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## **Weakening of the Senegalo-Mauritanian Upwelling System under climate change**

Adama Sylla, Juliette Mignot, Xavier Capet, Amadou Thierno Gaye

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## Abstract

Upwelling processes bring nutrient-rich waters from the deep ocean to the surface. Areas of upwelling are often associated with high productivity, offering great economic value in terms of fisheries. The sensitivity of spring/summer-time coastal upwelling systems to climate change has recently received a lot of attention. Several studies have suggested that their intensity may increase in the future while other authors have shown decreasing intensity in their equatorward portions. Yet, recent observations do not show robust evidence of this intensification. The senegalo-mauritanian upwelling system (SMUS) located at the southern edge of the north Atlantic system (12°N-20°N) and most active in winter/spring has been largely excluded from these studies. Here, the seasonal cycle of the SMUS and its response to climate change is investigated in the database of the Coupled Models Inter comparison Project Phase 5 (CMIP5). Upwelling magnitude and surface signature are characterized by several sea surface temperature and wind stress indices. We highlight the ability of the climate models to reproduce the system, as well as their biases. The simulations suggest that the intensity of the SMUS winter/spring upwelling will moderately decrease in the future, primarily because of a reduction of the wind forcing linked to a northward shift of Azores anticyclone and a more regional modulation of the low pressures found over Northwest Africa. The implications of such an upwelling reduction on the ecosystems and local communities exploiting them remains very uncertain.

**Keywords:** upwelling, climate change, climate models, Northeastern Tropical Atlantic

## 31 **1. Introduction**

32 The upwelling is an upward motion of sea water from intermediate depths (typically 50-200m)  
33 toward the ocean surface. It is an oceanographic phenomenon resulting from the friction of the wind  
34 on the ocean surface. Upwelled water masses are colder and richer in nutrients than the surface  
35 waters they replace. Upwellings therefore correspond to areas of very productive marine  
36 ecosystems and high fish resources. When the upwelling occurs along the coast, thus near fishermen  
37 and societies, their economic importance is very high (Gómez-Gesteira et al. 2008) . There are  
38 four major coastal upwelling systems (hereafter EBUSs for Eastern Boundary Upwelling Systems)  
39 in the global ocean, that are the Canary, Benguela, Humboldt and California systems. These areas  
40 cover less than 1% of the global ocean surface, but they contribute to about 8% of the global marine  
41 primary production and more than 20% of the global fish catches (Pauly and Christensen 1995) .  
42 EBUSs are situated in the tropics and subtropics. There, the trade winds blowing equator-ward  
43 parallel to the eastern borders of the ocean basins induce an Ekman transport from the coast to the  
44 open ocean, perpendicular to the wind stress forcing. This creates a transport divergence and  
45 thereby leads to an upwelling at the coast. This mechanism has been long considered as the main  
46 phenomenon that drive the upwelling systems on Earth (Sverdrup 1938). Yet, a divergent oceanic  
47 circulation may also be created at the surface by a cyclonic wind stress curl. In the eastern  
48 subtropical basins, the tendency for trade winds to slow down near the coast, the so-called wind  
49 drop-off (e.g. Bakun and Nelson 1991; Pickett 2003) induces a positive wind stress curl that also  
50 contributes to upwelling. The upwelling contributions of coastal Ekman divergence and Ekman  
51 pumping are difficult to compare. The former effect is strongly localized nearshore while the latter  
52 is more broadly distributed. Their relative importance thus depends on the choice of an across-shore  
53 length scale for how far offshore Ekman pumping is being considered that is often largely arbitrary

54 and studies do not agree yet (Nelson 1976; Halpern 2002; Pickett 2003; Bravo et al. 2016; Torres  
55 2003; Capet et al. 2004).

56         Among the four EBUSs mentioned above, we concentrate here on the southern part of the  
57 so-called Canary Upwelling System (CUS). This upwelling extends from the coast of West Africa  
58 at  $\sim 10^{\circ}\text{N}$  to that of the Iberian peninsula around  $45^{\circ}\text{N}$  (Aristegui et al. 2005). A comparison of the  
59 four major EBUSs shows that the northwest African coast is the most spatially and seasonally  
60 variable one in terms of primary production (Carr and Kearns 2003). This variability is still  
61 insufficiently documented and understood despite a number of recent studies including (Mason et  
62 al. 2011; Ndoye et al. 2014; Benazzouz et al. 2014; Faye et al. 2015). Based on arguments of  
63 seasonality, Aristegui et al. (2009) proposed that this system could be divided into 5 sub-regions. In  
64 the north, the Galician coast ( $42^{\circ}\text{N}$ - $45^{\circ}\text{N}$ ) and the Portuguese coast ( $37^{\circ}\text{N}$ - $42^{\circ}\text{N}$ ) are both  
65 characterized by a summer-time upwelling when the wind stress forcing reaches its maximum  
66 intensity. Further south, in the region of the Gulf of Cadiz ( $33^{\circ}\text{N}$ - $37^{\circ}\text{N}$ ) and the Moroccan coast  
67 ( $21^{\circ}\text{N}$ - $33^{\circ}\text{N}$ ) upwelling presents a reduced seasonality and is accompanied by the southward branch  
68 of the subtropical gyre, namely the offshore Canary current. At the southern end of the CUS the  
69 strength and position of the wind is highly influenced by the latitudinal migration of the  
70 Intertropical Convergence Zone (ITCZ) and the associated Azores high pressure area, both  
71 oscillating between their northernmost and southernmost positions in summer and winter  
72 respectively (Fig.1). This oscillation of pressure systems generates seasonal wind and SST  
73 fluctuations (Mittelstaedt 1991; Nykjær and Van Camp 1994). In winter, the ITCZ reaches its  
74 southernmost position between the geographic equator and  $5^{\circ}\text{S}$  thereby favoring strong northeastern  
75 trade winds along the coast of Guinea, Senegal and Mauritania ( $12^{\circ}\text{N}$ - $20^{\circ}\text{N}$ ) and thus the  
76 upwelling. This is largely responsible for the cold surface waters seen along the coast in Fig. 1. In  
77 summer, when the ITCZ reaches its northernmost position (around  $15^{\circ}\text{N}$ ), the winds in this region

78 are weak, and even reverse to south westerlies in the southernmost part of the region. In this season,  
79 the upwelling is absent along the coast of Senegal and much reduced offshore of Mauritania. In  
80 Senegal, this season is also marked by heavy monsoon rainfalls. The Senegalo-Mauritanian  
81 Upwelling System (SMUS) has thus a very specific seasonal behavior as compared to the northern  
82 part of CUS. This region is the most productive of the five sub-regions. It is the focus of our study.

83 In 1990, Bakun suggested that an increase of the greenhouse gases concentration in the  
84 atmosphere would lead to an intensification of upwelling winds in EBUSs. Given the strong  
85 linkages between upwelling and marine ecosystems (Blanchette et al. 2008; Fenberg et al. 2015),  
86 such evolution may have very important ecological implications. Precisely, Bakun (1990)  
87 hypothesized that inhibition of nighttime cooling and enhancement of daytime heating on land as a  
88 result of global warming would lead to the intensification of the continental thermal lows adjacent  
89 to upwelling regions. This intensification should lead to increased onshore-offshore atmospheric  
90 pressure gradients, accelerated alongshore winds, and thus intensified coastal upwelling circulation.  
91 This generic mechanism could in principle be applied to any of the coastal upwelling systems. In  
92 the CUS, several analyses have tested this hypothesis. Paleoclimatic reconstructions in the coastal  
93 upwelling area off Cape Ghir (30.5°N) have indeed shown a significant cooling of surface waters  
94 around the end of the 20<sup>th</sup> century, which may reflect an increase of the upwelling intensity  
95 (McGregor et al. 2007). Examining the variation of the northern part (near 30.5°N) of the CUS  
96 using SST observations datasets and meridional wind observations and reanalysis, Narayan et al.  
97 (2010) also concluded that the coastal upwelling intensity has been increasing over the second  
98 half of the 20<sup>th</sup> century (1960–2001). However other studies are at odds with Bakun (1990)'s  
99 conceptual view. In particular, Gómez-Gesteira et al. (2008) noted a long-term decrease of the CUS  
100 upwelling intensity between 20°N-32°N from 1967-2006 using the AVHRR SST and NOAA time  
101 series of zonal Ekman transport. Using reanalysis data sets, Pardo et al. (2011) and Santos et al.

102 (2012) have reached similar conclusions for the region spanning from the Iberian coast to Morocco.  
103 Finally, using different wind datasets and reanalyses, Barton et al. (2013) have found no statistically  
104 significant change of the annual mean wind intensity off Northwest Africa over the second half of  
105 the 20<sup>th</sup> century. Using a NCEP/NCAR reanalysis data for 60 years (Sydeman et al. 2014) found  
106 that summertime winds have intensified in the California, Benguela, and Humboldt upwelling  
107 systems and weakened in the Iberian system and this mentioned this change is equivocal in the  
108 Canary system.

109 The robustness of these previous results is most of the time limited by the lack of  
110 sufficiently long and continuous series of observations. In this context and despite their imperfect  
111 representation of fine oceanic structures, climate models provide an interesting alternative to  
112 explore the consequences of climate change on coastal upwellings. Mote and Mantua (2002) have  
113 used simulations of the 20<sup>th</sup> and 21<sup>st</sup> century (1990-2080) based on the NCAR-CSM (Boville and  
114 Gent 1998) and the HadCM3 (Gordon et al. 2000) climate models to compute upwelling indices for  
115 the four EBUSs including the CUS (over the latitude range 13°N –30°N). They did not find any  
116 significant change in the summer upwelling intensity and phasing of the seasonal cycle. More  
117 recently, Wang et al. (2015) have analyzed historical and future simulations of 22 CMIP5 Earth  
118 system models over the period 1950-2099. Focusing again on the summer season and the offshore  
119 wind-driven Ekman transport, they found that the CMIP5 models exhibit a strengthening of the  
120 upwelling intensity in the poleward sectors of all EBUSs except the California current system. On  
121 the other hand, they found a weakening of the *summertime* upwelling intensity at lower latitudes,  
122 that is between ~15°N and 25°N for the CUS. A distinction between the northern and southern parts  
123 of the CUS was also made in Cropper et al. (2014) for the recent period based on atmospheric  
124 reanalyses and in the meta-analysis of Sydeman et al. (2014). According to Wang et al. (2015) and  
125 as opposed to Bakun 's hypothesis, an increase in the land-sea thermal difference in this region in

126 summer is expected to strengthen the southwesterly monsoon circulation that drives downwelling-  
127 favourable winds in the subtropics. All these studies only relate to the summer season, when there is  
128 little upwelling along the senegalo-mauritanian coast.

129 To our knowledge, the future of the SMUS has not been investigated for the cold  
130 upwelling season, in spite of its primary importance to the functioning of the ecosystem (Zeeberg et  
131 al. 2008; Arístegui et al. 2009). In this paper, we study this system and its future, using climate  
132 models. In spite of their rather coarse resolution and biases in the tropical North Atlantic (Richter  
133 and Xie 2008; Wahl et al. 2011; Richter et al. 2014) such simulations (presented in section 2) are  
134 the only available tool to investigate the upwelling's future. We will define a number of indices to  
135 characterize the intensity of upwelling process and its thermal signature in such large scale models,  
136 and we will investigate the representation of these indices in the different simulations under present-  
137 day conditions. This approach will allow us to propose a first comparison of the different models  
138 assessing their ability to represent west African upwelling dynamics and its signature on SST  
139 (section 3). The response of upwelling indices to global warming is analyzed in section 4.  
140 Discussion and conclusion are offered in sections 5 and 6 respectively.

## 141 **2. Data**

### 142 2.1. Model data

143 This study is based on the CMIP5 (Coupled Model Inter-comparison Project Phase 5) data base.  
144 This database has already been extensively used for oceanic and climatic studies in the eastern  
145 tropical North Atlantic. Several authors have highlighted the general warm SST bias in this region  
146 (Breugem et al. 2008; Richter and Xie, 2008; Grodsky et al. 2012; Wahl et al. 2011) potentially due  
147 to a surface wind bias during spring (March-May) season (Chang et al. 2007; DeWitt, 2005; Richter  
148 et al. 2012) and the poor-representation of low levels clouds (Huang et al. 2007). In order to  
149 investigate how the seasonality and intensity of SMUS may change in the future, we have compared



150 the output of historical simulations with simulations of the 21<sup>st</sup> century for the RCP8.5 scenario.  
 151 The monthly sea surface temperature is available, for the historical period in 47 simulations (Table  
 152 1). In terms of oceanic variables, we have also used the sea surface height above geoid, the upward  
 153 ocean mass transport, the upper-ocean mixed layer depth and the atmospheric wind stress seen by  
 154 the ocean. When the latter was not available, we have used the atmospheric surface wind velocity  
 155 instead, which we converted offline into a wind stress following the procedure described below.  
 156 Regarding atmospheric variables, we have also used the sea level pressure and the precipitation rate.  
 157 Variables availability is indicated in Table 1 for each climate model that has been retained. The  
 158 reader is referred to the CMIP5 data base for more information on each of these models. For each  
 159 variable, we have constructed the average seasonal cycle over the period [1985- 2005] (hereafter  
 160 “present”) and the period [2080-2100] (hereafter “future”) from the historical (Taylor et al. 2012)  
 161 and the RCP8.5 (Riahi et al. 2011) simulations respectively. When several ensemble members are  
 162 available, we only selected the first one, in order to assign the same weight to all the models. To  
 163 facilitate the analyses, all outputs have subsequently been interpolated over a regular 1° grid using a  
 164 bilinear interpolation method as in Rykaczewski et al. (2015).

165 Some model (BCCBCC-CSM1-1, BCC-CSM1-1-m, HadCM3, HadGEM2-CC, HadGEM2-ES,  
 166 GISS-E2-H-CC, GISS-E2-R-CC, GISS-E2-H, GISS-E2-R) only provide near-surface winds, and  
 167 not the wind stress at the air-sea interface. The relationship between the wind speed and the wind  
 168 stress is usually written as a bulk formula under the form

$$169 \quad \tau_x = \rho_a C_d (uas^2 + vas^2)^{1/2} uas \text{ and } \tau_y = \rho_a C_d (uas^2 + vas^2)^{1/2} vas \text{ (1)}$$

170 with  $uas$  and  $vas$  the zonal and meridional wind components respectively,  $C_d$  the drag coefficient  
 171 and  $\rho_a$  the air density ( $\rho_a = 1.22 \text{ kg.m}^{-3}$ ). For the simulations listed above, the wind stress was  
 172 computed using Eq. (1) and  $C_d = 0.0014$  following Santos et al. (2012) (see appendix A for more  
 173 details). In these cases, the derived wind stress is used in both the historical and future simulations

174 to minimize biases coming from this offline computation. For all the other models, the oceanic  
175 wind stress provided in the data base is used. Note that CanCM4 and LASG-CESS models provide  
176 neither the wind stress nor the near-surface wind, even over the historical period. All missing data  
177 are marked with a dark blue band over Fig. (4, 5, 6) and a missing bar in Fig. (7, 8, 9, 10). Note also  
178 that not all models provide the oceanic mixed layer depth (see table 1). Given the potentially critical  
179 dependency of our results on estimations of the MLD, we only show computations involving this  
180 variable for models which provide it directly in the database. In other words, we do not attempt to  
181 recompute any mixed layer depth.

182 The Multi-Model Mean (MMM) is also systematically given for each index over both the historical  
183 and future period. The number of models incorporated in the calculation of a given MMM index  
184 varies according to availability of the input variables for that index, which can differ for present and  
185 future. When relevant, the significance of the sign of the MMM is tested statistically according to  
186 the number of available models and their spread (t-test of the mean considering that each model is  
187 independent). The significance of the changes in the future is also tested using a t-test that considers  
188 each available model as independent. Each significance is given at the 95% confidence level.

189 The analyses presented below frequently considers two sub-domains of the SMUS: a  
190 northern subdomain between 16°N and 20°N (hereafter nSMUS for “northern part of SMUS”) and  
191 a southern subdomain between 12°N and 15°N (hereafter sSMUS for “southern part of SMUS”).  
192 15°N is the latitude of Cape Verde. The latter is a well-known geomorphological irregularity with  
193 known dynamical implications in the ocean (Roy 1998; Alpers et al. 2013; Ndoye et al. 2014) and  
194 (Kounta et al. 2018). This separation nevertheless arose naturally from our analyses for reasons that  
195 will become progressively apparent to the reader. In particular, we find that the late 20<sup>st</sup> century  
196 upwelling projections differ for the two sub-domains.

197 2.2: Validation data sets

198 In order to evaluate the model outputs over the historical period, we have used several observation  
199 and reanalysis data sets. As for the climate models outputs, all the validation datasets have been  
200 interpolated on a regular  $1^\circ$  grid, when not originally provided on this grid. Except otherwise  
201 specified (Quikscat and AVISO data sets, see below), we use monthly climatologies built over the  
202 1985-2005 time period, i.e., consistent with the period defining present conditions in the climate  
203 simulations.

204 We have used the Extended Reconstructed Sea Surface Temperature data set (ERSST- v3b,  
205 Smith and Reynolds 2003) produced by NOAA at  $2^\circ$  spatial resolution. In order to account for  
206 uncertainty on SST observations, we have also used the gridded SST data set from the Met Office  
207 Hadley CentreHadISST (Rayner 2003). This data is provided on a  $1^\circ$  latitude-longitude grid from  
208 1870 to present.

209 To evaluate the model wind stresses, we have used the  $0.25^\circ$  resolution Quikscat wind stress  
210 climatology for the period 2000-2009 extracted from the <https://podaac.jpl.nasa.gov> database. The  
211 period 2000-2009 is approximately that of the satellite mission. It does not exactly match the  
212 “present” period considered in the climate models and this adds a slight uncertainty in the  
213 comparison with the climate models outputs. However, the inter-models differences shown below  
214 are larger than possible differences between the 1985-2005 and 2000-2009 wind stress  
215 climatologies. These direct observations have been compared to the TropFlux reanalysis. This data  
216 set combines the ERA-Interim reanalysis for turbulent and long-wave fluxes, and ISCCP  
217 (International Satellite Cloud Climatology Project) surface radiation data for shortwave fluxes. This  
218 wind stress product is described and evaluated in Praveen Kumar et al. (2013).

219 Meridional sea surface height (SSH) gradients play an important dynamical role in the SMUS  
220 as in other coastal regions (see Sec.3.3). To evaluate the models representation of SSH along the  
221 coast, we use the AVISO satellite altimetry product ([www.aviso.altimetry.fr](http://www.aviso.altimetry.fr); Ducet et al. 2000).

222 This data covers the period 1993-2013 at  $0.25^\circ$  spatial resolution. For comparison, we have also  
223 used the GODAS monthly SSH (<https://www.esrl.noaa.gov/psd/gridded/data.godas.html>), a  
224 monthly reanalysis provided at  $0.333^\circ$  latitude x  $1^\circ$  longitude of resolution. Again these two data  
225 sets are averaged over the period [1993-2005] and interpolated over a  $1^\circ$  longitude-latitude regular  
226 grid. Quantifying the effect of the SSH gradient on the geostrophic transport requires an estimation  
227 of the oceanic mixed layer depth (MLD). We use the climatology from de Boyer Montégut et al.  
228 (2004) available at the spatial resolution of  $2^\circ$ .

229 To analyse the influence of the large scale atmospheric circulation on the senegalo-mauritanian  
230 upwelling over the recent period, we have used sea level pressure fields from ERAI, originally  
231 available at  $0.75^\circ \times 0.75^\circ$  resolution (Dee et al. 2011). We have also used the monthly precipitation  
232 observations of the Global Precipitation Climatology Project available at  
233 <https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html>. The GPCP data cover the period 1979  
234 to 2017 at  $2.5^\circ$  resolution. These two fields are used to determine the mean latitudinal position of  
235 the ITCZ and the Azores Anticyclone (AA) during the upwelling season. The former corresponds to  
236 the average latitude of the precipitation maximum between November and May and between  $15^\circ\text{W}$   
237 and  $30^\circ\text{W}$ . The latter corresponds to the average latitude of the SLP maximum between November  
238 and May and between  $15^\circ\text{W}$  and  $30^\circ\text{W}$ .

### 239 **3. Characterization of the upwelling in climate models over the historical** 240 **period**

241 Our main objective is to determine the upwelling trends due to climate change in the SMUS. A  
242 prerequisite is to define indices that quantify upwelling intensity (and its changes over time) in the  
243 climate simulations despite the inability of the models to represent the fine-scale underlying process  
244 with realism, at least nearshore. In this section we propose a series of five indices, with varied  
245 complexity. Such an approach based on simple indices computed for CMIP5 simulations has been

246 applied successfully for example by Bellenger et al. (2014) in the context of ENSO. Note that in the  
247 following, brackets  $\langle \rangle$  denote a spatial average of the corresponding index, either along the  
248 longitude and/or the latitude.  $\langle \rangle_n$  and  $\langle \rangle_s$  correspond to spatial averages over the northern and  
249 southern SMUS sectors (nSMUS and sSMUS) delimited by the latitudes bands 16-20°N and 12-  
250 15°N respectively.

### 251 3.1 SST based characterization of upwelling

#### 252 3.1.a Amplitude of seasonal cycle ( $UI_{sst}^{seas}$ )

253 The annual cycle is generally the dominant timescale of temperature variability. In the tropics,  
254 however, seasons are less marked than at mid to high latitudes, and variability at this time scale is  
255 thus less energetic (e.g. Wang et al. 2015). As explained above, the SMUS is subjected to winter-  
256 time coastal cooling and it is therefore a subregional exception regarding seasonality. This feature is  
257 illustrated in Fig.2, which shows the amplitude of the SST seasonal cycle averaged zonally between  
258 16°W and 20°W (noted  $\langle UI_{sst}^{seas} \rangle$ ) for both SST data sets (first two columns), each individual  
259 simulations, as well as for the multi model mean of CMIP5 historical simulations (last column). The  
260 magnitude of the seasonal cycle is maximum between 12°N to 20°N in the two validation datasets  
261 because the seasonal upwelling contributes to wintertime cooling. This latitudinal range stands out  
262 between the northern sector (north of 20°N), where summer-time upwelling dominates and tends to  
263 compensate the effect of the solar flux seasonal cycle, and the southern sector (south of 12°N)  
264 where upwelling is very weak or absent.

265 The individual simulations generally reproduce an intensified  $\langle UI_{sst}^{seas} \rangle$  in the SMUS latitude  
266 band, but the maximum intensity and exact latitudinal position differ. The amplitude of  $\langle UI_{sst}^{seas} \rangle$  in  
267 the models ranges from roughly 3°C (IPSL-CM5B-LR) to more than 8°C (GFDL models). In

268 several models (BCC models, CSIRO-QCCCE, CSIRO-BOM, CMCC-CESM, CCSM4, EC-  
269 EARTH, LASG-IAP, LASG-CESS, MRI family, MIROC5, NorESM1 group)  $\langle UI_{sst}^{seas} \rangle$  is amplified  
270 only in the nSMUS and this feature affects the MMM. Conversely, marked seasonalities even extend  
271 north of 20°N in the FIO-ESM, the GFDL family and MIROC-ESM.

272 Only a few models appear to have  $\langle UI_{sst}^{seas} \rangle$  with realistic amplitude in the correct latitudinal band  
273 (12°N-20°N roughly). With respect to this criterion, the most realistic models appear to be CCCma,  
274 CMCC-CM, CMCC-CMS, CNRM, the HadGEM2 family and LASG-IAP. The CMIP5 Multi  
275 Model Mean (MMM) exhibits a realistic magnitude of the SST seasonal cycle, with a maximum of  
276 5°C to 6.5°C around 15°N. This is slightly weaker than observed, and this weakness is more  
277 pronounced in the sSMUS.

### 278 3.1.b: SST Upwelling Index ( $UI_{sst}^{cross}$ )

279 The difference of SST between the coast and the offshore ocean has been widely used to  
280 characterize upwelling intensity (e.g., Speth 1978; Mittelstaedt 1991; Santos et al. 2005; Gómez-  
281 Gesteira et al. 2008; Lathuilière et al. 2008; Marcello et al. 2011; Benazzouz et al. 2014). It is  
282 generally referred to as the SST Upwelling Index and abbreviated here as  $UI_{sst}^{cross}$ . Gómez-Gesteira  
283 et al. (2008) argue that  $UI_{sst}^{cross}$  lacks robustness because it is sensitive to a variety of processes and  
284 in particular the intensity of heat fluxes warming upwelling waters as they reach the surface.  
285 Remote modes of climate variability such as ENSO can also result in  $UI_{sst}^{cross}$  changes that are not  
286 related to coastal upwelling dynamics. Despite these caveats,  $UI_{sst}^{cross}$  is useful to measure the impact  
287 of upwelling on the SST zonal structure and to characterize the temporal variability of the system.  
288 For each latitude, we define the coastal SST ( $SST_{coast}$ ), as the SST at the ocean grid box closest to  
289 the coast (black points in Fig.1). The offshore SST ( $SST_{ocean}$ ) is the SST of the grid box located 5°  
290 away from the coast at the same latitude (magenta points in Fig.1).  $UI_{sst}^{cross}$  is then defined as:

$$UI_{sst}^{cross} = SST_{ocean} - SST_{coast} \quad (2)$$

Upwelling conditions thus correspond to positive values of this index. The 5° spacing between the coastal and offshore SST has been chosen following previous studies (eg. Cropper et al. 2014). Spacings of 7° or 9° have also been tested with no significant differences in the results (not shown).

Fig.3 (first two columns) shows the seasonal evolution of  $UI_{sst}^{cross}$  averaged in latitude bands over the nSMUS (noted  $\langle UI_{sst}^{cross} \rangle_n$ ) and the sSMUS ( $\langle UI_{sst}^{cross} \rangle_s$ ) over one climatological year in the two observation data sets. The general patterns are similar for the two datasets and consistent with previous studies (Nykjær and Van Camp 1994; Lathuilière et al. 2008; Santos et al. 2012).

In the nSMUS (Fig.3.a), the coastal upwelling is marked by positive winter values from November to June well apparent in both data sets (first two columns). In this latitude band, roughly two-third of the models exhibit a change of sign between summer and winter, with a consistent positive  $\langle UI_{sst}^{cross} \rangle_n$  in winter. On the contrary, some models (CanCM4, CanESM2, CMCC-CM, CMCC-CMS, HadGEM2-CC, HadGEM2-ES, HadGEM2-AO and IPSL-CM5A-LR) show positive values all year long, albeit for most of them with relatively weak values in summer indicating that the latitude band of permanent upwelling is displaced to the south compared to observations (where it is restricted to the north of Cape Blanc, 20°N). In spite of these differences, the multi-model mean (MMM) is remarkably consistent with the observations. It is significantly different from zero at the 95% of confidence level from November to May according to the student test with respect to the model configurations spread.

In the sSMUS (Fig.3b), the positive values of the  $\langle UI_{sst}^{cross} \rangle_s$  are only observed from December to May in both data sets. The models perform poorly in this latitudinal range. The multi-model mean is negative all year round, albeit with very weak values in winter, and only significantly different from zero during summer again as a consequence of the spread of the models behaviors.

314 The models tend to show either year-long negative  $\langle UI_{sst}^{cross} \rangle_s$  (no surface signature of the upwelling,  
 315 CESM group, FIO-ESM GFDL, GISS), a year long positive  $\langle UI_{sst}^{cross} \rangle_s$  (permanent signature of the  
 316 upwelling CNRM, LASG, NorESM1 families), or even a seasonality with the reversed sign (MPI  
 317 family). The HadGEM2-CC, HadGEM2-ES and HadGEM2-AO simulations stand out as the only  
 318 ones with a correct seasonality, yet with some deficiencies in terms of phasing and amplitude.

319 To conclude,  $\langle UI_{sst}^{cross} \rangle$  appears as a challenging indicator of the surface thermal signature of the  
 320 upwelling intensity. Specifically, the models fail to reproduce its seasonality in the southernmost  
 321 region of SMUS, probably due to a bias of the ITCZ position. The origin of this bias is further  
 322 explored below, in relation to the seasonality of the wind intensity and direction along the west  
 323 African coast. The diffuse nature of coastal upwelling in low-resolution models is also probably a  
 324 source of difficulty (see Discussion section 5).

### 325 3.2 Wind-based characterization of upwelling

#### 326 3.2.a: Contribution for the Ekman transport coastal divergence ( $UI_{wind}^{div}$ )

327 Assuming that the senegalo-mauritanian coast is orientated meridionally, which is nearly true in the  
 328 climate models, the Ekman transport can be quantified as:

$$329 \quad UI_{wind}^{div} = \frac{\tau_y}{\rho_w f} \quad (3)$$

330 where  $\tau_y$  is the meridional wind stress component at the grid box closest to shore,  $\rho_w$  the sea water  
 331 density, and  $f$  is the Coriolis parameter.  $\rho_w$  is chosen equal to  $1025 \text{ kg.m}^{-3}$  for our region of interest.

332  $UI_{wind}^{div}$  is expressed in  $\text{m}^2.\text{s}^{-1}$  and upwelling conditions correspond to positive  $UI_{wind}^{div}$ .

333 Fig.4 shows the climatological cycle of  $UI_{wind}^{div}$  averaged over the nSMUS ( $\langle UI_{wind}^{div} \rangle_n$ ) and  
 334 sSMUS ( $\langle UI_{wind}^{div} \rangle_s$ ) for the validation data sets, each model configuration, and the multi-model



335 mean. The seasonality of  $\langle UI_{wind}^{div} \rangle_n$  (Fig.4a) is modest: the wind index shows a seasonal reduction in  
336 summer, but without any change of sign. It nevertheless reaches such weak values that upwelling  
337 dynamics may become secondary for the SST budget, hence a plausible reason for the absence of  
338 upwelling signature on coastal SSTs as seen above. Most models appropriately capture this  
339 characteristic, albeit with a tendency to overestimation. A notable exception is the GFDL family for  
340 which the ITCZ may migrate too far to the north in summer, thereby leading to a wind seasonal  
341 cycle that is more strongly marked in nSMUS than in the other ones and the data. These models  
342 were also characterized by a systematic northward shift of the SST seasonal cycle pattern  
343 ( $\langle UI_{sst}^{seas} \rangle$ , Fig.2). As for the MMM, it is very close to the observations and it is significant at the  
344 95% level for all months of the year.

345 The seasonal cycle of  $\langle UI_{wind}^{div} \rangle_s$  (Fig.4.b) is more clearly marked than the one of  $\langle UI_{wind}^{div} \rangle_n$  in the  
346 data, and this is well reproduced in the models. Interestingly, in the southern sector the wind  
347 remains upwelling-favorable from October to June but the signature of the upwelling on observed  
348 SSTs ( $\langle UI_{wind}^{div} \rangle_s$ ) is restricted to the period December-May. This one-month lag between the cycle of  
349 the upwelling driver and that of its SST signature has been noted before (e.g., Nykjær and Van  
350 Camp 1994; Cropper et al. 2014). A plausible explanation is that the overall weakness of the winds  
351 at the beginning and end of the season makes upwelling particularly susceptible to counteracting  
352 effects, mean stratification and/or air-sea heat fluxes. The surface expression of the upwelling can  
353 also be delayed by the time it takes for the upwelling to draw deep waters to the surface.  
354 Furthermore, biannual baroclinic coastal trapped waves with amplitude  $\sim 10$ -20 m modulate the  
355 upwelling seasonal cycle (Kounta et al. 2018) and may also complicate interactions between wind  
356 forcing and the SST response. Again, the MMM is significant at the 95% confidence level all year

357 months and relatively close to the observations. Therefore, no systematic wind bias plagues the  
 358 climate simulations in the SMUS.

### 359 3.2.b: Effect of the offshore Ekman Pumping

360 The Ekman suction/pumping averaged along 16°W to 20°W is defined as:

$$361 \quad \langle UI_{wind}^{suc} \rangle = \left\langle \frac{1}{wf} \nabla \times \right\rangle_{longitude} \quad (4)$$

362 where  $\nabla \times$  is the curl of wind stress. This term is expressed in  $m.day^{-1}$ . Given that the wind stress  
 363 values that we use are located at the center of our working grid cells following several  
 364 interpolations, it is not entirely clear whether the nearshore integration bound should be right at the  
 365 coast or one grid cell away. We choose to include the grid cell closest to shore. Doing otherwise  
 366 slightly changes our results quantitatively but does not affect our conclusions (not shown).

367 In Fig.5 both observational dataset (first two columns of each panels) show that  $\langle UI_{wind}^{suc} \rangle_s$  is  
 368 maximum in winter (Fig.5.b) and  $\langle UI_{wind}^{suc} \rangle_n$  in spring (Fig.5.a). This mainly follows the seasonality  
 369 of the meridional wind stress intensity seen in Fig.4. Models generally yield stronger values than  
 370 the observations. This tendency is particularly marked in the CCSM4, CESM1, GFDL, MIROC,  
 371 and MRI groups. INM is the only simulation which exhibits negative (downwelling) values during  
 372 spring in the nSMUS. As observed for Ekman transport, the MMM is close to the observations and  
 373 significantly different from zero at the 95% level all year months.

### 374 3.3: A counteracting effect: the onshore geostrophic flow

375 As discussed by Marchesiello and Estrade (2010) and Ndoye et al. (2017) for example, the region  
 376 of southern CUS region is characterized by a southward gradient of sea surface height which drives  
 377 an onshore geostrophic flow. This flow can produce a convergence of water near the coast and  
 378 therefore counteract the upwelling. Jacox et al. (2018) unambiguously demonstrate that upwelling  
 379 indices accounting for this geostrophic contribution provide an improved characterization of the

380 local vertical transport. In order to quantify this effect, we have examined the SSH climatology  
 381 from the AVISO satellite data and the GODAS reanalysis, as well as in the climate models. The  
 382 SSH meridional gradient is computed as the difference between the SSH averaged over two regions  
 383 bordering the SMUS region to the north and to the south (black boxes in Fig.1). In the validation  
 384 data (Fig.6, first bars on the left), the SSH difference is indeed negative all year long, thus  
 385 potentially inducing an onshore geostrophic flow. It undergoes semi-annual oscillations with  
 386 maxima (resp. minima) in November -December and in June (resp. August-September and February  
 387 to April). This effect thus seems to be maximum at the beginning and at the end of the upwelling  
 388 season. It could explain the subtle time mismatch mentioned in section 3.2 between  $\langle UI_{sst}^{cross} \rangle$  and  
 389  $\langle UI_{wind}^{div} \rangle$  with the geostrophic coastward flow preventing the SST cold anomalies to develop at the  
 390 beginning and end of the upwelling season. In the climate models, the SSH difference is also always  
 391 negative, but its seasonality differs strongly among models and hardly mimics the data. The SSH  
 392 difference is strong all year long in many models (for example ACCESS1-0, ACCESS1-3, CESM1-  
 393 CAM5, CESM1-FASTCHEM, IPSL-CMA-LR, IPSL-CM5A-MR), which are in many cases those  
 394 with low  $\langle UI_{sst}^{seas} \rangle$  (Fig.2) and/or limited  $\langle UI_{sst}^{cross} \rangle$  (Fig.3). This correspondence could confirm the  
 395 counteracting effect of this geostrophic flow on the upwelling signature. In other models, the SSH  
 396 difference is relatively weak (eg: CNRM, GFDL, LASG-IAP, LASG-CESS). Averaging over this  
 397 diversity of model responses yields a weak seasonal cycle for the MMM but the timings for the two  
 398 seasonal minima (only August-September) and maxima are correct. The MMM is not significantly  
 399 different from zero.

400 Note that the choice of the offshore location where the cross-shore geostrophic transport should be  
 401 computed is a source of difficulty. It may seem natural to compute it as close to the shore as  
 402 possible (Jacox et al. 2018). In the models used here, doing so did not affect the results (not shown).

403 On the other hand, it should be kept in mind that in reality and in finer resolution simulations,  
404 alongshore pressure gradients are significantly affected by the transition between the open/deep  
405 ocean and the continental shelf. Viscous effects may also become important nearshore. Therefore,  
406 we have used reference points situated slightly offshore as several studies have done (Colas et al.  
407 2008; Marchesiello and Estrade, 2010;).

#### 408 3.4: Quantitative assessment of the upwelling rate

409 We now propose to integrate the terms diagnosed above that contribute to the upwelling of  
410 subsurface water so as to provide a bulk assessment of the upwelling rate. This integration is  
411 performed over the entire upwelling area defined as [12°N-20°N]/[16°W-20°W] and over the whole  
412 upwelling season running from November to May. Fig.7.a (green bars) shows the integrated effect  
413 of  $UI_{wind}^{div}$ . Here, we assume that all the water volume displaced zonally due to Ekman divergence  
414 along the coast is fed by upwelling, thereby neglecting any possible convergence/divergence of the  
415 alongshore flow (Yoon and Philander 1982). Under this simplifying assumption, the integrated  
416  $UI_{wind}^{div}$  terms leads to an upwelling of 1.2 to 1.4 Sv in the observations and reanalysis, while the  
417 models range from 1 Sv (CESM1-CAM5-1-FV2, CESM1-WACC, GISS family, HadCM3) to 2.25  
418 Sv (HadGEM2-AO). The integral of the  $UI_{wind}^{suc}$  term (yellow bars) computed from QuikSCAT and  
419 TropFlux both yield an upwelling of about 1.1 Sv to 1.3 Sv. In both wind data sets, the effect of  
420  $UI_{wind}^{div}$  slightly dominates over  $UI_{wind}^{suc}$  with a ratio between the two contributions of 1.6 (QuikSCAT)  
421 to 1.9 (TropFlux). The models tend to systematically overestimate  $UI_{wind}^{div}$ . The integrated value of  
422 the Ekman suction in the models ranges between 0.25-0.5 Sv (ACCESS1 and IPSL families) and  
423 about 2.75 Sv (HadGEM2-AO). The spread is larger for  $UI_{wind}^{suc}$ , with some models overestimating  
424 the data and others underestimating them.  $UI_{wind}^{div}$  dominates over  $UI_{wind}^{suc}$  in 22 out of the 45  
425 simulations for which the computation was possible (ACCESS, CSIRO-BOM, CNRM, CMCC,

426 IPSL, MPI). Similar ratios as in the data are found in several individual models (BCC, EC-EARTH,  
 427 CCSM4, CESM1). In the MMM, the contribution of  $UI_{wind}^{div}$  is  $1.62 Sv \pm 0.08$  while the  $UI_{wind}^{suc}$   
 428 induces  $1.08 Sv \pm 0.19$  of upwelling. Both components are 20-50% above observed values but  
 429 their ratio corresponds quite well to what is found in the data (1.5). No significant difference in  
 430 terms of the relative importance of the two Ekman processes was found when considering the  
 431 southern and the northern sectors separately (not shown).

432 The physical and biogeochemical responses to coastal divergence and Ekman suction differ in  
 433 important ways (Capet et al. 2004; Renault et al. 2016). As a first approach, the two may  
 434 nonetheless be added up to provide an estimate of upwelling strength. Jacox et al. (2018) have  
 435 recently suggested that the effect of Ekman processes should be estimated globally from the  
 436 integration of Ekman transport along the boundaries (north, west, and south) of the region of  
 437 interest. Comparison of this approach with the one proposed here has shown that both  
 438 methodologies generally yield very similar results (not shown). In the validation data sets, the  
 439 difference is less than 5%, with the Jacox et al. (2018)'s approach leading to slightly stronger  
 440 results, while the multimodel mean is weakened by approximately 10%. A final refinement consists  
 441 in accounting for the effect of the onshore geostrophic transport. The latter can be estimated as  
 442 follows:

$$443 \quad T_{geo} = MLD \cdot \frac{g}{f} (SSH_{north} - SSH_{south}) \quad (5)$$

444 where  $T_{geo}$  is the vertical transport (in Sv) due to the zonal current generated from the meridional  
 445 SSH gradient in the mean of Mixed Layer Depth (MLD, in meters) in the SMUS region,  $g$  is the  
 446 gravity coefficient ( $g = 9.81 m.s^{-2}$ ),  $SSH_{north} - SSH_{south}$  is the meridional difference of sea surface  
 447 height as computed in section 3.3.  $T_{geo}$  is thus an estimate of the zonally averaged geostrophic  
 448 transport in the SMUS region. The zonal averaging is such that the transport is centered at 18°W in

449 water depths of 2000m or more. As apparent in (5), we only account for the transport within the  
 450 mixed layer (see Marchesiello and Estrade, 2010).  $T_{geo}$  is counted negative eastward following the  
 451 sign convention used to quantify the upwelling. The result of this computation is shown in black in  
 452 Fig.7.a. All models and validation datasets show a negative (counteracting) contribution of this  
 453 term. Note though that, as for the Ekman divergence, a net geostrophic onshore flow at 18°W may  
 454 in part be linked to an intensification of alongshore currents ( $-\frac{v_{geo}}{y} > 0$ ), and not only counteracting  
 455 the upwelling. The MRI model is probably a case where this effect plays an important role (if not  
 456 the geostrophic onshore effect would produce an unrealistic downwelling transport of 6 Sv). In the  
 457 other models,  $T_{geo}$  ranges from 0.25 Sv (ACCESS-1-3) to 2 Sv (CNRM family and CESM1-CAM5).  
 458 The contribution of the geostrophic term is stronger in the MMM (1.59 Sv +/- 0.56, computed from  
 459 only 22 simulations) than in the validation datasets (0.54 Sv and 0.5 Sv respectively). This 3-fold  
 460 bias indicates a possibly important model deficiency that would warrant further investigation.

461 As a first approximation of the upwelling intensity, we thus consider the integrated sum of  
 462 the Ekman transport ( $UI_{wind}^{div}$ ), the Ekman suction ( $UI_{wind}^{suc}$ ) and the geostrophic flow in the mixed  
 463 layer  $T_{geo}$ . We refer to this sum (Ekman processes and geostrophic flow) as the total upwelling  
 464 index ( $UI_{total}$ ). This latter term is  $\sim 1.5 Sv$  to  $1.75 Sv$  in the two data estimations respectively).  
 465 Because of the very strong effect of the onshore geostrophic flow,  $UI_{total}$  is negative in MRI,  
 466 suggesting a downwelling. This result is inconsistent with the integrated vertical velocity over the  
 467 region (grey bar) and with the SST-derived indices shown in Fig.2 and Fig.3. This being said, MRI  
 468 is one of the models with weak upwelling signal in  $\langle UI_{sst}^{seas} \rangle$ . The downwelling effect of SSH may  
 469 thus indeed be relatively strong in this model but as indicated above, we suspect that the  
 470 approximations underlying the construction of  $UI_{total}$  are not fully valid in this model. In all the other  
 471 models,  $UI_{total}$  is positive and ranges between 0.25 Sv and  $\sim 2 Sv$  (Fig.7.b) which is lower than or in

472 the range of the estimates based on observations (1.5 Sv to 1.75 Sv). The MMM value (0.75 Sv +/-  
473 0.5) is more than 50% smaller.

474  $UI_{total}$  is approximatively equal to a direct estimation of the upwelling flux from vertical velocities  
475 diagnosed from the models ( $U_w$ , Fig. 7b, see appendix B for details) for about one third of the  
476 models. For the CNRM-CM5 and CESM models,  $UI_{total}$  is well above the effective model upwelling  
477 rate, suggesting again the role played by convergences of the meridional flow as a response to the  
478 Ekman divergence, and/or other neglected process. IPSL is the only family in which  $U_w$  is much  
479 stronger than estimation of upwelling transport. The MMM is roughly 0.61 Sv +/-0.2 which a  
480 difference less than 14% of total estimation.

#### 481 **4. Evolution of the senegalo-mauritanian upwelling as projected in global** 482 **warming scenarios.**

483 We now use the different indices introduced in section 3 to explore the evolution of the SMUS  
484 towards the end of the 21<sup>st</sup> century in the RCP8.5 simulations. We consider the indices on average  
485 over the whole upwelling season, running from November to May. The MMM values correspond to  
486 the average of the changes of each individual simulations. As introduced in section 2, the  
487 significance of these changes is evaluated with a student test of the mean, considering that each  
488 simulation is independent. The subscript  $s$  following the percentage of change in MMM marks the  
489 significance of changes at the 95% level. We also indicate the number of models that agree on the  
490 sign of a given projected change, as in Christensen et al. (2013) and Parvathi et al. (2017) for  
491 example.

492 Fig.8 shows the projected change of the SST-based upwelling signatures averaged in the  
493 two SMUS subregions, expressed in percent of the historical value. Most simulations exhibit a  
494 decrease of  $\langle UI_{sst}^{seas} \rangle_n$  (dark blue bars in Fig.8.a). The decrease lies between 5 and 15% of the

495 historical value for most of the models. The MMM shows a decrease of  $8 \pm 2\%$  and 92.5% of the  
 496 models agree on the sign of this change. Note that both the minimum and maximum climatological  
 497 SST are expected to rise under the effect of global warming. A separate analysis has shown that, on  
 498 average over the SMUS region, the sea surface temperature of the coldest climatological month  
 499 increases more than the one of the warm season (not shown). This explains the reduction of the  
 500 amplitude of the seasonal cycle and tends to suggest that it is indeed due to a reduction of the  
 501 upwelling. Yet, this attribution is ambiguous because other processes such as air-sea heat fluxes and  
 502 horizontal transports may be altered by climate change in a way that also impacts the SST seasonal  
 503 cycle. Most models also exhibit a decreasing winter  $\langle UI_{sst}^{cross} \rangle_n$  in the future, with the MMM relative  
 504 change being significantly negative. Because this index has very weak values in some historical  
 505 simulations during the upwelling season, the percentage of change may be very large. Therefore, we  
 506 show the percentage of change divided by 10 for clarity of the figure. The multi-model mean  
 507 change of  $\langle UI_{sst}^{cross} \rangle_n$  is  $65 \pm 23\%$ . Again, a majority of models (90%) agree on the sign of this  
 508 evolution. To conclude, in nSMUS, the SST upwelling signatures are well represented in the  
 509 CMIP5 simulations (section 3.1), and their evolution into the end of the 21<sup>st</sup> century provides some  
 510 consistent signs of upwelling reduction.

511 In sSMUS (Fig.8.b), the decrease of  $\langle UI_{sst}^{seas} \rangle_s$  reaches 10% to 15% in most of the models.  
 512 A few of them show a strengthening of this index in the future (CSIRO-QCCCE, CMCC-CESM,  
 513 EC-EARTH, GFDL-ESM2 and IPSL-CM5B-LR). Overall, the MMM projects a change of similar  
 514 amplitude as in the north ( $9 \pm 3\%$ ), with 87.5% of models agreeing on the sign of the change). In  
 515 this region, the projected change of  $\langle UI_{sst}^{cross} \rangle_s$  is generally negative as well but weaker than in the  
 516 north, and with less robustness (the index increases in 30% of the models). The MMM decreases by  
 517  $40 \pm 37\%$ . This strong relative change is a consequence of the small  $\langle UI_{sst}^{cross} \rangle_s$  value (close to zero)



518 in the MMM for the present period (Fig.3.b). Further analysis has shown that the MMM change is  
519 insignificant during the beginning of the upwelling season (November and January) and only  
520 marginally significant during the core upwelling season (December and February to April, figure  
521 not shown). Overall in the sSMUS, both SST signatures suggest an upwelling reduction as in the  
522 north but these results need to be considered with caution as large biases were found in the  
523 representation of the upwelling SST signal over the historical period and the projected changes are  
524 less consistent than in the north.

525 Fig.9.a displays the responses of wind upwelling indices in the nSMUS. While the  
526 representation of  $\langle UI_{wind}^{div} \rangle_n$  is very consistent among models over the historical period (Fig.4.b), the  
527 response of this index to climate change is approximately evenly split between increasing and  
528 decreasing signal for the future (green bars). Changes range from -5% to 5% roughly. The MMM  
529 exhibits an insignificant reduction of  $1.2 \pm 1.8\%$ . A closer check at the  $\langle UI_{wind}^{div} \rangle_n$  projected changes  
530 for individual models and individual months reveals noisy patterns with many sign changes from  
531 one month to the next (not shown). This lack of consistency among the winter model responses over  
532 this region was also noted by Rykaczewski et al. (2015). On the other hand,  $\langle UI_{wind}^{div} \rangle$  changes for  
533 summer indicate a more robust upwelling reduction consistent with the results of Wang et al. (2015)  
534 and Rykaczewski et al. (2015) (not shown). In this latitudinal band, the changes in Ekman suction  
535 are much more consistent: 89% of the models show a projected increase of  $\langle UI_{wind}^{suc} \rangle_n$  (yellow bars),  
536 favorable to an increase of the upwelling, in winter. The MMM intensification ( $18 \pm 7\%$ ) is not  
537 directly consistent with the SST-based results described above but Ekman pumping is known to  
538 exert limited second order effects on SST (Doi et al. 2009; Capet et al. 2004). A shift in the large  
539 scale atmospheric pressure system in the Atlantic (see Discussion below) is responsible for this  
540 situation where  $\langle UI_{wind}^{suc} \rangle$  increases while  $\langle UI_{wind}^{div} \rangle$  decreases.

541 In the sSMUS (Fig.9.b), the climate models show a much more consistent response of  
 542 the Ekman divergence, with a general agreement on a reduction of the winter-time offshore Ekman  
 543 transport in the future (CSIRO-QCCCE and MPI family excepted). The MMM indicates a  
 544 statistically significant reduction of  $9\pm 2\%$  of  $\langle UI_{wind}^{div} \rangle_s$ , with 92% models agreeing on the sign of  
 545 this change. The projected change of the  $\langle UI_{wind}^{suc} \rangle_s$  is also more homogeneous. A large group of  
 546 models indicate a reduction of  $\sim 10\%$ . The reduction reaches 15% or more in many other (for  
 547 example BCC, GFDL family and MIROC-ESM). Conversely,  $\langle UI_{wind}^{suc} \rangle_s$  increases moderately in  
 548 INM (where we have found a negative historical  $\langle UI_{wind}^{suc} \rangle_n$  in section 3.2.b) and MPI models. The  
 549 MMM reduction amounts to  $5\pm 2\%$ . The  $\langle UI_{wind}^{suc} \rangle_s$  and  $\langle UI_{wind}^{div} \rangle_s$  changes are qualitatively  
 550 consistent with the decrease of the SST-based indices seen in Fig. 8b.

551 As discussed above, the upwelling dynamics is *a priori* influenced by a combination of  
 552 coastal divergence, Ekman pumping, and across-shore geostrophic transport. Because of a poor  
 553 availability of MLD and SSH for future simulations, the evolution of geostrophic flow could only  
 554 be estimated in 18 simulations (Fig.10) and results are inconclusive: the downwelling associated  
 555 with this term increases in some simulations and decreases in others. The  $T_{geo}$  MMM change, which  
 556 is downwelling-favorable, is consequently very weak and lacks statistical robustness. As for the net  
 557 upwelling index  $UI_{total}$ , it exhibits a robust decrease of  $8\pm 7\%$  this tendency is found in 14  
 558 simulations out of the 18 ones in which the diagnostic could be computed. We have tried to  
 559 corroborate this result by examining the model vertical velocities. The projected changes for the  
 560 MMM vertical transport in the SMUS exhibit a very weak decrease in the future but this reduction  
 561 is not significant. More problematically, the differences between vertical transport and  $UI_{total}$  varies  
 562 widely from model to model for present-time conditions (Appendix B, Fig. 7b) and we are not able

563 to explain such differences. The reader is referred to Oyarzún and Brierley (2018) for a  
564 comprehensive analysis of the vertical velocities in climate models.

## 565 **5. Discussion**

566 The effects of climate change on upwelling systems has been the subject of ample research in the  
567 last decade but the sector of the Canary system situated south of  $\sim 20^\circ\text{N}$ , the SMUS, remains a blind  
568 spot in that respect. The focus of this work is the long-term winter/spring upwelling evolution of the  
569 SMUS. To this end, five upwelling indices based on sea surface temperature, surface wind stress  
570 and sea surface height have been defined, compared and combined. The two indices based on SST  
571 aim at describing the surface thermal signature of the SMUS upwelling, in space (inshore-offshore  
572 SST contrast) and time (seasonality of the upwelling). Although this is a restrictive view of the  
573 upwelling, these two indices have the advantage of being based on a well-observed variable so that  
574 they can be properly constrained by observations. The other three indices are based on the surface  
575 wind stress and meridional gradients of sea level. They aim at quantifying key mechanisms  
576 implicated in the generation of upwelling vertical velocities: coastal divergence of the Ekman  
577 transport, Ekman suction, and possible counteracting effects due to convergences of the geostrophic  
578 flow for the first four indices. Our work distinguishes two SMUS sectors, north and south of Cape  
579 Verde, located at  $\sim 15^\circ\text{N}$ . We have compared the four indicators defined above. Unfortunately,  
580 pairwise inter-model correlations were found to be insignificant, both on the whole SMUS region  
581 and on the northern and southern ones separately (not shown).

582 In the south, all indices are consistent with an upwelling reduction toward the end of  
583 the 21<sup>st</sup> century. This result is particularly robust as it can be found in the MMM as well as in a  
584 large fraction of individual CMIP5 models. Changes are most pronounced during the core of the  
585 upwelling period (February-April) but they are consistent over the entire period when upwelling  
586 winds are established (October-June). In the north, the evolution of the SST indices is also

587 consistent with a tendency to upwelling reduction in winter and spring but the dynamical indices of  
588 upwelling intensity indicate otherwise: we found insignificant changes in coastal Ekman divergence  
589 and an increase of Ekman suction. Our attempt to provide more elaborate assessments of the  
590 upwelling rates have remained inconclusive, in part because many of the CMIP5 simulations do not  
591 offer the model fields necessary to evaluate the changes in mixed layer geostrophic circulation  
592 (needed to compute  $UI_{total}$ ) or the true upwelling rates into the mixed layer (Appendix B). This is an  
593 important caveat given the magnitude of the upwelling compensation induced by geostrophic  
594 currents, which was confirmed by our analyses. Further progress will presumably require  
595 downscaling experiments as in Oerder et al. (2015).

596 In spite of their consistency, we regard the evolution of upwelling thermal signatures  $UI_{SST}^{seas}$   
597 and  $UI_{SST}^{cross}$  with caution. Indeed SST and its long-term evolution under changing climate conditions  
598 are determined by complex interactions between processes of different nature involving a wide  
599 range of time scales and nonlinear feedbacks between them (Wahl et al. 2011; Jia and Wu 2013).  
600 Several additional processes, not investigated here, are also possibly implicated in the SST  
601 response to the surface wind in general and in the upwelling regions in particular. The mesoscale  
602 turbulence (Gruber et al. 2011) and the chlorophyll concentration (Hernandez et al. 2017) are  
603 examples of possible candidates. The subsurface stratification also certainly influences the SST  
604 response but its investigation has been left for a future study. The oceanic horizontal resolution  
605 (between  $0.33^\circ$  and  $3.7^\circ$  in the models used here) and vertical resolution also strongly limit the  
606 representation of the bathymetry and thus the representation of the oceanic signature of the  
607 upwelling in these models. Using these SST indicators in future studies should take all these  
608 limitations into account. An important premise of the present study has been that upwelling in the  
609 eastern tropical Atlantic is a key determinant of regional SST and remains so in the future. In such  
610 conditions, it was reasonable to expect that changes in the wind regime leading to modifications of

611 the net volume of upwelled water would primarily control the SST evolution (colder resp. warmer  
612 when upwelling-favorable winds increase resp. decrease). In a system where SST spatial contrasts  
613 are so tightly controlled by the rate of wind-induced entrainment of subsurface water into the  
614 mixed layer, it indeed seems reasonable to assume that temporal SST evolution will mainly be  
615 determined by how this rate changes. This is actually observed on a broad range of scales, from  
616 synoptic (Ndoye et al. 2014), to inter annual (Roy 1989; Blanke et al. 2002), or even multi-decadal  
617 (Pardo et al. 2011; Seo et al. 2012). Nevertheless inter-model pairwise correlation of thermal and  
618 dynamical indices is in general not significant, primarily because of uncertainties in the SST-based  
619 indices.

620 On the other hand, climate models are imperfect representations of the real ocean. One  
621 important limitation is the representation of fine-scale processes such as coastal upwellings, whose  
622 typical across-shore scale  $L$  and intensity  $w$  are linked by the relation  $UI_{wind}^{div} \sim L..w$ . Because of their  
623 coarse resolution, the physics of coastal upwelling cannot be properly resolved in climate models,  
624 so  $L$  is determined by numerical considerations and we have  $L \sim O(dx)$  being grossly  
625 overestimated. An important consequence is that  $w$  is unrealistically small (e.g., see Capet et al.  
626 2008). Overall, the influence of cold water entrainment into the mixed layer is spread spatially and  
627 locally greatly diminished, hence the possibility for other processes (changes in cloud cover,  
628 changes in relative humidity and evaporative heat flux (Hourdin et al. 2015), changes in lateral  
629 advection of heat by Sverdrup transport (Xu et al. 2014; Small et al. 2015) to compete with  
630 upwelling modifications and control trends in SST, and its spatio-temporal contrasts. The SMUS  
631 mixed layer heat balance is relatively subtle (Faye et al. 2015) and its evolution may involve more  
632 than just upwelling changes, especially in modeling systems where upwelling is weakened by  
633 numerical limitations. For example, an increase of the poleward Sverdrup transport due to changes  
634 in wind stress curl (Fig.9) could produce a moderate warming along the African coast and explain

635 the tendency for SST based indices to decrease in CMIP models, despite the absence of significant  
636 changes in  $\langle UI_{wind}^{div} \rangle_n$ . Pending in-depth investigations of the present and future mixed layer heat  
637 budgets in CMIP runs (which would require model outputs rarely available from CMIP5 runs) we  
638 tend to place more confidence in the conclusions drawn from changes obtained for the dynamical  
639 indices ( $UI_{wind}^{div}$  and  $UI_{wind}^{suc}$ ).

640 To test the physical soundness of the trends emerging from the dynamical indices  
641 ( $UI_{wind}^{div}$  and  $UI_{wind}^{suc}$ ) and try to explain the differences between northern and southern SMUS we  
642 propose additional analyses pertaining to the main atmospheric pressure centers and their possible  
643 spatial shift between present and future conditions. The Azores high, the Sahara-Sahelian heat low  
644 and the ITCZ are a priori the dominant centers of action in the region (Fig.11.a) whose long-term  
645 changes (e.g, displacements) can influence the SMUS wind regime. Fluctuations in the position of  
646 the Azores anticyclone (AA) have recently been implicated in the synoptic/intra-seasonal variability  
647 of upwelling winds offshore of Senegal (Kounta, pers com). Its influence on upwelling historical  
648 trends for the period 1981-2012 has also been previously noted by Cropper et al. (2014). As for the  
649 ITCZ, its meridional migration may play a key role in the seasonal interruption of upwelling winds  
650 over the southern SMUS. Indicators of the meridional position of the AA and ITCZ over the eastern  
651 Atlantic are computed for historical and future conditions (Fig.12). In general accordance with the  
652 recent findings of Byrne et al. (2018) on the change of the zonally averaged ITCZ location, no clear  
653 tendency is found for the position of the ITCZ in our sector of interest: the historical and future  
654 positions of the ITCZ are virtually identical for the MMM given the resolution of the common  
655 regular grid used the analysis. This figure also confirms the fact that the ITCZ, diagnosed from the  
656 maximum precipitation, is generally located further south than in the GPCP observations in the  
657 region between 15°W and 30°W limits (see Richter and Xie 2008) and also (Siongco et al. 2015). In  
658 contrast, results unambiguously point to a northward migration tendency for the AA. Differences

659 between historical and future conditions are slightly above  $1^\circ$  towards the north for the MMM, with  
660 94.5% of the model agreeing on the migration direction. This result is consistent with those of Ma  
661 and Xie (2013), Sousa et al. (2017) and, more generally, with the projected expansion of the Hadley  
662 cell in a warming world (Lu et al. 2007).

663 A spatial view of these changes is given in Fig.11.b. The weak change of the ITCZ location  
664 in the eastern Atlantic is confirmed, although a northward shift is evident in the west, away from  
665 our region of interest. The northward shift of the AA manifests in the form of a positive SLP  
666 anomaly over the whole mid-latitudes and a negative anomaly further south. This latter anomaly is  
667 not purely zonal. It is more confined in the subtropical eastern Atlantic and appears as the oceanic  
668 prolongation of a clear cyclonic anomaly over land that is produced by anomalous warming. The  
669 presence of this anomalous heat low is one important aspect in the hypothesis of Bakun (1990) but  
670 its spatial structure is such that the upwelling wind response varies meridionally depending on the  
671 sector: its quasi-circular shape with a center at  $20^\circ\text{N}$ , i.e., the northern limit of the SMUS, implies  
672 very different wind anomalies north and south of this latitude. In the north, upwelling winds tend to  
673 intensify, as also found in Sousa et al. (2017). On the other hand, anomalous winds rotate to  
674 westerlies in the nSMUS and progressively to south-westerlies in the sSMUS at  $10^\circ\text{N}$ . This spatial  
675 structure of the wind anomalies differs from the annual mean picture emerging from Servain et al.  
676 (2014) for the historical period possibly because changes in atmospheric state are season dependent  
677 and trends in winter-spring do not reflect those for the annual mean. The SLP change is, on the  
678 other hand, consistent with the trends presented above for  $\langle UI_{wind}^{div} \rangle$  and their differences between  
679 the northern and southern SMUS. Considered over the entire Canary current system, the Bakun  
680 hypothesis does not appear to hold (as noted by Rykaczewski et al. 2015 for all four upwelling  
681 systems) because SLP modifications due to climate change are shaped in good part by the evolution

682 of the Hadley cell (zonally symmetric expansion and poleward migration of its descending branch)  
683 and by an intriguing Sahara-Sahelian heat low expansion that protrudes onto the ocean.

684 The wind anomaly also exhibits a notable curl in the nSMUS in agreement with the  $\langle UI_{wind}^{suc} \rangle_n$   
685 changes (Fig.9.a). The sSMUS evolution is *in fine* more influenced by regional subtleties of the  
686 SLP changes (the zonal and meridional structure of the anomalous low) than by the larger scale  
687 northward shift of the AA. Processes at even finer scale unresolved in the CMIP5 simulations may  
688 further modulate the evolution of the SMUS wind regime (Boé et al. 2011).

## 689 **6. Conclusion**

690 The focus of this work was the long-term upwelling evolution in the SMUS region. To this end, five  
691 upwelling indices based on sea surface temperature, surface wind stress and sea surface height have  
692 been defined, compared and combined. The two indices based on SST aimed at describing the  
693 surface thermal signature of the upwelling, in space (inshore-offshore SST contrast) and time  
694 (seasonality of the upwelling). The other three indices were based on the surface wind stress and  
695 anomalous SSH and aimed at quantifying key mechanisms for the upwelling.

696 Amplitude of the SST seasonal cycle in the upwelling region is generally well represented  
697 in the climate models. In 34 simulations out of 47, however, the index is underestimated in the  
698 southern part of the upwelling and the maximum is found north of 15°N. This bias is suggestive of  
699 difficulties to reproduce the tropical Atlantic climate in the models. The SST contrast between the  
700 open ocean and the coast shows a clear seasonal cycle and it even changes sign in summer. Again,  
701 models usually have largest biases in the southern part of the senegalo-mauritanian upwelling,  
702 where this seasonality is not as clearly defined as in the observations. Interestingly, despite biases in  
703 terms of intensity, the seasonality of the Ekman drift, diagnosed from the surface wind stress, is on



704 the contrary relatively well reproduced in the models, with a multi model mean very similar to the  
705 validation datasets in both SMUS subdomains.

706 We have also attempted to quantify the total volume of upwelled water. Our approach was  
707 based on the cumulative effect of two driving terms (Ekman drift and Ekman suction) and one  
708 damping term (geostrophic onshore flow). One strong limitation of this approach is the neglect of  
709 possible meridional/alongshore flow convergence. Specifically, a recent study has highlighted the  
710 seasonality of the slope current along West Africa (Kounta et al. 2018). This indirect estimation  
711 could also slightly overestimate the real upwelling volume by not distinguishing water upwelled  
712 through suction and divergence processes at the coast. Finally, the estimation of the geostrophic  
713 onshore flow was found to be inconsistent across models. Still, this estimation yields an upwelling  
714 ranging between  $\sim 1.5 Sv$  to  $1.75 Sv$  in the validation data. This amount tends to be underestimated  
715 in the MMM, where the volume of upwelled water is estimated to be  $0.75Sv \pm 0.5$ . The volume of  
716 upwelled water due to the diagnostic vertical velocity of the models yields an even weaker  
717 estimation:  $0.61Sv \pm 0.2$ , consistently with the approximations discussed above. Furthermore, note  
718 that these different averages are not performed over the exact same multi model ensemble due to  
719 restrictions data availability.

720 The response of the individual models to the RCP8.5 scenario is diverse in terms of  
721 amplitude but the general picture is a reduction of the upwelling toward the end of the 21<sup>st</sup> century.  
722 The response of the thermal indices is in general more consistent among models and it is only partly  
723 consistent with the response of the dynamical indices.  $\langle UI_{sst}^{seas} \rangle$  is clearly reduced in the future,  
724 because the warming trends is accentuated in winter (upwelling-season) when SST is coldest.  
725 Consistently with this picture of a reduced upwelling effect on the SST, the offshore-coastal SST  
726 contrast decreases in most models in both the northern and southern SMUS region. The meridional  
727 wind stress is also projected to decrease during the 21<sup>st</sup> century when averaged over the whole

728 southern SMUS (12°N-20°N). This projection of Ekman divergence is much more robust across  
729 models in the southern region, consistently with the large-scale anomalous SLP structure  
730 characterized by an anomalous heat low around 20°N (whose imprint extends over the ocean) and a  
731 northward shift of AA. A recent study in the Humboldt upwelling system in a subset of the CMIP5  
732 models database has suggested an increasing wind stress projected in the poleward upwelling part in  
733 winter (Oyarzún and Brierley 2018). However, the authors shows that an increasing stratification  
734 will in parallel limit the effect of the wind below the surface. Such effect has not been investigated  
735 here.

736 In the northern part of SMUS, models split roughly equally into those suggesting an increase  
737 and a decrease of the wind divergence. The large scale circulation indeed shows that the anomalous  
738 wind circulation is primarily westward. Consistently with these large scale circulation changes,  
739 Ekman suction increases in nSMUS and decreases in sSMUS. In nSMUS, this is not directly  
740 consistent with the thermal indices. Over the whole southern SMUS region (12°N-20°N), we found  
741 no significant change of the wind stress curl. Finally, the simple dynamical budget of the upwelling  
742 that we have proposed generally yields a weakening of the upwelling. The evolution of the vertical  
743 transport diagnosed from the vertical velocities computed online is inconclusive.

744 Our results generally show that the thermal indices are more sensitive to models deficiencies  
745 than the dynamical ones, which rely more on the large scale atmospheric circulation. Major efforts  
746 have recently been put on the reduction of tropical Atlantic biases (Richter 2015). Upcoming  
747 CMIP6 simulations are expected to benefit from them. While we have carried out a first exploration  
748 of the future of the SMUS upwelling with the CMIP5 models, we hope that the present study could  
749 be used as a benchmark framework to investigate the future of the coastal upwellings in the CMIP6  
750 simulations. Consequences of the present study and of the projected future of the upwelling for  
751 fishery activities in Senegal also still deserves extensive attention. Renault et al. (2016) have shown

752 that upwelling indices are not enough to evaluate primary production and fish stocks. Chavez and  
 753 Messié (2009) have even illustrated for example that Peru fish catch exceeds that from the other  
 754 EBUS by an order of magnitude even though primary production levels are similar. Downscaling  
 755 simulations may be needed to link the present projections to possible changes in primary production  
 756 and higher up in the trophic chain.

757 **Appendix A:** Wind stress estimation from wind speed data:

758 In principle, the drag coefficient  $C_d$  depends on both the atmospheric and the oceanic state and it is  
 759 variable. Large and Pond (1982) proposed the following scaling of  $C_d$  according to the wind speed  
 760  $V$ :

761 
$$10^3 C_d = 0.49 + 0.065V \quad \text{for } 10 < V < 25 \text{ m.s}^{-1} \quad (6)$$

762 
$$10^3 C_d = 1.14 \quad \text{for } 3 < V < 10 \text{ m.s}^{-1} \quad (7)$$

763 
$$10^3 C_d = 0.62 + 1.56V^{-1} \quad \text{for } V < 3 \text{ m.s}^{-1} \quad (8)$$

764

765 Gill (1982), (pp 29) proposed another scaling, based on results of Smith (1980) for large wind  
 766 speeds:

767 
$$C_d = 0.0011 \quad \text{for } V > 3 \text{ m.s}^{-1} \quad (9)$$

768 
$$10^3 C_d = 0.061 + 0.063V \quad \text{for } 6 < V < 22 \text{ m.s}^{-1} \quad (10)$$

769 The NOAA-TM-NMF S-SWFSC-231 rapport (1996) suggests to use the value of  $C_d = 0.0026$  with  
 770 monthly mean data while Santos et al. (2012) used  $C_d = 0.0014$  for their study of the Moroccan  
 771 upwelling zone (22°N-33°N).

772 Over the SMUS region, the maximum wind speed is around  $7 \text{ m.s}^{-1}$ . We compared the meridional  
 773 component of the wind stress computed online in the IPSL-CM5A-LR climate model to the wind

774 stress computed offline using the meridional wind speed component from the same simulations with  
775 the drag coefficient values from Santos et al. (2012) and the two methods (equation (9), (10)) drag  
776 coefficient listed above (figure not shown). We have chosen to test only the meridional wind  
777 component because it is the strongest one in the SMUS region and most directly associated to the  
778 upwelling intensity. We found that the scaling proposed in Gill (1982) underestimate the meridional  
779 wind stress amplitude north of roughly  $15^{\circ}\text{N}$ , in particular in summer north of  $20^{\circ}\text{N}$ , as well as in  
780 winter between  $12^{\circ}\text{N}$  and  $20^{\circ}\text{N}$  which the season and the latitude band of the SMUS. The  $C_d =$   
781  $0.0014$  scaling as used by Santos et al. (2012) yields the closest values to the observations. In all  
782 simulations where only the wind speed components are provided, the wind stress was thus  
783 computed using this latter scaling.

784 **Appendix B:** model upwelling from vertical velocities.

785 The net estimation of upwelling must in principle be consistent with the upward ocean mass  
786 transport diagnosed in the models (whenever available). However, this latter field is noisy (as  
787 recently emphasized by Oyarzún and Brierley 2018) and the depth at which the typical upwelling  
788 vertical velocity should be considered for comparison with upwelling indices is a difficult  
789 parameter to choose. Here, this depth has been chosen equal to the mixed layer depth averaged over  
790 the upwelling season (November to May) and over the SMUS region. In the models for which  $UI_{total}$   
791 could not be computed (see last column in Table 1) we choose the depth where  $U_w$  is maximum.

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Model		Available data		Depth (in $m$ ) used to compute $U_w$
	Acronym	Historical	RCP85	
1	ACCESS1-0	sst, wind stress, ssh wmo, psl, pr, uas, vas,	sst, wind stress, ssh wmo, psl, pr, uas, vas,	49
2	ACCESS1-3	sst, wind stress, ssh, wmo, psl, pr, uas, vas	sst, wind stress, ssh, wmo, psl, pr, uas, vas	54
3	bcc-csm1-1	sst, wind stress, ssh wmo, psl, pr, uas, vas	sst, wind stress, ssh wmo, psl, pr, uas, vas	30
4	bcc-csm1-1-m	sst, wind stress, ssh wmo, psl, pr, uas, vas	sst, wind stress, ssh wmo, psl, pr, uas, vas	30
5	CanCM4	sst, ssh psl, pr		
6	CanESM2	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	49
7	CSIRO-QCCCE	sst, wind stress, ssh psl, pr, uas, vas	sst, wind stress, ssh psl, pr, uas, vas	
8	CNRM-CM5	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	44
9	CNRM-CM5-2	sst, wind stress, ssh, mld wmo, psl, pr		40
10	CMCC- CM	sst, wind stress, ssh wmo, psl, pr,	sst, wind stress, ssh wmo, pr	30
11	CMCC-CMS	sst, wind stress, ssh wmo, psl, pr, uas, vas	sst, wind stress, ssh wmo, psl, pr, uas, vas	40
12	CMCC-CESM	sst, wind stress, ssh wmo, psl, pr, uas, vas	sst, wind stress, ssh wmo, psl, pr, uas, vas	50
13	CCSM4	sst, wind stress, ssh, mld, wmo, psl, pr, uas, vas	sst, wmo, psl, pr, uas, vas	49
14	CESM1-CAM5-1- FV2	sst, wind stress	sst	
15	CESM1-CAM5	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	41
16	CESM1-FASTCHEM	sst, wind stress, ssh, mld wmo, psl, pr		49
17	CESM1-WACCM	sst, wind stress, ssh, mld wmo, psl, pr	sst pr	42
18	CESM1-BGC	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	51
19	EC-EARTH	sst, wind stress psl, pr, uas, vas	sst, wind stress psl, pr, uas, va	
20	FIO-ESM	sst, wind stress psl, pr, uas, vas	sst, wind stress psl, pr, uas, vas	
21	GFDL-CM2p1	sst, wind stress, ssh psl, pr		

22	GFDL-CM3	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	32
23	GFDL-ESM2G	sst, wind stress, ssh, mld psl, pr, uas, vas	sst, wind stress, ssh, mld psl, pr, uas, vas	
24	GFDL-ESM2M	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	37
25	GISS-E2-H-CC	sst psl, pr, uas, vas	sst psl, pr, uas, vas	
26	GISS-E2-R-CC	sst, ssh wmo, psl, pr, uas, vas	sst, ssh wmo, psl, pr, uas, vas	30
27	GISS-E2-H	sst psl, pr, uas, vas	sst psl, pr, uas, vas	
28	GISS-E2-R	sst, ssh wmo, psl, pr, uas, vas	sst, ssh wmo, psl, pr, uas, vas	30
29	HadCM3	sst psl, pr, uas, vas		
30	HadGEM2-CC	sst psl, pr, uas, vas	sst psl, pr, uas, vas	40
31	HadGEM2-ES	sst psl, pr, uas, vas	sst psl, pr, uas, vas	40
32	HadGEM2-AO	sst, wind stress psl, pr, uas, vas	sst, wind stress psl, pr, uas, vas	
33	IPSL-CM5A-LR	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	49
34	IPSL-CM5A-MR	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	36
35	IPSL-CM5B-LR	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	56
36	INM	sst, wind stress, ssh, wmo, psl, pr, uas, vas	sst, wind stress, ssh, wmo, psl, pr, uas, vas	
37	LASG-IAP	sst, wind stress, ssh, wmo, psl, pr, uas, vas	sst, wind stress, ssh, wmo, psl, pr, uas, vas	50
38	LASG-CESS	sst, ssh psl, pr		
39	MRI-CGCM3	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	54
40	MRI-ESM1	st, wind stress, ssh, mld wmo, psl, pr, uas, vas	st, wind stress, ssh, mld wmo, psl, pr, uas, vas	56
41	MIROC-ESM	sst, wind stress, ssh psl, pr, uas, vas	sst, wind stress, ssh psl, pr, uas, vas	
42	MIROC5	sst, wind stress, ssh wmo, psl, pr, uas, vas	sst, wind stress, ssh wmo, psl, pr, uas, vas	50
43	MPI-ESM-LR	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	39
44	MPI-ESM-MR	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	39

45	MPI-ESM-P	sst, wind stress, ssh, mld wmo, psl, pr		39
46	NorESM1-ME	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	sst, wind stress, ssh, mld wmo, psl, pr, uas, vas	69
47	NorESM1- M	ssst, wind stress, ssh,mld wmo, psl, pr, uas, vas	ssst, wind stress, ssh,mld wmo, psl, pr, uas, vas	70

Table1: List of the CMIP5 simulations used in this study. The third and fourth columns list the variables that were available for our study: surface temperature (sst), wind stress, sea surface height (ssh, called zos in the CMIP5 database), mixed layer depth (mld, omlmax in the CMIP5 database), sea level pressure (slp), precipitation (pr), zonal and meridional surface wind components (uas and vas) and upward ocean mass transport (wmo). This latter variable (given in  $kg \cdot s^{-1}$  in the CMIP5 outputs) has been systematically converted into vertical velocity (in  $m \cdot s^{-1}$ ). For this conversion, we have used the average density of sea water in the region of the study [ $12^{\circ}N-20^{\circ}N$   $6^{\circ}W-20^{\circ}W$ ] estimated from the SODA reanalysis (Carton and Giese 2008). The last column indicate the depth which compute the direct estimation of the total upwelling transport (see Appendix B for details).

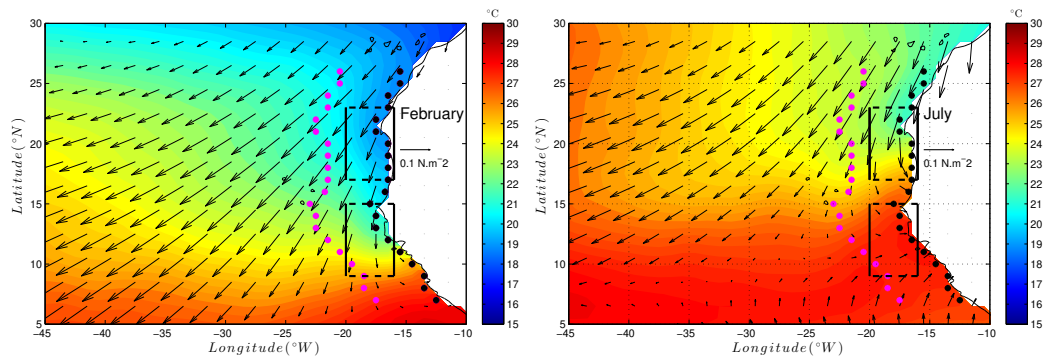


Figure 1: Colors: SST (HadISST) climatological mean (in  $^{\circ}C$ ) in February (left) and July (right) averaged over the period [1985-2005]. Vectors: Tropflux climatological wind stress for the same months respectively and the same period. The solid dots indicate the coastal (black) and offshore (magenta) locations used for the computation of the  $UI_{sst}^{cross}$  (see section 3.1.b). The black and magenta dots are separated by  $5^{\circ}$  of longitude. The black boxes represent the regions used for the computation of the SSH meridional gradient ( section 3.3).



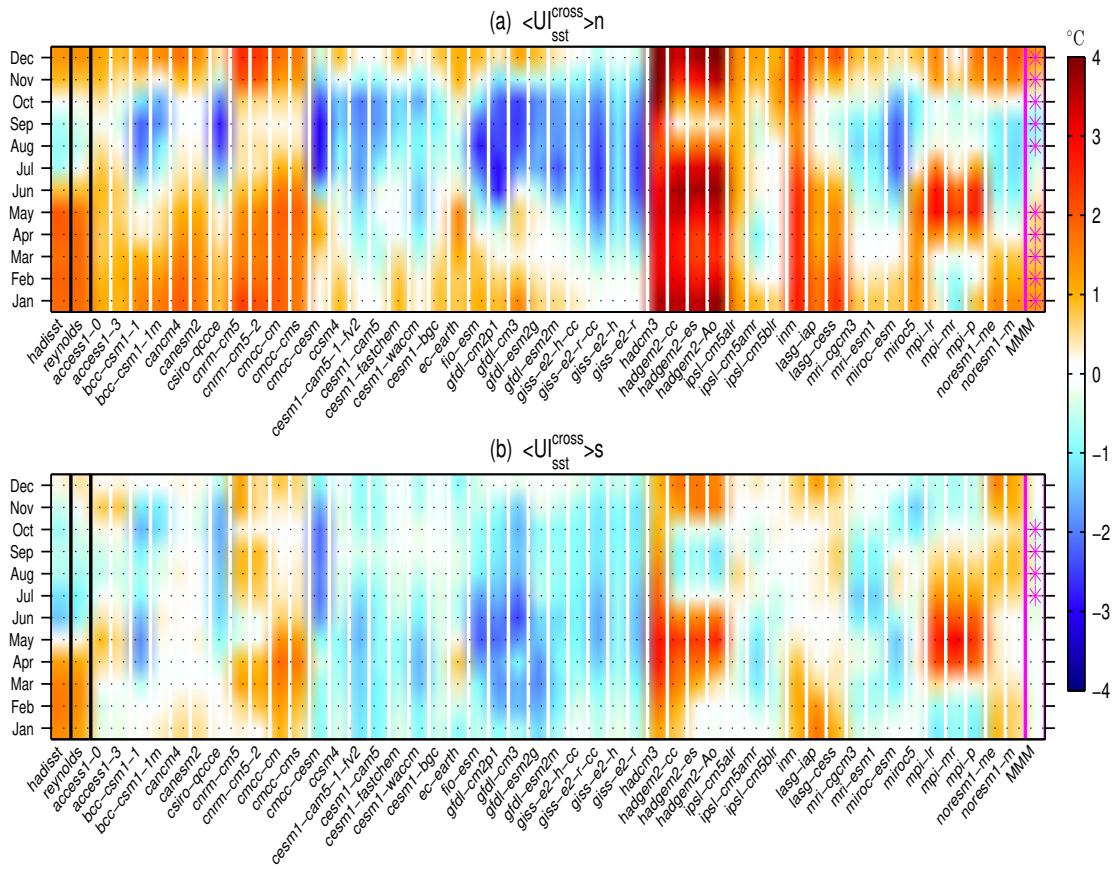


Figure 3: Panel a: monthly climatology [1985-2005] of  $UI_{sst}^{cross}$ , averaged over the northern part of the senegalo-mauritanian area (nSMUS,  $[16^{\circ}N - 20^{\circ}N]$ ), called  $\langle UI_{sst}^{cross} \rangle_n$ . Panel b: same as panel a for the southern zone (sSMUS,  $[12^{\circ}N - 15^{\circ}N]$ ),  $\langle UI_{sst}^{cross} \rangle_s$ . The first two columns on the left (highlighted in black) show the results for the two observational data sets, the other columns show the individual CMIP5 models and the last column in the right, highlighted in magenta, shows the Multi-Model Mean (MMM). Positive (negative) values correspond to upwelling (downwelling) conditions. The magenta stars in the last column mark the months for which the MMM is significant at the 95% level with respect to the multimodel spread.

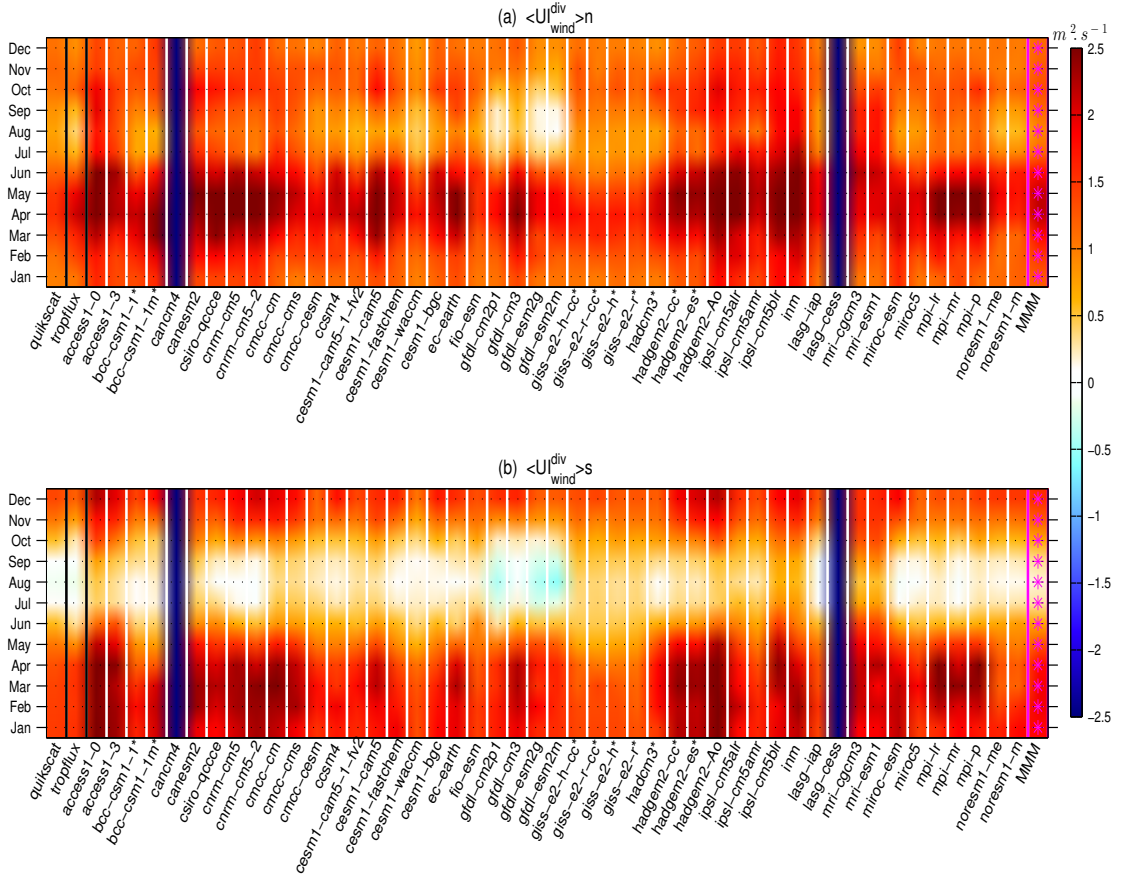


Figure 4: Monthly climatology of  $\langle UI_{wind}^{div} \rangle$  averaged over the northern part [ $16^\circ N$ -  $20^\circ N$ ] (panel a,  $\langle UI_{wind}^{div} \rangle_n$ ) and the southern part [ $12^\circ N$ -  $15^\circ N$ ] (panel b,  $\langle UI_{wind}^{div} \rangle_s$ ) of the senegalo-mauritanian upwelling region and computed over [1985-2005]. On both panels, the first two columns on the left (highlighted in black) show the index in the two validation data sets. The other columns show the individual CMIP5 simulations and the last column in the right (in magenta) shows the multi-model mean (MMM). Positive (negative) values correspond to upwelling (downwelling) conditions. The two dark blue columns stand for models for which neither the near surface wind intensity nor the wind stress was available (CanCM4 and LASG-CESS). The simulations for which the oceanic wind stress was not given directly in the CMIP5 data base and was thus computed offline following the methodology described in Appendix A are marked by a star following their name. See Fig. 3 for comments on the magenta stars



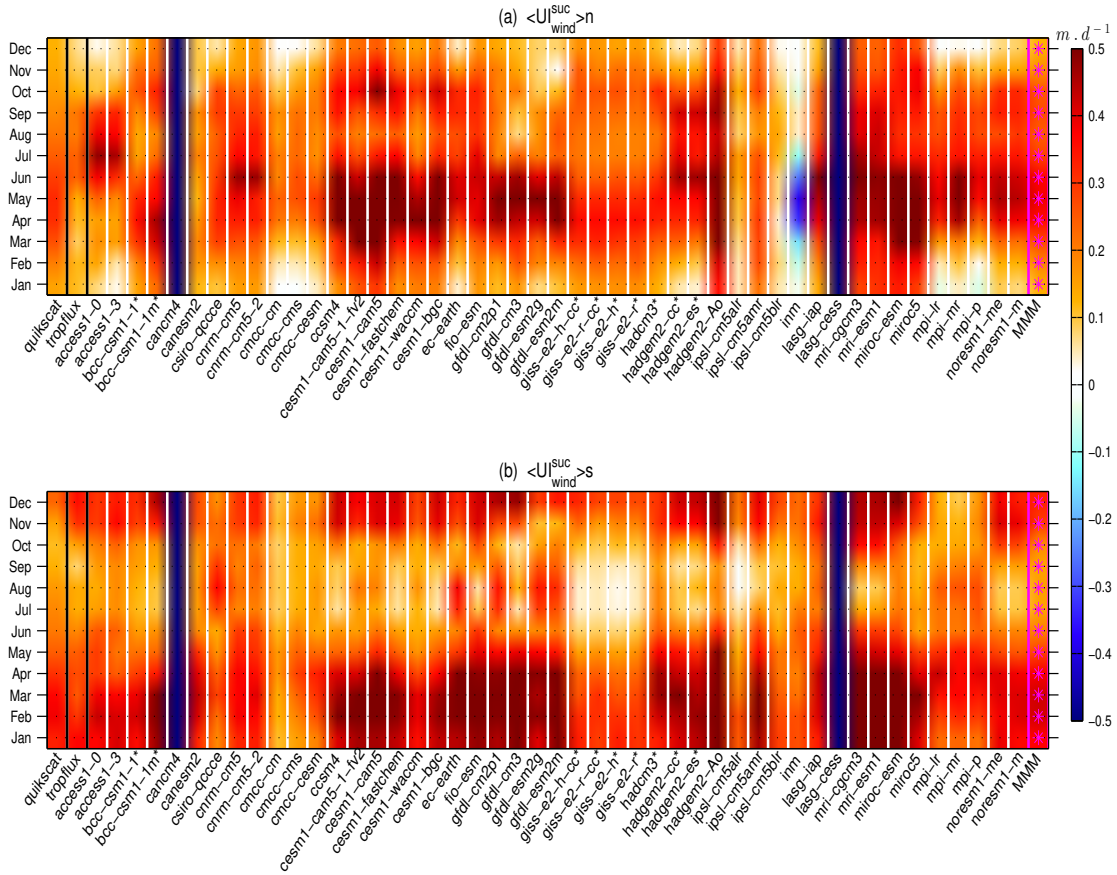


Figure 5: Monthly climatology of the Ekman suction index  $\langle UI_{wind}^{suc} \rangle$  averaged between  $[16^\circ W - 20^\circ W]$  and over the northern part  $[16^\circ N - 20^\circ N]$  (panel a,  $\langle UI_{wind}^{suc} \rangle_n$ ) and the southern part  $[12^\circ N - 15^\circ N]$  (panel b,  $\langle UI_{wind}^{suc} \rangle_s$ ) of the senegalo-mauritanian upwelling. The first two columns on the left (highlighted in black) show the two observational data sets, the other columns show the individual CMIP5 simulations and the last column in the right (in magenta) shows the (MMM) computed over  $[1985-2005]$ . The dark blue columns stand for the models which did not provide any wind data (CanCM4 and LASG-CES). See Fig. 3 for comments on the magenta stars.

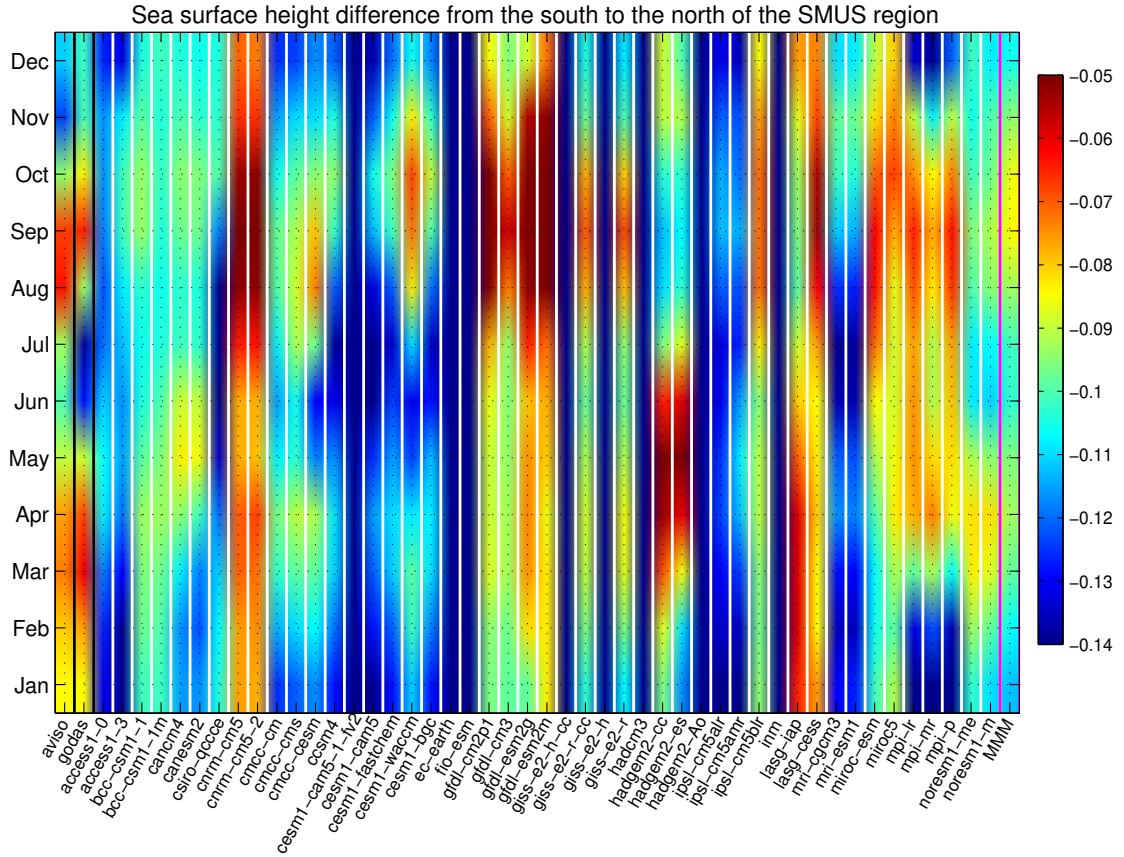


Figure 6: Monthly climatology of the meridional sea surface height difference (units:  $m$ ) between the region  $[9^{\circ}N-15^{\circ}N/16^{\circ}W-20^{\circ}W]$  and the region  $[17^{\circ}N-23^{\circ}N/16^{\circ}W-20^{\circ}W]$ . These regions are highlighted in Fig. 1 (black boxes). The first column on the left shows results from AVISO satellite data (period [1993-2005]), and the second one from the GODAS reanalysis [1985-2005]. The following columns show the results of the climate simulations for the period [1985-2005]. The last column on the right (highlighted in magenta) shows the (MMM). The dark blue columns stand for the simulations which did not provide the SSH data (CESM1-Cam5-1-fv2, EC-EARTH, FIO-ESM, GISS-E2-H-CC, GISS-E2-H, HadCM3, HadGem2-AO).

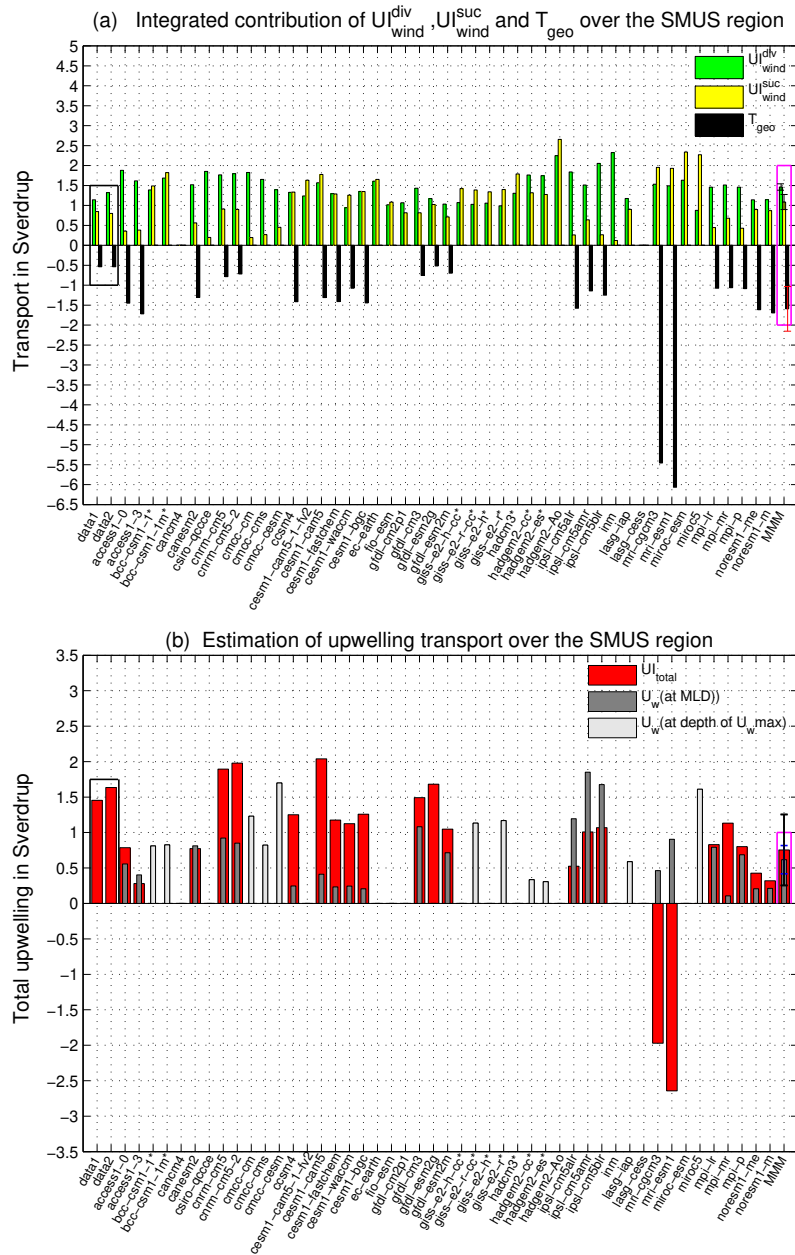


Figure 7: Panel (a): Estimate of the seasonal (November-May) integrated contribution (in Sverdrup) of the three dynamical indices ( $U_{wind}^{div}$ ,  $U_{wind}^{suc}$  and  $T_{geo}$ ) to the upwelling.  $U_{wind}^{suc}$  (yellow bars) was integrated between  $[12^{\circ}N - 20^{\circ}N]$  and  $[16^{\circ}W - 20^{\circ}W]$ ,  $U_{wind}^{div}$  (green bars) was integrated over the same latitude range.  $T_{geo}$  (black bars) was computed from Eq. (5) (see text). Data1 correspond to  $U_{wind}^{suc}$  and  $U_{wind}^{div}$  computed from Quikscat (2000-2009) and  $T_{geo}$  computed with the AVISO SSH product (1993-2005) and MLD from de Boyer Montegut. Data 2 represent  $U_{wind}^{suc}$  and  $U_{wind}^{div}$  computed from Tropflux (1985-2005), SSH computed from the GODAS reanalysis (1985-2005) and the same MLD used in data1. The following columns show the results for the individual climate simulations, and the last column (magenta) shows the MMM with confidence interval estimated from a student test of the mean given the dispersion of the individual models. Panel (b):  $U_{total}$  (red bars) shows an estimation of the total volume of upwelling water computed as the sum of the three contributions shown in panel (a). Data 1 and data 2 are the same as in panel (a). The dark grey bars display the volume of water effectively upwelled in the climate simulations computed as the integral of the vertical velocity diagnosed online over the upwelling region  $[12^{\circ}N - 20^{\circ}N] / [16^{\circ}W - 20^{\circ}W]$  and taken at MLD. The light grey bars represent the models for  $U_w$  was taken at the depth maximizing this quantity. The MMM is computed only for the simulations where both  $U_{total}$  and  $U_w$  computed at MLD are available. See text for details.

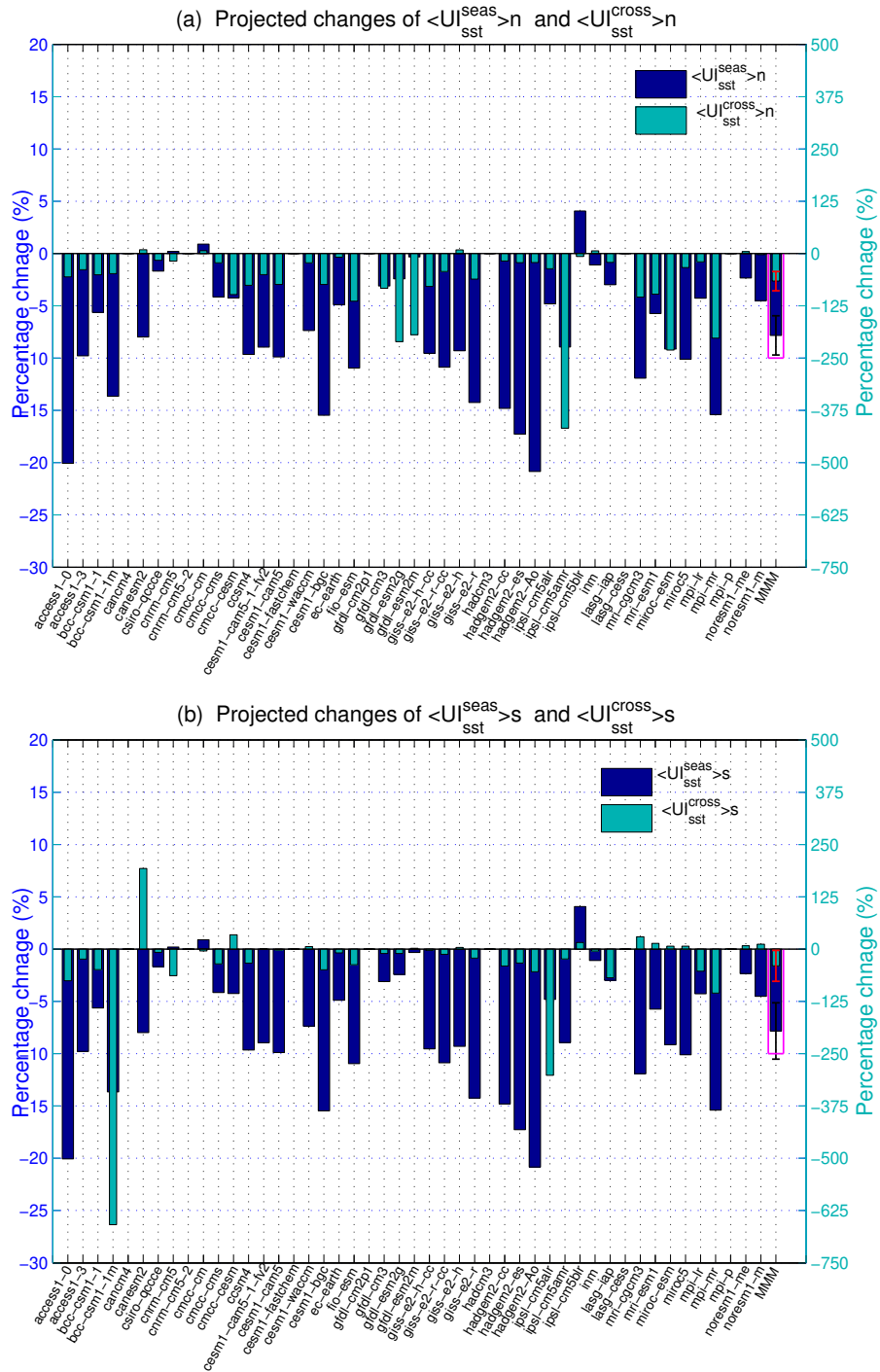


Figure 8: Projected changes of the indices of upwelling thermal signatures averaged over the northern region [ $16^{\circ}N$ -  $20^{\circ}N$ ] (top panel) and over the southern region [ $12^{\circ}N$ -  $15^{\circ}N$ ] (bottom panel). The dark blue bars show the projected changes (in %) of  $\langle UI_{sst}^{seas} \rangle_n$ . The changes are estimated as the difference between the future [2080-2100] and the historical [1985-2005] period and the percentage is estimated with respect to the historical value. The light blue bars show the projected changes of  $\langle UI_{sst}^{cross} \rangle_n$  averaged over the upwelling season (November - May). Models for which SST data was not available for the RCP8.5 scenario are marked by a empty space. The right column (in magenta) shows the percentage of change of the multi-model mean. The black and red whiskers bars indicate the 95% confidence interval of  $\langle UI_{sst}^{seas} \rangle$  and  $\langle UI_{sst}^{cross} \rangle$  MMM respectively.

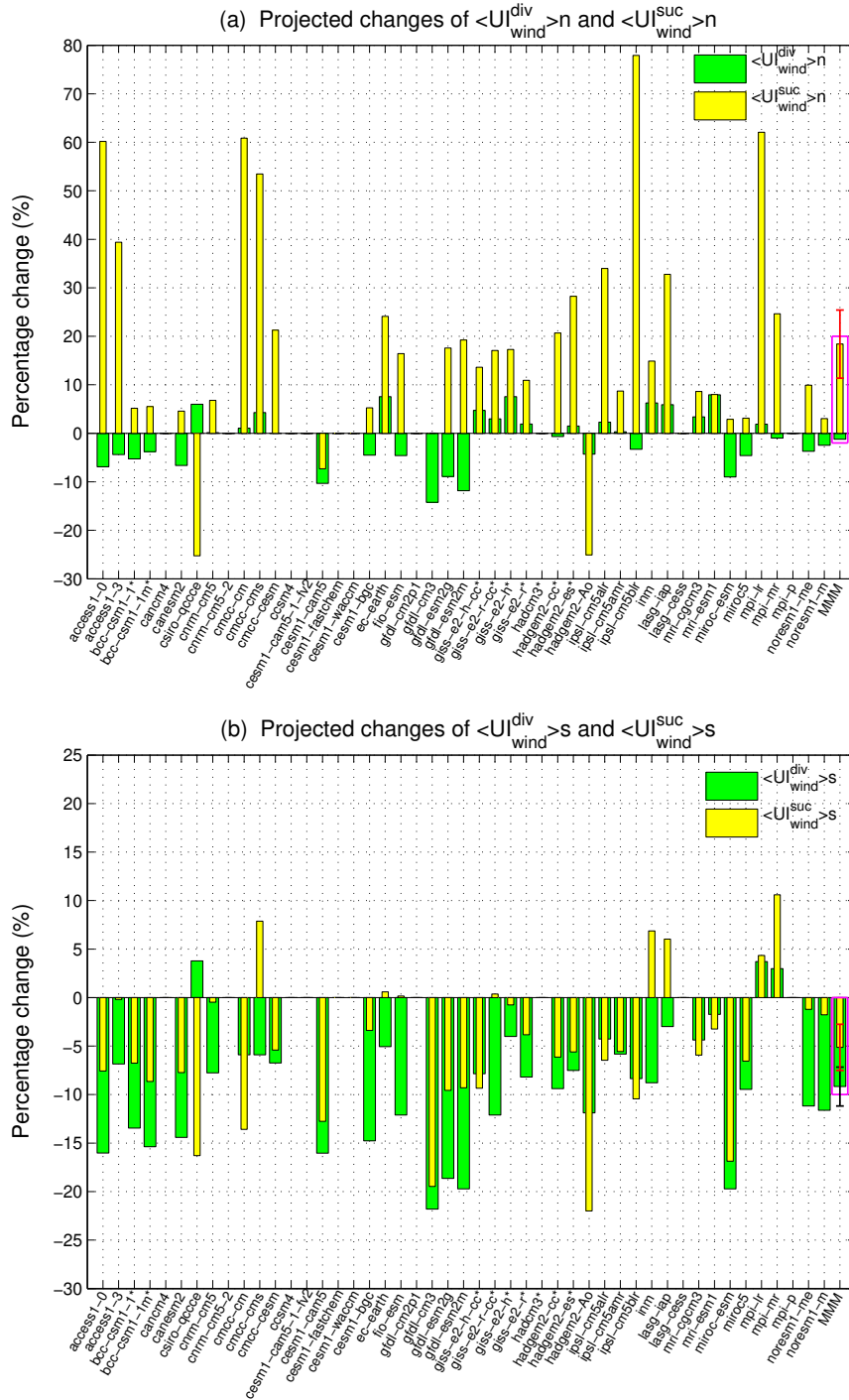


Figure 9: Projected changes of the dynamical indices of the upwelling integrated over the northern region [16°N- 20°N] (panel (a)) and over the southern region [12°N- 15°N] (panel (b)). The changes are estimated as the difference between the future [2080-2100] and the historical [1985-2005] period and the percentage is estimated with respect to the historical value. The green bars show the projected changes (in %) of  $\langle UI_{wind}^{div} \rangle_n$  averaged over the upwelling season (November-May). The yellow bars show the projected changes of  $\langle UI_{wind}^{suc} \rangle_n$  integrated over the longitude range [16°W- 20°W] and for the same climatological season. Models for which wind data was not available for the RCP8.5 scenario are marked by an empty space. The right column (in magenta) shows the percentage change of the multi-model mean. The black and red whiskers bars indicate the 95% confidence interval of  $\langle UI_{wind}^{div} \rangle_n$  and  $\langle UI_{wind}^{suc} \rangle_n$  respectively. Negative (positive) values exhibit a decrease (increase) of these upwelling indices.

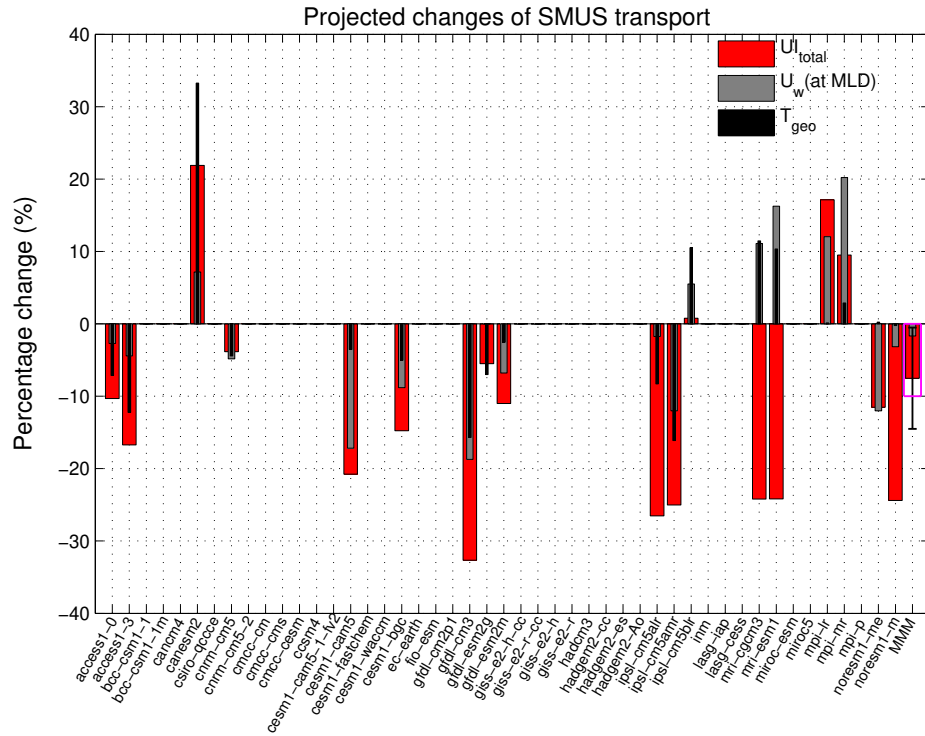


Figure 10: Projected change (in %) of geostrophic flux  $T_{geo}$ , the direct ( $U_w$ ) and indirect ( $U_t^{total}$ ) estimates of the total upwelling of water derived in Fig.7b and for models for which data was available. Here, in particular, we only consider models for which MLD was available both over the historical and the future period. As in Fig.9, the changes are computed as the difference between the future [2080-2100] and the historical [1985-2005] period, and the flux is averaged over the upwelling season (November-May). The grey and red bars show the projected changes of  $U_w$  (taken at MLD) and  $U_t^{total}$  estimates of the total volume of the upwelling respectively. The black bars show the projected change of the volume of upwelling water due to the geostrophic flux  $T_{geo}$ . The black whisker bar indicate the 95% confidence interval of  $U_t^{total}$  MMM.

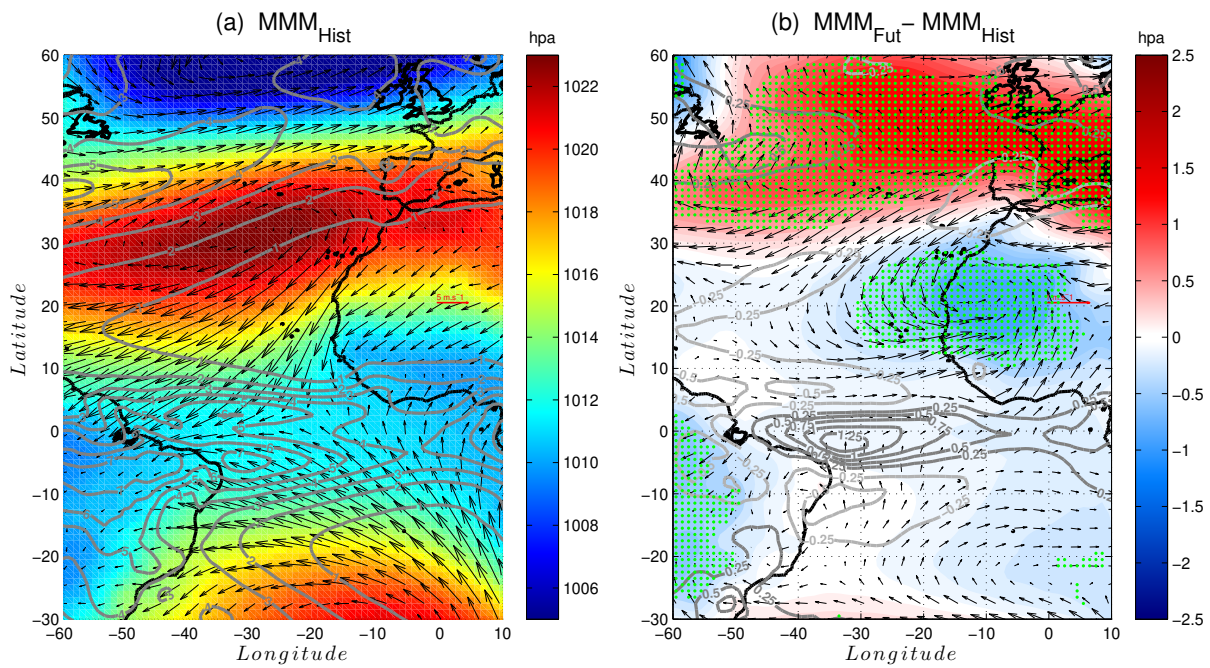


Figure 11: Panel (a): Climatology of sea level pressure (color), surface winds (vectors) and precipitation (contours) in the (MMM) averaged from November to May over the historical period [1985-2005]. Panel (b) shows the projected changes of these atmospheric variables. Regions where more than 75% models agree on the sign of the changes are indicated with green dots. These figures are computed from the 37 simulations which have at the same time the sea level pressure, precipitation and surface wind over both periods.



