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Fanny Chenillat, Serena Illig, Julien Jouanno, Founi Mesmin Awo, Gael Alory, et al.. How do climate modes shape the chlorophyll-a interannual variability in the tropical Atlantic?. *Geophysical Research Letters*, 2021, 48 (14), pp.e2021GL093769. 10.1029/2021GL093769 . hal-03646658

HAL Id: hal-03646658

<https://hal.science/hal-03646658>

Submitted on 19 Apr 2022

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1 **How do climate modes shape the chlorophyll-*a* interannual variability in the tropical Atlantic?**

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15
16 **Key points (3 – 140 characters each):**

- 17 • The leading Tropical Atlantic climate modes shape the spatio-temporal variability of
18 the Cold Tongue interannual surface Chlorophyll-*a*.
- 19 • The Atlantic Zonal Mode explains the interannual Chlorophyll-*a* variability in the
20 eastern part of the Cold Tongue with a peak in summer.
- 21 • The North Tropical Atlantic mode influences the interannual Chlorophyll-*a* variability
22 in the western Cold Tongue, from spring to winter.

23
24 **Keywords**

25 Tropical Atlantic; Equatorial Atlantic; Cold Tongue; chlorophyll-*a* concentration; interannual
26 variability; Climate modes; Remote sensing.

31 **Abstract**

32 Chlorophyll-*a* concentration (Chl-*a*) observed by satellite shows a marked seasonal and
33 interannual variability in the Tropical Atlantic. This study analyzes how the remotely-sensed
34 surface Chl-*a* responds to the leading boreal summer climate modes affecting the interannual
35 Tropical Atlantic variability over 1998-2018, corresponding to a positive Atlantic Multidecadal
36 Oscillation phase. We show that the Atlantic Zonal Mode (AZM) and the North Tropical
37 Atlantic Mode (NTAM) significantly drive the interannual surface Chl-*a* variability in the
38 equatorial Atlantic, with different timings and contrasted zonal modulation of the Cold
39 Tongue. The AZM involves remotely-forced wave propagations favoring upwelling in the east
40 and Chl-*a* modulation in the core of the Cold Tongue. Instead, the impact of the NTAM is
41 mainly in the west, in response to locally-forced pumping that modulates the western
42 extension of the Cold Tongue. Such conditions can affect the marine food web, inducing
43 significant variations for ecosystem functioning and fisheries.

44

45 **Plain language summary**

46 The Tropical Atlantic Ocean is characterized by strong year-to-year surface temperature
47 fluctuations which can be classified into basin-scale climate modes. In this study, we examine
48 to which extent these modes of variability have a signature on the surface Chlorophyll-*a*
49 concentration, a proxy of the biological activity at sea. Using two decades (1998-2018) of
50 ocean-color satellite observations, we show that the interannual surface Chlorophyll-*a*
51 modulation in boreal summer is impacted by these climate modes mainly along the equator.
52 They drive contrasted Chlorophyll-*a* fluctuations in time and space. The first climate mode,
53 the Atlantic Zonal Mode, involves distant mechanisms that propagate along the equator
54 favoring biological activity in the eastern equatorial basin. Conversely, the second mode, the
55 North Tropical Atlantic Mode, involves local mechanisms in the west of the equatorial band
56 where it favors biological activity. Such conditions can affect the marine food web, inducing
57 significant variations for the fisheries.

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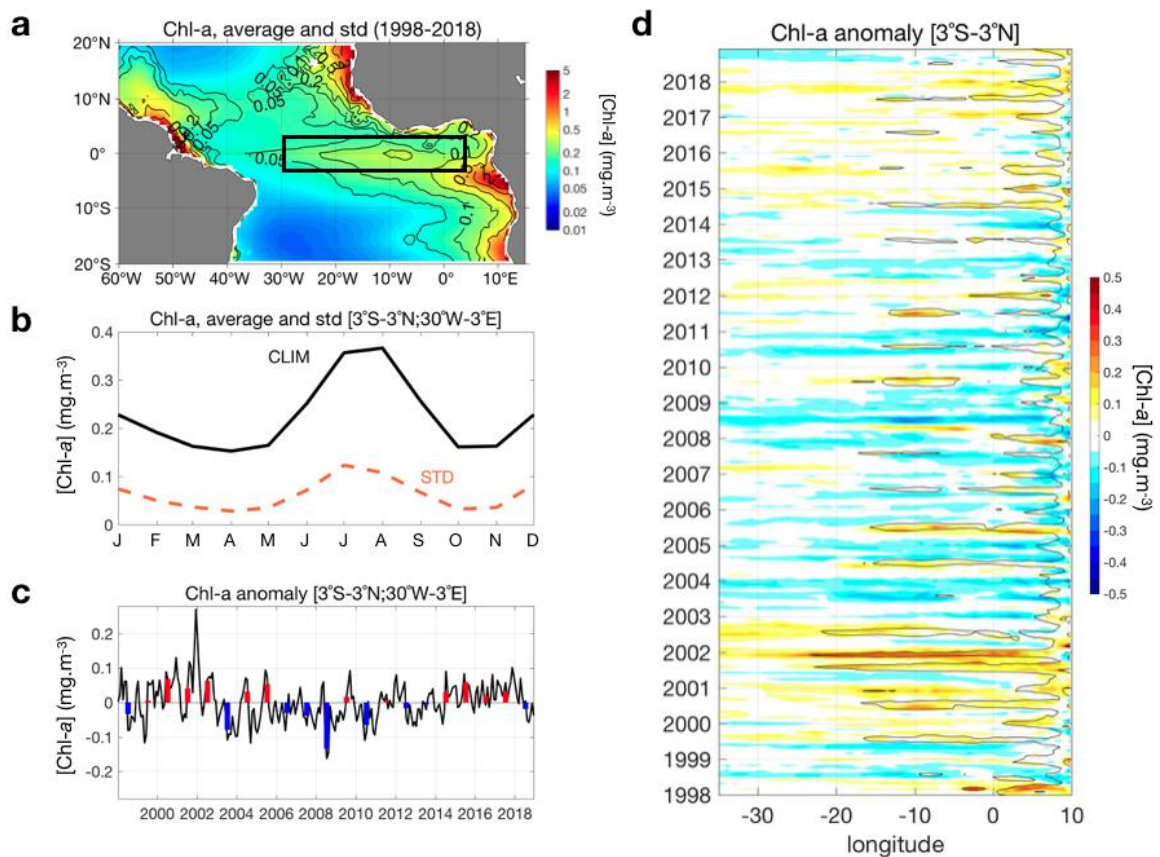
60 1. Introduction

61 The tropical Atlantic is characterized by a low-frequency variability of coupled ocean-
62 atmosphere processes leading to extreme events with critical societal impacts (high rainfall,
63 drought, hurricane; Foltz et al., 2019). These fluctuations which mainly concern the Sea
64 Surface Temperature (SST), occur at different time-scales, from interannual to multidecadal -
65 for the Atlantic Multidecadal Oscillation, AMO (Kerr 2000; Knight et al. 2006). The interannual
66 variability is governed by two leading climate modes, namely the Atlantic Zonal Mode (AZM)
67 and the North Tropical Atlantic Mode (NTAM) (Xie & Carton, 2004; Foltz & McPhaden, 2010).
68 The AZM, also known as the Atlantic Niño or the Atlantic equatorial mode, is associated in its
69 positive phase with anomalous warming (cooling in its negative phase) of boreal summer SST
70 confined to the eastern equatorial basin (**Fig.S1a**; Kushnir et al., 2006; Lübbecke et al., 2018).
71 In contrast, the NTA, also known as the Atlantic Meridional Mode or Dipole mode, is
72 characterized by an interhemispheric gradient of SST Anomalies (SSTA), with anomalous
73 warm conditions in the north tropical Atlantic region and weak negative SSTA south of the
74 equator during its positive phase, and conversely during its negative phase (**Fig.S1b**; Chiang
75 and Vimont, 2014). This interannual mode modulates the latitudinal position of the Atlantic
76 Intertropical Convergence Zone (ITCZ) (e.g., Xie & Carton, 2004). The NTAM usually peaks in
77 boreal spring (Kushnir et al., 2006) and, notably, still has a signature in boreal summer
78 (Richter et al., 2013). The imprint of these two interannual modes is directly modulated by
79 the multidecadal variability, namely the AMO as shown in Martin-Rey et al. (2018). In
80 particular, Vallès-Casanova et al. (2020) investigated the diversity of the Atlantic Niños and
81 showed significant variations of the onset and duration of the SSTA they attributed to
82 preconditioning driven by external forcing such as NTAM or Pacific El Niño.

83 This interannual variability of the tropical Atlantic is likely to impact the marine
84 ecosystem. SST variability can indeed reflect the modulation of nutrient inputs through the
85 thermocline depth displacements and result in the variability of chlorophyll-*a* concentration
86 (Chl-*a*) and marine ecosystems. Actually, in such a region where there is no light-limitation at
87 the surface, the primary production is predominantly driven by the nutrient availability in
88 response to physical drivers (Longhurst, 1993; Longhurst, 2007). To our knowledge, the most
89 extensive investigation of seasonal and interannual variability of Chl-*a* in the tropical Atlantic
90 was carried out by Grodsky et al. (2008), using a 10-year-long time-series of sea surface Chl-*a*.
91 Using an updated 21-year-long time-series of remote-sensed surface Chl-*a* (see Material &

92 Methods), we confirm their results (**Figs.1ab**): the richest surface waters are located along the
93 coasts due to coastal upwelling and river discharges and also in the equatorial Atlantic Cold
94 Tongue region where the equatorial upwelling prevails Seasonal and interannual equatorial
95 Chl-a variability peaks at [10°W, 0°N], spreading zonally from 0°E to 30°W (**Fig.1a**). It is
96 maximum in boreal summer ($0.35 \pm 0.1 \text{ mg.m}^{-3}$, **Fig.1b**). Surface Chl-a anomalies in the Cold
97 Tongue show marked interannual variability (**Fig.1c**). This is even more striking when looking
98 at the zonal extension of Chl-a anomalies along the whole equatorial waveguide (**Fig.1d**).
99 Interestingly, while on average surface Chl-a interannual variability peaks around 10°W, it can
100 extend over the whole equatorial Atlantic and peak at 25°W (e.g., in 2002).

101 Grodsky et al. (2008) hypothesized that interannual variability of the sea surface Chl-a
102 and SST in the eastern equatorial Atlantic is driven by the thermocline displacements,
103 associated with the interannual fluctuations of the equatorial zonal wind. However, the
104 broader effect on primary productivity is still unknown (Foltz et al., 2019). This leads us to
105 question the sources of such temporal and spatial variability of the sea surface Chl-a on an
106 interannual timescale in the entire equatorial Atlantic. We will focus our study on the boreal
107 summer period when seasonal and interannual variability of Chl-a is maximum (**Fig.1b**). In this
108 paper, using remote-sensed observations over the equatorial Atlantic region, we describe
109 how interannual climate modes shape the Chl-a variability in time and space.



111 **Figure 1:** Temporal and spatial variability of surface chlorophyll-a concentration (Chl-a, $\text{mg}\cdot\text{m}^{-3}$) from satellite observations over 1998-2018. (a) Mean (color shading) and interannual standard deviation (black contours). (b) Climatological monthly mean (black line) and interannual monthly standard deviation (STD, orange dashed line) averaged over the Cold Tongue area ($[3^{\circ}\text{S}-3^{\circ}\text{N}; 30^{\circ}\text{W}-3^{\circ}\text{E}]$, delineated by the black box in panel a). (c) Monthly (black line) and boreal summer (June-July-August average, red/blue bars) interannual anomalies in the Cold Tongue. (d) Longitude-time Hovmöller diagram of the equatorial interannual anomalies averaged within $[3^{\circ}\text{S}-3^{\circ}\text{N}]$.

120 2. Material and methods

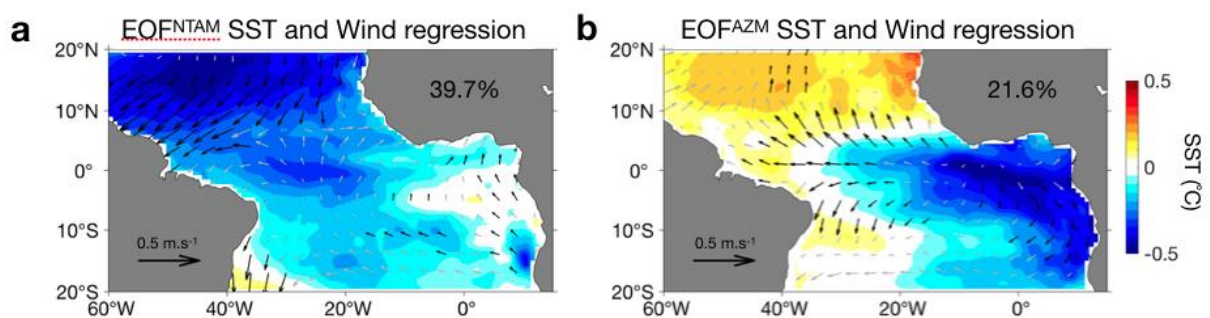
121 This study has been conducted using ocean-color satellite data over the 1998-2018
 122 period in the Tropical Atlantic region from 20°S to 20°N . Ocean-color, provided by E.U.
 123 Copernicus Marine Service Information, is a proxy for open-sea Chl-a (Fournier et al., 2015).
 124 We analyzed cloud-free Chl-a monthly maps (L4 product, CMEMS 2020a), based on the
 125 Copernicus-GlobColour processing chain. Concomitant monthly SST and near-surface
 126 atmospheric circulation described through zonal and meridional wind components come
 127 from the European Center for Medium-Range Weather Forecasts ERA-Interim reanalysis (Dee
 128 et al., 2011). Monthly remote-sensed Sea Surface Height (SSH), a good approximation for the

129 thermocline displacements, is available since 1993 from Aviso Ssalto/Duacs (Pujol et al., 2016)
130 and provided by CMEMS (2020b).

131 After removing the linear trend, monthly interannual anomalies are estimated relative to
132 the 1998-2018 climatology. We focus on the boreal summer Chl-a variability and following
133 Martin Rey et al. (2018), we average the data over the June-July-August (JJA) season.
134 Diagnostics consist primarily of Empirical Orthogonal Function (EOF) analyses and linear
135 regressions. The EOF analysis allows decomposing the Tropical Atlantic SSTA variability into a
136 set of independent spatial modes (the EOFs, **Fig.2**) associated with distinct temporal
137 fluctuations – the Principal Components (PCs, **Figs.3ef**). Classically, PCs are normalized such as
138 their standard deviation (STD) is 1 and the EOF spatial pattern conveys the unit of the
139 variable. To isolate the part of the Chl-a, wind, and SSH variability associated with each
140 leading EOF mode, we linearly regress (using a least mean square method) the interannual
141 time-series onto the standardized PCs (**Figs.3a-d**). The goodness of the linear fit is
142 systematically measured by the square of the correlation coefficient (r^2) between PCs and the
143 independent time-series. Throughout the study, the statistical significance of correlations is
144 estimated based on Student's t-test at the 95% confidence level.

145 Because we aim at understanding Chl-a interannual variability, we preferentially focus on
146 anomalous productive years corresponding to the occurrence of cold interannual events. We
147 thus present the cold phase of the leading EOF modes of the Tropical Atlantic SST, inversely to
148 the usual convention.

149



150 **Figure 2:** Maps of the two leading Empirical Orthogonal Function (EOF) of boreal summer
151 'June-July-August' (JJA) interannual Sea Surface Temperature (SST) variability (color shading,
152 °C) in the tropical Atlantic ([20°S-20°N, 60°W-15°E]) over the 1998-2018 period. The
153 percentages indicate the explained variance of each mode. JJA surface wind interannual
154 anomalies (arrows, m.s⁻¹) are regressed onto the respective standardized Principal
155 Component (PC, shown in **Fig.3ef**). Significant values (> 95% confidence level) are shown with
156 black vectors. NTAM: North Tropical Atlantic mode. AZM: Atlantic Zonal mode.

157 3. Climate modes and Chl-a variability

158

159 3.1 Tropical Atlantic interannual climate modes over a positive AMO phase

160 As expected, two leading modes of interannual climate variability emerge from the
161 analysis of boreal summer tropical Atlantic SSTA (**Fig.2**). While these patterns stray from the
162 classical climate modes (described in the introduction; **Fig.S1**), they are consistent with
163 Martin-Rey et al. (2018), who recently diagnosed a modulation of the leading climate mode
164 structures regarding the AMO phase. These results are coherent since EOFs analyses are
165 performed over the last two decades (1998-2018) and that the AMO is in a positive phase
166 since the 1990s (Wu et al., 2019).

167 The dominant mode explains 39.7% of the summer tropical Atlantic interannual
168 variability. It is characterized by strong northern cold (warm) SSTA, accompanied by
169 anomalous northeasterly (southwesterly) trades, contrasting with a near-zero positive
170 (negative) SSTA south of the equator, confined to the eastern tropical Atlantic (**Fig.2a**). Such
171 pattern resembles closely the NTAM pattern during positive AMO phases (Martin-Rey et al.,
172 2018).

173 The second leading mode explains 21.6% of the summer tropical Atlantic interannual
174 variability. It is characterized by strong cold (warm) SSTA in the eastern equatorial Atlantic
175 extending southeastwards along the African coast in the Benguela upwelling system (**Fig.2b**)
176 typical of an Atlantic Niña (Niño) event. Divergent (convergent) anomalous surface winds
177 overlie the eastern equatorial Atlantic SSTA event (Illig et al., 2020), associated with a
178 strengthening (relaxation) of the Trade winds in the western-central equatorial Atlantic and
179 coastal poleward (equatorward) winds along the Benguela current system. A weak
180 interhemispheric SSTA gradient is also depicted. This pattern is typical of the AZM during
181 positive AMO phases (Martin-Rey et al., 2018).

182 Note that, in comparison with Martin-Rey et al. (2018), we find that the NTAM explains
183 more variance than the AZM mode during a positive AMO phase. This can be attributed to the
184 temporal and spatial restriction of our analysis (Lübbecke et al., 2018), i.e., two decades over
185 20°S-20°N against 34 years over 30°S-30°N in Martin-Rey et al. (2018). A possible linear mix
186 between the two climate modes (Servain et al., 1999; Hormann and Brandt, 2009) or the
187 effect of climate change (Lübbecke et al., 2018) might also explain this difference.

188

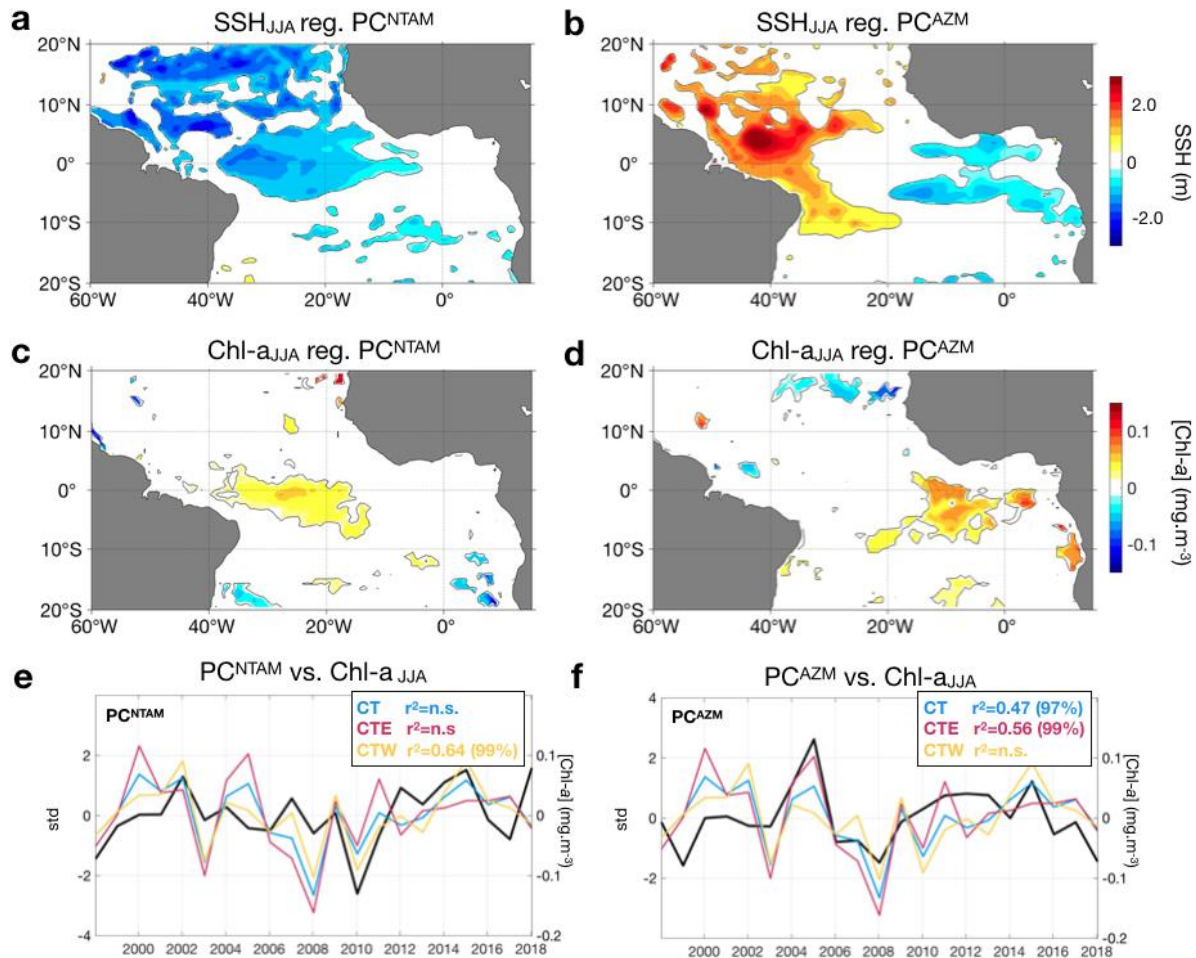
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190 3.2 NTAM and the western equatorial Chl-a modulation

191 The interannual SSH Anomaly (SSHA) signature associated with the cold phase of the
192 NTAM shows negative anomalies mostly in the northern Tropical Atlantic and in the western
193 part of the equatorial band, from 40°W to 5°W with a peak at 40-30°W (**Fig.3a**). Considering
194 the SSH as a proxy for the thermocline depth (Arnault et al., 1992), a cold-phase NTAM is
195 mainly associated with a shoaling of the thermocline where negative SSHA are located
196 (Martin-Rey et al., 2018). Such a thermocline rise is likely to be driven locally by a positive
197 Ekman Pumping anomaly associated with the increased surface wind-stress curl (**Fig.S2**). In
198 the western equatorial sector, the shoaling of the thermocline is expected to enhance the
199 equatorial upwelling and therefore increase the vertical nutrients flux into the euphotic layer
200 and enhance the Chl-a in the surface layer. This mechanism is consistent with the Chl-a
201 signature associated with the cold phase of the NTAM (**Fig.3c**). Positive Chl-a anomalies are
202 indeed observed in the western part of the equatorial fringe collocated with the SSHA. They
203 extend zonally from 40°W to 15°W with a peak near ~25°W. Such a close link between SSHA
204 and Chl-a patterns suggests control of Chl-a by vertical processes, though vertical nutrient
205 fluxes. There is, however, no significant modulation of the Chl-a anomalies in the northern
206 Tropical Atlantic (5-15°N), despite the significant productivity occurring yearlong (**Fig.1a**). The
207 apparent weak sensitivity of surface Chl-a in this region to climate modes would deserve
208 further attention.

209 Consistently, the interannual modulation of NTAM (PC^{NTAM}) and the Chl-a averaged in the
210 Cold Tongue region have a correlation of 0.41. Notably, this correlation substantially increases
211 and becomes statistically significant when only the western part of the Cold Tongue (CTW) is
212 considered ($r=0.64$, >99%) (**Fig.3e**), with a regression coefficient of $-0.12 \text{ mg}\cdot\text{m}^{-3}/^{\circ}\text{C}$.
213 Conversely, the eastern Cold Tongue Chl-a variability is not correlated with PC^{NTAM} . Hence, the
214 NTAM mode significantly explains the interannual boreal summer Chl-a variability in the
215 western equatorial band with strong positive anomalies in 2002 and 2015 and negative
216 anomalies in 2010 (also visible in **Figs.1cd**). Interestingly, 2002 has been investigated by
217 Hormann and Brandt (2009). They diagnosed 2002 as an abnormally warm year in the eastern
218 part of the Cold Tongue contrasting with cold anomalies in the western part (west of 25°W),
219 similarly to our cold-phase NTAM pattern (**Fig.1a**).

220



221 **Figure 3:** Spatio-temporal characteristics of the boreal summer (JJA) Sea Surface Height (SSH)
 222 and Chl-a associated with the NTAM and AZM interannual EOF modes of the Tropical Atlantic
 223 SST (presented in **Fig.2**). Regression maps of JJA SSH interannual anomalies (cm) onto the
 224 NTAM (a) and AZM (b) standardized Principal Components (PCs, shown in panels e-f (black
 225 lines)). Only significant values (> 95% confidence level) are shaded. (c-d) Same as panels a-b
 226 for JJA Chl-a anomaly ($\text{mg}\cdot\text{m}^{-3}$). NTAM (e) and AZM (f) EOF PCs (black lines) are compared with
 227 boreal summer surface Chl-a interannual anomalies averaged in different portions of the Cold
 228 Tongue: entire Cold Tongue (CT, $[3^{\circ}\text{S}-3^{\circ}\text{N}; 27^{\circ}\text{W}-3^{\circ}\text{E}]$, blue lines), eastern Cold Tongue (CTE,
 229 $[3^{\circ}\text{S}-3^{\circ}\text{N}; 15^{\circ}\text{W}-3^{\circ}\text{E}]$, pink lines), and western Cold Tongue (CTW, $[3^{\circ}\text{S}-3^{\circ}\text{N}; 27^{\circ}\text{W}-15^{\circ}\text{W}]$,
 230 orange lines). Squared correlation (r^2) and associated significance level (%) quantify the
 231 goodness of the linear fit between PCs and Chl-a time-series. n.s.: not significant
 232

233 3.3 AZM and the eastern equatorial Chl-a modulation

234 The cold phase of the AZM is associated with strong positive interannual SSHA in the
 235 western Tropical Atlantic, from 10°S to $\sim 15-20^{\circ}\text{N}$, which contrasts with the negative SSHA in
 236 the Gulf of Guinea (east of 20°W) (**Fig.3b**). The latter are divided into two SSH depressions
 237 nearly symmetric about the Equator and centered $\sim 4^{\circ}-5^{\circ}$ of latitude away from the Equator,
 238 typical of the signature of equatorial Rossby waves on the SSH. These upwelling waves are
 239 most likely associated with the reflection of eastward-propagating equatorial Kelvin waves

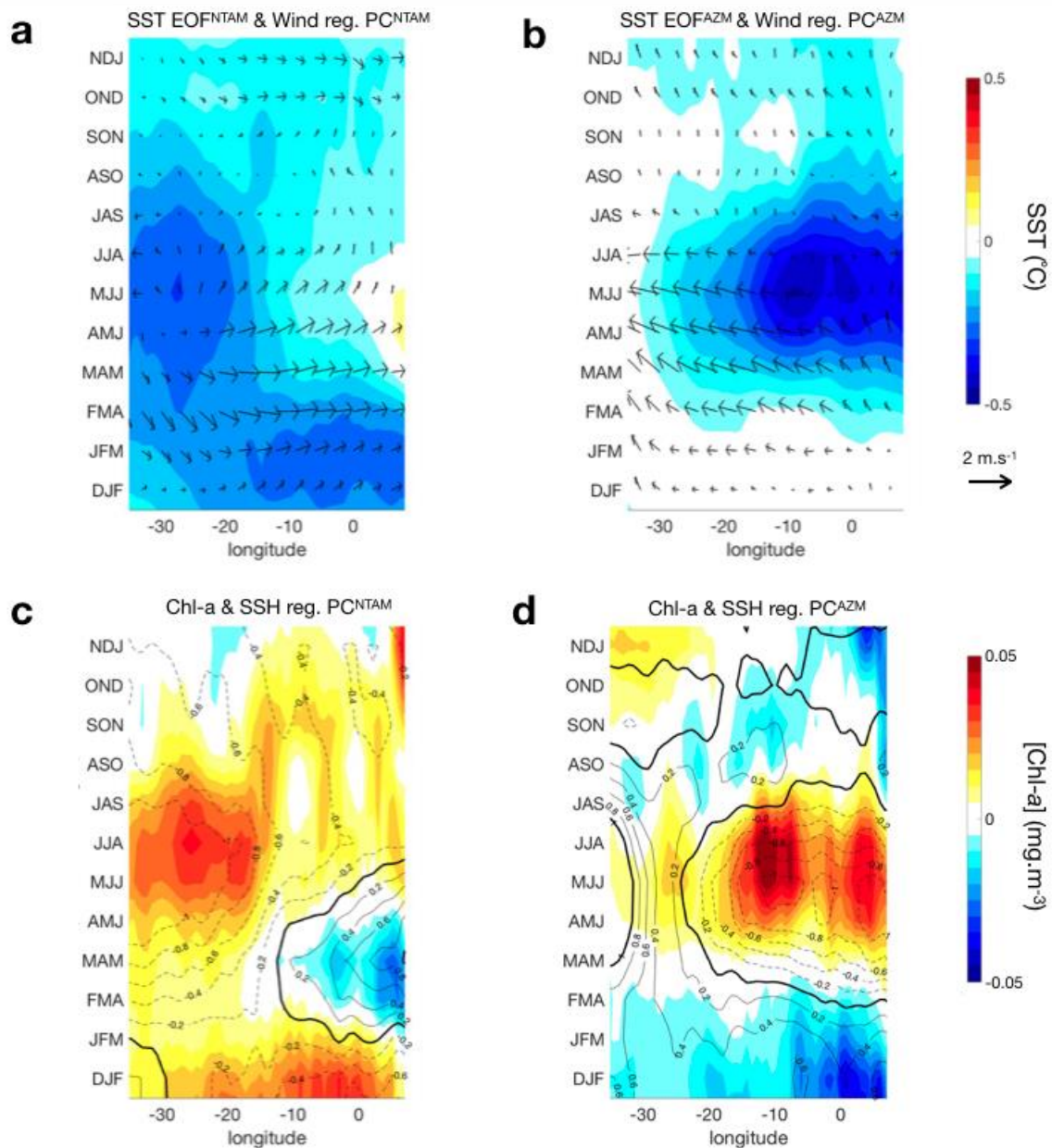
240 forced earlier in the western equatorial basin by the zonal wind fluctuations associated with
241 the AZM (Illig et al. 2004) (**Fig.2b**). Incidentally, part of the equatorial Kelvin wave energy
242 could also have been transmitted along the African coast as coastally-trapped waves, which
243 would explain the remaining negative SSHA depicted off Angola (Bachèlery et al., 2020).
244 These interannual SSHA fluctuations reflect a substantial shoaling (deepening) of the
245 thermocline in the eastern (western) tropical Atlantic. The rise of the thermocline would yield
246 a positive nutrient advection into the sunlit layer and should explain the positive surface Chl-a
247 anomalies in the eastern Cold Tongue and along the southern Angola Current (**Fig.3d**).

248 The interannual modulation of the AZM (PC^{AZM}) (**Fig.3f**) and the Chl-a averaged in the
249 whole Cold Tongue region are significantly correlated ($r=0.47$, $>95\%$). This correlation slightly
250 increases when only the eastern part of the Cold Tongue is considered ($r=0.56$, $>99\%$), with a
251 regression coefficient of $-0.10 \text{ mg}\cdot\text{m}^{-3}/^{\circ}\text{C}$, while the western Cold Tongue Chl-a variability is
252 not correlated with PC^{AZM} . Therefore, we conclude that the AZM interannual climate mode
253 significantly explains the interannual boreal summer Chl-a variability in the eastern equatorial
254 band with a strong positive event in 2005 and a negative event in 2008 (also visible in
255 **Figs.1cd**). Hormann and Brandt (2009) have also analyzed the abnormally cold year of 2005. In
256 contrast to 2002, during the boreal summer of 2005, the easterlies were stronger in the
257 western basin, and the Equatorial Under Current was embedded in a shallower thermocline
258 at 23°W . The thermocline was abnormally shallow in the whole eastern equatorial Atlantic
259 (Marin et al., 2009) and was associated with a weak equatorial Kelvin wave activity. Vallès-
260 Casanova et al. (2020) classified the year 2005 as a La Niña event, characterized by an early
261 onset and persistent cold anomalies along the equator, peaking at 10°W in June.

262

263 3.4 Spatio-temporal modulation under NTAM and AZM influence.

264 To diagnose the spatio-temporal modulation of the surface properties by the two leading
265 climate modes, we linearly regress the 3-month running-mean of SST, surface wind, SSH, and
266 Chl-a interannual anomalies onto the standardized PC^{NTAM} and PC^{AZM} (**Fig.3ef**). **Fig.4** presents a
267 longitude-time Hovmöller diagram of the regression coefficients averaged in the equatorial
268 band for the NTAM and AZM. A coherent longitude-time evolution of the key equatorial
269 properties anomalies emerges for both modes and can be interpreted in the light of the
270 climatological conditions (**Fig.S3**).



271 **Figure 4:** Phenology of equatorial surface properties associated with NTAM (left panels) and
 272 AZM (right panels) EOF modes of JJA interannual Tropical Atlantic SSTA. 3-month running-
 273 mean of SST, surface wind, SSH, and Chl-a interannual anomalies from preceding winter (DJF)
 274 to following winter (NDJ) are regresses onto the standardized PCs (shown in **Fig.3ef**) and
 275 averaged within [3°S-3°N]. SSTA (color shading, °C) and surface winds anomalies (arrows, m.s⁻¹)
 276 associated with JJA NTAM and AZM are displayed on top panels. On bottom panels, same
 277 for Chl-a (color shading, mg.m⁻³) and SSH (contours, cm) interannual anomalies.
 278

279 For the NTAM, the cold phase is associated with a cold equatorial SSTA in the western
 280 basin, that progresses from the end of the boreal winter. It extends from the Brazilian coast
 281 to ~15°W and peaks in early boreal summer (MJJ) at ~30°W, concomitant with anomalous
 282 wind divergence (**Fig.4a**). Collocated with this cold event in the western equatorial Atlantic,
 283 **Fig.4c** discloses a strong and persistent negative SSHA and a positive Chl-a anomaly in boreal

284 summer (MJJ and JJA). These physical and biogeochemical anomalies are concomitant and
285 steady, which suggests that they are associated with local vertical processes consistent with
286 the increased Ekman pumping diagnosed from the NTAM anomalous surface wind-stress curl
287 (**Fig.S2**). Concurrently, from boreal winter (JFM) to early summer (MJJ), relaxed easterlies are
288 observed along the whole equatorial band. They drive downwelling equatorial Kelvin waves
289 which reduce the equatorial upwelling in the eastern basin, leading to a strong zonal seesaw
290 pattern in SSH, SST, and Chl-a anomalies. Note that Hormann and Brandt (2009), who studied
291 a typical NTAM year (2002), also associated such a cold event with abnormally relaxed
292 easterlies in the western basin. They emphasized a strong Equatorial Under Current core
293 embedded in a shallow thermocline at 23°W which is consistent with the anomalous positive
294 Ekman Pumping triggered by the surface wind pattern discussed earlier. Interestingly, during
295 the following boreal fall, the cold anomaly slowly progresses eastwards east of 20°W,
296 accompanied by a negative SSHA and a positive Chl-a anomaly. Also, the NTAM boreal
297 summer western cold event is preceded by cold SSTA in the eastern basin in boreal winter
298 (DJF), associated with negative SSHA and positive Chl-a anomalies. This pattern is typical of
299 the boreal winter Atlantic Niño II mode from Okumura and Xie (2006), shown to act as
300 preconditioning to the subsequent boreal spring NTAM.

301 For the AZM, the cold phase is associated with a cold equatorial anomaly extending east
302 of ~20°W to the African coast. The cold event spreads slowly westwards from late winter and
303 is maximum in early boreal summer (MJJ) at ~10°W (**Fig.4b**, coherently with La Niña pattern
304 from Vallès-Casanova et al., 2020). This eastern cold anomaly is driven by an intensification of
305 easterly trades along the equator initiated in boreal winter (JFM) and persisting until the
306 following summer (JJA) (**Fig.4b**). Concomitant to this cold SSTA, a strong negative SSHA and a
307 positive Chl-a anomaly are depicted (**Fig.4d**). The latter can be attributed to the enhanced
308 upwelling of cold nutrient-rich deep water into the surface layer (Radenac et al., 2020), driven
309 by remotely-forced equatorial wave triggered by the reinforcement of easterly trades in the
310 western-central basin. This analysis also reveals a localized moderate positive Chl-a MJJ peak
311 at 25°W which seems to detach itself from the core eastern Chl-a anomaly and differs from
312 the SSTA pattern. After the summer, an abrupt termination of the interannual event (**Fig.4bd**)
313 is associated with an eastward propagation of moderate positive SSHA and negative Chl-a
314 anomalies that crosses the basin in ~2 months (from JAS to SON). The timing and
315 characteristics of this propagation are consistent with a Kelvin wave triggered by the

316 reflection of a downwelling Rossby wave forced upstream the relaxation of the easterlies.
317 Notably, similar to the NTAM signature, an eastern negative Chl-a anomaly associated with a
318 positive SSHA are observed at the beginning of the year. They may contribute to the
319 modulation of eastern boreal winter Cold Tongue bloom (also visible in **Fig.1bd**; Perez et al.,
320 2005; Grodsky et al. 2008; Radenac et al. 2020). The study of the winter bloom is beyond the
321 scope of this paper, but additional attention could be given to its interannual variability in
322 future research projects.

323

324 **4. Concluding Remarks**

325 This study aims to describe how interannual climate modes shape the Tropical Atlantic
326 Chl-a variability in time and space using a 21-year-long time-series remote-sensed
327 observations over the equatorial Atlantic region. As expected, two leading modes of
328 interannual climate variability emerge from our analysis of boreal summer (JJA) tropical
329 Atlantic SSTA: namely the AZM and NTAM. Both NTAM and AZM spatial patterns do not
330 strictly fit their canonical signature but are under the influence of a positive AMO phase
331 (1998-2018) (Martin-Rey et al., 2018).

332 We show that AZM and NTAM significantly drive the interannual surface Chl-a
333 variability in the equatorial Atlantic, with different timings and contrasted modulation in the
334 Cold Tongue. The NTAM influences preferentially the interannual Chl-a variability in the
335 western Cold Tongue, from boreal spring to fall (peaking in summer), while the AZM affects to
336 a large extent, the interannual Chl-a variability in the eastern Cold Tongue, during boreal
337 summer. Our results suggest that the significant surface wind anomalies in the western
338 equatorial Atlantic associated with each climate mode have distinct effects on the ocean
339 dynamics. The relaxation of easterlies in boreal winter associated with the cold phase of
340 NTAM yields a shoaling of the western equatorial thermocline through Ekman pumping, while
341 the AZM cold-phase intensification of easterlies in boreal spring triggers equatorial Kelvin
342 wave propagation responsible for an anomalous shoaling of the thermocline in the eastern
343 Cold Tongue region 1-2 months later. The locally- and remotely-forced vertical fluxes advect
344 nutrients into the euphotic layer and increase the primary production in the western and
345 eastern Cold Tongue region, respectively. This east/west contrast in the interannual Chl-a
346 fluctuations is confirmed by an EOF analysis on the Chl-a restricted over the 3°S-3°N
347 equatorial band (**Fig.S4**). The first leading mode is associated with a strong positive anomaly

348 along the equator (with concomitant peaks at 25, 18, and 10°W) with a temporal modulation
349 significantly correlated with PC^{NTAM} (r=0.53). The second leading mode highlights an east/west
350 dipole of Chl-a anomalies (peaking at 10°W and 30°W, respectively) with a temporal
351 modulation significantly correlated with PC^{AZM} (r=0.61). Additionally, similar east/west
352 partition of the equatorial Atlantic has already been documented (Longhurst, 2007; Perez et
353 al. 2005; Longhurst et al. 1995), but simply based on the seasonality of biogeochemical
354 properties, discriminating the variability in the Eastern Tropical Atlantic Province and the
355 Western Tropical Atlantic Province, associated with a large boreal summer variability and a
356 reduced boreal autumn peak, respectively.

357 This work provides new insights into the interannual variability of Chl-a, and its zonal
358 signature, in this under-explored region. The role of such climate variability on Chl-a is likely
359 to arise a myriad of biogeochemical implications. The challenge now is to understand how
360 these climate modes impact the interannual spatio-temporal variability of the planktonic
361 ecosystem, their consequences on the biological carbon pump and the higher trophic levels.
362 In-depth analyses using a coupled physical-biogeochemical model are under investigation to
363 diagnose biological-physical processes such as advective and diffusive nutrient fluxes and
364 their potential impact on subsurface Chl-a and plankton ecosystem dynamics.

365

366 Acknowledgements:

367 This work is part of the TRIATLAS European project (South and Tropical Atlantic climate-based
368 marine ecosystem prediction for sustainable management; H2020 grant agreement No
369 817578) which supported a post-doc fellowship funded by Institut de Recherche pour le
370 Développement (IRD-France). The reanalysis data were downloaded from ECMWF
371 (<http://www.ecmwf.int>).

372

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- 467 **Acronyms:**
- 468 AZM: Atlantic Zonal Mode
- 469 Chl-a: Surface Chlorophyll-a concentration
- 470 CT: Cold Tongue region ($[3^{\circ}\text{S}-3^{\circ}\text{N}; 30^{\circ}\text{W}-3^{\circ}\text{E}]$)
- 471 CTE: Eastern Cold Tongue area ($[3^{\circ}\text{S}-3^{\circ}\text{N}; 15^{\circ}\text{W}-3^{\circ}\text{E}]$)
- 472 CTW: Western Cold Tongue area ($[3^{\circ}\text{S}-3^{\circ}\text{N}; 27^{\circ}\text{W}-15^{\circ}\text{W}]$)
- 473 EOF: Empirical Orthogonal Function
- 474 NTAM: North Tropical Atlantic Mode
- 475 PDF: Probability Distribution Functions
- 476 SSH: Sea Surface Height
- 477 SSHA: Sea Surface Height interannual Anomalies
- 478 SST: sea Surface Temperature
- 479 SSTA: sea Surface Temperature interannual Anomalies
- 480 STD: standard deviation

Figure 1.

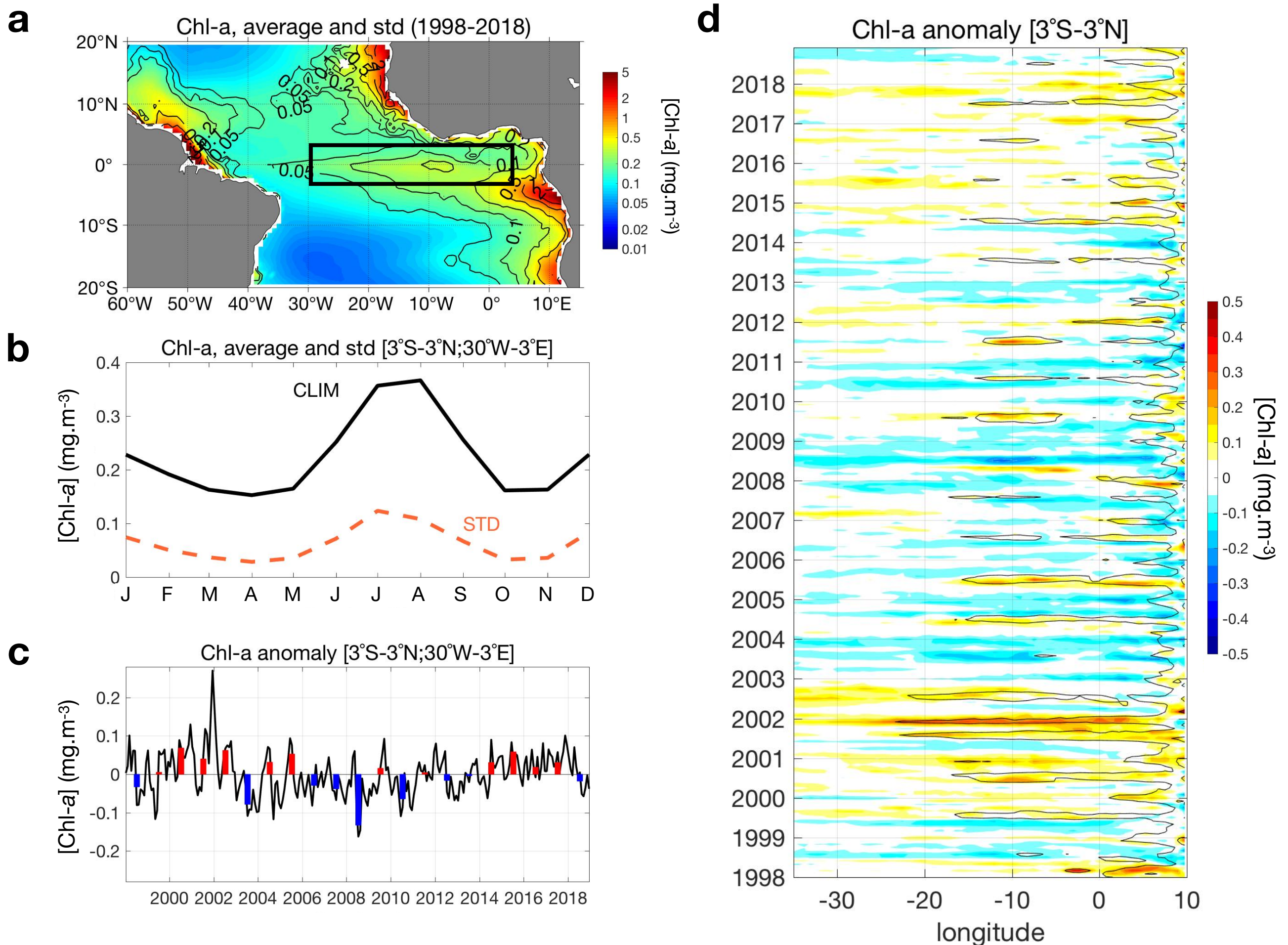


Figure 1: Temporal and spatial variability of surface chlorophyll-a concentration (Chl-a, $\text{mg}\cdot\text{m}^{-3}$) from satellite observations over 1998-2018. (a) Mean (color shading) and interannual standard deviation (black contours). (b) Climatological monthly mean (black line) and interannual monthly standard deviation (STD, orange dashed line) averaged over the Cold Tongue area ($[3^{\circ}\text{S}$ - 3°N ; 30°W - 3°E], delineated by the black box in panel a). (c) Monthly (black line) and boreal summer (June-July-August average, red/blue bars) interannual anomalies in the Cold Tongue. (d) Longitude-time Hovmöller diagram of the equatorial interannual anomalies averaged within $[3^{\circ}\text{S}$ - 3°N].

Figure 2.

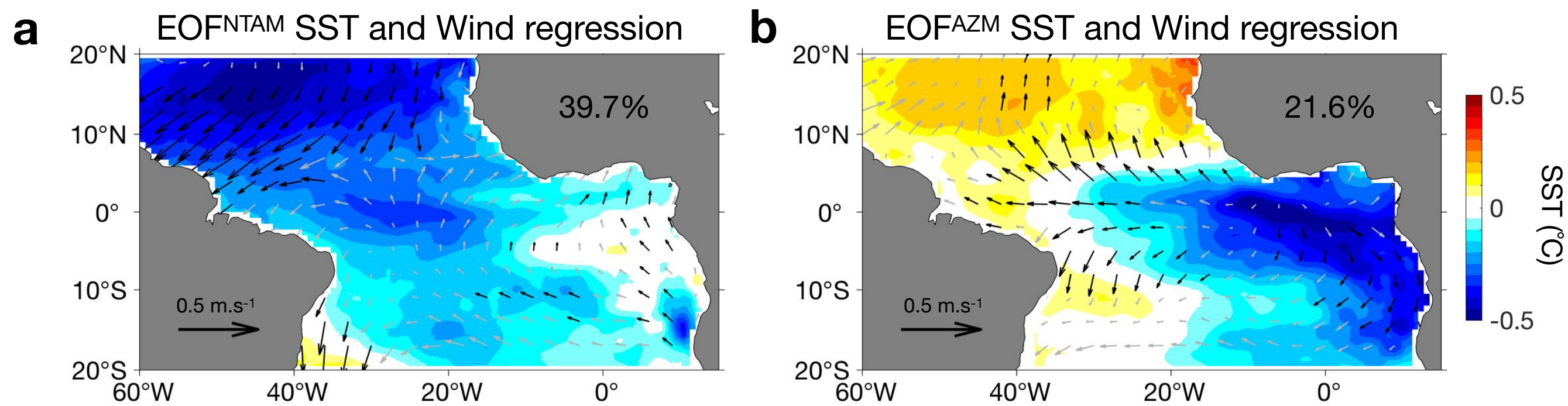


Figure 2: Maps of the two leading Empirical Orthogonal Function (EOF) of boreal summer 'June-July-August' (JJA) interannual Sea Surface Temperature (SST) variability (color shading, °C) in the tropical Atlantic ([20°S-20°N, 60°W-15°E]) over the 1998-2018 period. The percentages indicate the explained variance of each mode. JJA surface wind interannual anomalies (arrows, m.s⁻¹) are regressed onto the respective standardized Principal Components (PC, shown in Fig.3ef). Significant values (> 95% confidence level) are shown with black vectors. NTAM: North Tropical Atlantic mode. AZM: Atlantic Zonal mode.

Figure 3.

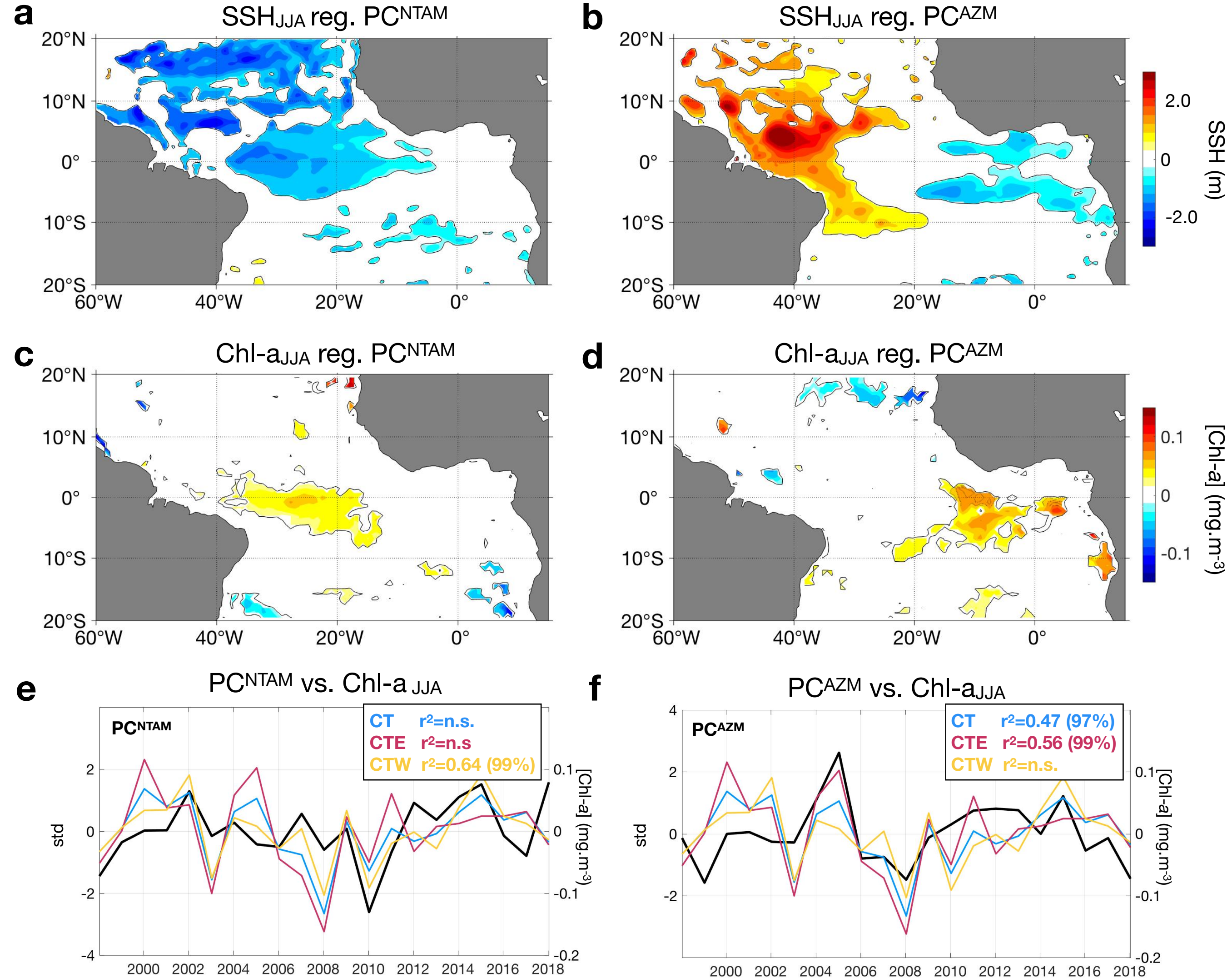


Figure 3: Spatio-temporal characteristics of the boreal summer (JJA) Sea Surface Height (SSH) and Chl-a associated with the NTAM and AZM interannual EOF modes of the Tropical Atlantic SST (presented in Fig.2). Regression maps of JJA SSH interannual anomalies (cm) onto the NTAM (a) and AZM (b) standardized Principal Components (PCs, shown in panels e-f (black lines)). Only significant values (> 95% confidence level) are shaded. (c-d) Same as panels a-b for JJA Chl-a anomaly (mg.m⁻³). NTAM (e) and AZM (f) EOF PCs (black lines) are compared with boreal summer surface Chl-a interannual anomalies averaged in different portions of the Cold Tongue: entire Cold Tongue (CT, [3°S-3°N; 27°W-3°E], blue lines), eastern Cold Tongue (CTE, [3°S-3°N; 15°W-3°E], pink lines), and western Cold Tongue (CTW, [3°S-3°N; 27°W-15°W], orange lines). Squared correlation (r^2) and associated significance level (%) quantify the goodness of the linear fit between PCs and Chl-a time-series. n.s.: not significant

Figure 4.

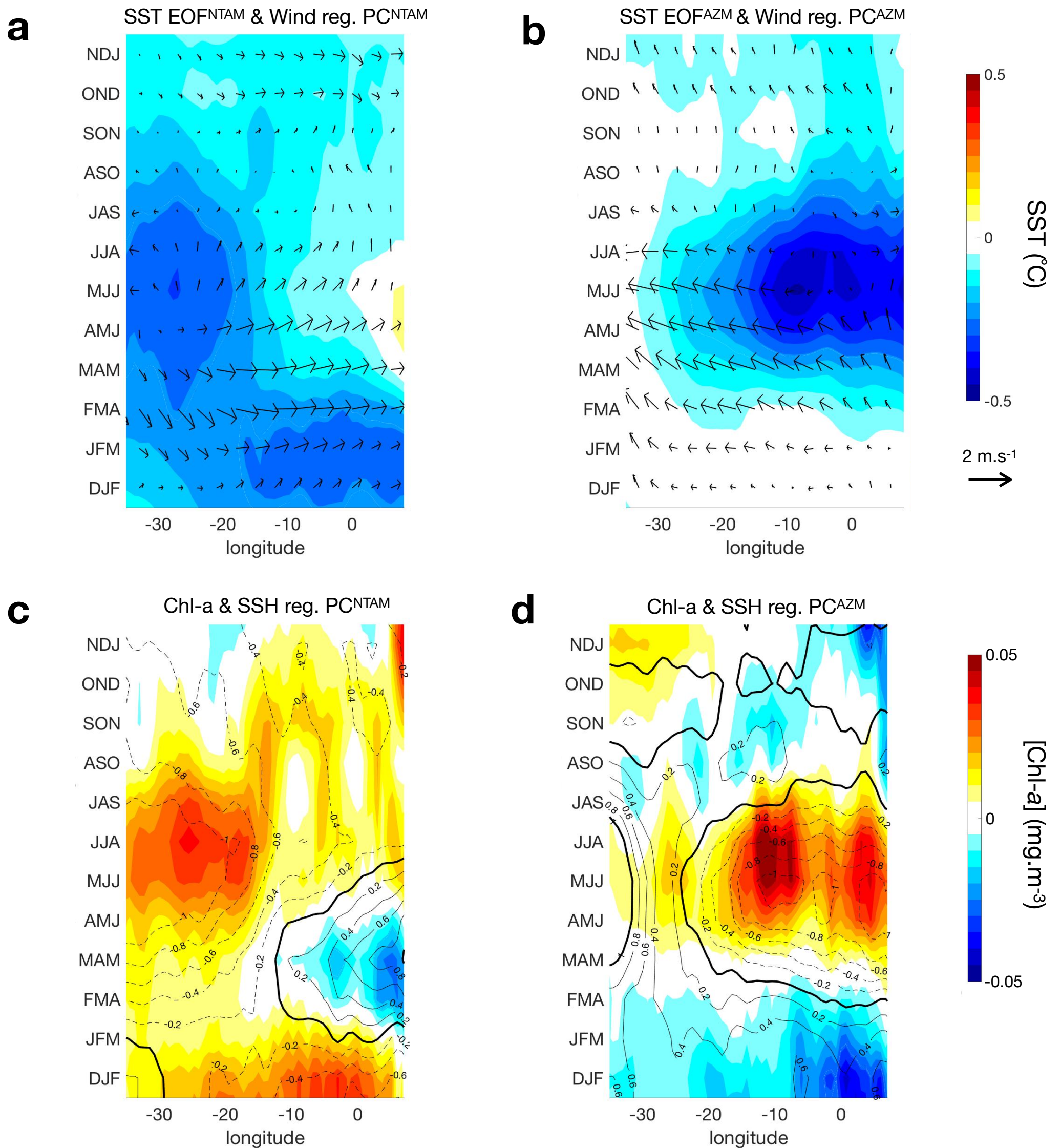


Figure 4: Phenology of equatorial surface properties associated with NTAM (left panels) and AZM (right panels) EOF modes of JJA interannual Tropical Atlantic SSTA. 3-month running-mean of SST, surface wind, SSH, and Chl-a interannual anomalies from preceding winter (DJF) to following winter (NDJ) are regressed onto the standardized PCs (shown in Fig.3ef) and averaged within [3°S-3°N]. SSTA (color shading, °C) and surface winds anomalies (arrows, m.s⁻¹) associated with JJA NTAM and AZM are displayed on top panels. On bottom panels, same for Chl-a (color shading, mg.m⁻³) and SSH (contours, cm) interannual anomalies.