTAC-obs (Fig. 8). In order to explain this feature, Figure 9 shows
485 the regressions between the Nino34 index during boreal winter of year 0 (e.g., ONDJ season)
486 and tropical SST, rainfall and 850-hPa wind anomalies during the preceding boreal winter and
487 spring (e.g., FMAM) in FTAC and FTAC-obs. As discussed in Section 3, boreal spring is the
488 season of El Niño onset and the results suggest that the SFM in the North Pacific plays a key-
489 role in ENSO onset in both observations and CTRL (Figs. S3 and S4). A robust association
490 between the SFM and El Niño onset is also found in the two AO decoupled experiments
491 (Figs. 9ab). But, this El Niño onset occurs during boreal winter of year -1, e.g., one season
492 before the El Niño onset in CTRL and observations, and one year before the El Niño peak
493 (Fig. 8), a feature, which has not been well documented in past studies (Frauen and
494 Dommenges 2012; Terray et al. 2016; Kajtar et al. 2017). Furthermore, warm SST anomalies
495 already cover the whole central and eastern equatorial PO during boreal spring of year 0 in
496 these two runs (Figs. 9cd), e.g. also one season in advance compared to observations or CTRL
497 in which this basin-wide SST anomalous pattern is seen in boreal summer (Fig. S3). The
498 associated rainfall and 850-hPa wind regression patterns during boreal spring of year 0 in
499 FTAC and FTAC-obs also describe an eastward shift of the convection center with positive
500 (negative) rainfall anomalies over the central (western) PO and westerly zonal wind
501 anomalies on the western side of the positive rainfall anomalies (Figs. 9ab). All these features
502 are fully consistent with the early El Niño onset in FTAC-obs and FTAC. Note, furthermore,
503 that the rainfall and 850-hPa regression patterns during boreal winter and spring preceding the
504 ENSO peak are very similar in the two AO decoupled experiments (Figs. 9ab). This suggests
505 that this early El Niño onset can be attributed to the common cancellation of the AO SST
506 variability in the two runs. This early El Niño onset also implies that the warm Nino34 SST
507 anomalies associated with El Niño are already well defined during boreal summer of year 0
508 consistent with the enhanced Nino34 SST variability during boreal summer found in the two
509 nudged experiments (Fig. 8). These results illustrate the important role played by AO SST
510 variability in generating spread in ENSO timing and amplitude through its influence on the
511 SFM. This is also consistent with several recent studies, which suggest that the SFM and its
512 modulation are an important source of spread in ENSO forecasts during boreal spring and
513 early summer (Ma et al. 2017; Ogata et al. 2019).

514 We now focus on the improved phase-locking and enhanced ENSO variability during
515 boreal winter found in FTAC-obs (Fig. 8). In order to illustrate the seasonal dependence in the
516 changes of ENSO variability in the different runs and isolate the role of the AO background

517 state, Figure S5 displays the seasonal differences of SST standard-deviation between FTAC-
518 obs and both CTRL and FTAC. Figs. S5a-c first confirm that almost all the tropical AO SST
519 variability has been removed in FTAC-obs. Outside the AO nudged region, the SST
520 variability changes in FTAC-obs relative to CTRL are mainly found in the tropical PO in the
521 form of an enhanced ENSO variability from boreal summer to winter (Figs. S5b-c). This is
522 consistent with the changes of Nino34 SST variability (Fig. 8). Furthermore, a large part of
523 these differences can be attributed to the corrected SST AO mean state as this seasonal pattern
524 of changes is also found in the differences of SST standard-deviation between FTAC-obs and
525 FTAC (Figs. S5d-f). In other words, decoupling the AO SST variability without restoring the
526 observed AO SST climatology leads only to a modest increase of ENSO variability,
527 especially during its peak phase (Fig. 8).

528 In order to understand why the AO background state has such an impact on ENSO, Figure
529 S6 displays the seasonal differences between FTAC-obs and CTRL climatologies of SST,
530 rainfall and Z20, and Figure 10 shows boreal spring differences of 850- and 200-hPa zonal
531 wind, velocity potential and stream function climatology between the same runs. The
532 differences between FTAC-obs and FTAC are similar as FTAC and CTRL have the same
533 mean state (not shown).

534 The correction of AO SST biases in FTAC-obs leads to drastic improvements of AO
535 rainfall and Z20 spatial distributions during all the seasons with enhanced precipitation in the
536 northwest tropical AO, reduced precipitation in the southeast tropical AO and, finally, a
537 deeper Z20 in the western AO (Figs. S6bc). These rainfall changes are consistent with the
538 imposed AO SSTs in FTAC-obs altering the regions that are above or below the threshold for
539 deep convection in FTAC-obs compared to CTRL or FTAC (Fig. S6a). An increase of
540 precipitation is also evident over the Amazon basin demonstrating that the response is not
541 purely local. These AO rainfall shifts may exert an influence on atmospheric teleconnections
542 because they alter diabatic heating (e.g., Gill 1980). However, surprisingly, while these
543 rainfall changes are significant during all seasons, the rectification of the tropical PO mean
544 state is mainly prominent during boreal spring and is characterized by a shift to an El Niño-
545 like mean state with a warmer (cooler) eastern (western) equatorial PO, a rainfall increase in
546 the central PO near the date line, a flatter thermocline (Figs. S6abc, first row) and, finally, a
547 slowdown of the mean Walker circulation across the tropical PO (Figs. 10ab). The westward
548 shift of rainfall over the AO may produce large heating anomalies at upper level centered over
549 central South America and extending in the eastern and central PO. The 850- and 200-hPa
550 velocity potential differences during boreal spring are in agreement with this hypothesis as

551 they show enhanced convergence (divergence) at 850-hPa (200-hPa) toward a large region
552 encompassing the central equatorial PO and the northwest AO and the opposite patterns
553 elsewhere in the Tropics (Figs. 10cd). The well-defined quadrupole structure of the 850- and
554 200-hPa stream functions over the eastern PO/western AO (Figs. 10ef) is further consistent
555 with the circulation expected from a Matsuno-Gill response (Gill 1980; Kucharski et al. 2009;
556 Li et al. 2016). These features directly link SST mean-state changes in the AO with the
557 rectification of the Pacific mean state during boreal spring. Such strong inter-basin
558 connectivity is not seen in the other seasons despite that AO rainfall changes are still
559 prominent (Fig. S6b). This may be related to the ITCZ and Walker cells in all oceanic basins
560 being closer to the Equator during boreal spring, thus providing ideal conditions for the
561 corrected rainfall and SST patterns in the tropical AO to influence the equatorial PO (Chang
562 et al. 2006; Tokinaga et al. 2019).

563 Not surprisingly, the rainfall and 850-hPa zonal wind variability is also significantly
564 enhanced and shifted eastward in the PO during boreal spring in FTAC-obs (Figs. 11ac),
565 while the related changes are weak in FTAC (Figs. 11bd). This is consistent with the eastward
566 shift of the mean PO SST and rainfall patterns during the same season in FTAC-obs (Figs.
567 S6ab). Furthermore, the changes of rainfall and 850-hPa zonal wind variability over the AO
568 are opposite in FTAC and FTAC-obs, with a large increase (decrease) of rainfall and 850-hPa
569 zonal wind variability over the tropical AO during boreal spring in FTAC (FTAC-obs) despite
570 the absence of AO SST variability in the two runs (Fig. 11). This again highlights strong
571 interactions between the biased AO SST mean-state and atmospheric variability in FTAC,
572 which may further perturb the ENSO onset. In other words, the El-Niño like changes of the
573 PO mean state in FTAC-obs provide more favorable conditions for El Niño to develop
574 through the Bjerknes feedback (e.g., a reduced equatorial SST gradient and a flatter
575 thermocline across the PO) and reduced atmospheric noise over the AO during boreal spring.
576 This finally leads to a much better seasonal phase-locking of the simulated ENSO and an
577 improved ENSO amplitude during its peak phase in FTAC-obs in comparison of CTRL and
578 FTAC (Fig. 8).

579 Finally, the AO decoupled experiments demonstrate that the AO SST variability
580 significantly modulates ENSO during its decaying phase. This is illustrated by the regression
581 analysis of the ONDJ Nino34 index with quarterly SST time series during the following year
582 (e.g., year +1) in FTAC-obs (Fig. 12). FTAC displays a very similar evolution (not shown).
583 The corresponding regression analyses for observations and CTRL are shown in Fig. S3. The
584 warm SST PO pattern lasts longer in FTAC-obs (and FTAC) than in CTRL and the ENSO

585 signal is still robust at the end of year +1 in this run (Fig. 12c). Overall, the results highlight
586 that the ENSO-induced AO SST anomalies in CTRL (Fig. S3b), which are on the other hand
587 very small in the FTAC-obs by design (Fig. 12), accelerate the transition from El Niño to La
588 Niña (in CTRL) during year +1.

589 4.c AO indirect effect on ISMR

590 The AO “indirect” effect on ISMR, mediated by the ISM-ENSO relationship, is now
591 explored again with the help of the AO decoupled experiments.

592 Figure 13ab shows the ISMR seasonal cycle and monthly standard deviations from
593 observations, CTRL and the two AO decoupled experiments. The ISMR seasonal cycle is not
594 altered in the nudging experiments, and they replicate the same results of CTRL (Fig. 13a).
595 Observed ISMR variability is characterized by a double peak at the onset and withdraw of
596 ISM (Fig. 13b). CTRL underestimates ISMR variability in the pre-monsoon season and
597 overestimates it during June-August. The nudging experiments reduce the simulated ISMR
598 variability during the last half of ISM, especially FTAC-obs (Fig. 13b), despite ENSO is
599 stronger during boreal summer in FTAC and FTAC-obs (Fig. 8).

600 To assess the AO “indirect” effect, Figure 13c shows the observed and simulated lead-lag
601 correlations between ISMR and Nino34 SST quarterly time series in a three-year window
602 from the beginning of year -1 (preceding the ISMR year) to the end of year +1 (following the
603 ISMR year). Consistent with Fig. 4a, there are significant positive correlations one year
604 before ISM in observations. The sign of the correlation reverses during the pre-monsoon
605 season of year 0 and the correlation gets significantly negative during boreal summer and
606 winter of year 0. These negative correlations fade away during year +1. Thus, the most
607 favorable conditions for a strong ISM are during the transitions from an El Niño during year -
608 1 to a La Niña event in year 0 in agreement with our analysis in Section 3.

609 CTRL is able to reproduce realistically the significant negative correlation between the
610 ISMR and ENSO during boreal summer of year 0 and the decrease of amplitude of this
611 negative correlation during year +1. However, the model shows large discrepancies from
612 observations with a negative correlation during several months before ISM (Fig. 13c). This
613 bias is again consistent with the results of Section 3. However, the relative success of the
614 model in reproducing the observed simultaneous relationships between ISMR and ENSO is
615 important as the analysis of the nudged experiments can then provide more insights on the
616 precise role of the AO in this system.

617 In this respect, the lead-lag correlations between ENSO and ISMR are significantly
618 different between the two nudging experiments and CTRL, with a consistent weakening of
619 the simultaneous inverse relationship between ISM and ENSO (Fig. 13c), despite Nino34
620 SST variability is higher during boreal summer in FTAC and FTAC-obs compared to CTRL
621 (Fig. 8). In FTAC-obs, the simultaneous negative correlation between ISMR and Nino34 SST
622 is not even statistically significant at the 90% confidence level, despite FTAC-obs simulates a
623 stronger ENSO than FTAC. As the warm SST mean-state bias affecting the tropical AO in
624 both CTRL and FTAC is removed in FTAC-obs, these results are consistent with the AGCM
625 experiments performed in Kucharski et al. (2007, 2009) in which a weakening of the
626 monsoon-ENSO relationship is simulated in response to a cooling trend of the tropical AO
627 and allow us to isolate the specific role of the SST AO mean-state biases on the simulated
628 monsoon-ENSO relationship. However, as the simultaneous correlation between ISMR and
629 Nino34 SST is also reduced in FTAC, the reduced AO SST variability in the two runs may
630 also play an important role in the weakening of the inverse relationship between ISM and
631 ENSO in the nudged experiments.

632 Furthermore, ISMR evolution has different flavors during the ENSO decaying year in each
633 dataset (Figure S7). First, we find more rainfall over the IO during the pre-monsoon season
634 (not shown) and over India and the Arabian Sea during the monsoon of year +1 in
635 observations (Fig. S7a). This is consistent with the warm local SSTs (Fig. S3a) and the fast
636 demise of ENSO induced upper-level subsidence during year +1 in observations (Figs. S7a
637 and 14a). During boreal summer of the ENSO decaying year, the significant 200-hPa velocity
638 potential anomalies are restricted to a regional dipole opposing upper-level divergence over
639 the Arabian Sea to upper-level convergence over the western PO in observations (Fig. 14a).
640 These features are physically consistent with the positive (negative) correlation between
641 precipitation in the Indian (northwest PO) region during ISM of year +1 and the Nino34 index
642 in observations (Fig. S7a). This confirms that with the increase (decrease) of the SST over the
643 Nino34 region during the preceding boreal winter, the ISM of the next year is enhanced
644 (suppressed) or vice versa in observations (Fig 13c; Yang et al. 2007).

645 CTRL is not able to replicate these precipitation and 200-hPa velocity potential anomalies
646 during year +1 (Figs. S7b and 14b). In CTRL, there is a significant negative correlation
647 between precipitation over the Indian region during the monsoon season of year +1 and
648 Nino34 SST (Fig. S7b). This negative ISMR anomaly is consistent with both the persistent
649 ENSO signal (Fig. S3b) and the (significant) positive 200-hPa velocity potential anomalies
650 over India during boreal summer of year +1 in CTRL (Figs. S7b and 14b).

651 The effect of removing the AO SST variability on these correlations can now be analyzed
652 (Figs. 14c and S7c), keeping in mind that the ENSO-related SST signal persists even longer
653 and is stronger in FTAC-obs than in CTRL (Fig. 12). Despite this enhanced ENSO forcing,
654 the correlation of Nino34 SST with ISMR during year +1 has vanished and is not significant
655 in FTAC-obs (Fig. S7c). This is consistent with the weakening of the simultaneous ISMR-
656 Nino34 correlation in FTAC-obs (Fig. 13c). The origin of this paradox can be seen from Fig.
657 S7c, which shows that there is a stronger negative correlation between precipitations over the
658 tropical AO during year +1 with Nino34 SST (during the preceding boreal winter) in FTAC-
659 obs compared to CTRL. Moreover, these negative correlations are also shifted westward in
660 FTAC-obs consistent with the corrected AO mean state in this simulation. This stronger
661 relationship between ENSO and the AO in FTAC-obs during year +1 is further confirmed by
662 the associated 200-hPa velocity potential signal (Fig. 14c). In the 200-hPa velocity potential
663 anomalous pattern during boreal summer of year +1 in FTAC-obs, there is a significant
664 positive correlation over the AO, while the correlation over the Indian region is near zero.
665 This implies that the upper-level divergent flow is mainly from the central PO to the TNA
666 instead toward the Indian region, which results in suppression of rainfall in the AO and a near
667 normal ISM in FTAC-obs. FTAC shows a similar evolution (not shown), but the ENSO-
668 induced subsidence over the AO during boreal summer of the ENSO decaying year is weaker,
669 presumably due to the biased AO SST mean-state in FTAC.

670 In a nutshell, these last results illustrate that, by artificially removing the SST variability
671 over the tropical AO in the nudged experiments, the ENSO signal is stronger and persists
672 longer during the ENSO decaying year, but the associated upper-level divergent winds will
673 flow mainly from the central PO to the tropical AO, resulting in rainfall suppression in the
674 AO, but only in a weak ENSO forcing on ISM during the ENSO decaying year.

675

676 **5. Conclusions and discussion**

677 In this study, we use dedicated coupled experiments to isolate both the “direct” and
678 “indirect” effects of AO SSTs on ISMR. The “direct” effect refers to the AO forcing on ISMR
679 in the absence of others dominant forcings like ENSO (Kucharski et al. 2009). The “indirect”
680 effect refers to the AO forcing on ENSO (Rodriguez-Fonseca et al. 2009; Ding et al. 2012;
681 Ham et al. 2013ab; Jiang and Li 2021), which may subsequently affect ISMR through ENSO
682 teleconnections. Furthermore, with the help of these experiments, we also identify the role of
683 AO SST mean-state biases on the simulated ISMR-ENSO relationship.

684 First, we found that the “direct” effect of AO SST variability on ISMR is insignificant.
685 Overall, the results highlight that ISMR is a major player in the tropical atmospheric
686 circulation in the absence of ENSO, even forcing an AMM SST pattern during boreal fall
687 after ISM, rather than the reverse. Several studies have suggested that the AZM may provide
688 a direct remote forcing on ISMR through a Gill-Matsuno mechanism with a Kelvin wave
689 transporting the signal to the IO (Kucharski et al. 2009; Pottapinjara et al. 2014), but we were
690 not able to isolate this “direct” effect on ISMR in our simulations or observations during
691 recent decades, even when ENSO is removed (Figs. S2 and 7), which is consistent with the
692 modeling results of Ding et al. (2012). However, it is known that current CGCMs struggle to
693 represent this relationship, partly due to their common strong AO SST mean-state biases
694 (Barimalala et al. 2012; [Voldoire et al. 2019](#)). Taking into account the warm SST mean-state
695 bias in the southeastern AO found in SINTEX-F2 (see Fig. 1a), the realism and amplitude of
696 the AZM simulated by this CGCM [areis](#) severely biased (Fig. 3b) and this may also
697 deteriorate the simulated ISMR-AZM relationship in our model.

698 The global-scale effects of the corrections of tropical AO SST biases are readily apparent
699 when comparing our two AO decoupled experiments with each other and with a long free
700 simulation performed with the same model. When the AO SST mean-state bias (present in
701 CTRL and FTAC) is corrected in FTAC-obs, the rainfall pattern in the AO is shifted
702 northwestward (see Fig. 1f). The associated changes in diabatic heating produce a Matsuno-
703 Gill atmospheric response centered over the eastern equatorial Pacific and generate a trans-
704 basin (e.g., Pacific/Atlantic) atmospheric see-saw with upward motion over a large region
705 encompassing the central and eastern equatorial Pacific and the western tropical AO and
706 descending motion elsewhere in the Tropics. The overall effect is the emergence of an El
707 Niño-like mean-state pattern in FTAC-obs when these AO SST biases are corrected, which is
708 mostly significant during boreal spring (see Fig. S6). This also corresponds to the onset
709 period of El Niño events in observations and the simulations analyzed here.

710 The comparison of FTAC and FTAC-obs with CTRL suggests that the “indirect” influence
711 of the tropical AO SST variability on ISMR is significant. The main effect of AO SST
712 variability is to modulate the amplitude and length of ENSO events, especially during their
713 onset and decaying phases. First, AO SST variability plays a key-role in ENSO developing
714 phase in agreement with the results of Ham et al. (2013ab). Without AO SST variability,
715 ENSO onset, while still seasonally phase-locked and linked to the SFM over the North Pacific
716 (Vimont et al. 2003; Boschhat et al. 2013; Terray et al. 2016), occurs one season before during
717 boreal winter. This finding supports the idea that AO SST variability is also a possible source

718 of ENSO spread, especially during its developing year. This may reduce ENSO predictability
719 and contributes to the spring ENSO predictability barrier (Ma et al. 2017; Ogata et al. 2019).
720 Furthermore, when AO SST mean-state biases are also corrected (e.g., in FTAC-obs), this
721 leads to ~~an~~ increased ENSO amplitude during its peak phase (e.g. boreal winter) as well,
722 demonstrating nonlinear interactions between the mean state during boreal spring and ENSO
723 amplitude.

724 Our results also confirm that AO SST variability modulates the length of the ENSO
725 decaying phase, as ENSO is still active up to the end of the ENSO decaying year in the FTAC
726 and FTAC-obs experiments (Terray et al. 2016; Kajtar et al. 2017). During boreal winter,
727 ENSO influences the subtropical and tropical AO through the atmospheric bridge (Jiang and
728 Li 2019). These ENSO-induced SST AO anomalies may then feedback negatively on ENSO
729 and fasten the transition from El Niño to La Niña. This discharging capacitor effect of the AO
730 (Wang et al. 2017) serves as a phase-reversal mechanism for the ENSO cycle as for the IO
731 (Xie et al. 2009; Terray et al. 2016, 2021).

732 All these findings are consistent with recent modelling studies that have identified a tight
733 physical linkage between AO and PO variability on decadal timescales (Kucharski et al.
734 2011) or in a global warming context (McGregor et al 2014) and also demonstrate that the
735 prominent AO SST biases play a significant role in modulating the simulated Pacific Walker
736 circulation at both the seasonal and longer time scales in current CGCMs (Kajtar et al 2018;
737 McGregor et al. 2018; Li et al. 2020).

738 Finally, the absence of AO SST variability weakens the simultaneous inverse relationship
739 between ISM and ENSO despite ENSO is stronger during boreal summer and persists longer.
740 This result is opposite to the one found in similar IO decoupling experiments in which ENSO
741 is also stronger and more persistent (Terray et al. 2021). The origin of this paradox lies
742 mainly in the modulation of the Walker circulation when SST variability is removed in one of
743 the two oceanic basins, especially during boreal summer of the ENSO decaying year (see Fig.
744 14). The upper-level divergent wind flows mainly from the PO to the AO, resulting in rainfall
745 suppression in the AO, but in a weaker forcing on ISMR in the AO decoupled experiments
746 (Fig. 14c). On the other hand, both the stronger ENSO amplitude and the enhanced upper-
747 level convergence towards the Indian domain act in concert and result in a much stronger
748 inverse ENSO-ISM relationship in similar IO decoupled experiments (Terray et al. 2021).

749 In conclusion, while the AO “direct” effect on ISMR is insignificant in our coupled model,
750 we highlight that the AO “direct” effect on ENSO is significant in many aspects including the

751 ENSO triggering mechanism and AO capacitor effect, which ~~also~~ have also a significant
752 “indirect” impact on ISMR mediated by the ENSO teleconnections. We hope that these
753 encouraging results will promote the interest of performing similar IO, PO and AO decoupled
754 experiments with other coupled models in order to verify if the insignificant “direct” effect of
755 AO SSTs on ISMR found here is model dependent or not.

756

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985 **Figure captions**

986 **Figure 1:** **a)** SST annual means difference (unit: °C) between CTRL and ERAi; **b)** Z20
987 annual means difference (unit: m) between CTRL and SODA; **c)** Rainfall annual means
988 difference (unit: mm/day) between CTRL and GPCP; **d)** SST annual means difference (unit:
989 °C) between FTAC-obs and CTRL; **e)** Z20 annual means difference (unit: m) between FTAC-
990 obs and CTRL; **f)** Rainfall annual means difference (unit: mm/day) FTAC-obs and CTRL.
991 The nudging domain for FTAC and FTAC-obs (FTAC and FTAC-obs) is shown in blue
992 (purple) in panel a. See Section 2 and Table 1 for more details.

993 **Figure 2:** Empirical Orthogonal Function (EOF) 1st and 2nd modes for detrended SST
994 anomalies with monthly means removed ~~12-monthly anomaly of detrended SST~~ obtained
995 from ERAi and CTRL. **a)** and **b)** EOF1 and EOF2 from ERAi SST detrended anomalies
996 (1979-2015 period), respectively. **c)** and **d)** EOF1 and EOF2 from CTRL SST detrended
997 anomalies (11-210 period), respectively. The number in parentheses for each panel gives the
998 % of SST variance described by the EOF mode.

999 **Figure 3:** **a)** Monthly means of ATL3 SST (unit: °C) index from ERAi (blue) and CTRL
1000 (orange); **b)** Monthly standard deviations of ATL3 SST (unit: °C) from ERAi (blue) and
1001 CTRL (orange).

1002 **Figure 4:** **a)** Quarterly SST time series (from ERAi) during the preceding (e.g. year -1) and
1003 simultaneous (e.g. year 0) ISM years, regressed against the ISMR index (e.g. JJAS ISM
1004 rainfall from GPCP). Unit for the SST regression coefficient is °C by mm/day. **b)** Same as **a)**,
1005 but for regression using quarterly rainfall and 850-hPa wind time series. Units for the rainfall
1006 and 850-hPa wind regression coefficients are mm/day by mm/day and m s^{-1} by mm/day,
1007 respectively. **c)**, Same as **a)**, but for quarterly 200-hPa velocity potential time series. Unit for
1008 the 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression
1009 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1010 test (Ebisuzaki 1997) with 999 samples are contoured (SST or 200-hPa velocity potential) or
1011 shown (rainfall and 850-hPa wind). Quarterly time series refer to the seasons February-
1012 May, June-September, October-January and so on.

1013 **Figure 5:** **a)** Quarterly SST time series during year 0 regressed against the ISMR index (e.g.
1014 JJAS ISM rainfall) in CTRL. Unit for the SST regression coefficient is °C by mm/day. **b)**
1015 Same as **a)**, but for regression using quarterly rainfall and 850-hPa wind time series in CTRL.
1016 Units for the rainfall and 850-hPa wind regression coefficients are mm/day by mm/day and m
1017 s^{-1} by mm/day, respectively. **c)**, Same as **a)**, but for quarterly 200-hPa velocity potential time

1018 series in CTRL. Unit for the 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by
1019 mm/day. Regression coefficients reaching the 90% significance level according to a phase-
1020 scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity potential)
1021 or shown (rainfall and 850-hPa wind).

1022 **Figure 6:** Empirical Orthogonal Function (EOF) 1st and 2nd modes for 12-monthly anomaly
1023 of detrended SST obtained from FTFC and FTFC-obs experiments. **a)** and **b)** EOF1 and
1024 EOF2 from FTFC SST detrended anomalies (11-110 period), respectively. **c)** and **d)** EOF1
1025 and EOF2 from FTFC-obs SST detrended anomalies (11-50 period), respectively. The
1026 number in parentheses for each panel gives the % of SST variance described by the EOF
1027 mode.

1028 **Figure 7:** **a)** Quarterly SST time series during year 0 regressed against the ISMR index (e.g.
1029 JJAS ISM rainfall) in FTFC-obs. Unit for the SST regression coefficient is $^{\circ}\text{C}$ by mm/day. **b)**
1030 Same as **a)**, but for quarterly 200-hPa velocity potential time series in FTFC-obs. Unit for the
1031 200-hPa velocity potential regression coefficient is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression
1032 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1033 test (Ebisuzaki 1997) are contoured.

1034 **Figure 8:** Monthly standard deviations of Nino34 SST (unit: $^{\circ}\text{C}$) from ERAi (blue), CTRL
1035 (orange), FTAC (green) and FTAC-obs (red).

1036 **Figure 9:** **a)** Quarterly rainfall and 850-hPa wind time series during boreal winter and spring
1037 regressed against the Nino34 index during the following boreal winter (e.g. ONDJ Nino34
1038 SST) in FTAC-obs. **b)** Same as **a)**, but in FTAC. **c)** Quarterly SST time series during boreal
1039 winter and spring regressed against the Nino34 index during the following boreal winter in
1040 FTAC-obs. **d)** Same as **c)**, but in FTAC. Unit for the SST regression coefficient is $^{\circ}\text{C}$ by $^{\circ}\text{C}$.
1041 Units for the rainfall and 850-hPa wind regression coefficients are mm/day by $^{\circ}\text{C}$ and m s^{-1} by
1042 $^{\circ}\text{C}$, respectively. Regression coefficients reaching the 90% significance level according to a
1043 phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured (SST or 200-hPa velocity
1044 potential) or shown (rainfall and 850-hPa wind).

1045 **Figure 10:** **a)** 850-hPa zonal wind boreal spring means differences (unit: m s^{-1}) between
1046 FTAC-obs and CTRL; **b)** 200-hPa zonal wind boreal spring means differences (unit: m s^{-1})
1047 between FTAC-obs and CTRL; **c)** 850-hPa velocity potential boreal spring means differences
1048 (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **d)** 200-hPa velocity potential boreal spring
1049 means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **e)** 850-hPa stream
1050 function boreal spring means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-obs and CTRL; **f)**

1051 200-hPa stream function boreal spring means differences (unit: $10^6 \text{ m}^2 \text{ s}^{-1}$) between FTAC-
1052 obs and CTRL.

1053 **Figure 11:** **a)** Rainfall boreal spring standard deviation differences (unit: mm/day) between
1054 FTAC-obs and CTRL. **b)** Same as **a)**, but for FTAC and CTRL. **c)** 850-hPa zonal wind boreal
1055 spring standard deviation differences (unit: m s^{-1}) between FTAC-obs and CTRL. **d)** Same as
1056 **c)**, but for FTAC and CTRL.

1057 **Figure 12:** **a)** Boreal spring SST time series regressed against the Nino34 index during the
1058 preceding boreal winter (e.g. ONDJ Nino34 SST) in FTAC-obs; **b)** Boreal summer SST time
1059 series regressed against the Nino34 index during the preceding boreal winter in FTAC-obs; **c)**
1060 Boreal winter SST time series regressed against the Nino34 index during the preceding boreal
1061 winter in FTAC-obs. Unit for the SST regression coefficient is $^{\circ}\text{C}$ by $^{\circ}\text{C}$. Regression
1062 coefficients reaching the 90% significance level according to a phase-scrambling bootstrap
1063 test (Ebisuzaki 1997) are contoured.

1064 **Figure 13:** **a)** Monthly means of ISMR (unit: mm/day) from GPCP (blue), CTRL (orange),
1065 FTAC (green) and FTAC-obs (red); **b)** Monthly standard deviations of ISMR (unit: mm/day)
1066 from GPCP (blue), CTRL (orange), FTAC (green) and FTAC-obs (red); **c)** Lead-lag
1067 correlations between ISMR and quarterly Nino3.4 SSTs starting from the beginning of the
1068 previous year (e.g. year - 1) to the end of the following year of the monsoon (e.g. year +1) in
1069 observations (blue), CTRL (orange), FTAC (green) and FTAC-obs (red). X-axis indicates the
1070 lag (in 4 months interval) for a 36 months period starting one year before the developing year
1071 of ISMR (e.g. year 0) and Y-axis is the amplitude of the correlation. Thus, the coefficients
1072 corresponding to -1, 0, +1 lags refer, respectively, to the correlations between ISMR in year 0
1073 (e.g. JJAS ISM rainfall) and February-May, June-September and October-January Niño-3.4
1074 SSTs, also during year 0, and so on. Circles indicate correlations that are above the 90%
1075 significance confidence level according to a phase-scrambling bootstrap test (Ebisuzaki
1076 1997).

1077 **Figure 14:** **a)** Boreal spring and summer 200-hPa velocity potential time series regressed
1078 against the preceding boreal winter Nino34 SST in ERAi. **b)** Same as **a)**, but for CTRL. **c)**
1079 Same as **a)**, but for FTAC-obs. Unit for the 200-hPa velocity potential regression coefficient
1080 is $10^5 \text{ m}^2 \text{ s}^{-1}$ by mm/day. Regression coefficients reaching the 90% significance level
1081 according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are contoured.

1082 **Figure S1:** **a)** SST seasonal standard deviations (unit: $^{\circ}\text{C}$) computed from ERAi. **b)** SST
1083 seasonal standard deviation differences (unit: $^{\circ}\text{C}$) computed between CTRL and ERAi.

1084 **Figure S2: a)** Quarterly (residual) SST time series during the preceding (e.g. year -1) and
1085 simultaneous (e.g. year 0) ISM years regressed against the ISMR index (e.g. JJAS ISM
1086 rainfall) in ERAi and GPCP after the simultaneous linear effect of the Nino34 index has been
1087 removed from the SST time series by a linear regression method. Unit for the SST regression
1088 coefficient is °C by mm/day. **b)** Same as **a)**, but for CTRL.

1089 **Figure S3: a)** Quarterly SST time series during years 0 and +1 regressed against the Nino34
1090 index during boreal winter (e.g. ONDJ Nino34 SST) in ERAi. **b)** Same as **a)**, but in CTRL.
1091 Unit for the SST regression coefficient is °C by °C. Regression coefficients reaching the 90%
1092 significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997) are
1093 contoured.

1094 **Figure S4: a)** Quarterly rainfall and 850-hPa wind time series during years 0 and +1
1095 regressed against the Nino34 index during boreal winter (e.g. ONDJ Nino34 SST) in GPCP
1096 and ERAi. **b)** Same as **a)**, but in CTRL. Units for the rainfall and 850-hPa wind regression
1097 coefficients are mm/day by °C and m s^{-1} by °C, respectively. Regression coefficients reaching
1098 the 90% significance level according to a phase-scrambling bootstrap test (Ebisuzaki 1997)
1099 are shown.

1100 **Figure S5: a)** SST standard-deviation differences (unit: °C) between FTAC-obs and CTRL
1101 during boreal spring (e.g. FMAM); **b)** SST standard-deviation differences (unit: °C) between
1102 FTAC-obs and CTRL during boreal summer (e.g. JJAS); **c)** SST standard-deviation
1103 differences (unit: °C) between FTAC-obs and CTRL during boreal winter (e.g. ONDJ). **d)**, **e)**
1104 and **f)** Same as **a)**, **b)** and **c)**, but for SST standard-deviation differences between FTAC-obs
1105 and FTAC.

1106 **Figure S6: a)** SST seasonal mean climatological differences (unit: °C) between FTAC-obs
1107 and CTRL; **b)** Rainfall seasonal means differences (unit: mm/day) between FTAC-obs and
1108 CTRL; **c)** Z20 seasonal means differences (unit: m) between FTAC-obs and CTRL.

1109 **Figure S7: a)** Correlation coefficients between the Nino34 index during boreal winter (e.g.
1110 ONDJ) and boreal summer (e.g. JJAS) rainfall during the following year in ERAi and GPCP.
1111 **b)** Same as **a)**, but in CTRL. **c)** Same as **a)**, but in FTAC-obs. Correlation coefficients above
1112 the 90% significance confidence level according to a phase-scrambling bootstrap test
1113 (Ebisuzaki 1997) are contoured.

1114

1115 **Table captions**

1116 **Table 1:** Summary of the numerical experiments with their main characteristics, including
1117 length, nudging domain and SST climatology used for the nudging in the AO or PO
1118 decoupled experiments. The nudged experiments are the Forced Tropical Pacific Climatology
1119 (FTPC), the Forced Tropical Pacific observed Climatology (FTPC-obs), the Forced Tropical
1120 Atlantic Climatology (FTAC) and, finally, the Forced Tropical Atlantic observed Climatology
1121 (FTAC-obs) runs. See text for more details. For the FTPC and FTPC-obs experiments only
1122 ocean grid-points in the PO are included in the correction or smoothing areas and, similarly,
1123 for the FTAC and FTAC-obs experiments. The different correction domains are also
1124 displayed in Figure 1a. The observed daily SST climatology used in the FTPC-obs and
1125 FTAC-obs experiments is derived from the AVHRR only daily Optimum Interpolation SST
1126 version 2 (OISSTv2) dataset for the 1982-2010 period (Reynolds et al. 2007).