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1 **Projection of upwelling-favorable winds in the Peruvian upwelling system under the RCP8.5 scenario using**
2 **a high-resolution regional model**

3

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15

16 **Abstract.** The Peruvian upwelling system (PUS) is the most productive Eastern Boundary Upwelling System
17 (EBUS) of the world ocean. Contrarily to higher latitude EBUSs, there is no consensus yet on the response of up-
18 welling-favorable winds to regional climate change in this region. Global climate models are not able to reproduce
19 the nearshore surface winds, and only a few downscaling studies have been performed by using relatively coarse-
20 grid atmospheric models forced by idealized climate change scenarios. In the present study, the impact of climate
21 change on the PUS upwelling-favorable winds was assessed using a high resolution regional atmospheric model to
22 dynamically downscale the multi-model mean projection of an ensemble of thirty-one CMIP5 global models under
23 the RCP8.5 worst-case climate scenario. We performed a 10-year retrospective simulation (1994-2003) forced by
24 NCEP2 reanalysis data and a 10-year climate change simulation forced by a climate change forcing (i.e.
25 differences between monthly-mean climatologies for 2080-2100 and 1989-2009) from CMIP5 ensemble added to
26 NCEP2 data. We found that changes in the mean upwelling-favorable winds are weak (less than 0.2 m s⁻¹). Sea-
27 sonally, summer winds weakly decrease (by 0-5%) whereas winter winds weakly increase (by 0-10%), thus
28 slightly reinforcing the seasonal cycle. A momentum balance shows that the wind changes are mainly driven by
29 the alongshore pressure gradient, except in a local area north of the Paracas peninsula, downstream the main up-

30 welling center, where wind increase in winter is driven by the shoreward advection of offshore momentum. Sensi-
31 tivity experiments show that the North-South sea surface temperature gradient plays an important role in the wind
32 response along the north and central coasts, superimposed onto the South Pacific Anticyclone large-scale forcing.
33 A reduction (increase) of the gradient induces a wind weakening (strengthening) up to 15% (25%) off the northern
34 coast during summer. This local mechanism is not well represented in global climate models projections, which
35 underlines the strong need for dynamical downscaling of coastal wind in order to study the impact of climate
36 change on the Peruvian upwelling ecosystem.

37

38 **Keywords**

39 Regional climate change - Peruvian upwelling system - Upwelling-favorable wind - Ocean-atmosphere interac-
40 tions

41

42 **1. Introduction**

43 The Peruvian Upwelling System (PUS) is located on the northern part of the Humboldt Current System, one of the
44 four major Eastern Boundary Upwelling Systems (EBUSs) in the global ocean. It extends approximately between
45 4°S and 16° S along the western coast of South America and produces more fish catch per unit area than any other
46 region (e.g. Chavez et al., 2008). There, as in other EBUSs, equatorward alongshore winds generate an offshore
47 transport of coastal surface waters, which are replaced by subsurface colder and nutrient-rich waters. The cold sur-
48 face waters influence the regional climate by creating a persistent low cloud cover (e.g. Mechoso et al., 2014) and
49 the nutrient enrichment produced by the upwelling sustains a very high biological productivity (e.g. Cushing,
50 1990).

51

52 As other coastal upwelling systems, the PUS is expected to be impacted by the global warming associated to the
53 continued increase of greenhouse gases in the atmosphere. Gutierrez et al. (2011) suggests that the coastal up-
54 welling may have increased off Central and South Peru, driving a negative sea surface temperature (SST) trend be-
55 tween 1950 and 2010. The cooling is consistent with a wind increase found in the ERA40 reanalysis, suggesting a
56 possible intensification of wind-driven upwelling. Unfortunately, instrumental records of multi-decadal nearshore
57 wind variability are scarce in the PUS. Using in situ wind observations from the International Comprehensive
58 Ocean Data Set (ICOADS; Worley et al., 2005), Bakun (1990) found an increase in nearshore wind stress in the
59 period 1950-1990 in all EBUSs including the PUS, which led him to propose the 'Bakun hypothesis': the increase

60 of upwelling-favorable winds is driven by an increased cross-shore pressure gradient (due to low pressure over
61 land and high pressure over sea), generated by an enhanced land-sea thermal contrast under climate change. How-
62 ever, Tokinaga and Xie (2011) showed that the alongshore wind positive trend in ICOADS observations was partly
63 biased because of the increase in anemometer height of ship masts. Furthermore, Narayan et al. (2010) highlighted
64 the large discrepancies between the PUS wind stress trends from ICOADS, NCEP/NCAR and ERA-40 reanalyses.

65

66 On the other hand, only a few studies based on global climate models have analyzed the response of upwelling-fa-
67 vorable winds to climate change in EBUSs, including the Humboldt Current System. Using the Coupled Model In-
68 tercomparison Phase 5 (CMIP5) ensemble, Rykaczewski et al. (2015) compared surface winds during the 2071-
69 2100 period under the Representative Concentration Pathway (RCP) 8.5 to those during the 1861-1890 historical
70 period. They found an intensification of upwelling-favorable winds in the poleward parts of EBUSs and a weaken-
71 ing in their equatorward parts. These changes were associated with a poleward migration of the high pressure sys-
72 tems (e.g. the South Pacific Anticyclone (SPA) for the Humboldt Current System). However, the PUS was not
73 considered in their analysis as the latitudinal domain was limited to 16°S-44°S. Using the ensemble mean of thirty-
74 one CMIP5 global climate models (see Table 1) for the 2080-2100 period relative to 1989-2009 period under the
75 RCP8.5 scenario, the projected 10-meter wind changes are weaker (< 0.5 m/s) off the Peru coasts (3°S-18°S) than
76 off Chile (18°S-40°S) (Fig.1a). Note that the coastal wind changes are not robust among the CMIP5 ensemble
77 members, except south of 16°S and north of 6°S (red dots in Fig. 1a). Seasonally, alongshore winds tend to de-
78 crease in the north of Peru (5°S-9°S) in summer and tend to increase in winter south of 12°S (Fig. 1b). The relative
79 changes are weak ($< 5\%$), except near 5°S in summer ($\sim 30\%$). The marked differences between the wind changes
80 off Peru and Chile suggest that the PUS winds could be influenced by other mechanisms than those related to the
81 migration and intensification of the SPA (e.g. Rykaczewski et al., 2015).

82

83 Furthermore, only a few regional studies focused on the impact of the climate change on the PUS coastal winds us-
84 ing downscaling. Goubanova et al. (2010) used a statistical downscaling method to obtain surface wind at ~ 50 km
85 resolution from the coarse-resolution (~ 300 km) CMIP3 IPSL-CM4 global model (Marti et al., 2010). They found
86 that surface alongshore winds off Peru changed only during summer (5–10% weakening) under the idealized cli-
87 mate scenarios simulating a doubling ($2\times\text{CO}_2$) and a quadrupling of CO_2 ($4\times\text{CO}_2$) concentrations with respect to
88 preindustrial conditions. On the other hand, Belmadani et al. (2014) used a global atmospheric model with a pro-
89 gressively refined horizontal resolution of $0.5^\circ \times 0.5^\circ$ in the Humboldt Current system, forced by IPSL-CM4 SST
90 from the same scenarios as downscaled by Goubanova et al. (2010). They found that winds off Peru weakened

91 both during summer (20-30%) and winter (5-10%). The weaker winds were associated to the SPA southward shift
92 and to the convective anomalies associated to the increased precipitation in the tropics during summer.

93

94 Nevertheless, the scope of these studies is limited in several ways: the wind projections remain relatively coarse-
95 grid ($\sim 50\text{--}100$ km), only one CGCM (IPSL-CM4) was downscaled under idealized scenarios (2xCO₂ and
96 4xCO₂). Moreover, mesoscale SST patterns can have an impact on the wind intensity (e.g. Chelton et al. 2001;
97 Small et al. 2008). Chamorro et al. (2018) have recently shown that pronounced coastal SST alongshore gradients
98 which appear under El Niño conditions reinforce coastal winds off Peru. This mechanism, which may play a role if
99 latitudinal SST gradients are modified by climate change, has not been considered in previous studies.

100

101 In the present work, we expand on previous studies by assessing the coastal wind projections off Peru following a
102 different approach. We use a high resolution regional atmospheric model forced by the multi-model mean (MMM)
103 of thirty-one CMIP5 models under the worst-case Representative Concentration Pathway 8.5 (RCP8.5) scenario,
104 for which the radiative forcing reaches 8.5 W m^2 by the year 2100 (van Vuuren et al. 2011). Using the regional
105 model, we are able to study the mechanisms governing local wind changes and relate them to previous findings.
106 Section 2 presents the CMIP5 ensemble, the regional settings, the data used to evaluate the model's realism and the
107 methods. Section 3 presents the model results: an evaluation of surface winds in the present period, the projected
108 wind and SPA changes, and a surface momentum budget. A summary and discussion of the main results are pre-
109 sented in section 4.

110

111 **2. Model, data and methods**

112

113 **2.1 The regional atmospheric model**

114 We used the Weather Research and Forecasting (WRF, Skamarock and Klemp, 2008) model version 3.7.1 with
115 three nested domains (Fig. 2a). The largest domain (d01) encompasses the tropical-subtropical South Pacific ocean
116 ($42^\circ\text{S}\text{--}26^\circ\text{N}$; $101^\circ\text{E}\text{--}58^\circ\text{W}$) with a horizontal resolution of 105 km and 210×78 grid points. This so-called parent
117 domain has a wide extension as it was also used to carry out climate change projections in the South Pacific Con-
118 vergence Zone (Dutheil et al., 2019). The second domain (d02) covers the coasts of Peru and Northern Chile
119 ($30^\circ\text{S}\text{--}10^\circ\text{N}$; $101^\circ\text{W}\text{--}63^\circ\text{W}$) with a horizontal resolution of 21 km and 200×220 grid points. The innermost do-
120 main (d03) covers nearly all of the Peruvian coastal segment ($17^\circ\text{S}\text{--}5^\circ\text{S}$) with a horizontal grid spacing of 7 km and

121 225 x 216 grid points. The high resolution of the latter allows to better represent the Andes and the coastal orogra-
122 phy (Fig. 2b). 32 levels in the vertical were used in the parent domain, while 60 levels in the vertical were used in
123 the second and innermost domain, with 21 levels in the first ~1000 m above sea surface, following Oerder et al
124 (2016). The model time steps were set to 300, 45 and 15 s for the domains of 105, 21 and 7 km resolution, respec-
125 tively. The three different configurations are hereafter referred to as WRF105, WRF21 and WRF7.

126 The set of physical parameterizations used in the model are similar to those used in Chamorro et al. (2018). They
127 include the Dudhia (1989) and Mlawer et al. (1997) schemes for shortwave and longwave radiation respectively,
128 the Hong and Lim (2006) microphysics scheme, the Betts-Miller-Janjić parameterization for convection (Janjić
129 1994), the Mellor–Yamada– Nakanishi–Niino (MYNN) Level 2.5 planetary boundary layer (Nakanishi and Niino
130 2009) along with the MYNN surface layer parameterization, and the Noah land surface model (Chen and Dudhia
131 2001).

132

133 **2.2 Historical and climate change experiments**

134 First, a control simulation was carried out for the retrospective period 1994-2003 in the three nested domains in
135 one-way offline nesting mode (i.e. a finer-grid-resolution run is made as a subsequent run after the coarser-grid-
136 resolution run). The initial and lateral boundary conditions for the finer-grid run are obtained from the coarse grid
137 run. Input from terrestrial fields (e.g. terrain, land use, etc.) and masked surface fields (such as soil temperature and
138 moisture) are obtained from WRF geographical datasets (available in [http://www2.mmm.ucar.edu/wrf/users/down-
139 load/get_sources_wps_geog.html](http://www2.mmm.ucar.edu/wrf/users/download/get_sources_wps_geog.html)). The parent domain is forced by lateral open boundary conditions from the 6-
140 hourly NCEP2 reanalysis data at 1° (Kanamitsu et al., 2002) and by daily Optimum Interpolation Sea Surface
141 Temperature (OISST) data at 0.25° (Reynolds et al. 2007) at the air-sea interface. In this simulation, the atmo-
142 spheric CO₂ concentration is increased from 358 to 375 ppm (levels representative of the retrospective period).

143

144 Second, a climate change simulation was performed for a ten-year period (2086-2095) in the three nested domains
145 in one-way offline nesting mode, following the pseudo global warming downscaling method (PGW-DS, Kimura
146 and Kitoh, 2007). The ensemble mean from thirty-one CMIP5 global models (see Table 1) of monthly anomalies
147 of the seasonal cycle were added to the 6-hour NCEP2 lateral atmospheric boundary conditions, following the
148 same methodology as in Dutheil et al. (2019). The monthly anomalies of the ensemble-mean seasonal cycle are
149 differences between the monthly ensemble-mean seasonal cycle computed over the period 2080-2100 under the
150 RCP8.5 scenario and the monthly ensemble-mean seasonal cycle computed over the historical period 1989-2009.
151 The ensemble-mean RCP8.5 historical annual-mean SST anomaly displays a strong surface warming of 3-4°C

152 along the Peru coasts (Fig 3a). The monthly anomalies were added to the NCEP2 6-hourly air temperature, water
153 vapor mixing ratio, geopotential height, zonal and meridional wind at each vertical level of the model, and also to
154 the daily OISST sea surface temperature (Reynolds et al., 2007). The land parameters were kept identical to the
155 control simulation. In this simulation, the atmospheric CO₂ concentration is increased from 810 to 890 ppm. These
156 levels are representative of the RCP8.5 projections at the end of the twenty-first century for the period 2086-2095
157 (Van Vuuren et al., 2011).

158

159 The PGW-DS method adds a climate perturbation signal to a present-day conditions in order to obtain future cli-
160 mate conditions at a higher resolution. This approach also allows to remove the present-day biases of CMIP5 mod-
161 els and cancels the variable part of errors from individual models when multi-model mean anomalies are used
162 (Kawase et al. 2008). Using the ensemble mean of a large ensemble of coupled model simulations (e.g. 30) reduces
163 the uncertainty caused by the low-frequency variability and highlights the impact of external greenhouse gas
164 forcing (Jebri et al., 2020), even in the case of a relatively short integration period (e.g. ten years), regardless of the
165 selected control period (Kawase et al., 2009). However, this approach has some limitations that should be kept in
166 mind when interpreting the results. The intra-seasonal and interannual variabilities in the lateral and surface bound-
167 aries conditions for the 2086-2095 period are identical to those of the 1994-2003 period. In other words, the
168 anomaly related to the strong 1997-1998 El Niño is encountered in 2083-2084 under the RCP8.5 scenario. Thus,
169 the possible change of El Niño variability under the RCP8.5 scenario and its local impact in the PUS are not simu-
170 lated in this study.

171

172 In addition, three sensitivity experiments were performed using WRF21 for ten-year periods under RCP8.5 sce-
173 nario. They were forced by the same multi-model mean lateral atmospheric boundary conditions as previously de-
174 scribed, and by SST anomalies from CESM1-CAM5, CNRM-CM5 and CSIRO-Mk3-6-0 climate models instead
175 of the multi-model mean. These simulations were used to assess the sensitivity of the coastal wind projections to
176 the along-shore SST gradient. These particular climate models simulated distinct alongshore SST gradient changes
177 in RCP8.5 conditions with respect to the historical period: around zero, negative and positive (see CGCM numbers
178 8, 11 and 12 respectively in Fig. 3b). The three downscaled simulations are called WRF-cesm1, WRF-cnrm and
179 WRF-csiro.

180

181 **2.3 Monthly momentum budget**

182 Changes in the momentum budget was analyzed following the method described in Oerder et al. (2016) and
 183 Chamorro et al. (2018). We used the following vectorial relation:

$$184 \quad \langle \mathbf{V} \rangle_{M+1} - \langle \mathbf{V} \rangle_M = \sum_{F_n \in \{\text{forces}\}} \Delta t \left([F_n]_{M+1} - [F_n]_M + \langle F_n \rangle \right) \quad (1)$$

185 where, $\langle \mathbf{V} \rangle$ is the wind monthly-mean vector, Δt is the model time step, and F_n are the four momentum terms: ad-

186 vection $(\vec{v} \cdot \vec{\nabla}) \vec{v}$, vertical mixing $\frac{\partial}{\partial z} \left(\frac{\vec{\tau}}{\rho} \right)$, Coriolis $-f \vec{k} \times \vec{v}$ and pressure gradient $-\frac{1}{\rho} \vec{\nabla} P$, such as the left

187 side of Eq. (1) represents the change of the wind monthly-mean \mathbf{V} (from month M to month $M+1$) and the right

188 side of the equation represents the sum of the forces contributions (F) to the momentum balance. The angle bracket

189 $([\])$ in Eq. (1) is the double time averaging operator defined as:

$$190 \quad [F_n] = \frac{1}{N+1} \sum_{p=0}^N \left(\sum_{k=1}^p F_n \right) \quad (2)$$

191 with N the number of model time steps during one month. We computed the alongshore (i.e. parallel to the WRF
 192 model smoothed coastline) component of Eq. (1) in four coastal boxes (between 5-8°S, 8-13°S, 13-14°S and 14-
 193 17°S) of one-degree width along the coast of Peru.

194

195 **2.4 Observational data**

196 Surface wind observations from the QuikSCAT satellite-borne scatterometer (CERSAT, 2002; hereafter QSCAT)
 197 at 1/2° resolution were used to evaluate the realism of the modeled wind over the period 2000-2003. The product
 198 was downloaded from <http://www.ifremer.fr/cersat>.

199 The Pearson correlation coefficient (r) was used to measure the strength of a linear association between two vari-
 200 ables, and a p-value less than 0.01 was used to determine the statistical significance of the relationship.

201

202 **3. Results**

203 **3.1 Evaluation of the mean and seasonal cycle of present-day coastal winds**

204 In order to evaluate the impact of the model resolution on surface winds, the 2000-2003 annual mean surface wind
 205 from WRF105, WRF21 and WRF7 are shown in domain d03 in Figs. 4a-c. Observations (Fig. 4d) indicate that
 206 surface winds blow north-westward in a direction approximately parallel to the coastline. The wind is strongest
 207 offshore ($\sim 6-7.5$ m s⁻¹) and decreases shoreward down to $\sim 4.5-5$ m s⁻¹. Offshore winds are slightly overestimated

208 (~1 m s⁻¹) in the model simulations, regardless of the spatial resolution. The wind intensification off the Paracas
209 peninsula (14°S) is too strong in WRF21 (Fig. 4b) but likely more realistic in WRF7 (Fig. 4c). The near-shore
210 wind decrease from ~200 km towards the coast (a.k.a. ‘wind drop-off’, *e.g.* Capet et al., 2004) off the central coast
211 (7°S-12°S) is relatively well reproduced in WRF7, whereas it is overestimated in WRF105 and underestimated in
212 WRF21 (Fig. 4e). Note that the wind drop-off in the 0–50 km coastal band is not well known due to the so-called
213 ‘satellite blind zone’.

214

215 Figure 5 shows the WRF7 and QuikSCAT mean annual cycle of alongshore wind averaged in a coastal band of
216 ~100 km, for the period 2000-2003 (when satellite observations are available). Observed winds are weakest in
217 summer, reaching a minimum of ~ 3-4 m s⁻¹ off the center and north coast (5°S-12°S), and strongest in winter,
218 reaching a maximum of ~ 7 m s⁻¹ near 15°S (Fig. 5a). Overall, WRF7 reproduces well the seasonal cycle of along-
219 shore winds in good agreement with satellite observations, replicating the local maximum value near 15°S in win-
220 ter. However, some discrepancies are found, such as an overestimation (~ 1 m s⁻¹) of the alongshore wind intensity
221 off central Peru (8°S-15°S) in summer and a slight underestimation (<1 m s⁻¹) in winter-spring (Fig. 5c). Note that
222 QuikSCAT data does not include information of wind in the satellite blind zone, which may explain some of the
223 discrepancies.

224

225 Overall, these comparisons show that WRF7 reproduces the mean state and seasonal variability of coastal winds
226 off Peru with a fair degree of realism. The surface wind bias in the coastal region due to coarse horizontal resolu-
227 tion has been an important limitation of previous modeling studies (*e.g.* Belmadani et al., 2014; Oerder et al.,
228 2015).

229

230 **3.2 Climate-driven changes in coastal wind**

231 The projected wind changes simulated are now characterized by the different model configurations. We focus on
232 WRF21 and WRF7 wind changes, because of WRF105 too coarse spatial resolution. WRF21 and WRF7 show
233 weak (and very similar) changes in the mean surface wind field (Figs. 6 a-b, d-e). Wind velocity increases slightly
234 in the offshore region (400 km from the coast) by about 0.2 m s⁻¹. In the nearshore region, wind velocity decreases
235 slightly (<0.2 m s⁻¹) off the north and south coast of Peru, and increases slightly off the central coast (9°S-13°S).
236 The increase in the nearshore region (11–14°S) is larger in WRF7 (~0.2 m s⁻¹) than in WRF21 (~0.1 m s⁻¹). Ex-
237 amination of the alongshore wind seasonal cycle (Figs. 6 c, f) reveals more pronounced differences among the
238 coastal regions. The winds tend to weaken in spring and summer (by ~5 %) off the north (5°S-9°S) and south

239 (14°S-17°S) coasts, and alongshore winds tend to intensify in fall and winter off the central coast by ~5 % in
240 WRF21 and ~10 % in WRF7. Overall, the relative changes do not exceed 5-10% which contrasts with the stronger
241 change (e.g. up to a 30 % reduction in summer near 5°S) in the CMIP5 ensemble mean (Fig. 1b). Besides, the reso-
242 lution increase from 21 km to 7 km did not strongly modify the results, except the slightly stronger winter wind in-
243 tensification off central Peru (9°S-14°S).

244

245 **3.3 Changes in the structure of the SPA**

246 As suggested by previous studies (e.g. Rykaczewski et al., 2015; Belmadani et al., 2014), changes in the SPA are
247 expected to impact the nearshore winds in the PUS. In this section the characteristics of the SPA are investigated in
248 the WRF105 simulation (which provides lateral boundary conditions to WRF21). The SPA reaches its southern-
249 most position and is weakest in summer (Fig. 7a). It is shifted southwestward and sea level pressure (SLP) at its
250 center increases by ~2 hPa under RCP8.5 scenario with respect to the retrospective period (1994-2003). This coin-
251 cides with a slight SLP decrease (~-1 hPa) along the Peru coasts. In contrast, the SPA is located at its northernmost
252 position and is more intense during winter (Fig. 7b). Its position does not change much under RCP8.5 (black lines
253 in Fig. 7b) but SLP increases by ~2 hPa at its center. These changes may impact the alongshore pressure gradient
254 off Peru, a potential driver of alongshore wind changes.

255

256 **3.4 Climate-driven changes in the momentum budget**

257 In order to investigate the dynamical processes associated with the wind change off Peru, a monthly alongshore
258 momentum budget (see section 2.3) is performed in four coastal boxes of ~100 km width between 5-8° S, 8-13°S,
259 13-14°S and 14-17°S using WRF7 outputs (see boxes in Fig. 8a) for summer (Fig.8) and winter (Fig.9). The main
260 equatorward force in the alongshore wind direction (i.e. equatorward) is the alongshore pressure gradient, (light
261 blue bars in Fig. 8b,9b), that is mainly balanced by the vertical mixing term (i.e. friction over the sea surface, yel-
262 low bars in Fig. 8b,9b) during both seasons. This balance takes place along most of the coasts except between 13-
263 14°S (off Pisco bay). There, the equatorward pressure gradient is much weaker: advection of momentum drives the
264 coastal wind and is balanced by friction. Note that the alongshore component of the Coriolis force is weaker than
265 the other terms as the wind direction is mainly alongshore.

266

267 In RCP8.5 summer conditions, the pressure gradient decreases all along the coast (light blue bars in Fig. 8c). The
268 decrease is stronger between 14–17°S than in other coastal sectors. The alongshore pressure gradient decrease is

269 consistent with the SPA poleward shift during summer (Fig.7a). The pressure gradient weakens more between 5°S-
 270 8°S than between 8°S-13°S, suggesting that at low latitudes (5°S-8°S) another process may take place over the ef-
 271 fect of the SPA poleward displacement (see section 3.6). The pressure gradient decrease is mainly equilibrated by a
 272 change in Coriolis force near 13°S-17°S, due to the change of wind direction toward the land (Fig.8a). This direc-
 273 tion change could be attributed to the meridional shift of the SPA (Fig.7a).

274

275 In RCP8.5 winter conditions, the pressure gradient tends to increase between 13°S–17 °S and less between 8-13
 276 °S, in contrast with summer conditions (Fig.9c). This is in line with the large scale SLP intensification during this
 277 season (Fig.7b). Advection of momentum also tends to increase near Pisco Bay (13°S-14°S) but also along a large
 278 coastal sector (8°S-13°S), playing a stronger role than the pressure gradient on the wind acceleration.

279

280 **3.5 Local wind intensification to the north of Paracas peninsula (13°S-14°S)**

281 The coastal area north of the Paracas Peninsula and Pisco area (13°-14°S) is located downstream the main up-
 282 welling area off Peru and stands out as a segment where the momentum balance is different from other coastal ar-
 283 eas (Figs. 8b, 9b). Furthermore it is the region of maximum wind winter intensification under RCP8.5 conditions
 284 (Figs. 6c, f). For these reasons, the momentum budget deserves a particular examination here (it is also interesting
 285 to mention that this area sustains very important tourism and aquaculture activities; e.g. Aguirre-Velarde et al.,
 286 2019). In order to investigate in more details the respective contributions (zonal, meridional and vertical advection)
 287 to the total meridional momentum advection term driving the wind change in this area, we computed these terms
 288 offline from the velocities in present and RCP8.5 future conditions, using the following:

$$289 \quad V_{ADV} = -U \frac{\partial V}{\partial X} - V \frac{\partial V}{\partial y} - W \frac{\partial V}{\partial Z} \quad (3)$$

290 where V_{ADV} is the total advection of meridional momentum, $-U \frac{\partial V}{\partial X}$ is the zonal advection term, $-V \frac{\partial V}{\partial y}$ is

291 the meridional advection term, and $-W \frac{\partial V}{\partial Z}$ is the vertical advection term (of the meridional momentum).

292

293 We found that downwind of the Paracas peninsula, the advection term accelerates the meridional wind in late fall-
 294 early austral winter (May-August) present conditions (Fig. 10a). Similar results have been reported in modelling
 295 studies of expansion fans near capes (e.g. Perlin et al., 2011; Soares et al. 2018). The meridional wind acceleration

296 is mainly due to meridional advection (Fig. 10b), while zonal advection is weak, excepted in the inner bay where
297 the intensities are comparable (not shown). The contribution of the vertical advection term is negligible every-
298 where (not shown). Expectedly, the large-scale pressure gradient forced by the SPA is unable to generate a strong
299 pressure gradient in and slightly north of the Paracas bay (Figs. 8b, 9b) as this area is shielded by topographic ob-
300 stacles blocking the pressure force, as illustrated by a much weaker wind change in WRF21 (0.5 vs 1.0 m s⁻¹; Fig.
301 6cf).

302

303 Under RCP8.5 conditions, the acceleration increase due to the advection term increase (Fig. 10c) is mainly related
304 to the change in meridional advection ($-V \cdot \partial V / \partial y$; Fig. 10d), in line with the increase in alongshore velocity off-
305 shore of the bay (Fig. 9a). Small-scale positive changes in zonal and vertical advection occur only nearshore just
306 north of Paracas bay and around the peninsula, respectively (not shown).

307

308 **3.6 Sensitivity of coastal wind change to alongshore SST gradient change**

309 In the following, the sensitivity of wind change to alongshore SST gradient changes is investigated as in Chamorro
310 et al. (2018) under El Niño conditions. These authors showed that an equatorward alongshore temperature gradi-
311 ent, such as the one encountered in the presence of anomalously warm waters in the north of Peru, may generate a
312 pressure gradient and an equatorward wind anomaly. As shown in Figure 3b, the slight decrease of the alongshore
313 SST gradient (~ -0.2 °C per 1000 km, i.e. the SST increase is slightly weaker in the north than in the south of Peru)
314 may contribute to weaken the alongshore pressure gradient and thus the wind intensity during summer (Fig. 8),
315 particularly in the north of Peru, and possibly to compensate part of the wind increase during winter (Fig. 9).

316

317 Figure 11 shows Hovmöller latitude-time diagrams of the absolute (colours) and relative (contours) change in
318 alongshore wind for the three WRF sensitivity experiments described in section 2.2. The simulation WRF-cnrm
319 was forced by a stronger negative alongshore SST gradient change than the ensemble mean (~ -0.5 °C per 1000 km,
320 see bar n°11 in Fig.3b). It shows a weakening of the alongshore wind in the north coast with a relative wind change
321 of -15% and -10% in summer and spring respectively (Fig. 11a). The winter wind increase in central Peru (9°S-
322 13°S) is also weaker than in the ensemble-mean simulation (Fig.6c). In contrast, the simulation WRF-csiro was
323 forced by a positive alongshore SST gradient change (~ 0.6 °C per 1000 km, see bar n°12 in Fig.3b), of opposite
324 sign than the ensemble mean. This results in a sustained intensification of the alongshore winds in the north of
325 Peru (5°S-10°S) with maxima in relative wind change of +30% and +20% in summer and spring respectively (Fig.
326 11b). The third simulation, WRF-cesm1, forced by a very weak (~ 0.05 °C per 1000 km, see bar n°8 in Fig.3b)

327 coastal alongshore SST gradient change, shows alongshore wind relative changes less than 10% (Fig. 11c). The
328 patterns are quite similar to those of the control simulation (Fig.6c), however the wind slackening in summer is
329 weaker than that of the control simulation, and the wind increase in winter is stronger than that of the control simu-
330 lation. Note that the alongshore wind changes in the southern sector (14°S-18°S) do not vary as much in the three
331 sensitivity experiments and in the control simulation, which suggests that the wind changes off southern Peru are
332 less forced by SST changes, and likely rely more on large scale changes associated with the SPA (which are simi-
333 lar in the four simulations).

334

335 These results confirm that alongshore wind projections in the north and central coast off Peru are sensitive to
336 changes in alongshore SST gradient, as during El Niño events (Chamorro et al., 2018). The seasonal response of
337 the alongshore winds to the local alongshore SST gradients is further investigated in detail in the northern, central
338 and southern coastal sectors. In most cases, SST gradient changes are weaker than 1°C per 1000 km in the northern
339 sector, but strong SST and wind changes may occur during summer (Fig.12a). SST gradient and wind change oc-
340 cur all year long in the central sector (Fig.12b). The wind response to local SST change is stronger in summer-
341 spring ($r=0.84-0.87$) than in fall-winter ($r=0.53-0.7$). Indeed, as the SPA is located further south in summer (Fig.
342 7a), the local pressure gradient driven by the alongshore SST gradient may play a more important role than the
343 large-scale one associated with the SPA. As shown in Fig.11, the relation between changes in alongshore SST gra-
344 dient and coastal wind is non-existent in the south sector ($r= 0.01$, Fig. 12c).

345

346 **4. Discussion and conclusions**

347 **4.1 Summary**

348 Regional dynamical downscaling using a high resolution regional atmospheric model was performed to study the
349 impact of climate change on upwelling-favorable surface winds in the PUS. The WRF regional model was applied
350 using three nested domains with 105, 21 and 7 km of horizontal resolution. We performed a control simulation
351 forced by the NCEP2 reanalysis for the period 1994-2003, and a climate change simulation where the ensemble-
352 mean monthly anomalies from thirty-one CMIP5 global models forced under RCP8.5 were added to the NCEP2
353 lateral boundary conditions and SST. The high- resolution atmospheric simulation (WRF7) reproduced well the in-
354 tensity and seasonal variability of alongshore winds, as well as the mean wind drop-off near the coast. The down-
355 scaling of the RCP8.5 ensemble resulted in weak changes in the annual-mean alongshore winds off Peru with re-
356 spect to the control simulation. In general, winds tended to decrease weakly in spring-summer ($\sim 0-5\%$) and to in-
357 crease weakly in fall-winter ($\sim 5-10\%$), which corresponds to a slight intensification of the seasonal cycle. The

358 WRF7 simulation was able to represent a local relatively strong intensification of the nearshore winds between
359 ~10-14°S during winter thanks to its fine grid.

360

361 A momentum balance shows that over most of the coastal region the alongshore pressure gradient is the dominant
362 force driving the surface wind acceleration. Under the climate change scenario, the pressure gradient decreases in
363 summer and increases in winter, driving the weakening and/or intensification of the winds off the coast respec-
364 tively. As the nearshore region between ~13-14°S located north of the Paracas-Pisco bay area is protected by
365 coastal topography, horizontal advection of momentum plays a crucial role on the wind dynamics. In particular, the
366 enhanced offshore momentum during winter is advected to the Pisco bay area, resulting in a strong local wind in-
367 tensification (~10 %). Note that the wind remains nevertheless weaker downwind than upwind of the cape.

368

369 RCP8.5 changes in alongshore winds seem to be driven for a good part by changes in the large-scale atmospheric
370 circulation, which may impact the alongshore pressure gradients off Peru. The decrease of the RCP8.5 PUS coastal
371 winds during summer could be associated to the SPA south-westward displacement, even though the SPA becomes
372 slightly more intense (+2 hPa, Fig. 7a). In contrast, the coastal winds intensification during winter is consistent
373 with the strong SPA intensification (+4 hPa, Fig. 7b). These results suggest that the projected coastal winds off
374 Peru are sensitive to the changes in the position and intensity of the SPA, in agreement with previous studies (Bel-
375 madani et al., 2014).

376

377 Sensitivity simulations using different SST conditions than the ensemble-mean indicate that the alongshore SST
378 gradient acts as a local forcing superimposed onto the SPA large-scale forcing. It is able to modulate significantly
379 coastal winds off central-north Peru (e.g. like the intensification of coastal winds due to an increased north-south
380 SST difference during El Niño studied by Chamorro et al., 2018). Long-term changes in the alongshore SST gradi-
381 ent off Peru are likely related to the evolution of El Niño under climate change: we could speculate that in case of
382 more frequent or more intense Eastern Pacific El Niño events (Cai et al., 2014), a stronger SST gradient and more
383 intense coastal winds may be expected, whereas more frequent Central Pacific events may trigger the opposite situ-
384 ation (Yeh et al., 2009). In addition, the alongshore SST gradient off Peru could also have played a role in the re-
385 cent SST trends described by Gutierrez et al (2011), since the cooling off the Central-South coast and the warming
386 off the North coast (i.e. strengthening of the alongshore SST gradient) could have forced a coastal wind intensifica-
387 tion, which in turn would force a coastal cooling. So, this may imply a negative feedback on the SST warming off

388 the coast (Echevin et al., 2020) and a positive feedback on coastal upwelling under climate change. However the
389 strength of these feedbacks remain to be investigated.

390

391 **4.2 Roles of the land-sea thermal contrast and low-latitude precipitation**

392 Bakun (1990) and Bakun et al. (2010) proposed that the land-sea thermal contrast may increase under climate-
393 change and induce alongshore wind increase. This would result from a stronger heating of the land mass, driving a
394 low pressure over land and enhancing the land-sea pressure gradient and the related along-shore geostrophic wind.
395 Wang et al. (2015) found a marked correlation between alongshore wind increase and land-sea thermal contrast at
396 higher latitudes in different EBUS (note that most of the Peru upwelling system was not included in the studied lat-
397 itude range) suggesting a link between nearshore winds and thermal contrast in those areas. Using a 50 km resolu-
398 tion model in the Peru region, Belmadani et al. (2014) found an increase in land-sea thermal contrast all along the
399 Peru-Chile coasts and a reduction of the alongshore pressure gradient and coastal winds off Peru. Our 7-km model
400 (WRF7) simulated an increase of the land-sea thermal gradient by ~20-25% (i.e. the same order of magnitude,
401 shown in Fig.8c from Belmadani et al., 2014) during summer and by 30-50% during winter (Figure not shown).
402 Thus, while we can not rule out a potential role of the land-sea thermal contrast in winter, it is clearly not the domi-
403 nant process during summer as winds decrease along most of the coastal range (Fig. 6).

404

405 Besides, Belmadani et al. (2014) suggested that part of the summer wind decrease in the north of Peru may be
406 caused by an increase of precipitation associated with convection, which produces surface wind convergence and
407 equatorward wind reduction. Our simulations also show an increase in precipitation in summer (not shown), but it
408 occurred more than 200 km away from the Peru coast, so that its contribution to the coastal winds decrease in sum-
409 mer may not be as strong as in Belmadani et al. (2014) simulations. Note that this precipitation effect depends on
410 the amplitude of the RCP8.5 SST anomalies in the north of Peru, and are not directly linked to the alongshore SST
411 gradient.

412

413 **4.3 Role of large-scale versus local processes on CMIP5 coastal wind projections**

414 We showed that changes in the ensemble-mean large-scale conditions (SLP) and local conditions (SST alongshore
415 gradient) strongly impact on the coastal winds in our downscaling approach. Here we investigate whether the dy-
416 namical relation between coastal wind change and SST gradient change may also be valid for the individual mem-
417 bers of the CMIP5 ensemble. In addition, we analyze the contribution of the large scale SLP change to the wind

418 changes. To this aim, we computed the large-scale sea level pressure (SLP) meridional gradient change between
419 the tropical and subtropical regions and the alongshore SST gradient change off the Peru coast (computed as the
420 difference of the SST averaged between 5-9°S and 9-13°S in a 200-km coastal band) for the 31 CMIP5 global
421 models (Fig. 3b). The large-scale SLP meridional gradient was computed as the difference between the SLP maxi-
422 mum value in the subtropical eastern Pacific (i.e. the highest pressure in the center of the SPA) and the SLP mini-
423 mum value in the tropical Pacific (between 0 and 20°S) at the same longitude of the center of the SPA (Fig 13a).
424 We found that the coastal wind change off Peru coast from CMIP5 global models correlates significantly ($r=0.57$,
425 $p<0.01$) with intermodels large-scale SLP meridional gradient change. The correlation is stronger in winter
426 ($r=0.68$) and not significant in summer (Fig 13b). This is consistent with Rykaczewski et al. (2015) who found
427 similar results for the southern Humboldt system (16°S-40°S). On the other hand, CMIP5 wind change and SST
428 gradient change correlations are not significant ($p>0.01$; Fig. 13c). Global models are thus not able to reproduce
429 the nearshore wind variability driven by the alongshore SST gradient change, likely due to a poor representation of
430 the nearshore wind structure at coarse-resolution. This makes a strong argument in favor of the dynamical down-
431 scaling approach that was used.

432

433 **4.4 Limitations and future work**

434 A first limitation of our methodology lies in the construction of the SST field forcing WRF. Anomalies of the SST
435 seasonal cycle from each CMIP5 model were interpolated from their ocean model grid onto a common 1°x1° grid,
436 regardless of the resolution and coastline of the CGCM ocean model. It must be noted that differences between
437 CGCM coastline shapes and positions can be quite important. The common 1°x1° SST anomalies were then lin-
438 earlyly extrapolated in the nearshore region to force our high-resolution atmospheric climate change simulations.
439 The two methodological steps could introduce unrealistic SST patterns that can impact on the coastal wind. This
440 predicament can be solved by using a regional ocean-atmosphere coupled model with the same resolution in the
441 ocean and atmosphere (e.g. as in Oerder et al., 2016, 2018) in which the ocean dynamics could correctly represent
442 the nearshore SST field. Such an approach would also allow ocean-atmosphere feedbacks, which could modify the
443 nearshore surface temperature and winds.

444

445 A second obvious limitation is the use of a single set of boundary conditions for RCP8.5 (the ensemble-mean in
446 the present case). In the presence of a large variety of large-scale SLP gradients (Fig.13b), it would be interesting
447 to select a limited number of models and investigate the impact of different large-scale conditions (e.g. different

448 SLP evolutions) from different large grid-scale climate models. This would be needed to improve the consensus on
449 the wind projections in the Peruvian upwelling system for the end of the 21st century.

450

451 **4.5 Conclusions**

452 Using a dynamical downscaling method, we found that the climate-driven large-scale changes associated with the
453 SPA impact coastal wind changes in the Peruvian Upwelling System but that these changes can be modulated by
454 coastal topography and by alongshore sea surface temperature gradients. We have also shown that dynamical
455 downscaling at scales of the order of 10 km is mandatory to tackle such subtle effects. In addition, this study has
456 examined in details the impact of the RCP8.5 scenario in the PUS, the northern part of the Humboldt Current Sys-
457 tem that has been largely ignored in recent studies about future wind changes in EBUSs (e.g. Sydeman et al., 2014;
458 Rykaczewski et al., 2015; Wang et al., 2015; Oyarzun and Brierley (2019) have systematically ignored the coastal
459 sector north of 16°S). Overall, our results are partly in line with previous studies based on different and more ex-
460 treme climate scenarios: we confirm the moderate weakening of coastal winds off Peru in summer, found in previ-
461 ous studies (Goubanova et al. 2011; Belmadani et al. 2014). However, we found a wind increase during winter off
462 central Peru, contrasting with previous findings.

463

464 Such weakening of the alongshore wind in summer (and strengthening in winter) results in an enhanced seasonal
465 cycle of the coastal winds off central Peru. This may impact on the productivity of the Peruvian upwelling system.
466 Indeed the intensity of the wind controls the upwelling of nutrient-replete waters and the intensity of surface pro-
467 ductivity (Echevin et al. 2008). Thus regional climate change may induce a less productive summer season because
468 of an enhanced nutrient limitation and a less productive winter season because of an enhanced light limitation due
469 to a thicker mixed layer (Echevin et al., 2008). Thus, a straightforward perspective of the present work is to study
470 the impact of the downscaled RCP8.5 wind changes on the Peruvian ecosystem by forcing a regional biophysical
471 model following the approach of Echevin et al. (2020). Furthermore, the wind weakening off Northern Peru in
472 summer could generate more frequently coastal warming conditions, known as ‘Coastal El Niño’ (Takahashi and
473 Martínez 2017; Garreaud 2018; Echevin et al., 2018) in the future, which should be investigated using a regional
474 ocean-atmosphere coupled model. These issues will be the subject of future works.

475

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648

649 **7. Tables**

650 Tabla 1. Name and characteristics of the CMIP5 models. The numbers are used in Figures 3 and 13.

651

652 **8. Figure captions.**

653 Figure 1. Projected changes in a) mean wind speed (in m s^{-1} , colors) and wind direction (arrows), and b) annual
654 cycle of upwelling-favorable alongshore wind (color shading in m s^{-1} , contours for relative change (in %)) off
655 Peru-Chile, for the 2080-2100 period relative to the 1989-2009 period by the multi-model mean of 31 CMIP5
656 global models under the RCP8.5 scenario. Regions where more than 75% models agree on the sign of wind speed
657 change are indicated with red dots in Figure 1a. The alongshore wind is averaged in a 100-km coastal band in Fig-
658 ure 1b.

659

660 Figure 2. a) The three model domains used in the WRF simulations: d01, d02 and d03, with horizontal resolution
661 of 105, 21 and 7 km, respectively. b) Model topography (in meters) above sea level for the innermost domain with
662 7 km of resolution.

663

664 Figure 3. Projected changes for the 2080-2100 period relative to the 1989-2009 under the RCP8.5 scenario for a)
665 the mean sea surface temperature (in $^{\circ}\text{C}$) from the multi-model mean of 31 CMIP5 models, and for b) the along-
666 shore SST gradient (in 10^{-3}C km^{-1}) off Peru (5-9°S minus 9-13°S averaged within a 200-km coastal band indi-

667 cated in Fig. 3a by red boxes) from individual climate models (see Table 1) and from the multi-model mean (with
668 blue whisker indicating the intermodel standard deviation). Red bars correspond to the CMIP5 models used to run
669 sensitivity experiments.

670

671 Figure 4. Mean surface wind (in m s^{-1}) for the period 2000-2003 from a) WRF105, b) WRF21, c) WRF7 simula-
672 tions, and d) QuikSCAT satellite observations. e) Mean wind speed cross-shore profile averaged alongshore be-
673 tween 7°S and 13°S . The vertical dotted line in e) indicates the offshore distance of the satellite blind zone.

674

675 Figure 5. Mean annual cycle of alongshore wind (averaged in a coastal band of ~ 100 km) from a) WRF7 and b)
676 QuikSCAT satellite over the period 2000-2003. c) Map of the differences between model and observations
677 (WRF7-QuikSCAT). High resolution model data was interpolated on the $1/4^{\circ}$ QuikSCAT grid. Units are in m s^{-1} .

678

679 Figure 6. Mean surface wind (in m s^{-1}) under the RCP8.5 scenario (2086-2095 period) from a) WRF21 and d)
680 WRF7. Projected change (2086-2095 minus 1994-2003) in mean surface wind from b) WRF21 and e) WRF7. Pro-
681 jected relative change (color shading in m s^{-1} and contours in %) in the mean seasonal cycle of alongshore winds
682 (in a coastal band of ~ 100 km) from c) WRF21 and f) WRF7.

683

684 Figure 7. WRF105 mean sea level pressure (in hPa) for the retrospective period (1994-2003, white contours), un-
685 der the RCP8.5 scenario (2086-2095, black contours), and sea level pressure change (2086-2095 minus 1994-2003,
686 color shading) for a) summer (JFM) and b) winter (JAS).

687

688 Figure 8. a) Projected surface wind change (in m s^{-1} , 2086-2095 minus 1994-2003) under RCP8.5 scenario and b)
689 monthly alongshore momentum budget for the historical period (1994-2003) over coastal boxes indicated in Fig
690 8a, in summer (January-March). c) Projected change (in m s^{-1} , 2086-2095 minus 1994-2003) in the monthly
691 alongshore momentum budget under RCP8.5 summer conditions. ADV, PGF, COR and MIX are acronyms used
692 for advection, pressure gradient, Coriolis and vertical turbulent mixing forces contribution, respectively.

693

694 Figure 9. Same as Fig. 8 but for winter (July-September)

695

696 Figure 10. a) Total advection of meridional momentum in May-August for the retrospective period (1994-2003)
697 and b) the meridional advection term ($-V \cdot \partial V / \partial y$). Arrows in a) indicate the averaged surface winds (in m s^{-1}). Pro-
698 jected change in c) total advection of meridional momentum and in d) meridional advection term in May-August
699 under RCP8.5 conditions. Units are in 10^{-4} m s^{-2} .

700

701 Figure 11. Projected change (in m s^{-1} , 2086-2095 minus 1994-2003, color shading) in the mean seasonal cycle of
702 alongshore winds (in a coastal band of $\sim 100 \text{ km}$) under RCP8.5 conditions from WRF21 sensitivity experiments
703 forced by SST from a) CNRM-CM5, b) CESM1-CAM5 and c) CSIRO-Mk3-6-0 (see red bars in Fig.3b). Black
704 contours indicate the absolute wind change as a percentage (%) of the value of the alongshore wind over the retro-
705 spective period (1994-2003).

706

707 Figure 12. Scatter plots of monthly mean coastal wind (in %) and alongshore SST gradient change (in $10^{-3} \text{ }^\circ\text{C}$
708 km^{-1}) from the three sensitivity experiments (WRF-cnrm, WRF-csiro and WRF-cesm1 with square, circles and di-
709 amond marks, respectively) and from the ensemble-mean climate change simulation (stars). Alongshore wind in
710 the a) North ($4\text{-}9^\circ\text{S}$), b) Central ($9\text{-}14^\circ\text{S}$) and c) South ($14\text{-}18^\circ\text{S}$) of Peru are shown. Data and linear fits for each
711 seasons (Fall-Winter and Spring-Summer) are shown in blue and red colors respectively. Linear fits for the annual
712 mean values are shown in black. The change in alongshore SST gradient is computed for the north, central and
713 south coasts as the mean coastal SST difference between $4^\circ\text{S}\text{-}9^\circ\text{S}$ minus $9^\circ\text{S}\text{-}14^\circ\text{S}$, $9^\circ\text{S}\text{-}14^\circ\text{S}$ minus $14^\circ\text{S}\text{-}19^\circ\text{S}$,
714 and $14^\circ\text{S}\text{-}18^\circ\text{S}$ minus $18^\circ\text{S}\text{-}22^\circ\text{S}$, respectively. Note the change of horizontal and vertical scales in the different
715 panels.

716

717 Figure 13. a) Sea level pressure (in hPa) for the 2080-2100 period (black contours), for the historical 1989-2009
718 period (white contours), and sea level pressure change (2080-2100 minus 1989-2009, in hPa, color shading). SLP
719 is computed from the multi model mean of 31 CMIP5 models under the RCP8.5 scenario. Scatter plots of b) mean
720 coastal wind change along $7\text{-}11^\circ\text{S}$ (in %) versus the large-scale meridional sea level pressure gradient change (in
721 $10^{-4} \text{ hPa km}^{-1}$), and c) mean coastal wind change versus the alongshore SST gradient change (in $10^{-3} \text{ }^\circ\text{C km}^{-1}$)
722 off Peru coast. Each number (from 1 to 31) indicates a particular CMIP5 model value under the RCP8.5 scenario
723 (see Table 1 and Figure 3). Black, red and blue numbers correspond to annual, summer-spring and fall-winter

724 mean values, respectively. Linear regression lines for annual, summer-spring and fall-winter mean values are
725 shown in black, red and blue, respectively.

Figure 1.

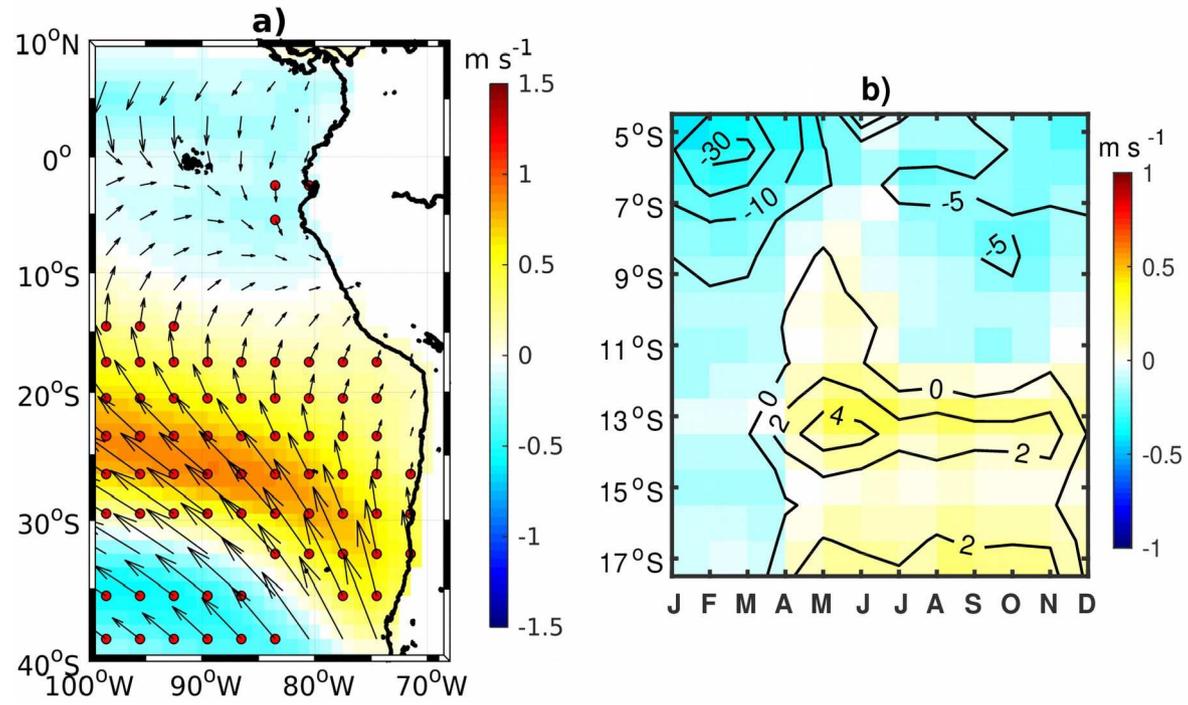


Figure 2.

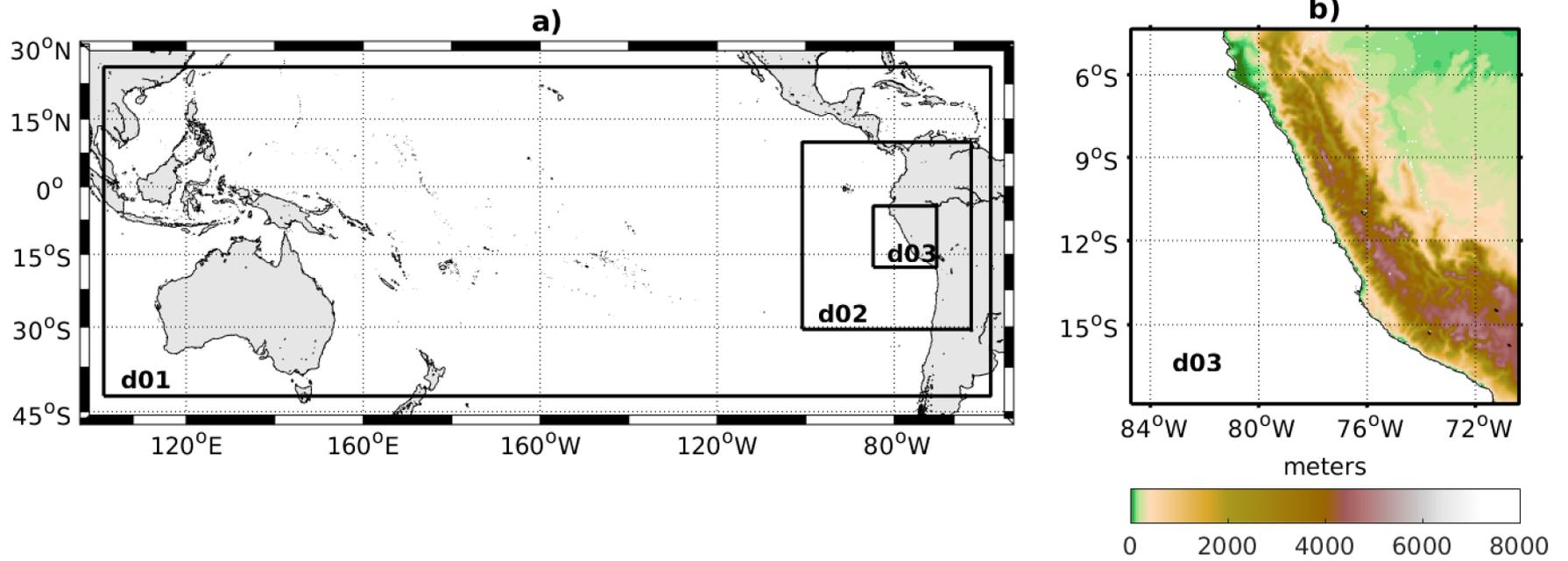


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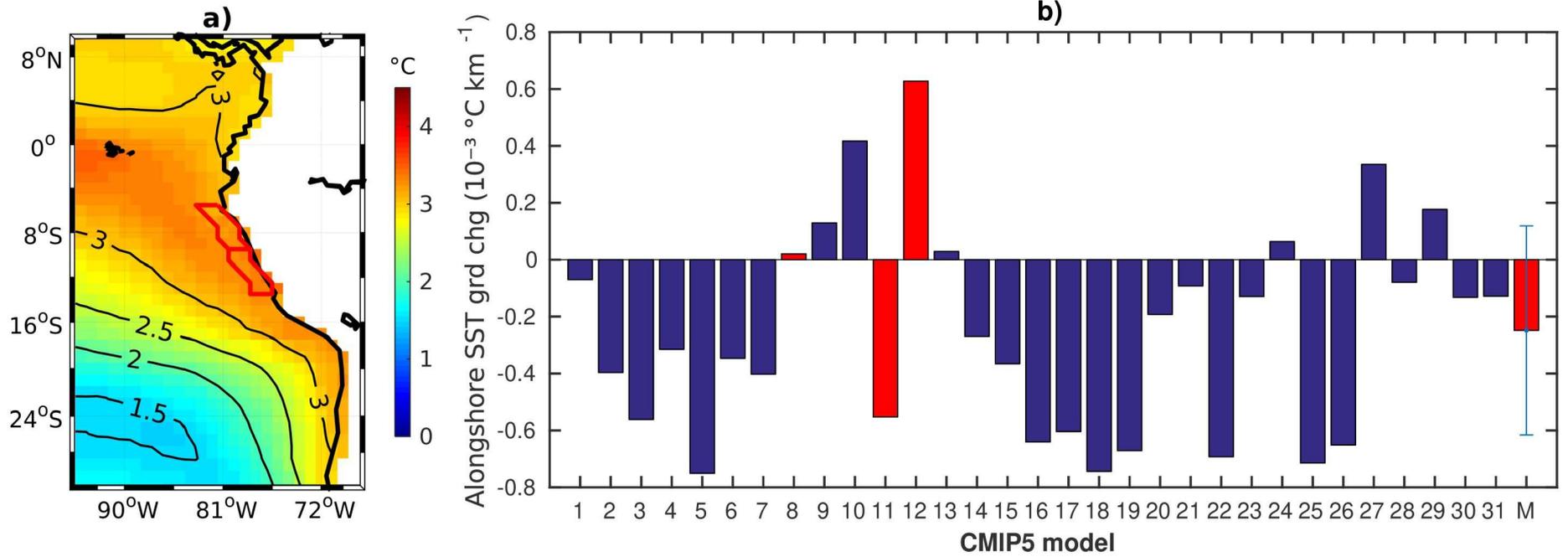


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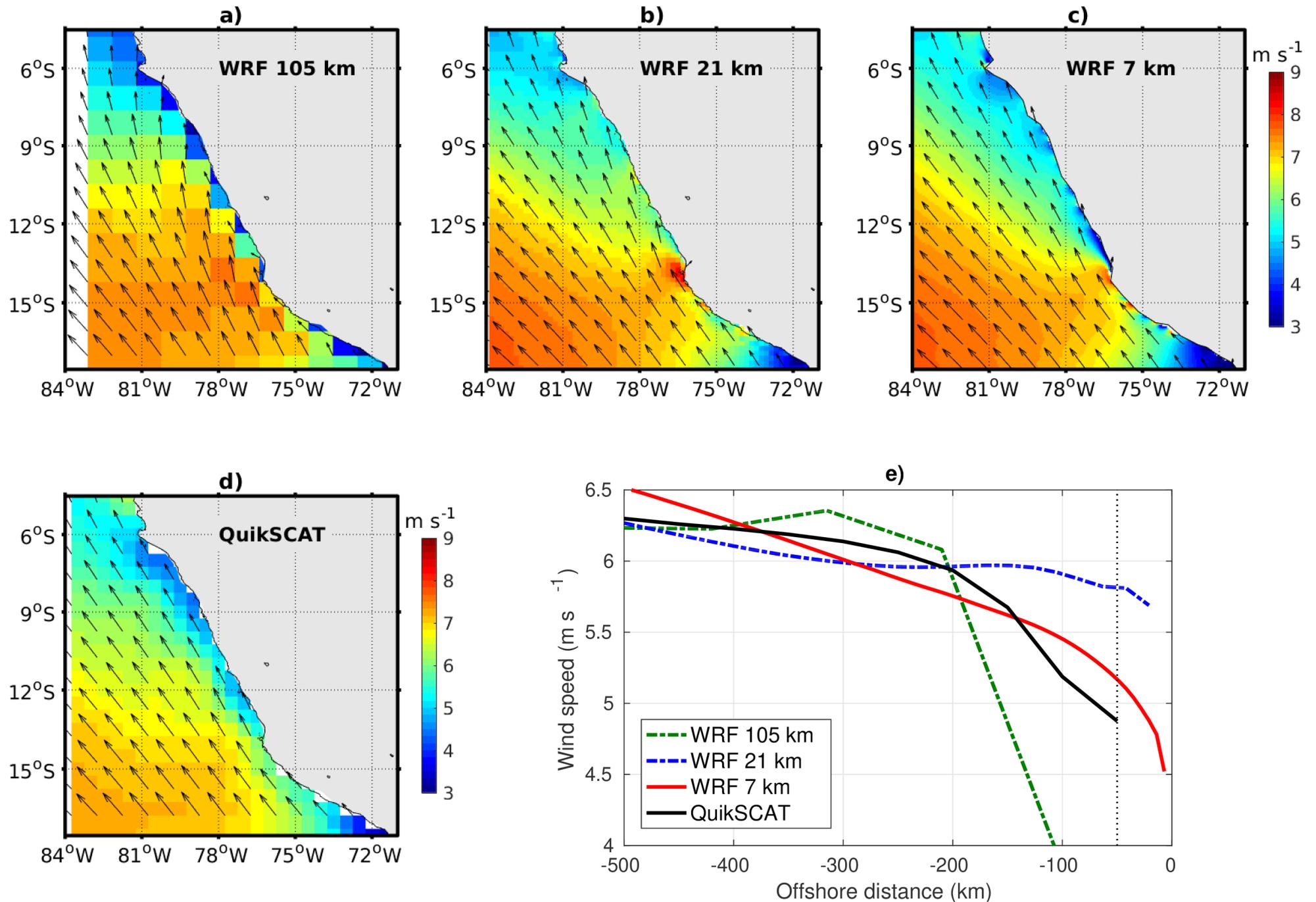


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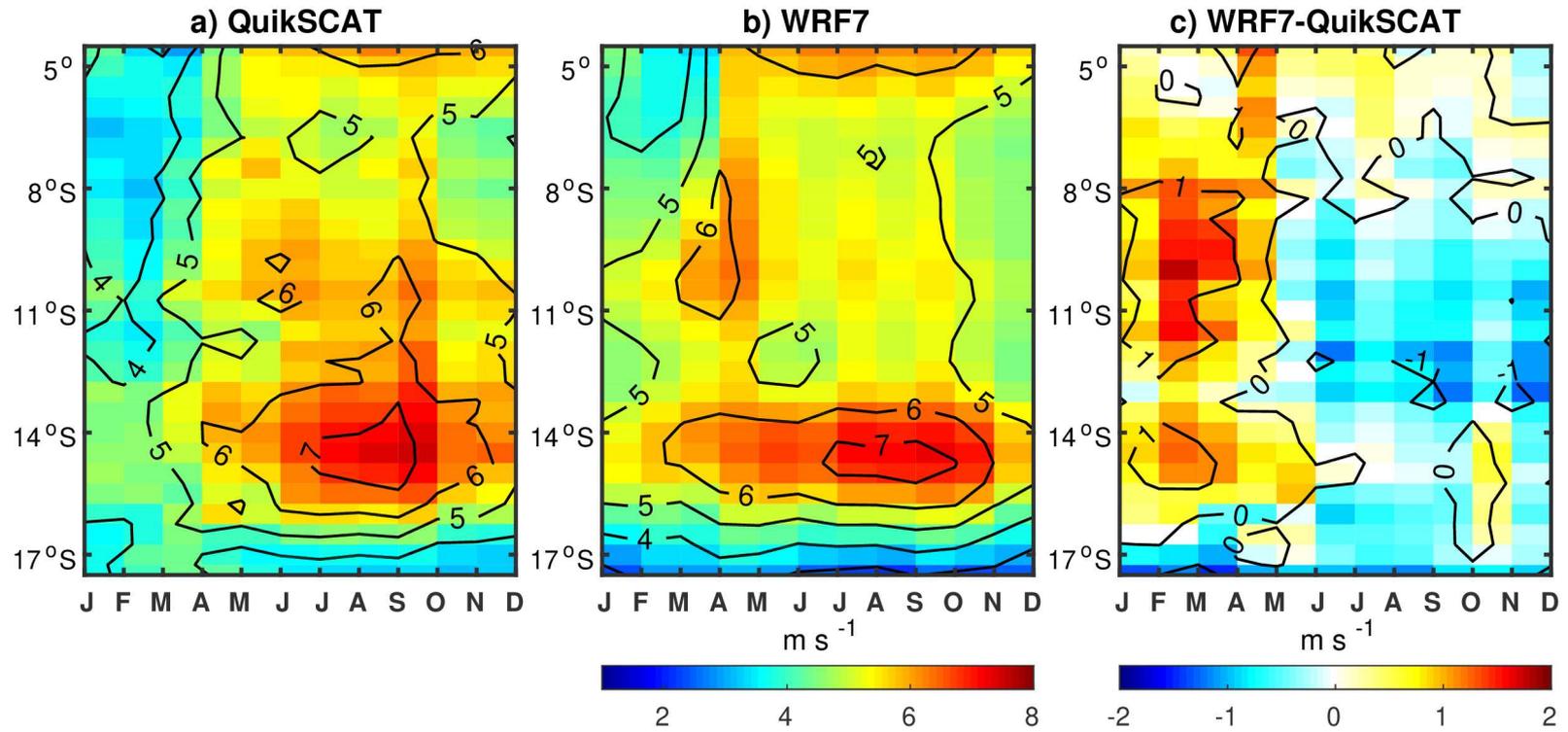


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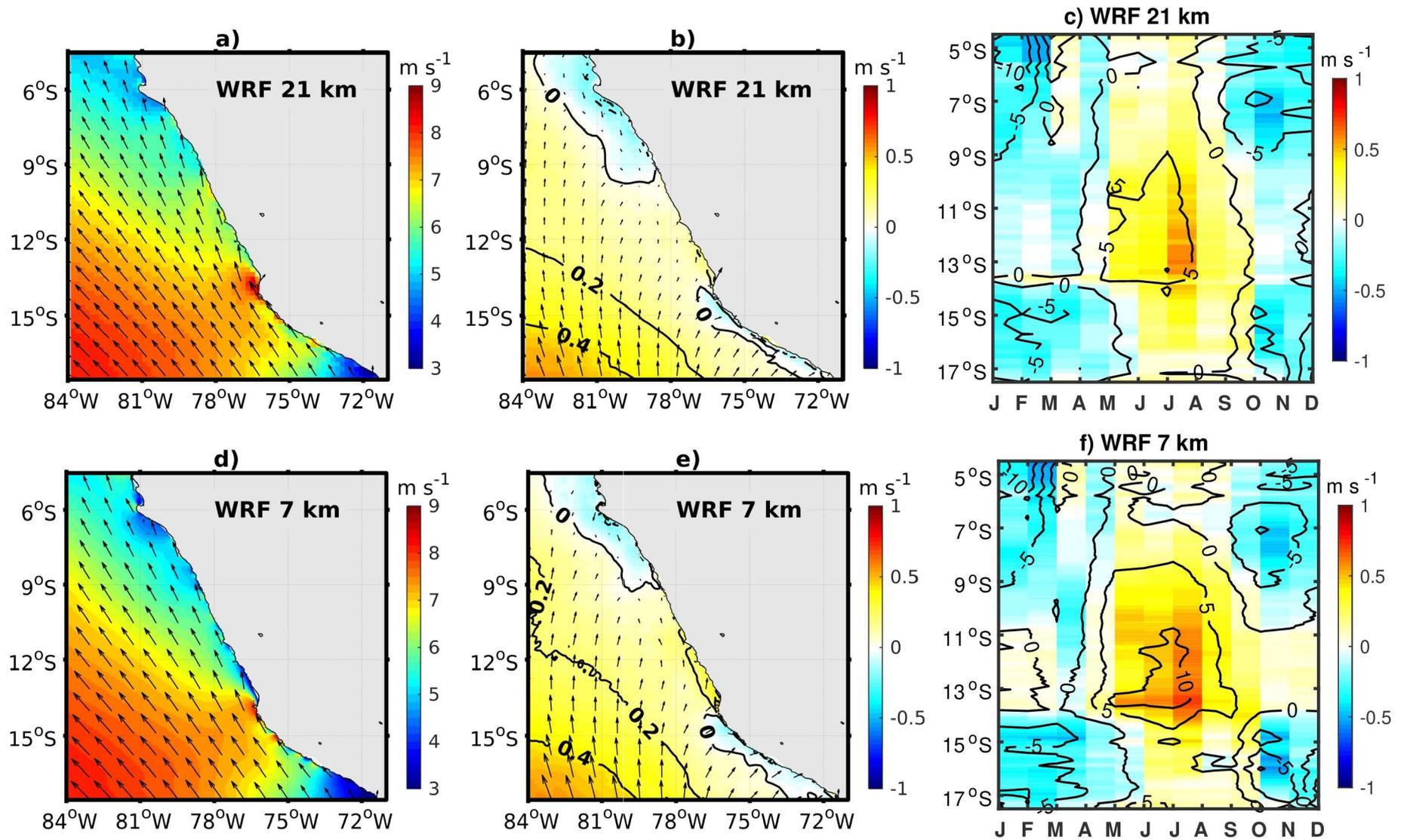


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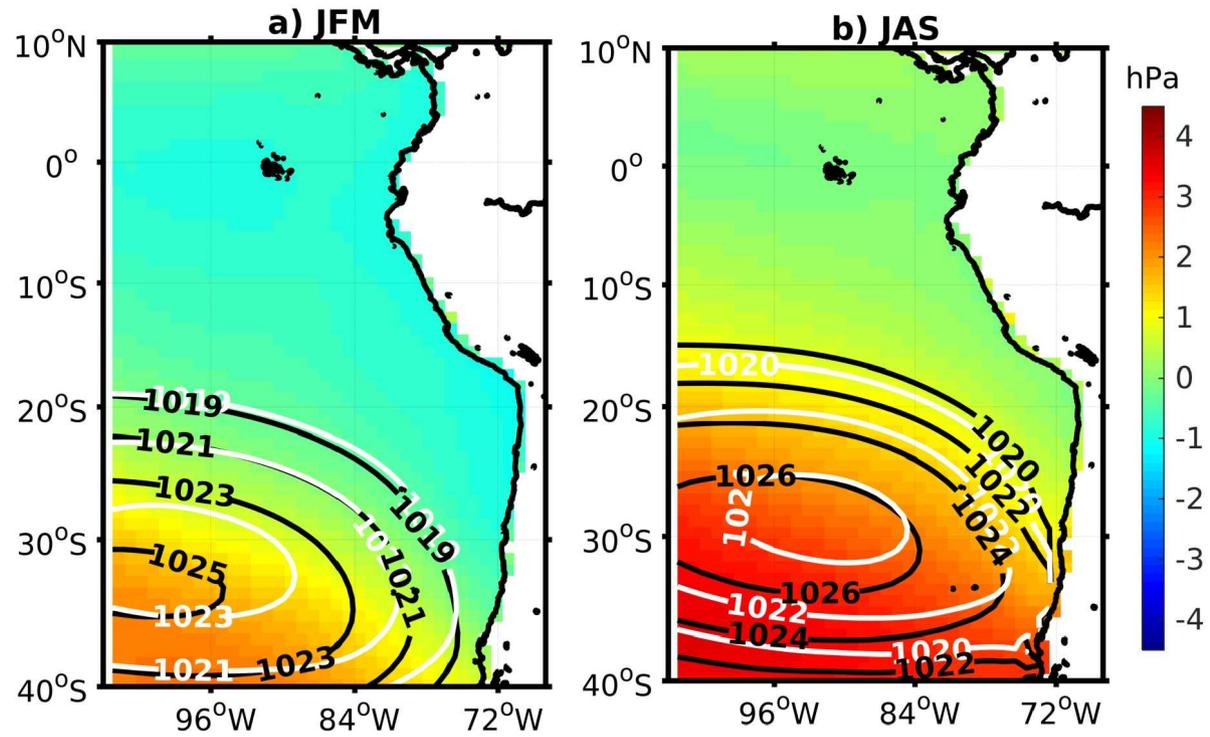


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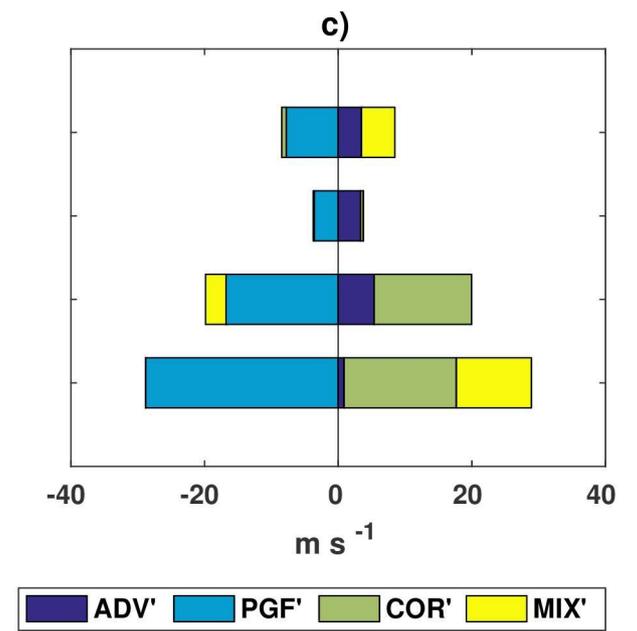
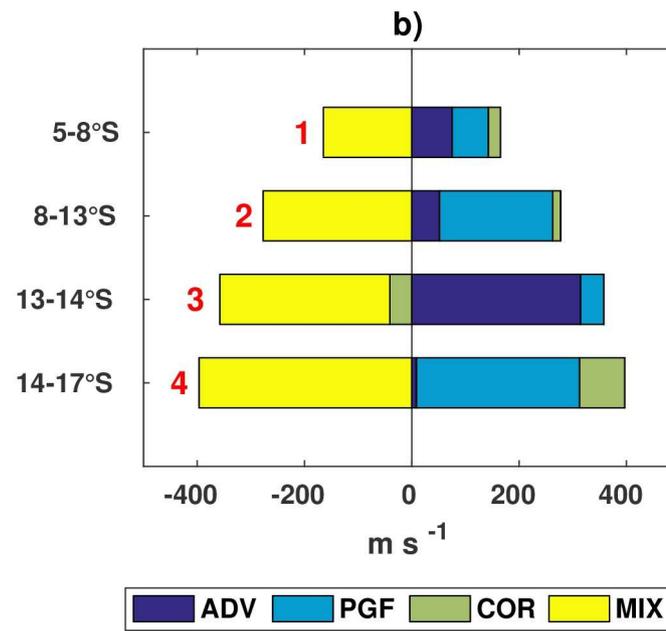
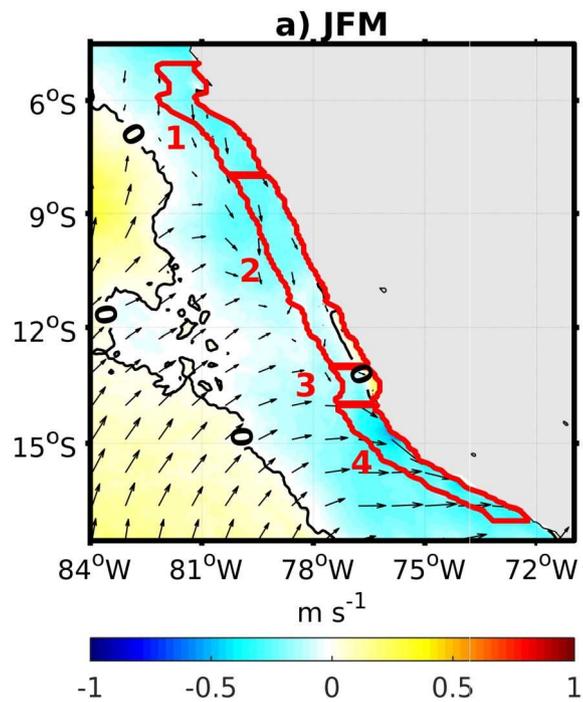


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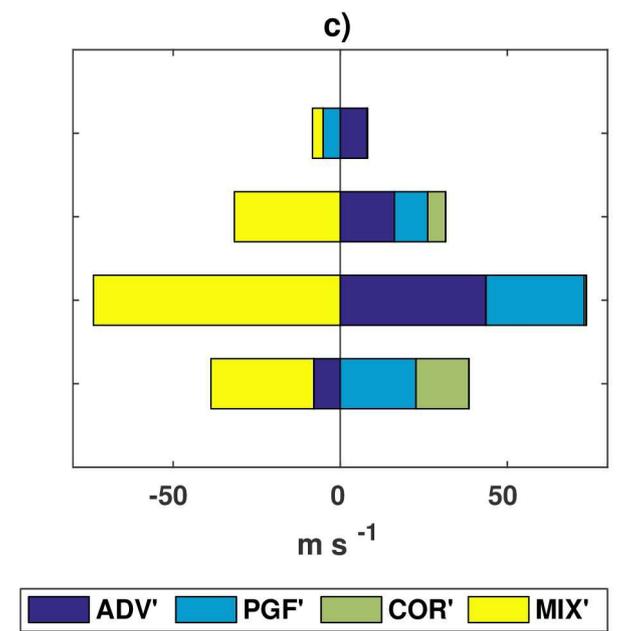
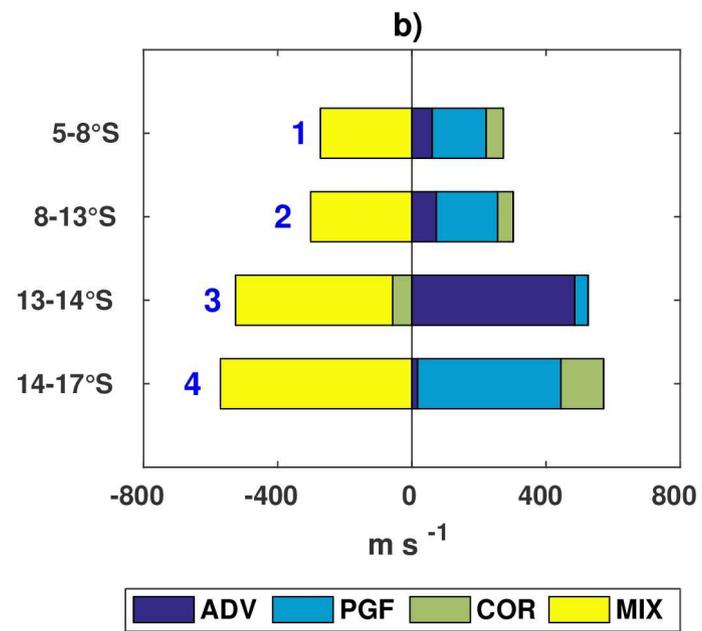
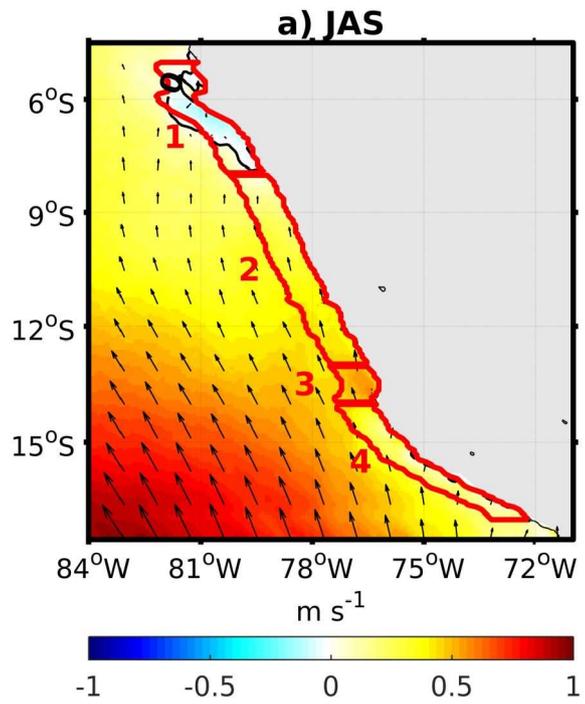


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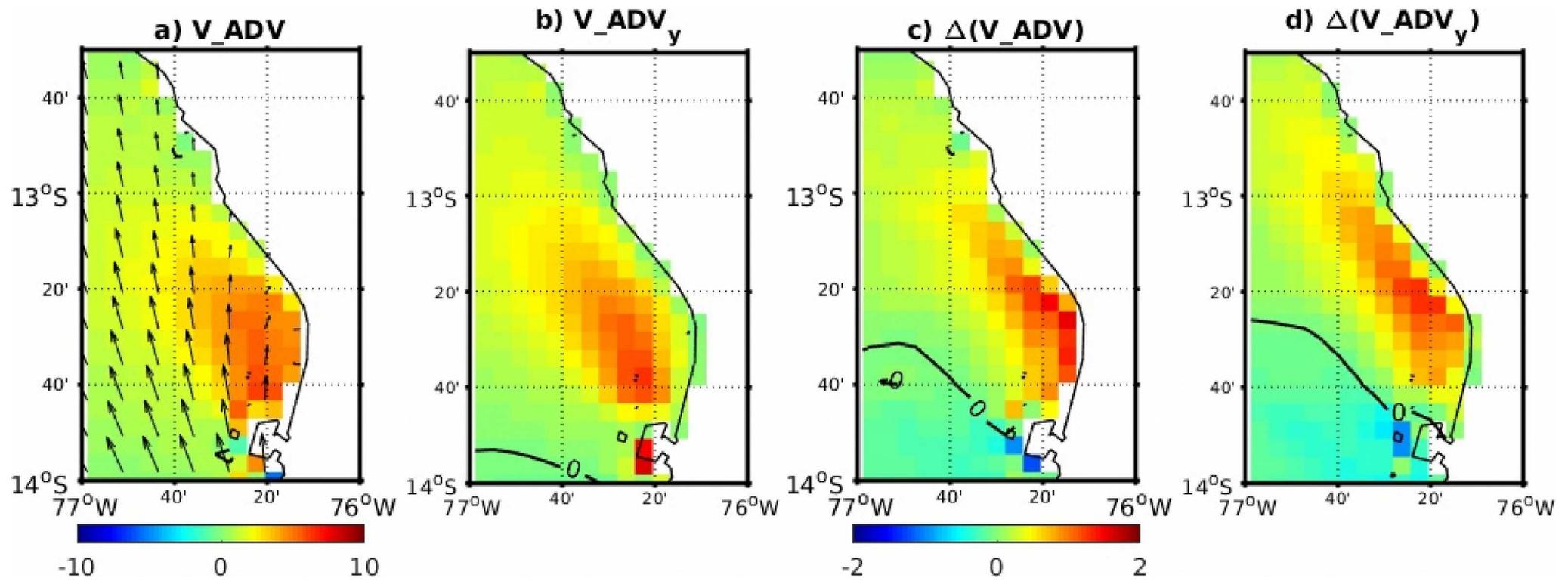


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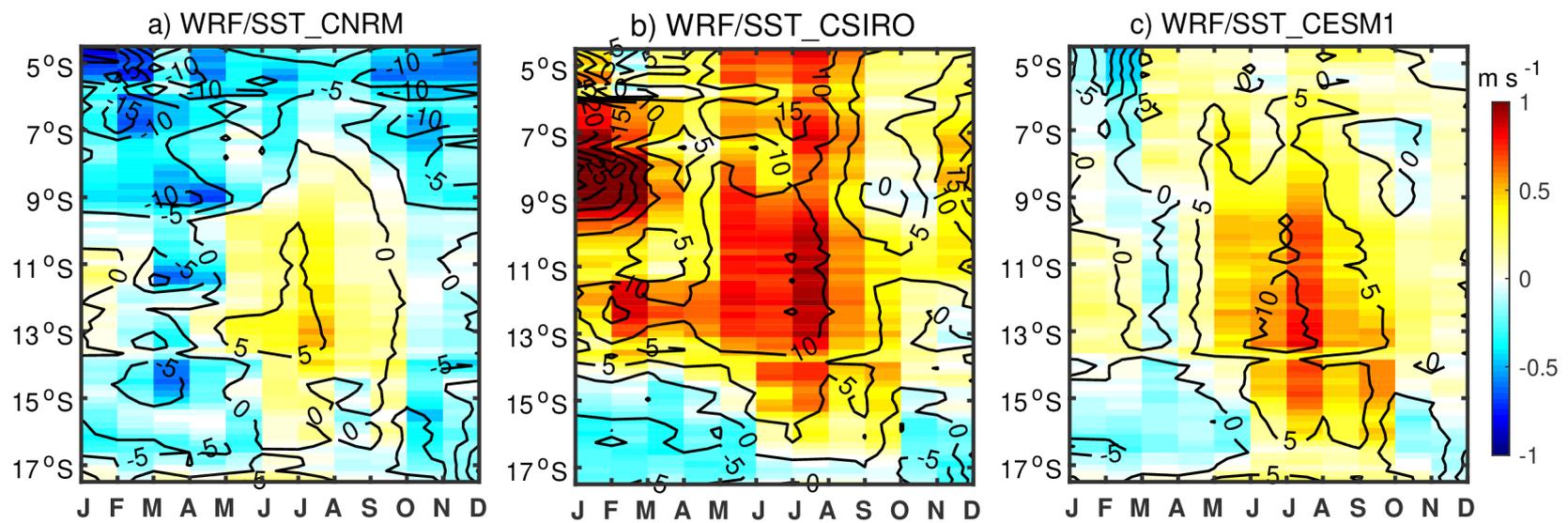


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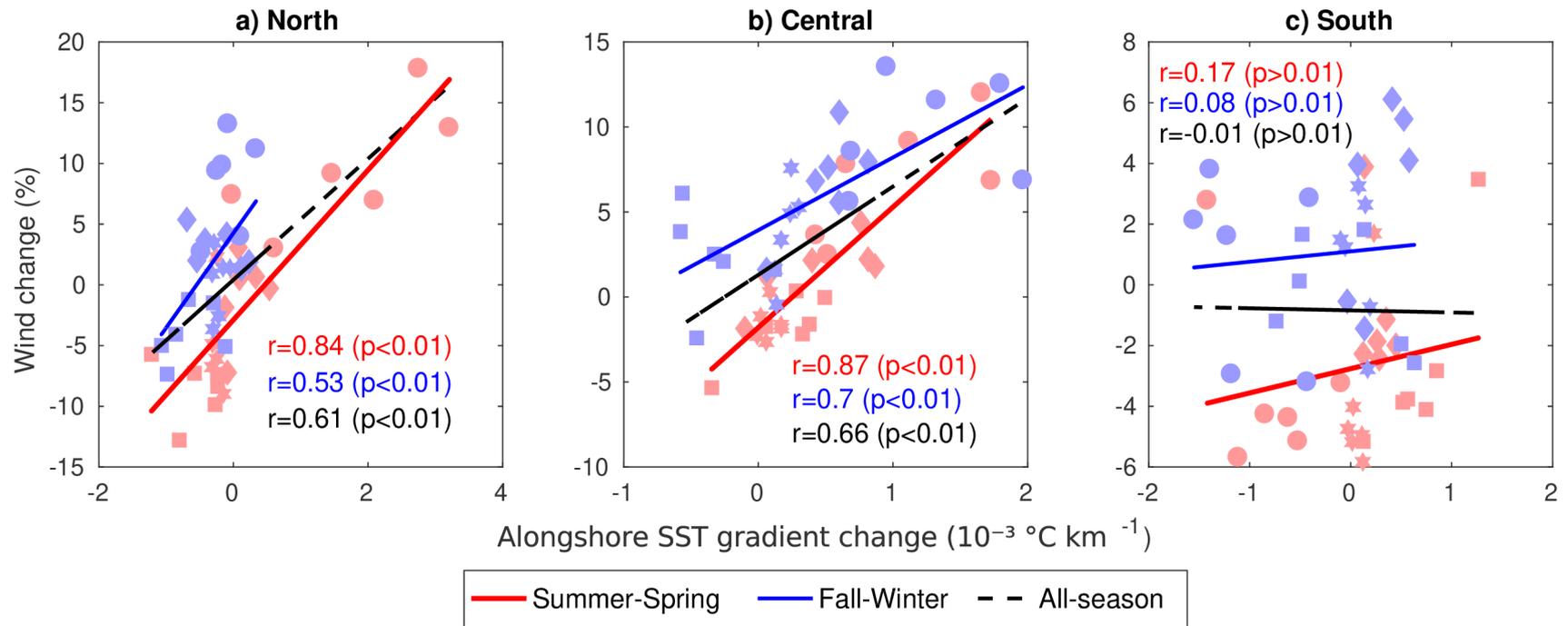


Figure 13.

