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1 **Is the boreal spring tropical Atlantic variability**  
2 **a precursor of the Equatorial Mode?**

3  
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18  
19  
20 **Abstract**

21  
22 The Equatorial Mode (EM) governs the tropical Atlantic inter-annual variability during  
23 boreal summer. It has profound impacts on the climate of adjacent and remote areas.  
24 However, predicting the EM is one of the most challenging and intriguing issues for the  
25 scientific community. Recent studies have suggested a possible connection between the  
26 boreal spring Meridional Mode (MM) and the EM through ocean wave propagation. Here,  
27 we use a set of sensitivity experiments with a medium-resolution ocean model to  
28 determine the precursor role of a MM to create equatorial SST variability. Our results  
29 demonstrate that boreal summer equatorial SSTs following a MM, are subject to two  
30 counteracting effects: the local wind forcing and remotely-excited oceanic waves. For a  
31 positive MM, the anomalous easterly winds blowing along the equator, shallow the  
32 thermocline, cooling the sea surface via vertical diffusion and meridional advection.  
33 Anomalous wind curl excites a downwelling Rossby wave north of equator, which is  
34 reflected at the western boundary becoming an equatorial Kelvin wave (KW). This  
35 downwelling KW propagates eastward, deepening the thermocline and activating the  
36 thermocline feedbacks responsible for the equatorial warming. Moreover, the local wind  
37 forcing and RW-reflected mechanism have a significant and comparable impact on the  
38 equatorial SST variability. Changes in the intensity and persistence of these distinct  
39 forcings will determine the equatorial SST response during boreal summer. Our results  
40 give a step forward to the improvement of the EM predictability.

41  
42 **Key words:** tropical Atlantic, Meridional Mode, Equatorial Mode, ocean waves, SST  
43 variability

## 46 **1. Introduction**

47

48 The Equatorial Mode (EM) or Atlantic Niño is an air-sea coupled mode that controls the  
49 inter-annual tropical Atlantic variability during boreal summer (Zebiak 1993; Lübbecke  
50 et al. 2018). During its positive phase, the EM is characterized by an anomalous warming  
51 in the eastern equatorial Atlantic, accompanied by a reduction of the climatological trade  
52 winds (Lübbecke et al. 2018). The EM is thought to be generated by ocean dynamics  
53 (Keenlyside and Latif 2007; Lübbecke and McPhaden 2013; Polo et al. 2015a; Jouanno  
54 et al. 2017; Martín-Rey et al. 2019), mainly driven by the Bjerknes feedback (Bjerknes  
55 1969). The Bjerknes mechanism implies the propagation of oceanic waves as a response  
56 of an anomalous wind burst in the western equatorial Atlantic (Keenlyside and Latif  
57 2007). Indeed, several authors have underlined the potential role of oceanic waves in the  
58 development and decay of the EM (Carton and Huang 1994; Polo et al. 2008a; Lübbecke  
59 et al. 2010), shaping its distinct structure and timing (Martín-Rey et al. 2019). In recent  
60 decades, additional mechanisms as air-sea fluxes (Nnamchi et al. 2015, 2016), equatorial  
61 deep jets (Brandt et al. 2011) or meridional advection of subsurface temperature (Richter  
62 et al. 2013) have been also proposed to generate the equatorial SST variability.

63

64 The EM has profound impacts on the climate of adjacent and remote areas. Fluctuations  
65 in the onset and intensity of West African (Losada et al. 2012a; Rodríguez-Fonseca et al.  
66 2015), Indian (Kucharski et al. 2008; 2009) and East-Asian Monsoon (Jin and Huo 2018)  
67 have been found associated with the EM. Moreover, the EM influences the precipitation  
68 regime over Europe and Mediterranean Sea (Losada et al. 2012b; Mohino and Losada  
69 2015) and is able to favour the development of a next winter ENSO during certain decades  
70 (Rodríguez-Fonseca et al. 2009; Martín-Rey et al. 2014; 2015; Polo et al. 2015b). Thus,  
71 predicting the EM is one of the most challenging and intriguing issues for the scientific  
72 community.

73

74 The EM emerges as an internal mode of tropical Atlantic variability, however, it is also  
75 subject to multiple external forcings (Ruiz-Barradas et al. 2000; Czaja et al. 2002), which  
76 provide additional sources for its predictability. In this sense, the North Atlantic  
77 Oscillation and ENSO phenomenon are the main remote forcings (Latif and Grötzner  
78 2000; Ruiz-Barradas et al. 2000; Handoh et al. 2006). Interestingly, within the tropical  
79 Atlantic basin, a possible connection between the boreal spring Meridional Mode (MM)  
80 and EM has been suggested (Servain et al. 1999; Murtugudde et al. 2001; Andreoli and  
81 Kayano 2003; Foltz and McPhaden 2010a). The MM is characterized by an anomalous  
82 inter-hemispheric SST gradient associated with anomalous winds blowing to the warmer  
83 hemisphere (Nobre and Shukla 1996). Several mechanisms have been reported to explain  
84 the interaction between these modes. Servain et al. (1999) stated that MM and EM can be  
85 connected at decadal and short inter-annual time scales, via the latitudinal migration of  
86 the Inter Tropical Convergence Zone (Murtugudde et al. 2001; Andreoli and Kayano  
87 2003). The meridional advection of north tropical Atlantic subsurface temperature  
88 anomalies (Richter et al. 2013) or the discharge of ocean heat content into the equatorial

89 wave guide (Huang and Shukla 1997; Zhu et al. 2012) could also link the tropical and  
90 equatorial Atlantic variability.

91

92 Remarkably, a dynamical mechanism based on remotely-excited oceanic waves could  
93 also mediate the MM-EM connection (Foltz and McPhaden 2010a). During the  
94 development of the MM, an anomalous wind stress curl triggers a Rossby wave north of  
95 the equator that propagates westward and is reflected at the western boundary. As a  
96 consequence, it becomes a Kelvin wave that propagates along the equatorial wave guide,  
97 impacting in the equatorial SST anomalies (Foltz and McPhaden 2010b; Burmeister et al.  
98 2016). This wave-reflected mechanism has been considered as part of a destructive  
99 interaction between the MM and EM (Foltz and McPhaden 2010a) or as responsible of  
100 the inconsistent relationship between ENSO and EM (Lübbecke and McPhaden 2012).  
101 However, recent studies highlight the potential role of the wave-reflected mechanism in  
102 creating equatorial SST variability during boreal summer (Foltz and McPhaden 2010b;  
103 Burmeister et al. 2016). Moreover, Martín-Rey et al. (2019) demonstrate that this ocean  
104 wave propagation shapes the distinct structure and timing of the EM. Under this context,  
105 disentangling the potential impact of the MM in the development of the Equatorial Mode  
106 becomes necessary.

107

108 The present study aims to shed light about the precursor role of the MM to generate  
109 equatorial Atlantic variability during the following summer. We will determine the  
110 mechanisms associated with the development and decay of the MM, with a special focus  
111 on the wave activity. Moreover, we will assess the relative contribution of the MM-  
112 associated wind forcing and ocean wave propagation in the generation of equatorial SST  
113 anomalies, and in turn, of an EM event. For this purpose, a set of sensitivity experiments  
114 with the medium-resolution ocean NEMO model have been performed and analysed.

115

## 116 **2. Data and Methodology**

117

### 118 ***2.1 Observations model description***

119 To validate the ability of the model to reproduce the boreal spring tropical Atlantic  
120 variability, observed SSTs are considered from the HadISST dataset (Rayner et al. 2003)  
121 for the period 1960 to 2011.

122

123 The tropical Atlantic configuration of the ocean NEMO model (Madec 2008) has been  
124 used (Faye et al. 2015). The horizontal resolution is  $\frac{1}{4}^\circ$  with a tripolar grid and 46 z-  
125 levels. The model is forced with observed meteorological atmospheric variables (air  
126 temperature and air humidity at 2m, surface wind at 10m, shortwave and longwave  
127 radiation and precipitation) from the DRAKKAR forcing sets, version DFS4.4 (Brodeau  
128 et al. 2010). The air-sea fluxes and wind stress are computed interactively using the bulk  
129 formula and providing them as outputs by NEMO model. In order to investigate the  
130 mechanisms underlying the development of the MM and its potential connection to the  
131 EM, a set of sensitivity experiments have been conducted:

132

- 133 • **INTER:** 52-year simulation forced with inter-annual air-sea fluxes for the period  
134 1960-2011 (for more details, see Martín-Rey et al. (2019)).  
135
- 136 • **MM-REF:** 1.5-year simulation forced with the composite air-sea fluxes  
137 associated with a typical Meridional Mode event from July (year -1) to December  
138 (year 0) (see Methods for more details). This simulation reproduces the growing  
139 and decaying phase of the MM pattern, which allows us to assess the impact of  
140 the MM in the following summer equatorial variability.  
141
- 142 • **MM-WAVE:** 1.5-year simulation forced with the composite air-sea fluxes  
143 associated with a typical Meridional Mode event from July (year -1) to June (year  
144 0). This simulation is similar to MM-REF, but here the atmospheric forcings act  
145 only until late spring and they are set to climatological values from July. MM-  
146 WAVE allows for isolating the contribution of ocean wave propagation in the  
147 generation of boreal summer equatorial SST anomalies.  
148
- 149 • **MM-WIND:** 1.5-year simulation forced with the composite air-sea fluxes  
150 associated with a typical Meridional Mode event from July (year 0) to December  
151 (year 0). The atmospheric forcings have been set to climatological values from  
152 July (year -1) to June (year 0). This simulation is similar to MM-REF, but here  
153 the atmospheric forcings act only from late spring until winter months. Thus,  
154 MM-WIND allows for isolating the role of the local wind forcing to create  
155 equatorial SST anomalies during summer months.  
156

157 The above-mentioned simulations start from the same initial conditions, taken from a  
158 stabilized climatological run. The climatological simulation has been run for 10 years  
159 using the climatological DFS4.4 forcings from the total period 1958-2011. For the MM-  
160 WAVE and MM-WIND experiments, a smooth linear transition has been applied to the  
161 atmospheric forcings during June, to minimize the initial shock of the model. Our  
162 experimental design allows for determining the relative role played by the MM-associated  
163 thermodynamic (surface wind) and dynamic processes (oceanic waves) in the  
164 development of equatorial SST variability and, in turn, in the generation of a EM event,  
165 during next summer.  
166

167 The output variables used throughout the study are: SST, sea surface height (SSH), the  
168 isotherm of 16°C as a proxy of the thermocline depth (D16), horizontal currents and wind  
169 stress.  
170

## 171 **2.2 Methods**

172

173 Seasonal anomalies have been computed by subtracting the seasonal cycle of the whole  
174 period (1960-2011) for observations and INTER simulation. A high-pass 7-year cut-off  
175 Butterworth filter (Butterworth 1930) has been applied to the anomalies to isolate the  
176 inter-annual variability and remove the global warming trend.

177 The leading modes of boreal spring (March-April-May-June) SST variability have been  
 178 computed using the Principal Component Analysis (PCA, Figure 1a-b). PCA decomposes  
 179 the anomalous SST space-time field  $Y(n_s, n_t)$  into a number of *modes of variability* that  
 180 maximize its variance. Each mode of variability is formed by a spatial structure  
 181 (Empirical Orthogonal Function, EOF) and a time series (Principal Component, PC),  
 182 which explain a fraction of the total variance of the original SST field (von Storch and  
 183 Zwiers 2001). A typical Meridional Mode has been constructed as a composite of 28  
 184 events, defined as those years in which the MM time series exceeds  $\pm 1$ std over the whole  
 185 period 1960-2011 (Figure 1c). The composite of its associated air-sea fluxes is used to  
 186 force the sensitivity experiments described in Section 2.1.

187  
 188 For the analysis of the MM-REF, MM-WAVE and MM-WIND experiments, the 5-day,  
 189 monthly and seasonal anomalies are computed by subtracting the associated seasonal  
 190 cycle from the climatological simulation. To better assess the wave activity, a band-pass  
 191 Butterworth filter that retains the 60 days – 540 days frequency, has been applied to the  
 192 5-day SSH, D16 and wind stress anomalies.

193  
 194 Several climate indices have been defined to characterize the key regions of the MM and  
 195 EM patterns. North Tropical Atlantic (NTA) index is referred to the averaged anomalous  
 196 fields in the area  $[50^\circ\text{W}-15^\circ\text{W}, 10^\circ\text{N}-20^\circ\text{N}]$ , while the equatorial region is characterized  
 197 by the Atl3 index  $[20^\circ\text{W}-0^\circ, 3^\circ\text{N}-3^\circ\text{S}]$ .

198  
 199 The calculation of the Ekman velocity has been done according to the following formula:  
 200

$$w_E = -\text{curl}(\tau/\rho f) = \partial(\tau_x/\rho f)/\partial y - \partial(\tau_y/\rho f)/\partial x$$

201  
 202 where  $\tau_x$  and  $\tau_y$  are the zonal and meridional components of the wind stress,  $\rho$  is the sea  
 203 water density and  $f$  is the Coriolis parameter.  
 204

### 205 206 **2.3 Heat budget analysis**

207  
 208 To explore the air-sea interactions responsible for the development and decay of the MM,  
 209 a heat budget analysis has been carried out in the tropical Atlantic. Our model allows the  
 210 interactive calculation of the heat budget in the mixed layer using the bulk formula. The  
 211 different terms of equation [1] are provided as outputs by the model. The temporal  
 212 variations of the mixed layer temperature are explained by the contribution of diverse  
 213 terms (Peter et al. 2006; Martín-Rey et al. 2019):  
 214

$$215 \partial_t \langle T \rangle = \underbrace{-\langle u \partial_x T \rangle - \langle v \partial_y T \rangle + \langle D_l(T) \rangle}_{a} - \underbrace{\frac{1}{h} \frac{\partial h}{\partial t} (\langle T \rangle - T_{z=-h}) - \langle w \partial_z T \rangle - \frac{1}{h} (\kappa_z \partial_z T)_{z=-h}}_{b} + \underbrace{\frac{Q_s(1-F_{z=-h}) + Q^*}{\rho_0 c_p h}}_{c} \quad [1]$$

216  
 217  
 218  
 219

220 with  $\langle * \rangle = \frac{1}{h} \int_{-h}^0 * dz$  where  $h$  is the depth of the mixed layer;  $T$  and  $T_h$  are the temperature  
221 of the mixed layer and below the mixed layer respectively;  $u, v$  and  $w$  are the zonal,  
222 meridional and vertical currents, respectively;  $D_l$  is the lateral diffusion operator and  $\kappa_z$   
223 is vertical mixing coefficient. Notice that the net surface heat fluxes,  $Q_{net}$ , is decomposed  
224 into the non-solar (latent, sensible and longwave)  $Q^*$  and solar (shortwave)  $Q_s$  heat  
225 fluxes. As not all the incident shortwave solar radiation will penetrate in the mixed layer,  
226 the function  $F_{z=-h}$  is included, which describes the fraction of shortwave fluxes absorbed  
227 in this layer and thus contributing to the mixed layer heating. Finally,  $\rho_0$  is the seawater  
228 density and  $C_p$  is the seawater specific heat capacity coefficient.

229

230 The equation [1] shows that the tendency of the temperature in the mixed layer (left) can  
231 be expressed as the sum of atmospheric and oceanic contributions. The atmospheric term  
232 is referred to air-sea fluxes (c), while the oceanic component is associated with horizontal  
233 terms (a, zonal and meridional advection and lateral diffusion) and vertical processes (b,  
234 turbulent mixing, vertical advection and entrainment). This approach has been found to  
235 be a very useful method to investigate the air-sea interactions involved in the tropical  
236 Atlantic variability modes (Polo et al. 2015a; Martín-Rey et al. 2019).

237

### 238 **3. Results**

239

#### 240 ***3.1 Tropical Atlantic inter-annual variability in boreal spring***

241

242 The Meridional Mode (MM) emerges as the second mode of tropical Atlantic SST  
243 variability during boreal spring in both observations and INTER simulation for the period  
244 1960-2011 (Figure 1a-b). It accounts for the 24.6% and 24.8% of the total variance in  
245 observations and model simulation respectively. The inter-hemispheric SST gradient is  
246 well captured by the model ( $r = 0.96$ ), although the equatorial cooling displays a westward  
247 extension reaching South American coast compared to the observations (Figure 1a-b).  
248 This can be due to the equatorial cold bias of the model, associated with a thinner  
249 equatorial mixed layer and reduced thermocline slope (Martín-Rey et al. 2019). The inter-  
250 annual variability of the MM is also well reproduced by the INTER simulation over the  
251 whole period ( $r=0.9$ , Figure 1c).

252

253 As a first attempt to evaluate the possible connection between MM and EM, lead-lag  
254 correlation between the time series (PC2 fixed in MAMJ) of the MM and Atl3 SST index  
255 from JFMA to DJFM has been computed (Figure 1d). Significant negative correlations  
256 are found from previous winter to next summer, being maximum at lag 0 (MAMJ, up to  
257 0.8). Our results show that the entire evolution of the MM is strongly linked to equatorial  
258 SST anomalies. In particular, the correlation between MM and boreal summer (JJAS)  
259 Atl3 index reaches 0.4-0.5, which suggests that the MM would explain around 20% of  
260 the summer equatorial SST variability and thus an important contribution to the  
261 generation of the EM (Figure 1d).

262

### 263 *3.2 Air-sea interactions involved in the development of the MM*

264

265 To gain further understanding of the physical processes controlling the development of  
266 the MM and its potential link to the EM, a sensitivity experiment based on a typical MM  
267 pattern has been performed, hereinafter MM-REF (see section 2.2). This typical MM has  
268 been computed as a composite MM SST pattern, which has a good agreement (not shown)  
269 with the leading mode of boreal spring variability from INTER simulation (Figure 1b).  
270 MM-REF reveals that the development of the MM starts in boreal winter (JFM) with a  
271 reduction of the north-easterly trades, associated with an anomalous cyclonic circulation  
272 (Figure 2a). These anomalous winds persist until late spring (Figure 2b), decreasing the  
273 latent heat loss and warming the mixed layer (Figure 2d). In contrast, the trades intensify  
274 in the south tropical Atlantic from boreal winter to spring (Figure 2a-b), enhancing the  
275 evaporation and cooling the sea surface (not shown). Our results corroborate the role of  
276 the air-sea fluxes to lead the large-scale structure of the MM, with a negligible  
277 contribution of oceanic terms (Amaya et al. 2016; Chang et al. 1997, Wagner et al. 1996;  
278 Carton et al. 1996).

279

280 This asymmetric SST structure causes a meridional sea surface pressure gradient (not  
281 shown), according with the Wind-Evaporation-SST feedback (Amaya et al. 2016), given  
282 rise to anomalous cross-equatorial winds blowing to the warmer hemisphere (Figure 2a-  
283 c). Consequently, the easterlies are reinforced along the equator from winter to spring,  
284 shallowing the thermocline and activating the ocean processes responsible for the surface  
285 cooling (Figure 2a-c,e). During summer months, the MM-wind forcing decays and the  
286 equatorial band warms up, due to vertical diffusion and meridional advection respectively  
287 (Figure 2d-e). Notice that during the entire MM evolution, the air-sea fluxes tend to damp  
288 the equatorial SST anomalies (blue line, Figure 2e). Thus, our results provide further  
289 evidence of the essential role of ocean dynamics to control the equatorial SST variations  
290 (Foltz et al. 2003; Peter et al. 2006; Polo et al. 2015a; Martín-Rey et al. 2019).

291

### 292 *3.3 Wave propagation in the MM evolution*

293

294 During the development of the MM event, the atmospheric wind forcing originates an  
295 anomalous zonal SSH gradient at the equator (shaded, Figure 3a). This east-west SSH  
296 dipole shifts its phase during summer months (shaded, Figure 3b), resembling the Kelvin  
297 and Rossby wave signature of the delayed oscillator mechanism (Suarez and Schopf  
298 1988). This anomalous configuration suggests the existence of ocean wave activity during  
299 the development and decay of the MM, according to previous findings (Foltz and  
300 McPhaden 2010a; Burmeister et al. 2016). However, a better characterization of the ocean  
301 wave propagation, as well as the mechanisms responsible of the excitation of Rossby and  
302 Kelvin waves, is required.

303

304 Figure 3a reveals that an anomalous anticyclonic circulation produces a negative wind  
305 stress curl north of the equator (purple vectors), causing an anomalous Ekman pumping  
306 (positive Ekman velocity, black contours) and downwelling conditions around 30°W-



307 20°W from January to June (white box). This vertical Ekman velocity (dark grey line,  
308 Figure 3c) is produced by an anomalous convergence of horizontal currents (purple line),  
309 which causes the SSH to rise (orange line) and the thermocline to deepen from January  
310 to May (magenta line, Figure 3c). The simultaneous alteration of off-equatorial surface  
311 and subsurface suggests the excitation of a baroclinic ocean wave. To better visualize the  
312 wave activity during the MM, time-longitude diagrams of filtered 5-day SSH anomalies  
313 at 2°N-4°N and along the equator are displayed in Figure 4.

314

315 During the growing phase of the MM, an anomalous wind burst in the western equatorial  
316 Atlantic [40°W-30°W], triggers a set of upwelling Kelvin waves, uKW1 and uKW2 (pink  
317 arrows), propagating eastward from February to May and May to July respectively  
318 (Figure 4b). The uKW1 and uKW2 resemble 2<sup>nd</sup> baroclinic modes (1.27 m/s and 1.45  
319 m/s respectively; Illig et al. (2004); Polo et al. (2008b)) that shallow the thermocline and  
320 favour the mixed layer cooling by vertical processes from February to July (Figure 2e).  
321 Notice that uKW1 is reflected at the African coast, returning as an upwelling Rossby  
322 wave (uRW1, ~ 0.59 m/s, 2<sup>nd</sup> baroclinic mode) along 2°N-4°N (pink arrow, Figure 4c).

323

324 During the decaying phase of the MM, a downwelling Rossby wave (dRW) is excited  
325 north of the equator (Figure 4a), associated with an anomalous negative wind stress curl  
326 (Figure 3c). The dRW propagates to the west as a 2<sup>nd</sup> baroclinic mode (~0.56 m/s) and is  
327 reflected at the western boundary in June-July (Figure 4a), becoming an equatorial  
328 downwelling Kelvin wave (dKW~1.4m/s, Figure 4b). From July to September, as the  
329 dKW propagates eastward, the thermocline deepens setting up the favourable conditions  
330 to warm the equator by vertical diffusion and meridional advection (Figure 2e).

331

332 Our results confirm the existence of RW-reflected mechanism during the MM evolution  
333 proposed by previous studies (Foltz and McPhaden 2010a). Furthermore, we demonstrate  
334 that oceanic waves modulate the development and decay of the Meridional Mode.  
335 Locally-excited equatorial Kelvin waves contribute to generate the MM-associated  
336 equatorial cold tongue, while remotely-excited Rossby wave tends to counteract these  
337 equatorial SSTs during summer months. This is in agreement with Martín-Rey et al.  
338 (2019) who suggested the crucial role of oceanic waves in shaping the distinct structure  
339 and timing of the EM. Nevertheless, it is worth mentioning that we can interpret our  
340 results as a two-way contribution of the MM to the equatorial SST anomalies during  
341 boreal summer. On the one hand, local easterly winds produce a shallower equatorial  
342 thermocline, favouring the surface cooling via vertical diffusion. On the other hand, the  
343 north-equatorial wind curl triggers a downwelling RW boundary-reflected into a KW that  
344 deepens the thermocline and sets up the favourable conditions to warm the equator. The  
345 competition between both phenomena will determine the equatorial SST anomalies  
346 during boreal summer, and thus, the potential relation between the MM and EM.

347

348 Under this context, a quantification of the relative contribution of the oceanic waves,  
349 compared to the local wind forcing on the equatorial SST variability is necessary and will  
350 be assessed in next section.

351

352 *3.4 Impact of the MM-associated ocean wave and wind forcing in the equatorial*  
353 *Atlantic variability*

354

355 The boreal summer equatorial SST anomalies following the peak of the MM are subject  
356 to two distinct contributions: the local wind forcing and the remotely-forced ocean waves.  
357 To isolate the relative contribution of each forcing, two additional sensitivity experiments  
358 have been performed, MM-WAVE and MM-WIND (for more details, see Section 2.2).  
359 In the MM-WAVE, the atmospheric forcing is suppressed from June, allowing the free  
360 propagation of the RW-reflected into the equatorial dKW (Figure 4) during summer  
361 months. Complementary, MM-WIND experiment only considers the atmospheric forcing  
362 from June, which allows for determining the impact of the local wind in the generation  
363 of equatorial SST variability.

364

365 Figure 5 presents the summer (August) SST anomalies in the equatorial Atlantic for the  
366 three sensitivity experiments. MM-WAVE and MM-WIND clearly illustrates a different  
367 and opposite impact over the equatorial SSTs. While MM-WAVE shows a strong  
368 equatorial warming in August (up to  $0.5^{\circ}\text{C}$ ), MM-WIND presents a homogeneous surface  
369 cooling ( $\sim 0.2^{\circ}\text{C}$ - $0.3^{\circ}\text{C}$ ). The temporal evolution of the 5-day SST and SSH anomalies in  
370 the Atl3 region reveals that the equatorial band warms up two times faster in MM-WAVE  
371 respect to MM-REF ( $0.165^{\circ}\text{C}/\text{month}$  vs  $0.062^{\circ}\text{C}/\text{month}$ , Figure 5c), associated with the  
372 early arrival of the dRW at the eastern equatorial Atlantic (thin pink line, Figure 5c).  
373 Indeed, the dKW propagation is more than two time faster in absence of local wind ( $\sim 3.4$   
374 m/s in MM-WAVE respect to 1.4 m/s in MM-REF). In MM-REF, the anomalous  
375 easterlies blow against the wave propagation, establishing a competitive interaction  
376 between both effects (thin purple line, Figure 5a). When the intensified equatorial winds  
377 act alone during summer months, they are able to create an equatorial cooling up to  $0.2^{\circ}\text{C}$   
378 (orange line, Figure 5a).

379

380 This competition between the surface winds and ocean waves is clearly illustrated in the  
381 heat budget analysis of the eastern equatorial Atlantic (Figure 6). In absence of wind  
382 forcing, the propagation of the dKW deepen the thermocline, enhancing the contribution  
383 of vertical processes (dark green line) responsible to warm the equator from July to mid-  
384 August (dark blue line, Figure 6b). The horizontal terms has a negligible effect, while the  
385 air-sea fluxes tend to damp the equatorial warming (light green and blue lines, Figure 6b).  
386 In contrast, when both the atmospheric forcing is included, the positive temperature trend  
387 exhibits large variations (dark blue line, Figure 6a) due to the dominant effect of  
388 horizontal advection (light green line, Figure 6a) added to the air-sea fluxes (light blue  
389 line, Figure 6c). Both phenomena tend to counteract the impact of the ocean dynamics  
390 activated by the dKW propagation (dark green line, Figure 6a).

391

392 Our results bring to light the competition between two distinct forcings to create the  
393 equatorial Atlantic variability during a MM event. Both, the MM-associated local surface  
394 wind and remotely-excited oceanic waves control equatorial SST anomalies during next

395 summer. We have demonstrated for the first time that oceanic waves and wind forcing  
396 have a significant and comparable impact in the generation of equatorial SST variability  
397 (up to 0.2°C, Figure 5a), strong enough to generate an EM event during boreal summer.  
398

399 It is worth-mentioning that our results are referred to a typical MM event in which the  
400 local wind counteracts the ocean wave contribution, given rise to with quasi-neutral  
401 equatorial SST conditions during boreal summer (Figure 2c and Figure 5a). However,  
402 changes in the amplitude or persistence of the atmospheric and/or oceanic forcing could  
403 modify the boreal summer equatorial SST response.  
404

#### 405 **4. Discussion and conclusions**

406  
407 We have investigated the precursor role of the Meridional Mode in generating equatorial  
408 Atlantic variability and potentially an Equatorial Mode event during next summer. For  
409 this purpose, we have performed a set of sensitivity experiments with the ocean NEMO  
410 model. The main conclusions achieved in the present study are:

- 411 • The model reproduces quite well the spatial structure and inter-annual variability  
412 of the Meridional Mode (MM). The MM shows a strong connection with the  
413 equatorial SST anomalies during its entire evolution.  
414
- 415 • The inter-hemispheric SST gradient of the MM pattern is driven by  
416 thermodynamic processes (air-sea fluxes), while equatorial SST anomalies are  
417 controlled by ocean dynamics (vertical diffusion and horizontal advection).  
418
- 419 • Oceanic waves contribute actively to the development and decay of the MM,  
420 shaping the equatorial SST anomalies:  
421
  - 422 ○ *During the growing phase of the MM:* anomalous wind burst in the  
423 western equatorial Atlantic trigger a set of upwelling Kelvin waves (KW)  
424 that propagate eastward during winter and spring, shallowing the  
425 thermocline and favoring the equatorial cooling via vertical diffusion and  
426 meridional advection.  
427
  - 428 ○ *During the decaying phase of the MM:* an anomalous negative wind  
429 stress curl north of the equator, originates an anomalous convergence of  
430 surface currents and an Ekman pumping during boreal spring.  
431 Consequently, the SSH elevates and the thermocline deepens, exciting a  
432 downwelling Rossby wave that is reflected in the western boundary,  
433 becoming an equatorial dKW in boreal summer. The dKW deepens the  
434 thermocline, activating the vertical processes responsible to warm up the  
435 equator.  
436

437

438           ○ Two distinct forcings are responsible of the boreal summer equatorial  
439 SSTs following a MM event: the local surface wind and the remotely-  
440 excited oceanic waves. Both contributions show a significant and  
441 comparable effect over the equatorial Atlantic SST variability. In  
442 absence of local wind forcing, the equatorial band warms up two times  
443 faster due to the early arrival of the dKW that activates the vertical  
444 processes. However, the surface wind contribution shallows the  
445 equatorial thermocline, favouring the surface cooling and counteracting  
446 the oceanic wave effect.

447  
448

449 In the present paper, we provide further evidence of the precursor role of the MM to  
450 generate an EM during next summer. The equatorial SST variability is subject to two  
451 distinct and competitive contributions that determine the equatorial SST response during  
452 boreal summer. Remarkably, Figure 7 illustrates that there are some MM events followed  
453 by the same-sign EM. The Meridional Mode event of 1966 was followed by a strong  
454 equatorial warming, a positive Equatorial Mode-like pattern (Figure 7b-c). Similarly, for  
455 the negative MM event of 2009 (Figure 7d-e). According to our results, during those MM  
456 events, the reflected-RW mechanism was stronger than the local wind, being able to  
457 transfer the north tropical SST anomalies to the equatorial band (Burmeister et al. 2016).  
458 Thus, changes in the amplitude and strength of the MM-associated surface wind could  
459 modulate the equatorial SSTs during next summer, and then the MM-EM connection.

460

461 Our results give a step forward in the better understanding of the processes controlling  
462 the Meridional Mode and its connection to the equatorial Atlantic variability. Moreover,  
463 this study provides evidence about the precursor role of the Meridional Mode to generate  
464 an Equatorial Mode event during next summer, which could be very useful to improve  
465 the predictability of the EM. Nevertheless, further research is still required to better  
466 understand the interaction between the MM and EM, conciliating the distinct proposed  
467 mechanisms as well as the possible multidecadal modulation by natural variability and  
468 Global Warming.

469

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475 by the MetOffice Hadley Centre, from its website  
476 at <https://www.metoffice.gov.uk/hadobs/hadisst/>. The data from the INTER, MM-REF,  
477 MM-WIND and MM-WAVE simulations are available from the authors upon request.

478 **Figure Caption**

479

480 **Figure 1. Observed and simulated Meridional Mode and its connection with**  
481 **equatorial SST variability.** (a-b) Regression maps of the anomalous SSTs (in °C) on the  
482 Principal Component (PC) of the second mode (EOF2) of inter-annual tropical Atlantic  
483 variability in boreal spring (March-April-May-June) for the period 1960-2011 from  
484 observations and INTER simulation. (c) Principal components of the observed (solid  
485 orange line) and modelled (dashed pink line) Meridional Mode for the period 1960-2011.  
486 Grey lines denote the  $\pm 0.5$ std threshold. (d) Lead-lag correlation between the 4-month  
487 averaged SST anomalies in the Atl3 region [20°W-0°, 3°N-3°S] and the Meridional Mode  
488 (PC2 of TA SST in MAMJ) for observations (pink dots) and INTER (green dots) during  
489 the period 1960-2011. The PC2 of the Meridional Mode is fixed in MAMJ (Lag 0), thus,  
490 negative lags (from JFMA to MAMJ) indicate Atl3 leading while positive lags (from  
491 MAMJ to DJFM) imply the leadership of Meridional Mode over the equatorial SST  
492 anomalies. Statistically significant scores at 95% confidence level according to a t-test  
493 are shown in filled dots.

494

495 **Figure 2. Air-sea interactions underlying the development of the Meridional Mode.**  
496 (a-c) Anomalous simulated SSTs (shaded, in °C) and observed surface wind forcing  
497 (vectors, in m/s) from boreal winter (DJF), spring (MAM) and summer months (JJA) for  
498 the MM-REF simulation. Composite surface winds used to force MM-REF come from  
499 DFS4.4 dataset. (d-e) Temporal evolution of the heat budget terms in North Tropical  
500 Atlantic, NTA [50°W-15°W, 5°N-20°N] and Atl3 [20°W-0°, 3°N-3°S] regions (green  
501 boxes in (c)) from previous fall (year -1) to winter months (year 0) following the  
502 Meridional Mode. The heat budget terms displayed are the tendency of the mixed layer  
503 temperature (purple line), the air-sea fluxes (blue line) and oceanic horizontal (green line)  
504 and vertical processes (red line). The data comes from MM-REF simulation.

505

506 **Figure 3. Excitation of ocean waves during the development of the Meridional Mode.**  
507 (a-b) Anomalous temporal variations of SSH (shaded, in m), wind stress (purple vectors,  
508 in N/m<sup>2</sup>) and vertical Ekman velocity (black contours, m/s) for boreal winter-spring  
509 (January-May, a) and summer months (July-September, b) from MM-REF simulation.  
510 (c) Seasonal evolution of monthly variations of SSH (solid orange line, in cm/day), D16  
511 (solid pink line, in m), Ekman vertical velocity (solid black line, in m/day) and divergence  
512 of the horizontal currents (solid purple line in m<sup>3</sup>/day) averaged [30°W-20°W, 3°N-6°N]  
513 (white box in (a)). The monthly variations are computed as the difference between the  
514 last 5-day mean data respect to the first 5-day mean data of each month.

515

516 **Figure 4. Equatorial wave propagation associated with the Meridional Mode.**  
517 Time-longitude diagrams of anomalous filtered 5-days SSH anomalies (in cm) at 2°N-  
518 4°N (a,c) and along the equator (b). Notice that the 2°N-4°N diagram has been duplicated  
519 and the x-axis is reversed to better visualize the propagation and coastal-boundary  
520 reflection of the Rossby and Kelvin waves along the equatorial Atlantic. The propagation

521 of Kelvin and Rossby waves are highlighted with white and pink arrows, for  
522 downwelling and upwelling conditions respectively.

523

524 **Figure 5. Impact of the wave activity in the equatorial Atlantic variability.** (a-c)  
525 Anomalous SST (shaded, in °C) and surface wind forcing (vector, in m/s) in the equatorial  
526 Atlantic in August for MM-REF, MM-WAVE and MM-WIND. (d-e) Time series of  
527 anomalous 5-day SST (solid thick lines, in °C) and SSH (solid thin lines, in cm) in the  
528 Atl3 region for MM-REF, MM-WAVE and MM-WIND from January to December.

529

530 **Figure 6. Heat budget analysis in the eastern equatorial Atlantic.** (a-c) Anomalous 5-  
531 day heat budget terms (temperature trend, air-sea fluxes, vertical and horizontal  
532 processes, in °C/day) in MM-REF, MM-WAVE and MM-WIND simulations from  
533 January to December.

534

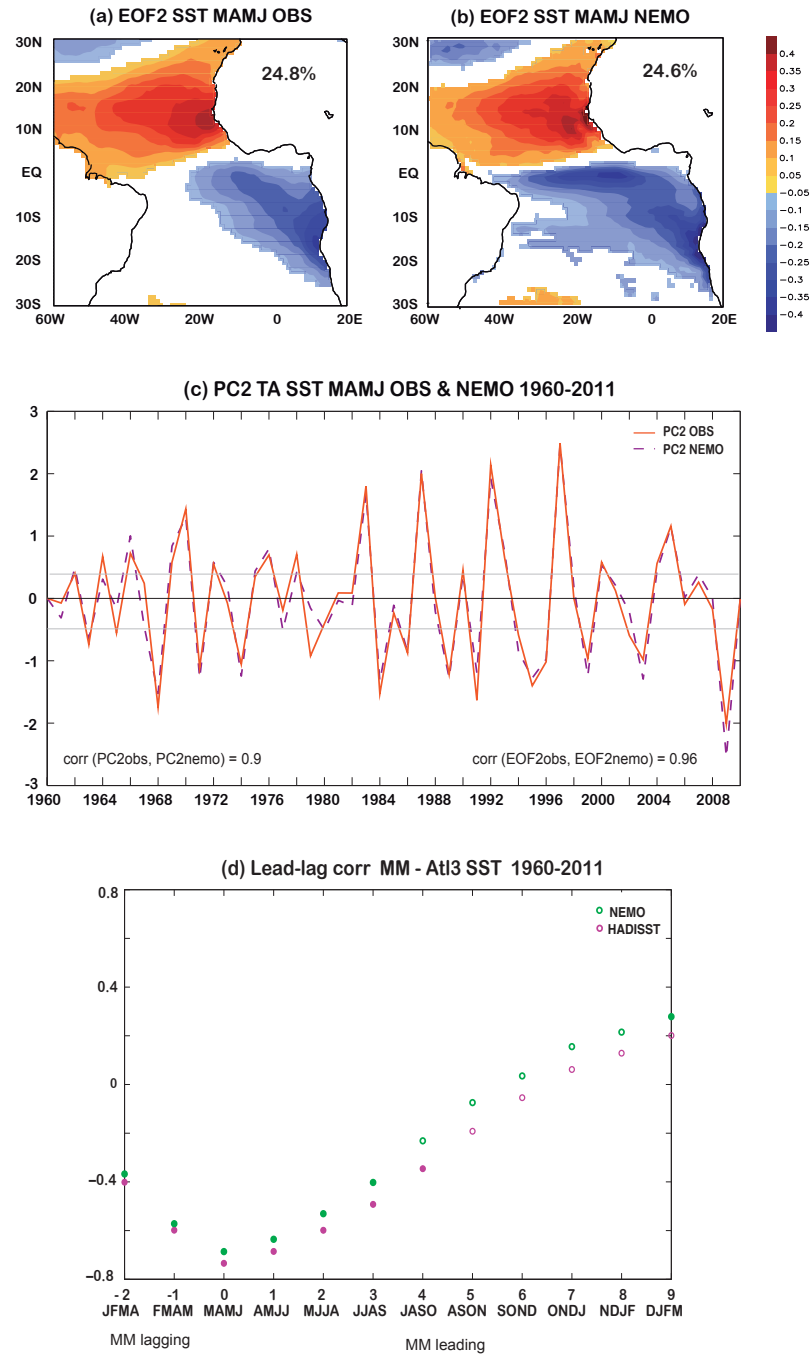
535 **Figure 7. Meridional Mode followed by Equatorial Mode events.** (a) Time series of  
536 Meridional Mode (PC2) and Equatorial Mode (Atl3) during 1960-2011 for observations  
537 and INTER simulation. (b-e) Observed anomalous tropical Atlantic SSTs in boreal spring  
538 (MAMJ) and summer (JJAS) months are displayed for two Meridional Mode events,  
539 1966 and 2009. In both cases, pronounced equatorial SST anomalies during summer  
540 months followed the development of a same-sign Meridional Mode event.

541

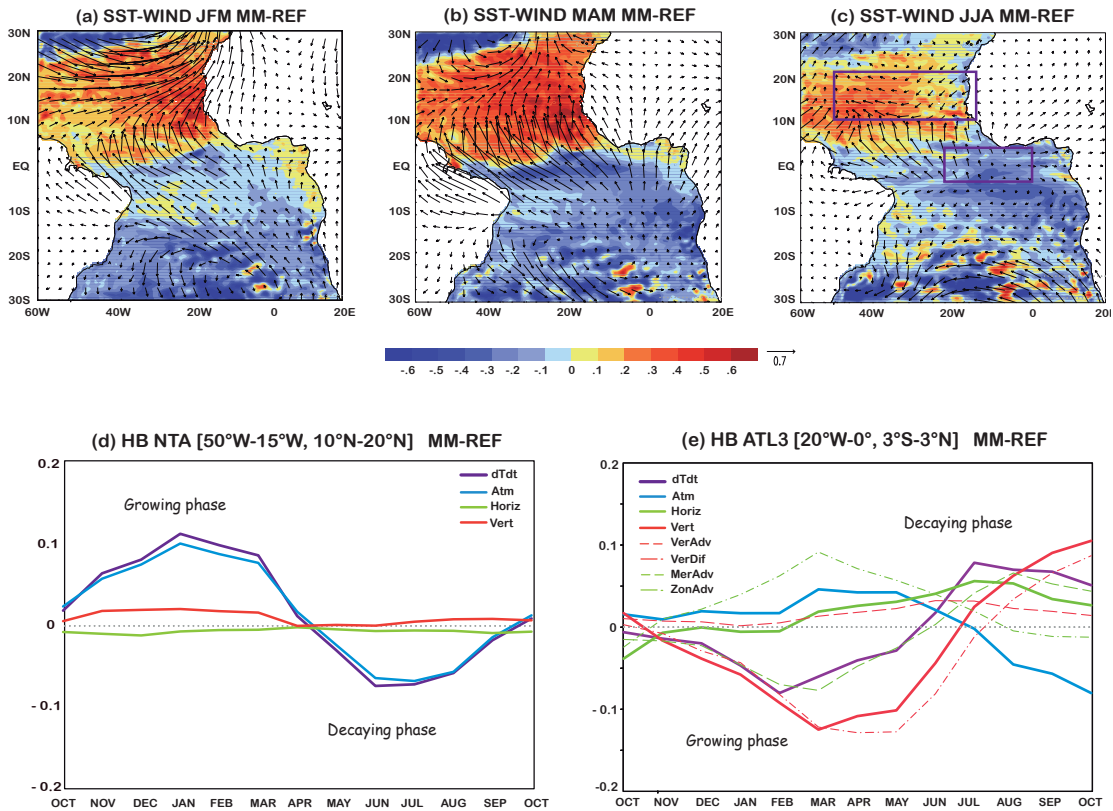
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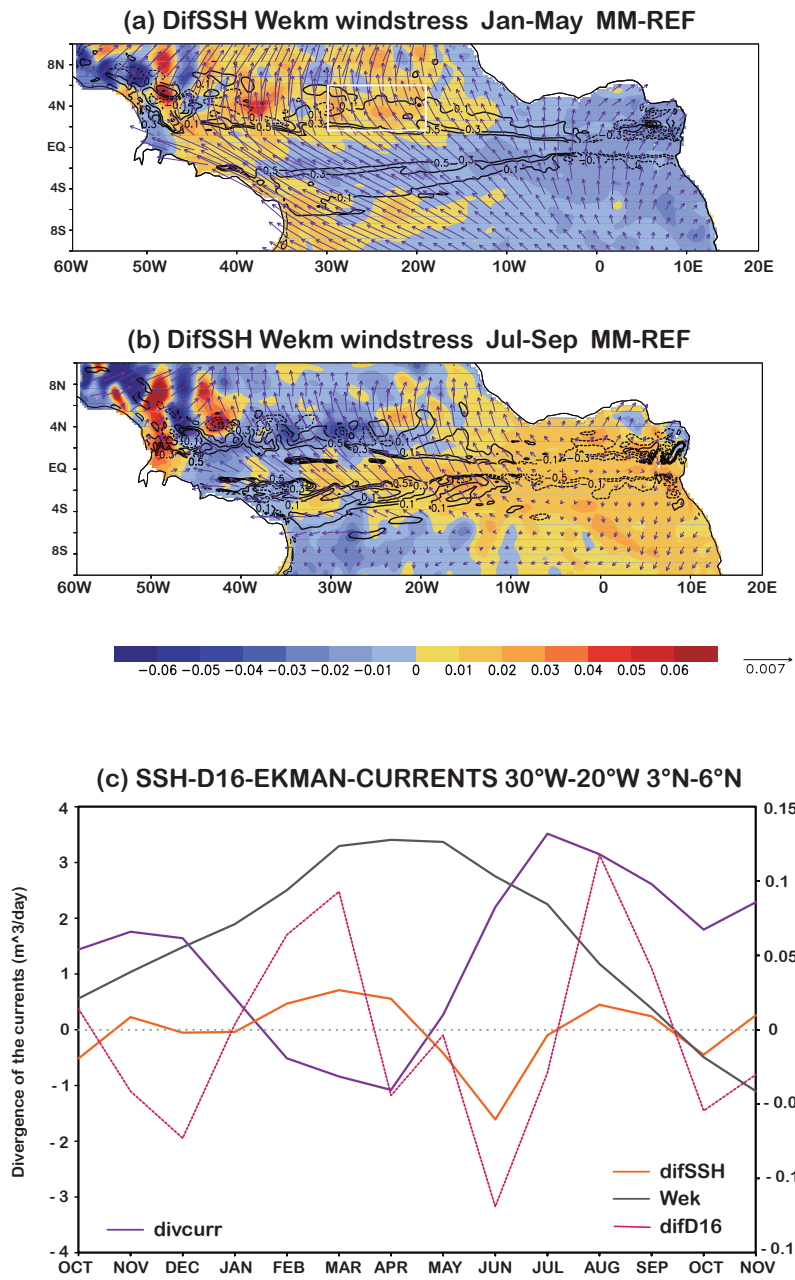


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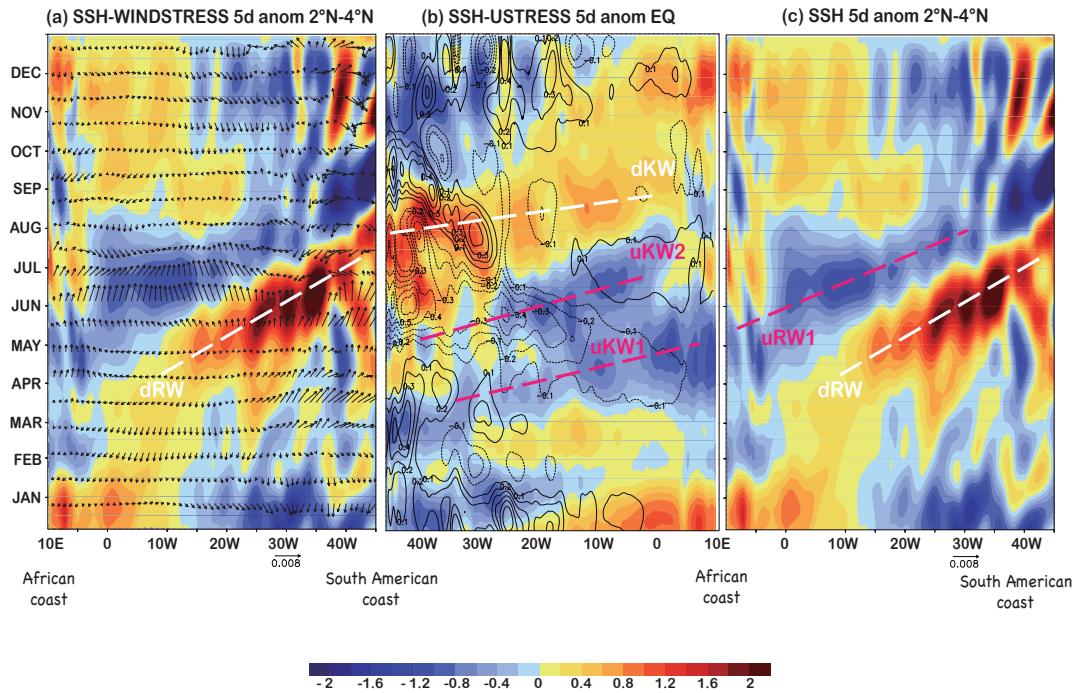
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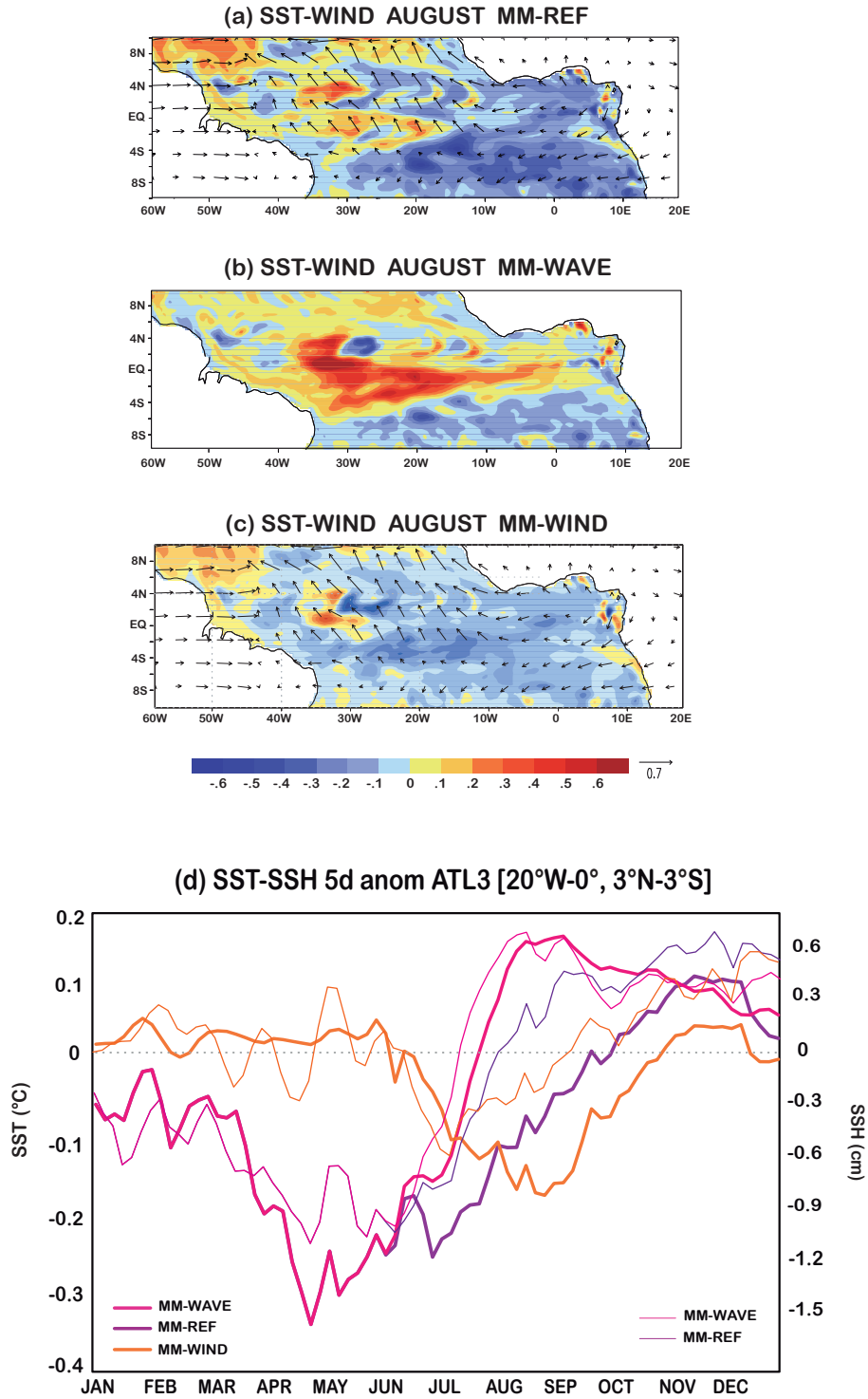
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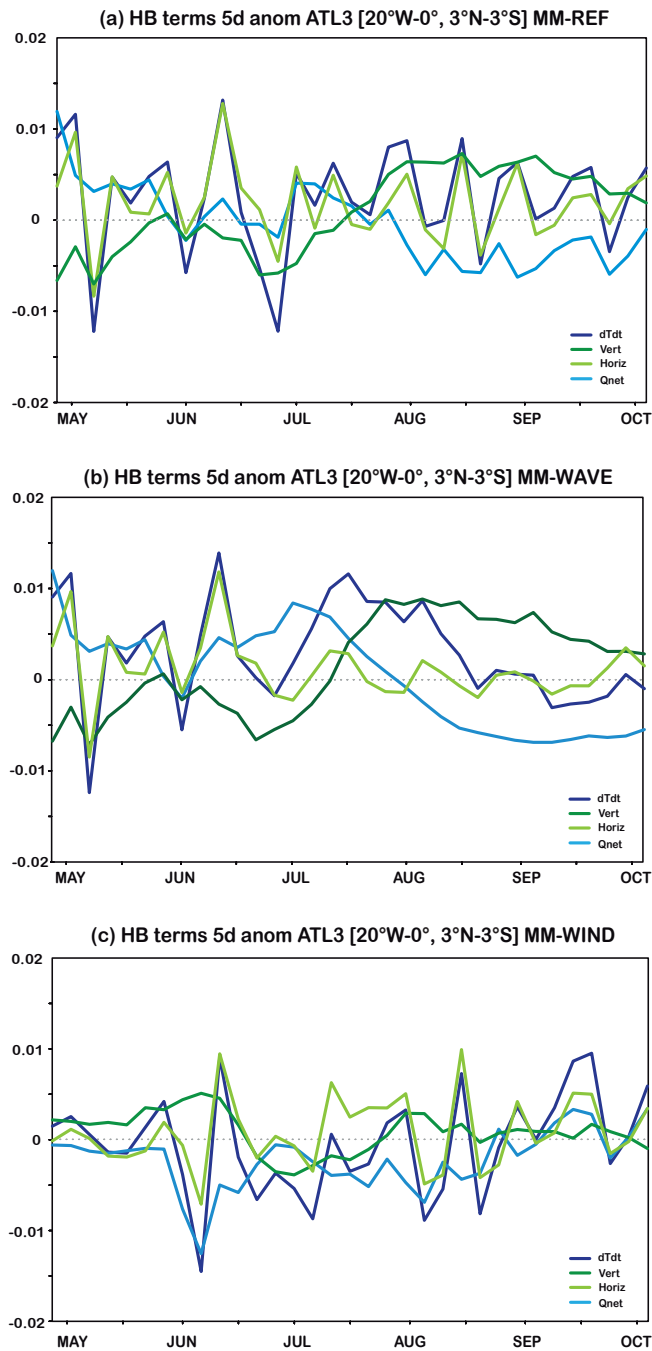


**Figure 4. Equatorial wave propagation associated with the Meridional Mode.**

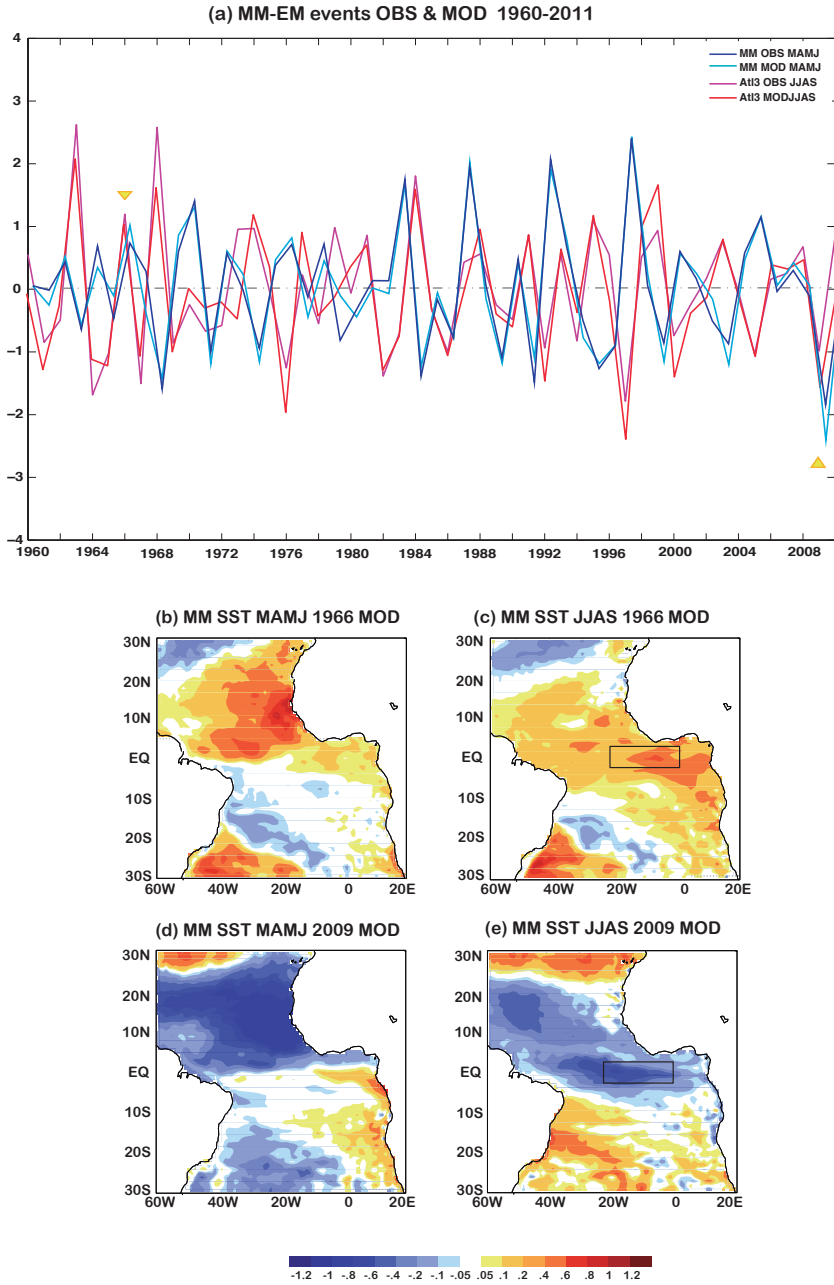
Time-longitude diagrams of anomalous filtered 5-days SSH anomalies (in cm) at  $2^{\circ}\text{N}$ - $4^{\circ}\text{N}$  (a,c) and along the equator (b). Notice that the  $2^{\circ}\text{N}$ - $4^{\circ}\text{N}$  diagram has been duplicated and the x-axis is reversed to better visualize the propagation and coastal-boundary reflection of the Rossby and Kelvin waves along the equatorial Atlantic. The propagation of Kelvin and Rossby waves are highlighted with white and pink arrows, for downwelling and upwelling conditions respectively.



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## References

- Amaya, D. J., M. J. DeFlorio, A. J. Miller, and S.-P. Xie, 2016: WES feedback and the Atlantic Meridional Mode: observations and CMIP5 comparisons. *Clim. Dyn.*, 1-15.
- Andreoli, R. V., and M. T. Kayano, 2003: Evolution of the equatorial and dipole modes of the sea-surface temperature in the Tropical Atlantic at decadal scale. *Meteorology and Atmospheric Physics*, **83**, 277-285.
- Bjerknes, J., 1969: Atmospheric teleconnections from the equatorial Pacific. *Mon. Wea. Rev.*, **97**, 163-172.
- Brandt, P., A. Funk, V. Hormann, M. Dengler, R. J. Greatbatch, and J. M. Toole, 2011: Interannual atmospheric variability forced by the deep equatorial Atlantic Ocean. *Nature*, **473**, 497.
- Brodeau, L., B. Barnier, A. M. Treguier, T. Penduff, and S. Gulev, 2010: An ERA40-based atmospheric forcing for global ocean circulation models. *Ocea. Mod.*, **31**, 88-104.
- Burmeister, K., P. Brandt, and J. Lübbecke, 2016: Revisiting the cause of the eastern equatorial Atlantic cold event in 2009. *J. Geophys. Res.: Oceans*, **121**, 4777-4789.
- Butterworth, S., 1930: On the theory of filter amplifiers. *Experimental wireless and the wireless engineer* **7**, 536-541.
- Carton, J. A., and B. Huang, 1994: Warm Events in the Tropical Atlantic. *J. Phys. Ocea.*, **24**, 888-903.
- Czaja, A., P. Van der Vaart, and J. Marshall, 2002: A Diagnostic Study of the Role of Remote Forcing in Tropical Atlantic Variability. *J. Climate*, **15**, 3280-3290.
- Faye, S., A. Lazar, B. Sow, and A. Gaye, 2015: A model study of the seasonality of sea surface temperature and circulation in the Atlantic North-eastern Tropical Upwelling System. *FrPhy*, **3**, 76.
- Foltz, G. R., and M. J. McPhaden, 2010a: Interaction between the Atlantic meridional and Niño modes. *Geophys. Res. Lett.*, **37**, L18604.
- , 2010b: Abrupt equatorial wave-induced cooling of the Atlantic cold tongue in 2009. *Geophys. Res. Lett.*, **37**, n/a-n/a.
- Foltz, G. R., S. A. Grodsky, J. A. Carton, and M. J. McPhaden, 2003: Seasonal mixed layer heat budget of the tropical Atlantic Ocean. *J. Geophys. Res.: Oceans*, **108**, 3146.
- Handoh, I. C., G. R. Bigg, A. J. Matthews, and D. P. Stevens, 2006: Interannual variability of the Tropical Atlantic independent of and associated with ENSO: Part II. The South Tropical Atlantic. *International Journal of Climatology*, **26**, 1957-1976.

- Huang, B., and J. Shukla, 1997: Characteristics of the Interannual and Decadal Variability in a General Circulation Model of the Tropical Atlantic Ocean. *J. Phys. Ocea.*, **27**, 1693-1712.
- Illig, S., and Coauthors, 2004: Interannual long equatorial waves in the tropical Atlantic from a high-resolution ocean general circulation model experiment in 1981–2000. *J. Geophys. Res.: Oceans*, **109**, n/a-n/a.
- Jin, D., and L. Huo, 2018: Influence of tropical Atlantic sea surface temperature anomalies on the East Asian summer monsoon. *Q.J.R. Met. Soc.*, **144**, 1490-1500.
- Jouanno, J., O. Hernandez, and E. Sanchez-Gomez, 2017: Equatorial Atlantic interannual variability and its relation to dynamic and thermodynamic processes. *Earth Systems Dynamics*, **8**, 1061-1069.
- Keenlyside, N. S., and M. Latif, 2007: Understanding Equatorial Atlantic Interannual Variability. *J. Climate*, **20**, 131-142.
- Kucharski, F., A. Bracco, J. H. Yoo, and F. Molteni, 2008: Atlantic forced component of the Indian monsoon interannual variability. *Geophys. Res. Lett.*, **35**, L04706.
- Kucharski, F., A. Bracco, J. H. Yoo, A. M. Tompkins, L. Feudale, P. Ruti, and A. Dell'Aquila, 2009: A Gill–Matsuno-type mechanism explains the tropical Atlantic influence on African and Indian monsoon rainfall. *Q.J.R. Met. Soc.*, **135**, 569-579.
- Latif, M., and A. Grötzner, 2000: The equatorial Atlantic oscillation and its response to ENSO. *Clim. Dyn.*, **16**, 213-218.
- Losada, T., B. Rodríguez-Fonseca, and F. Kucharski, 2012b: Tropical influence on the summer Mediterranean climate. *AtScL*, **13**, 36-42.
- Losada, T., B. Rodriguez-Fonseca, E. Mohino, J. Bader, S. Janicot, and C. R. Mechoso, 2012a: Tropical SST and Sahel rainfall: A non-stationary relationship. *Geophys. Res. Lett.*, **39**, L12705.
- Lübbecke, J., and M. J. McPhaden, 2012: On the Inconsistent Relationship between Pacific and Atlantic Niños\*. *J. Climate*, **25**, 4294-4303.
- , 2013: A Comparative Stability Analysis of Atlantic and Pacific Niño Modes\*. *J. Climate*, **26**, 5965-5980.
- Lübbecke, J., C. W. Böning, N. S. Keenlyside, and S.-P. Xie, 2010: On the connection between Benguela and equatorial Atlantic Niños and the role of the South Atlantic Anticyclone. *J. Geophys. Res.: Oceans*, **115**, C09015.
- Lübbecke, J., B. Rodríguez-Fonseca, I. Richter, M. Martín-Rey, T. Losada, I. Polo, and N. Keenlyside, 2018: Equatorial Atlantic variability - modes, mechanisms and global teleconnections. *Wiley Interdisciplinary Reviews: Climate Change*.
- Madec, G., 2008: NEMO ocean engine, Note du Pole de modélisation.

- Martín-Rey, M., B. Rodríguez-Fonseca, and I. Polo, 2015: Atlantic opportunities for ENSO prediction. *Geophys. Res. Lett.*, **42**, 6802-6810.
- Martín-Rey, M., B. Rodríguez-Fonseca, I. Polo, and F. Kucharski, 2014: On the Atlantic–Pacific Niños connection: a multidecadal modulated mode. *Clim. Dyn.*, **43**, 3163-3178.
- Martín-Rey, M., I. Polo, B. Rodríguez-Fonseca, A. Lazar, and T. Losada, 2019: Ocean dynamics shapes the structure and timing of tropical Atlantic variability modes *Geophys. Res. Lett.*, **submitted**.
- Mohino, E., and T. Losada, 2015: Impacts of the Atlantic Equatorial Mode in a warmer climate. *Clim. Dyn.*, **45**, 2255-2271.
- Murtugudde, R. G., J. Ballabrera-Poy, J. Beauchamp, and A. J. Busalacchi, 2001: Relationship between zonal and meridional modes in the tropical Atlantic. *Geophys. Res. Lett.*, **28**, 4463-4466.
- Nnamchi, H., J. Li, F. Kucharski, I.-S. Kang, N. S. Keenlyside, P. Chang, and R. Farneti, 2015: Thermodynamic controls of the Atlantic Niño. *Nature Communications*, **6**, 8895.
- , 2016: An Equatorial–Extratropical Dipole Structure of the Atlantic Niño. *J. Climate*, **29**, 7295-7311.
- Nobre, P., and J. Shukla, 1996: Variations in sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America. *J. Climate*, **9**, 2464-2479.
- Peter, A.-C., and Coauthors, 2006: A model study of the seasonal mixed layer heat budget in the equatorial Atlantic. *J. Geophys. Res.: Oceans*, **111**, C06014.
- Polo, I., B. Rodríguez-Fonseca, T. Losada, and J. García-Serrano, 2008a: Tropical Atlantic Variability Modes (1979–2002). Part I: Time-Evolving SST Modes Related to West African Rainfall. *J. Climate*, **21**, 6457-6475.
- Polo, I., A. Lazar, B. Rodríguez-Fonseca, and S. Arnault, 2008b: Oceanic Kelvin waves and tropical Atlantic intraseasonal variability: 1. Kelvin wave characterization. *J. Geophys. Res.: Oceans*, **113**, C07009.
- Polo, I., A. Lazar, B. Rodríguez-Fonseca, and J. Mignot, 2015a: Growth and decay of the equatorial Atlantic SST mode by means of closed heat budget in a coupled general circulation model. *Frontiers in Earth Science*, **3**, 37.
- Polo, I., M. Martín-Rey, B. Rodríguez-Fonseca, F. Kucharski, and C. Mechoso, 2015b: Processes in the Pacific La Niña onset triggered by the Atlantic Niño. *Clim. Dyn.*, **44**, 115-131.
- Rayner, N. A., and Coauthors, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.: Atmosphere*, **108**, 4407.



- Richter, I., S. K. Behera, Y. Masumoto, B. Taguchi, H. Sasaki, and T. Yamagata, 2013: Multiple causes of interannual sea surface temperature variability in the equatorial Atlantic Ocean. *Nature Geoscience*, **6**, 43-47.
- Rodríguez-Fonseca, B., I. Polo, J. García-Serrano, T. Losada, E. Mohino, C. R. Mechoso, and F. Kucharski, 2009: Are Atlantic Niños enhancing Pacific ENSO events in recent decades? *Geophys. Res. Lett.*, **36**, L20705.
- Rodríguez-Fonseca, B., and Coauthors, 2015: Variability and Predictability of West African Droughts: A Review on the Role of Sea Surface Temperature Anomalies. *J. Climate*, **28**, 4034-4060.
- Ruiz-Barradas, A., J. A. Carton, and S. Nigam, 2000: Structure of Interannual-to-Decadal Climate Variability in the Tropical Atlantic Sector. *J. Climate*, **13**, 3285-3297.
- Servain, J., I. Wainer, J. P. McCreary, and A. Dessier, 1999: Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic. *Geophys. Res. Lett.*, **26**, 485-488.
- Suarez, M. J., and P. S. Schopf, 1988: A Delayed Action Oscillator for ENSO. *J. Atmos. Sci.*, **45**, 3283-3287.
- von Storch, H., and F. Zwiers, 2001: Statistical Analysis in Climate Research. *Cambridge University Press*, 484
- Zebiak, S. E., 1993: Air–Sea Interaction in the Equatorial Atlantic Region. *J. Climate*, **6**, 1567-1586.
- Zhu, J., B. Huang, and Z. Wu, 2012: The Role of Ocean Dynamics in the Interaction between the Atlantic Meridional and Equatorial Modes. *J. Climate*, **25**, 3583-3598.