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# Impacts of the Mesoscale Ocean-Atmosphere Coupling on the Peru-Chile Ocean Dynamics: The Current-Induced Wind Stress Modulation

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- 1 **Impacts of the mesoscale ocean-atmosphere coupling on the Peru-Chile**  
2 **ocean dynamics :**  
3 **impact of the current-induced wind stress modulation**
- 4 **Vera Oerder · François Colas · Vincent Echevin ·**  
5 **Sebastien Masson · Florian Lemarié**
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## 1 Introduction

Momentum, heat and water fluxes at the air-sea interface are responsible for intense energy transfers between the atmosphere and the ocean. Processes behind these transfers vary according to the horizontal scales (*e.g.* Xie, 2004). For example, at the basin scale, negative (positive) correlations found by Liu et al (1994) between the SST and the surface wind (solar irradiance, respectively) suggested a forcing of the ocean by the atmosphere, with strong winds and thick cloud cover cooling the underlying water. This contrasts with the observations obtained above smaller scale oceanic structures that suggested an influence of the ocean on the atmosphere. Indeed, satellite data showed an imprint of the oceanic mesoscale ( $\sim 10\text{-}100$  km) on the atmospheric fields, in particular on the surface wind stress (*e.g.* Small et al, 2008).

This atmospheric response is created by both the Sea Surface temperature (SST) and the surface current mesoscale anomalies. On one hand, the surface current affects the wind stress ( $\vec{\tau}_s$ ) because it depends on the relative motion between the ocean and the atmosphere (Dewar and Flierl, 1987), *i.e.* the difference between the absolute surface wind velocity and the oceanic surface current. The intensity of the mesoscale current anomaly can locally be much stronger than the mean current and represent more than 10 % of the surface wind speed (Cornillon and Park, 2001). Consequently, wind stress satellite data show a current-induced mesoscale modulation that represents up to 20 % of the mean wind stress in regions with strong currents like the Gulf Stream (Chelton et al, 2004). In particular, above an oceanic coherent eddy, the wind stress is enhanced on the eddy side where the current's direction is opposite to the wind, and reduced on the eddy side where the current is downwind. This current-induced wind stress anomaly creates a wind stress curl anomaly centered on the eddy (see Fig. 1 from Zhai and Greatbatch, 2007; Gaube et al, 2015).

On the other hand, mesoscale SST anomalies modify the fresh water and heat fluxes at the air-sea interface (Bourras et al, 2004; Frenger et al, 2013) and also affect the wind stress (*e.g.* Chelton et al, 2001) leading to mesoscale anomalies of the wind stress intensity ( $\|\vec{\tau}_s\|'$ ) proportional to the SST anomalies ( $SST'$ ). The underlying mechanism may vary with the studied region. For example, in the South East Pacific, it has been shown that these surface wind stress anomalies result from atmospheric vertical mixing modifications due to

33  $SST'$  (Oerder et al, 2016). Such SST-induced wind stress modulation above SST fronts creates wind stress curl  
34 structures proportional to the crosswind SST gradient (Chelton et al, 2001).

35 Thus, the atmospheric response to the oceanic mesoscale modifies the air-sea exchanges that could further  
36 generate a feedback on the mesoscale ocean dynamics. This is the object of the present study. In the present  
37 paper, we focus on the current-induced effect, the SST-induced effect being investigated in a companion paper  
38 (Oerder et al, 2017).

39 It has been suggested that the eddies could be directly affected by the wind stress response to the surface  
40 current (Dewar and Flierl, 1987). Indeed, in the Southern hemisphere, the negative (positive) wind stress curl  
41 observed by Gaube et al (2015) above the warm (cold) anomalies associated with the anticyclones (cyclones)  
42 are responsible for an Ekman suction (pumping, respectively) that could damp the eddies.

43 The oceanic impact of the wind stress modulation by the surface current has been first examined using  
44 oceanic models forced by wind stress computed online using a surface wind velocity forcing and the modeled  
45 surface current velocity (*e.g.* Zhai and Greatbatch, 2007; Eden and Dietze, 2009; Munday and Zhai, 2015). These  
46 studies evidenced a reduction of the mesoscale activity, which is coherent with an eddy damping effect. However,  
47 using a forced oceanic model presents a limitation based on the difficulty to obtain a correct absolute surface  
48 wind velocity forcing (either from satellite data or from an atmospheric model). Indeed, surface winds derived  
49 from satellite data are obtained from an inversion of the wind stress observations inferred from the backscatter  
50 of the signal on the ocean surface. Thus, they correspond to a surface wind relative to the moving ocean surface  
51 (Cornillon and Park, 2001). In addition, the current-induced wind stress modulation may impact the absolute  
52 surface wind velocity through atmospheric boundary layer processes. Current-induced mesoscale variations are  
53 not present in wind velocity products (either observed or simulated) with a coarse spatial resolution, nor when the  
54 absolute wind is obtained from an atmospheric model which computes the surface stress from the absolute wind  
55 only. Besides, a high resolution absolute wind product, either observed or resulting from an atmospheric model  
56 (usually forced by remotely-sensed SST), would contain the imprint of the real mesoscale oceanic structures.  
57 This oceanic mesoscale (*e.g.* present in the SST forcing of the atmospheric model) would necessarily differ from  
58 the mesoscale structures simulated by the ocean model. Therefore, in both cases, the absolute wind forcing is

59 not coherent with the oceanic mesoscale. An appropriate way obtain adequate absolute wind conditions at the  
60 air-sea interface is thus to use an ocean-atmosphere coupled model. Two recent studies with a regional coupled  
61 model in the California Upwelling System (Seo et al, 2015; Renault et al, 2016) take into account the surface  
62 current in the wind stress calculation and confirm an impact on the ocean mesoscale revealed by a significant  
63 Eddy Kinetic Energy (EKE) reduction.

64 In the present work, a regional coupled model is configured to simulate the current-wind stress interaction in  
65 the South East Pacific and to study its impact on the ocean dynamics. This region hosts one of the most  
66 intense Eastern Boundary Upwelling System (EBUS) which are of particular interest because of their intense  
67 biological activity (Carr and Kearns, 2003). In EBUS, the mesoscale activity is particularly relevant because it  
68 contributes to a large part of the heat transport (Colas et al, 2012) and also structures the nearshore primary  
69 production (Lathuilière et al, 2010; Gruber et al, 2011). The upwelling characteristics (nearshore vertical flux,  
70 spatial extension of the upwelling zone, etc.) depend on the wind stress intensity and its spatial variations in  
71 the coastal zone (Capet et al, 2004). The wind stress also forces an alongshore, equatorward surface coastal  
72 current (Strub et al, 1998) with a mesoscale ( $\sim 150$  km) cross-shore extension that could modulate the wind  
73 stress intensity and affect the coastal dynamics. As previous studies find a strong impact of the current-stress  
74 interaction on the mesoscale activity, we pay a particular attention to the EKE generation mechanisms and  
75 the coherent eddies evolution. Our modeling approach allows us to study for the first time the impact on the  
76 3D eddy structure. The methodology is first described in Sec. 2, and the impact on the ocean mean state and  
77 mesoscale variability is analyzed in Sec. 3). Then, we focus on the current-wind stress interaction above the  
78 simulated coherent eddies (Sec. 4). Finally, we discuss our results and conclude our study in Sec. 5.

## 79 **2 Methodology**

### 80 2.1 Regional ocean-atmosphere coupled model

#### 81 *2.1.1 Atmospheric model*

82 The atmospheric dynamics is modeled using the 3.6 version of the Weather Research and Forecasting (WRF)  
83 regional model, with the ARW solver (Advanced Research WRF; Skamarock and Klemp, 2008). WRF is based  
84 on the fully compressible non-hydrostatic Euler equations with a C-grid and terrain-following mass vertical  
85 coordinates. We use the configuration described in Oerder et al (2016) with a  $1/12^\circ$  horizontal resolution.  
86 Our domain extends from  $10^\circ\text{N}$  to  $30^\circ\text{S}$  and from  $100^\circ\text{W}$  to  $60^\circ\text{W}$ . The planetary boundary layer physics  
87 are parameterized by the Mellor–Yamada–Nakanishi–Niino (MYNN) scheme (Nakanishi and Niino, 2009) and  
88 surface fluxes are computed using the MYNN surface scheme. The  $3/4^\circ$  resolution ERA-interim reanalysis (Dee  
89 et al, 2011) provides 6-hour averages for the initial and boundary conditions. For further details on the model  
90 configuration the reader is referred to Oerder et al (2016).

#### 91 *2.1.2 Oceanic model*

92 The ocean dynamics is modeled using the 3.4 version of the Nucleus for European Modeling of the Ocean  
93 (NEMO; Madec, 2008) in a regional configuration with the same  $1/12^\circ$  horizontal grid as the atmospheric  
94 model. It has 75 z-levels in the vertical, with 25 levels in the upper 100 m. Near the surface, the vertical  
95 resolution is 1 m and decreases with depth, reaching 200 m at the bottom. Tracer and momentum horizontal  
96 advection is parameterized by an upstream-biased (UBS) third order scheme (Farrow and Stevens, 1995; Webb  
97 et al, 1998). A total variance dissipation (TVD) scheme (Lévy et al, 2001) for tracers and a second order centered  
98 scheme for momentum are used in the vertical direction. The vertical mixing parameterization is based on a  
99 turbulent kinetic energy closure scheme (Blanke and Delecluse, 1993; Madec, 2008). At the open boundaries,  
100 the barotropic dynamics is processed using a Flather scheme (Flather, 1976), and relaxation towards boundary  
101 values is imposed to the baroclinic fields (Engedahl, 1995). The  $1/2^\circ$  resolution Simple Ocean Data Assimilation  
102 (SODA; Carton and Giese, 2008) reanalysis in its 2.1.6 version provides 5-day average boundary conditions

103 over the 2002-2008 period. An initial state already containing a developed mesoscale circulation is used to  
 104 reduce the spin-up duration in the coupled simulation. This initial state is obtained from a forced ocean-only  
 105 simulation over the 2000-2001 period using the NEMO configuration described above. The atmospheric forcing  
 106 for the spin-up phase is from QSCAT scatterometer (Dunbar et al, 2006) for wind stress and ERA-interim for  
 107 heat and fresh water fluxes.

### 108 2.1.3 Ocean-atmosphere coupling and simulations

109 The Ocean Atmosphere Sea Ice Sol (OASIS3-MCT, Valcke et al, 2013) coupler is used to couple WRF and  
 110 NEMO at a  $1 \text{ h}^{-1}$  frequency as described in Oerder et al (2016). The surface wind stress is computed by WRF  
 111 in the surface layer scheme using NEMO's surface velocity, and then sent to NEMO through OASIS. Thus, the  
 112 wind stress fields used by WRF and NEMO are identical (unlike in Seo et al, 2015). Routines in WRF 3.6 (and  
 113 subsequent versions) were modified (Lemarié, 2015) so that the wind stress ( $\vec{\tau}_s$ ) can be computed using either  
 114 the 10 m absolute wind ( $\vec{v}_s^a$ ) or the 10 m relative wind ( $\vec{v}_s^a - \vec{v}_s^o$ ),  $\vec{v}_s^o$  being the surface oceanic current velocity:

$$\vec{\tau}_s = \rho_a C_d \|\vec{v}_s^a\|^2 \frac{\vec{v}_s^a}{\|\vec{v}_s^a\|} \quad (1)$$

115 or

$$\vec{\tau}_s = \rho_a C_d \|\vec{v}_s^a - \vec{v}_s^o\|^2 \frac{\vec{v}_s^a - \vec{v}_s^o}{\|\vec{v}_s^a - \vec{v}_s^o\|} \quad (2)$$

116 with  $\rho_a$ , the surface air density and  $C_d$ , the drag coefficient. The drag coefficient  $C_d$  is computed in WRF surface  
 117 scheme using the roughness length and the stability function. Two simulations are compared in the following  
 118 sections. They are named Abs and Rel, and differ only by the wind stress computation using equation (1) for  
 119 Abs and (2) for Rel. The simulations are run over the period 2002-2008. The first year is considered the spin-up  
 120 so results are analyzed over 2003-2008.



#### 121 2.1.4 Simulation realism and modeled current-stress interaction characteristics

122 This coupled model configuration has been shown to realistically capture mesoscale air-sea interactions (Oerder  
 123 et al, 2016). The wind stress modulation by the surface current is characterized by a negative correlation between  
 124 wind stress curl and surface current vorticity (Chelton et al, 2004; Renault et al, 2016). The wind stress curl  
 125  $(\vec{\nabla} \wedge \vec{\tau}_s) \cdot \vec{k}$  (hereafter  $\text{curl}(\vec{\tau}_s)$ ) presents mesoscale variations induced both by the crosswind projection of the  
 126 SST anomaly gradient  $\text{grad}_{cw}(SST') = \|\vec{\nabla}(SST')\|\sin(\theta)$  (with  $\theta$  the counterclockwise angle between  $\vec{\nabla}(SST')$   
 127 and  $\vec{\tau}_s$ ) and by the surface current vorticity. Indeed,  $\text{curl}(\vec{\tau}_s)$  follows the relation (*e.g.* Chelton et al, 2001) :

$$\text{curl}(\vec{\tau}_s') \propto \text{grad}_{cw}(SST') \quad (3)$$

128 with a correlation of  $\sim 0.6$  and a regression coefficient  $s_1^c = 0.46 \cdot 10^{-2} \text{ N m}^{-2} \text{ } ^\circ\text{C}^{-1}$  (Fig. 1). We estimate the  
 129 current-induced effect, assuming that the SST and the current-induced effect sum-up linearly, by computing the  
 130 residual curl ( $\text{curl}(\vec{\tau}_s)_{res} = \text{curl}(\vec{\tau}_s) - s_1^c \text{grad}_{cw}(SST')$ ) which represents the part of the wind stress curl not  
 131 proportional to  $\text{grad}_{cw}(SST')$ .  $\text{curl}(\vec{\tau}_s)_{res}$  is negatively correlated ( $-0.64$ ) to the surface current vorticity (Fig.  
 132 1).

133 A detailed comparison of the modeled and observed, atmospheric and oceanic dynamics in the South East  
 134 Pacific can be found in Oerder et al (2016). The surface current velocity and wind stress averaged over the  
 135 2003-2008 period are shown on Fig. 2. The model (Rel simulation) simulates a coastal equatorward current of  
 136  $0.1\text{-}0.2 \text{ m s}^{-1}$ , a well-known characteristic of the South-East Pacific circulation (*e.g.* Strub et al, 1998). West of  
 137  $80^\circ\text{W}$ , between  $5^\circ$  and  $20^\circ\text{S}$ , the offshore current is westward. Its mean intensity is  $\sim 6 \text{ cm s}^{-1}$  in the model,  $\sim 10$   
 138 % weaker than the OSCAR product. The modeled wind stress presents a spatially homogeneous bias of  $\sim -0.015$   
 139  $\text{N m}^{-2}$  (it is removed in Fig. 2a for comparison with the observations). It reaches a maximum ( $\sim 0.2 \text{ N m}^{-2}$ )  
 140 nearshore south of  $25^\circ\text{S}$  and offshore between  $20^\circ\text{S}$  and  $5^\circ\text{S}$ , a pattern also found in the observations. Offshore  
 141 and south of  $25^\circ\text{S}$ , the modeled wind stress is underestimated by  $\sim 30\%$ . In spite of this discrepancy, we  
 142 consider this simulation realistic enough to carry out a detailed analysis of the simulated mesoscale processes.

## 143 2.2 Eddies characteristics

### 144 2.2.1 Coherent eddies tracking

145 Coherent oceanic eddies are detected using 1-day average surface fields, following the method described by  
146 Kurian et al (2011), and used by Colas et al (2012) and Liang et al (2012). This method is based on closed  
147 contours identification of the Q parameter, the second invariant of the horizontal velocity gradient tensor (Isern-  
148 Fontanet et al, 2003). Closed contours are fitted by circles and a shape error is defined as the closed-contour  
149 area deviation from the circle area. Coherent eddies are defined as the largest circular (*i.e.* shape error < 20%)  
150 closed-contours fitted by a circle with a radius ranging from 40 km to 120 km. The eddy center and radius  
151 are those of the fitted circle. Eddies are tracked by comparing eddy centers and properties between consecutive  
152 times. The nearest eddies with the minimum change in amplitude and surface are assigned to common tracks.  
153 Only eddies detected for at least 15 days are retained for the analysis.

### 154 2.2.2 Composites

155 To study coherent eddies characteristics, composites of the sea temperature anomaly associated to cyclones  
156 and anticyclones are computed using the following methodology. For each eddy detected on daily-averaged field,  
157 temperature anomalies relative to a 90-day running mean are calculated. The mean surface wind direction above  
158 the eddy is also computed using a 90-day running mean. The temperature anomaly field is then rotated around  
159 the eddy center, so that the mean wind direction is South-North for each eddy. Zonal and meridional axes are  
160 rescaled so that the eddy radius is equal to the averaged cyclone or anticyclone radius. Finally, all temperature  
161 anomaly fields are averaged to create the composite eddy. The same method is applied for the current velocity  
162 and vorticity.

### 163 3 Impacts of the current-stress interaction on the ocean dynamics

164 The Abs and Rel simulations are compared to investigate the ocean response to the current-stress interaction.  
165 We first examine the impact of this process on the ocean mean state, and then on mesoscale variability by  
166 quantifying the impact on the eddy kinetic energy (EKE) and its generation terms.

#### 167 3.1 Regional circulation

168 Offshore of a  $\sim 400$ -km-wide coastal band, between the equator and  $30^\circ\text{S}$ , the differences in the large-scale fields  
169 are weak. The observed and modeled mean SST differ by less than  $0.2^\circ\text{C}$ , and the wind stress by less than 0.2  
170 % (not shown). In the nearshore region, the surface wind  $\vec{v}_s^a$  is weaker ( $\sim 5 \text{ m s}^{-1}$ ) than offshore and the mean  
171 surface current is stronger ( $\sim 20 \text{ cm s}^{-1}$ ). As both the surface wind and the surface current are alongshore  
172 and equatorward, we could expect the wind stress computed from the relative wind (Rel) to be weaker than the  
173 wind stress computed from the absolute wind (Abs). A wind stress is calculated offline using Abs surface fields  
174 (surface wind and current) and following equation (2). Between  $7^\circ\text{S}$  and  $13^\circ\text{S}$ , this offline wind stress is up to  
175 8 % weaker than the Abs wind stress computed online (Fig. 3a). One could expect a similar difference between  
176 Rel and Abs wind stress, however, it is much weaker ( $<3\%$ ). This is due to a partial compensation of the wind  
177 stress weakening in Rel by a stronger surface wind (up to 3 % stronger, not shown).

178 Due to the weak modification of the alongshore wind stress (Fig. 3a), the mean oceanic circulation is virtually  
179 unchanged. The modeled poleward undercurrent is consistent with *in situ* ADCP observations (Chaigneau et al,  
180 2013). It extends between 30 m and 230 m depth and between the coast and 150 km offshore (Fig. 3b), as in  
181 the observations albeit with a weaker intensity. The mean current velocities differ by less than  $2 \text{ cm s}^{-1}$  in the  
182 simulations (Fig 3c). The isotherms shoaling near the coast (Fig 3b), characteristic of the upwelling, is well  
183 reproduced by the model. The thermal structure is barely modified by the coupling, with differences less than 2  
184 % (Fig 3c).

### 185 3.2 Eddy Kinetic Energy

186 We now focus on the impact on the mesoscale activity. Both simulations show a depth-integrated EKE minimum  
 187 nearshore (Figs. 4a-c), two maxima (north of 18°S, between  $\sim 50$  and 300 km from the coast, and south of 26°S)  
 188 which is consistent with estimations of surface EKE from observations (*e.g.* Colas et al, 2012; Oerder et al, 2015).  
 189 The EKE is maximum at the surface and decreases with depth (Fig. 4d). Its structure changes with the distance  
 190 from the coast: it decreases offshore in the surface layer (above 50 m depth) and increases in the subsurface  
 191 (Fig. 4d). This indicates a "barotropization" of the eddy kinetic energy related to the inverse energy cascade of  
 192 the eddies moving offshore (*e.g.* Haney et al, 2001; Colas et al, 2013).

193 The integrated EKE is weaker in Rel than in Abs both nearshore (-15 % within 200 km from the coast) and  
 194 offshore (-20 % between 200 and 500 km). The EKE reduction is stronger near the surface: the mean surface EKE  
 195 is reduced by  $\sim 30$  % over the first 500 km (Fig. 4c). Vertical profiles shows that the EKE difference between  
 196 the two simulations reduces with depth. However, as a consequence of the eddy energy "barotropization", this  
 197 difference at depth increases with offshore distance: at  $\sim 150$  m depth, it is 2.3 stronger 400 km from the coast  
 198 than 100 km from the coast (Fig. 4d). Interestingly, this leads to a slightly more barotropic EKE in Rel than in  
 199 Abs.

### 200 3.3 EKE production

201 To explain the EKE reduction due to the current-stress interaction, the main conversion terms contributing to  
 202 the depth-integrated EKE generation are analyzed (*e.g.* Marchesiello et al, 2003):

$$F_e K_e = \frac{1}{\rho_0} (\overline{u'(z=0)\tau'_x} + \overline{v'(z=0)\tau'_y}) \quad (4)$$

$$K_m K_e = - \int_Z^0 (\overline{u'u' \frac{\partial \bar{u}}{\partial x}} + \overline{u'v' \frac{\partial \bar{u}}{\partial y}} + \overline{u'w' \frac{\partial \bar{u}}{\partial z}} + \overline{v'u' \frac{\partial \bar{v}}{\partial x}} + \overline{v'v' \frac{\partial \bar{v}}{\partial y}} + \overline{v'w' \frac{\partial \bar{v}}{\partial z}}) dz \quad (5)$$

$$P_e K_e = - \int_Z^0 \frac{g}{\rho_0} \overline{\rho'w'} \quad (6)$$

with  $\bar{\cdot}$  the temporal average and  $\cdot'$  the anomaly relative to the temporal average.  $u$ ,  $v$  and  $w$  are the zonal, meridional and vertical velocities,  $\rho$  is the water density,  $\rho_0$  is a reference density (mean water density), and  $\tau_x$  and  $\tau_y$  are the zonal and meridional wind stress.  $P_e K_e$  represents the energy transfer from eddy potential energy to EKE (baroclinic conversion), and  $K_m K_e$  is the transfer from mean kinetic energy to EKE (barotropic conversion).  $F_e K_e$  corresponds to the EKE creation by the wind stress anomalies. These terms are computed from 5-day-average model outputs. Anomalies  $\cdot'$  are defined relatively to a 90-day running mean.

The EKE generation mainly occurs in a  $\sim 200$ -km wide coastal strip, with a maximum located 75-100 km offshore (Fig. 5). This is consistent with previous studies using forced-ocean simulations in the South-East Pacific (Belmadani et al, 2012) and the California (Marchesiello et al, 2003) regions. 70 % of the total EKE generation is due to  $P_e K_e$  within 200 km from the coast (Fig. 5a). There, the vertical shear of the alongshore velocity generated by the equatorward surface current and the poleward undercurrent (Fig. 3b) leads to baroclinic instabilities, converting potential energy to EKE (Colas et al, 2012). The  $K_m K_e$  cross-shore profile shows that eddies transfer energy to the alongshore mean flow ( $K_m K_e < 0$ ) within 50 km from the coast. Between 50 and 150 km offshore, barotropic instabilities transfer kinetic energy from the mean coastal currents to EKE ( $K_m K_e > 0$ ). This transfer represents  $\sim 15$  % of the total EKE generation (Fig. 5b). As the two other terms,  $F_e K_e$  is maximum near the coast, where the current anomalies are largest (Fig. 5c). It contributes to  $\sim 25$  % of the EKE generation within 200 km from the coast.

Overall, the total EKE generation (i.e.  $F_e K_e + K_m K_e + P_e K_e$ ) is 30 % weaker in Rel than in Abs in a 200-km-wide coastal band. This reduction is mainly due to a weakening of  $F_e K_e$  (Fig. 5c). This is consistent with Duhaut and Straub (2006), who showed that the wind stress work computed with a relative wind (Eq. 2) is always smaller than the one computed with absolute surface wind (Eq. 1). It is generally considered that the main wind work contribution to the ocean circulation is the wind work associated with the geostrophic current (e.g. Wunsch, 1998) :

$$F_e K_{eg} = \frac{1}{\rho_0} (\overline{u'_g(z=0)\tau'_x} + \overline{v'_g(z=0)\tau'_y}) \quad (7)$$

228 with  $u_g$  and  $v_g$  the zonal and meridional geostrophic velocities, respectively. Indeed, the wind work on ageostrophic  
 229 currents can be approximated by the work on Ekman currents that is dissipated through vertical turbulence in  
 230 the Ekman layer, and, thus, it does not contribute to the ocean circulation (Wang and Huang, 2004). Our results  
 231 show that  $\sim 90$  % of the  $FeKe$  reduction is attributed to the  $FeKeg$  reduction (Fig. 5c). Note that  $FeKeg$  and  
 232  $FeKe$  are negative in Rel offshore of 75 km from the coast, indicating an EKE dissipation by the wind work.

233 The averaged differences (from the coast to 500km offshore) in  $K_m K_e$  ( $0.03 \text{ cm}^3 \text{ s}^{-3}$ ) and  $P_e K_e$  ( $0.04 \text{ cm}^3$   
 234  $\text{s}^{-3}$ ) between the two simulations are much smaller than in  $F_e K_e$  ( $0.18 \text{ cm}^3 \text{ s}^{-3}$ ).  $P_e K_e$  is, on average, increased  
 235 by  $0.02 \text{ cm}^3 \text{ s}^{-3}$  (10 % of the total change in EKE generation) in Rel between 50 - 100 km from the coast, and  
 236 reduced by  $0.08 \text{ cm}^3 \text{ s}^{-3}$  (30 %) between 100 and 200 km offshore ((Fig. 5a). Note that a longer simulation  
 237 period would be necessary to ensure the robustness of these values.

238 In summary, the main impact of the current-stress interaction is to reduce the surface and depth-integrated  
 239 EKE by 30 % and 20 % respectively. This EKE reduction can partly be attributed to a 30 % weakening of  
 240 the EKE sources in the coastal region. A part can also be due to the fact that mesoscale structures, created  
 241 nearshore through instabilities of the alongshore current system, evolve differently during their lifetime under  
 242 the action of this air-sea interaction. To investigate this issue, we now analyze the coherent eddies detected in  
 243 the simulations.

## 244 4 Impacts on the coherent eddies

245 To examine the impact of the current-stress interaction on the coherent eddies, we first compare the eddy  
 246 characteristics (Sec. 4.1) before studying the processes responsible for the differences between the two simulations  
 247 (Sec. 4.2).

### 248 4.1 Eddies characteristics

249 In both simulations, the cyclones and anticyclones have the same mean radius (differences are less than 2 %;  
 250 Table 1) and cyclones have a longer lifetime ( $\sim 20$  %) than anticyclones. The mean eddy lifetime is  $\sim 15$  %

251 shorter in Rel than in Abs, which can be related to an eddy damping by the current-wind stress interaction.  
252 This shorter lifetime may explain why fewer eddies ( $\sim 10\%$ , Table 1) were found in Rel as part of short-lived  
253 ( $\lesssim 20$  days) eddies population in Abs may have its lifetime becoming smaller than the detection threshold (15  
254 days; Sec. 2.2.1) in Rel.

255 More anticyclones ( $\sim 10\%$ ) than cyclones are found in Abs. This is at odds with estimations from satellite  
256 altimetry (Chaigneau et al, 2008, 2009; Pegliasco et al, 2015) which do not find a significative difference between  
257 the number of cyclones and anticyclones in the South East Pacific. The asymmetry in the cyclones-anticyclones  
258 number is reduced in Rel ( $\sim 3\%$ , Table 1), indicating a more realistic eddy distribution. This is related to  
259 the fact that there are more (179) short-lived ( $<20$  days) anticyclones than cyclones in Abs and the lifetime  
260 reduction in Rel makes an important part of these eddies undetected. As a result, the eddy number diminution  
261 between Abs and Rel is larger for the anticyclones (277) than for the cyclones (127).

262 The cyclonic cores are located near the surface while the anticyclones can be separated in two categories  
263 according to their core depth (at surface or subsurface; Colas et al, 2012; Pegliasco et al, 2015). An anticyclone  
264 is considered a surface eddy when its temperature anomaly maximum is located above 100 m depth. Half of  
265 the detected anticyclones are subsurface eddies. Note that some of the subsurface anticyclones having a weak  
266 signature at the surface may not be detected as our method is based on the surface current fields. However, in  
267 the context of the present study, we do not see this as a limitation as eddies with a very weak surface current  
268 signature do not interact strongly with the atmosphere. 159 (118) fewer surface (subsurface) anticyclones are  
269 found in Rel than in Abs. Again this difference is likely attributed to the longer lifetime ( $\sim 30\%$ ) of subsurface  
270 anticyclones compared to the surface ones.

271 To investigate further the eddy differences between the simulations, we now analyze their 3D structures.  
272 Composites of the temperature anomaly are shown in Fig. 6. Cyclones are associated with a cold anomaly  
273 reaching a minimum at  $\sim 150$  m depth (Figs.6a,d), which is consistent with observations (Chaigneau et al,  
274 2011). In Rel, the cyclones temperature anomaly is reduced by  $\sim 10\%$  at the surface and  $\sim 30\%$  at the  
275 core depth (Fig.6g). Below 150 m, the difference between the two simulations decreases with depth. The surface  
276 anticyclones present positive temperature anomaly at all depths, with a maximum around 50 m depth (Figs.6b,e).

277 In contrast, the subsurface anticyclones are characterized by a cold anomaly in the near-surface and a warm  
278 anomaly (Figs.6c,f) reaching a maximum around 400 m (in agreement with Pegliasco et al, 2015).

279 As for cyclones, the surface anticyclones anomaly reduction in Rel is weak at the surface ( $\sim 5\%$ ) and reaches  
280 a maximum at the eddy core depth ( $\sim 45\%$  at 100 m depth; Fig.6h). Interestingly, the situation is different for  
281 the subsurface anticyclones (Fig.6i). In Rel, their anomaly is cooler at all depths, leading to an intensification  
282 (by  $\sim 35\%$ ) and a deepening (down to  $\sim 100$  m) of the near-surface cold anomaly. Below, as for the surface  
283 anticyclones, the warm anomaly is significantly reduced in the depths range of the eddy core ( $-55\%$  at 150 m  
284 depth).

285 The vorticity anomaly structure shows a maximum intensity near the surface for the three eddy types (Fig.  
286 7). This is consistent with the eddy vorticity composites in Colas et al (2012). In Abs, the surface vorticity  
287 anomaly is stronger for the surface anticyclones and the cyclones ( $\sim -0.14$  and  $\sim -0.12$ , respectively) than for  
288 the subsurface anticyclones ( $\sim -0.095$ ). The vorticity anomaly extends deeper in the subsurface anticyclones:  
289 anomaly stronger than 0.04 reaches 400 m depth, while it goes to 200 m depth in the surface anticyclones and  
290 250 m depth in the cyclones. In Rel, vorticity anomalies are  $\sim 30-35\%$  weaker for all eddy types in the surface  
291 layer (0-50 m depth). This difference decreases abruptly below: at 400m depth, the vorticity anomaly is reduced  
292 by  $\sim 10-15\%$ . The surface anticyclones still have a stronger surface vorticity anomaly than the subsurface ones  
293 but by only  $\sim 20\%$  (Fig.7b,c). In contrast, at 400 m depth, the vorticity anomaly in the subsurface anticyclones  
294 is  $\sim 40\%$  larger than in the surface ones.

295 Overall, both the temperature and vorticity anomalies associated to the eddies are reduced under the action  
296 of the current-stress coupling. This reduction is stronger around 100-150 m depth for the temperature and near  
297 the surface for the vorticity.

#### 298 4.2 Current-induced stress anomalies effect

299 The eddy damping effect due to the current-stress interaction reduces the eddies intensity and shortens their  
300 averaged lifetime by 15%. To further investigate this effect, we now focus on the eddies with a lifetime longer  
301 than 40 days. We first examine the temporal evolution of the eddy surface characteristics (Fig. 8). The eddy



302 amplitude is derived from the sea surface height (SSH) anomaly following Chaigneau et al (2008), Samelson  
 303 et al (2014) and Pegliasco et al (2015). Cyclones (anticyclones) amplitude is  $\sim 20\%$  ( $10\%$ , respectively) weaker  
 304 in Rel on the first day of detection. This is likely due to the action of current-stress interaction on the eddy  
 305 structure during its early life phase, before it is actually detected as a coherent eddy. To better compare the  
 306 eddy evolution in Abs and Rel, we normalize the eddy amplitude by its first day value. The eddy evolution is  
 307 divided in two periods: growth during the first  $\sim 25$  days, and decay after (Fig. 8a). This is consistent with  
 308 Samelson et al (2014) and Pegliasco et al (2015) who distinguish an eddy growth phase during the first half  
 309 of its lifetime and a decay phase in the second half. Both the growth and decay phases are affected by the  
 310 current-stress interaction in Rel. For the cyclones (anticyclones), the growing is reduced by  $\sim 15\%$  ( $\sim 25\%$ ,  
 311 respectively). The decay is  $\sim 50\%$  stronger for both cyclones and anticyclones. As a result, after 40 days the  
 312 eddy amplitude is  $\sim 10\%$  weaker in Rel. Thus, not only the coherent eddies are weaker in Rel than in Abs at  
 313 their initial detection, but they are also clearly damped during their entire lifetime

314 This contrasts with Renault et al (2016) who suggest no impact of the current-stress interaction on growth  
 315 and decay phases, while during a “mid-life phase” the coupling process is responsible for a slow decay. This  
 316 discrepancy may be due to their choice to normalize the time axis for the eddy evolution, so that changes in the  
 317 eddy lifetime are not taken into account.

318 The surface amplitude evolution of the surface and the subsurface anticyclones has also been compared (not  
 319 shown). For both anticyclones types, eddy damping is consistently stronger in Rel.

320 The eddy damping also impacts the mean surface vorticity (Fig 8b). After 40 days, the cyclones (anticy-  
 321 clones) vorticity are  $\sim 5\%$  ( $12\%$ , respectively) weaker in Rel.

322

323 Previous studies suggested that two dynamical processes can explain the eddy damping by the current-stress  
 324 interaction: wind work (*e.g.* Zhai and Greatbatch, 2007) and Ekman pumping (*e.g.* Dewar and Flierl, 1987). We  
 325 first focus on the role of the wind work  $F_e K_{eg}$ . We showed that, near the coast (within a 75 km band) the wind  
 326 work transfers eddy kinetic energy from the atmosphere to the ocean, whether or not current-stress interaction  
 327 is taken into account (Sec.3.3 ; Fig. 5c). On the other hand further offshore (from  $\sim 75$  km),  $F_e K_{eg}$  exhibits

328 ubiquitous negative values in Rel, which corresponds to an EKE sink. To better understand this mechanism, we  
 329 construct composite average above coherent eddies in Abs (Fig. 9) and in Rel (Fig. 10). In Abs, a wind stress  
 330 anomaly is created above the eddies. It is only due to the SST anomaly following the relation:

$$\|\vec{\tau}_s\|' \propto SST' \quad (8)$$

331 with a very high correlation of 0.97 (0.88) for negative (positive) wind stress anomaly above the cyclones  
 332 (anticyclones, respectively; Figs. 9a-d). A more detailed study of the mesoscale SST-induced wind stress response  
 333 is presented in (Oerder et al, 2017). **XXX (PB REF ICI) XXX** It can be noted that, over both cyclones  
 334 and anticyclones, the wind stress anomaly is along (opposed) the current anomaly on the right (left) side of the  
 335 eddy. As a consequence,  $(F_e K_{eg})$  is necessarily positive (negative) on the right (left) side of the eddy (Fig. 9c,d).  
 336 This is confirmed by examining the composites (Figs. 9e,f): wind work tends to inject energy on the right side  
 337 and to dissipate it on the left side.

338 In Rel, the wind stress is affected by the SST and the current mesoscale anomalies. The wind stress anomaly  
 339 composites (Fig. 10a,b) display very different patterns than in Abs. A bilinear regression shows that the com-  
 340 posites follow the relation

$$\|\vec{\tau}_s\|' = s_1^T SST' - s_2^T \vec{v}_g' \cdot \frac{\vec{\tau}_s}{\|\vec{\tau}_s\|} \quad (9)$$

341 with high correlation of 0.99 for the cyclones and 0.97 for the anticyclones. The regression coefficients are  $s_1^T = 6.8$   
 342  $(7.9) \cdot 10^{-3} \text{ N m}^{-2} \text{ }^\circ\text{C}^{-1}$  and  $s_2^T = 1.2(0.6) \cdot 10^{-2} \text{ N s m}^{-2} \text{ m}^{-1}$  for the cyclones (anticyclones, respectively). The  
 343 wind stress direction is largely altered by the surface current (Figs. 10a,b) that tends to create a wind stress gyre  
 344 in an opposed sense of rotation than the eddy (*e.g.* Cornillon and Park, 2001). The composite above cyclones  
 345 (Fig. 10a) shows that the current-induced effect on the wind stress direction dominates the SST-induced effect.  
 346 As the wind stress anomaly above cyclones is clearly an anticyclonic gyre,  $F_e K_{eg}$  is expected to be negative all  
 347 over the eddy rim (Fig. 10c), which is confirmed by the composite (Fig. 10e).

348 Results above anticyclones are slightly different. The wind stress anomaly composite has a positive curl, as the  
 349 current-induced effect creates a cyclonic wind stress gyre (Fig. 10b). However, the positive SST anomaly creates

350 a wind stress anomaly in the wind current direction (Fig. 9a) which strengthens the current-induced wind stress  
 351 on the eddy left side and weakens it on the right side. As a result, the total stress anomaly almost vanishes  
 352 on the right side (Fig.10b). The wind stress cyclonic gyre created by the current should theoretically induce a  
 353 negative  $F_e K_{eg}$  on both sides of the anticyclonic eddy (Fig. 10d). This contrasts with the  $F_e K_{eg}$  composite (Fig.  
 354 10f) that has a negative pattern on the left side only. On the right side, the SST-induced effect compensates  
 355 the current-induced effect. Note that removing the SST effect by subtracting the patterns in Fig. 9f and 10f  
 356 consistently results in a negative wind work over both eddy sides (not shown). Overall, our results show a very  
 357 different action of the wind work on the coherent eddies when the current-stress interaction is taken into account:  
 358 the wind work always dissipates energy above the eddies and thus actively contributes to damp them.

360 We now turn to the impact of the eddy-induced Ekman pumping on the eddy damping. In Abs, eddy-induced  
 361 wind stress variations (Fig 9a,b) create a wind stress curl and an Ekman pumping anomaly proportional to  
 362  $\text{grad}_{cw}(SST')$  (Fig 11a,b). Correlations between the Ekman pumping and  $\text{grad}_{cw}(SST')$  composites are 0.95 and  
 363 0.88, with a regression coefficient of -0.15 and -0.19  $\text{m}^2 \text{s}^{-1} \text{ } ^\circ\text{C}^{-1}$  for the cyclones and anticyclones, respectively.  
 364 Consequently, the SST-induced Ekman pumping corresponds to a dipole (Fig 11a,b) with upwelling on the left  
 365 (right) side of the cyclone (anticyclone) and downwelling on the right (left) side of the cyclone (anticyclone).  
 366 The corresponding vertical velocities do not exceed  $6.5 \text{ cm day}^{-1}$ . In Rel, Ekman pumping anomalies  $W'_{Ek}$  result  
 367 from the superposition of the two effects, following the relation:

$$W'_{Ek} = s_1^{Ek} \text{grad}_{cr}(SST') - s_2^{Ek} \text{curl}(\vec{v}_o) \quad (10)$$

368 with  $s_1^{Ek} = -0.085 (0.14) \text{ m}^2 \text{ s}^{-1} \text{ } ^\circ\text{C}^{-1}$ ,  $s_2^{Ek} = 2.10 \cdot 10^{-4} (3.10 \cdot 10^{-4}) \text{ m}^{-1}$  for cyclones and anticyclones, respectively. A  
 369 correlation of 0.98 between the left and right terms of equation (10) is found for both cyclones and anticyclones.  
 370  $W'_{Ek}$  composites in Rel (Fig 11c,d) are dramatically different from those in Abs (Fig 11a,b). Anomalies in Rel  
 371 are centered on the composite eddy, indicating that the current-induced Ekman pumping ( $-s_2^{Ek} \text{curl}(\vec{v}_o)$ ) is  
 372 overwhelmingly dominating the SST-induced effect. The corresponding vertical velocity reaches  $16 \text{ cm day}^{-1}$ .

373 The current-stress effect also induces subsurface vertical velocities within the eddy core (Fig. 12). The mean  
374 vertical velocity (between 0 and 400 m depth) is 90 % (65 %) stronger in Rel than in Abs in the cyclones  
375 (anticyclones, respectively). The Ekman downwelling (upwelling) in the cyclones (anticyclones) is responsible  
376 for the enhanced damping of the cold (warm) temperature anomaly in Rel (Fig. 6). Note that the vertical  
377 circulation anomaly induced by the Ekman pumping is stronger around 50-100 m depth (Fig. 12), which could  
378 explain why the eddy temperature damping is stronger at depth than at the surface (Fig. 6).

## 379 **5 Discussions and conclusion**

380 We study the impact of the wind stress-surface current coupling on the mesoscale circulation by comparing  
381 two coupled simulations that differ by their surface wind stress computation: using the absolute surface wind  
382 velocity (Abs) or the relative surface wind velocity (Rel). The regional ocean circulation is almost unchanged  
383 because the surface absolute wind velocity is modified in Rel in such a way that the mean wind stress remains  
384 very similar in both simulations. Nevertheless, the surface EKE is reduced by  $\sim 30\%$  in Rel with respect to Abs.  
385 This reduction is due to a weakening of the coastal EKE generation (through a decrease of the wind work) and  
386 a more efficient offshore damping of coherent eddies in Rel. The eddy damping results from energy dissipation  
387 by the wind work and eddy-induced Ekman pumping.

### 388 5.1 Impact on the mean circulation and eddy kinetic energy

389 The mean coastal ocean currents in the South East Pacific differ by less than  $2 \text{ cm s}^{-1}$  between Abs and  
390 Rel because of the unchanged mean wind stress. These results are consistent with contemporary studies in  
391 the California Current System (CCS) showing little impact on the mean regional circulation (Seo et al, 2015;  
392 Renault et al, 2016). However, the impact on the mean circulation may be stronger in other regions with much  
393 stronger currents where wind stress modulations by surface current were clearly evidenced (Chelton et al, 2004).  
394 Yet there is an ubiquitous EKE reduction, due to the current-stress interaction, that is comparable to those  
395 obtained for the CCS (surface EKE reduced by 42 % in Seo et al (2015) and integrated EKE reduced by 27

396 % in Renault et al (2016)). Examining the EKE generation terms in the nearshore region, we find a significant  
 397 decrease of the energy input by the wind work and indistinguishable modifications of  $P_e K_e$  and  $K_m K_e$  when the  
 398 current-stress interaction is active. This is consistent with Renault et al (2016) but contrasts with the increase  
 399 in  $P_e K_e$  in Seo et al (2015).

400 The strong impact of the current-stress interaction on the EKE intensity calls for a better representation of  
 401 this effect in ocean simulations. The computational cost of a high-resolution ocean-atmosphere coupled models  
 402 over large domains remains very high, so alternative approaches need to be found for ocean-forced simulations.  
 403 Several studies attempted to simulate this effect by using a surface wind velocity forcing to compute a wind  
 404 stress (following equation 2, Sec. 1). These works show an EKE reduction from 10 % (Zhai and Greatbatch,  
 405 2007) to 50 % (Eden and Dietze, 2009) with  $F_e K_e$  decreasing from 17 % (Zhai and Greatbatch, 2007) to 35  
 406 % (Duhaut and Straub, 2006), which is consistent with our results. However, in the coastal region, we show  
 407 that, for a given fixed absolute wind velocity, the wind stress reduction using Eq.(2) instead of (1) is much  
 408 higher (up to 8%) than in a coupled framework (less than 3 %). This indicates that using (2) in a forced ocean  
 409 model would lead to unrealistic ocean dynamics. In an attempt to mimic the current-induced wind velocity  
 410 modulation in an oceanic forced simulation, Renault et al (2016) suggest to modify the computation of the wind  
 411 stress (Eq. 2) following the relation :  $\vec{\tau}_s = \rho_a C_d \|\vec{v}_s^a - (1 - s_w) \vec{v}_1^o\|^2 \frac{\vec{v}_s^a - (1 - s_w) \vec{v}_1^o}{\|\vec{v}_s^a - \vec{v}_1^o\|}$ .  $s_w$  is determined by examining  
 412 the wind velocity response to the surface current anomalies in a coupled ocean-atmosphere simulation. This  
 413 method presents several limitations. First, it requires to run a preliminary coupled simulation to estimate  $s_w$ ,  
 414 and this has an important computational cost. Second, it ignores a possible dependency of the current-induced  
 415 wind velocity modulation from the atmospheric model parameterizations. Last, it assumes that  $s_w$  is constant  
 416 in time and space. Therefore, adequately representing the modulation of the surface wind conditions by the  
 417 oceanic mesoscale without using a fully coupled model remains an open question.

## 418 5.2 Impacts on the coherent eddies

419 Our results show that the mesoscale-induced wind stress variations above the coherent eddies can be approx-  
 420 imated by the linear superposition of the SST-induced and the current-induced effects (Eq. 9). The term

421  $s_1^\tau SST' \sim 10^{-3} \text{ N m}^{-2}$  is dominant compared to  $s_2^\tau \vec{v}_g^\lambda \cdot \frac{\vec{\tau}_s}{\|\vec{\tau}_s\|} \sim 3 \cdot 10^{-4} \text{ N m}^{-2}$ , which explains why relation  
 422 (8) is clearly identified in the South-East Pacific (Oerder et al, 2016) and other regions (e.g. O'Neill et al,  
 423 2010b). On the other hand, the current-stress interaction has a stronger impact on the wind stress direction,  
 424 leading to a larger wind stress curl anomaly above eddies than the SST-induced effect. We show that the term  
 425  $s_1^{Ek} \text{grad}_{cw}(SST')$  is 2.5 times weaker than the term  $-s_2^{Ek} \text{curl}(\vec{v}_o^\lambda)$  ( $6.5 \text{ cm day}^{-1}$  and  $16 \text{ cm day}^{-1}$ , respectively).  
 426 This is consistent with Gaube et al (2015) showing a SST-induced Ekman pumping being two to three times  
 427 weaker than the current-induced Ekman pumping. Note that the wind stress divergence is little affected by the  
 428 current-induced effect (not shown). This might explain why correlations between the wind stress divergence and  
 429 the downwind SST gradient are always higher than the correlations between wind stress curl and cross-wind  
 430 SST gradient (relation 3, O'Neill et al, 2010a).

431 Above cyclones, the current-induced effect on the wind stress direction is largely dominant compared to the  
 432 SST effect, and it generates a clear wind stress anticyclonic anomaly. Above anticyclones, the relative importance  
 433 of the two effects is somewhat different. The SST-induced wind stress modification, that strengthens (weakens)  
 434 the current-induced wind stress anomaly on the left (right, respectively) side of the eddy, compensates almost  
 435 totally the current effects on the right side of the eddy (Fig. 10b). Two reasons might explain why the SST-  
 436 induced effect is more important for the anticyclones. First, the surface current anomaly is on average stronger  
 437 for cyclones than for anticyclones because many of the latter are subsurface eddies with deep cores. Second, the  
 438 regression coefficient  $s_2^\tau$  is twice stronger for the cyclones than for the anticyclones, while  $s_1^\tau$  is almost unchanged.  
 439 This indicates that, for a given intensity of the surface current anomaly, the current-induced effect on the wind  
 440 stress is twice stronger in the case of cyclones. Understanding the mechanisms driving this asymmetric response  
 441 is beyond the scope of our study.

442

443 Overall, the wind stress response to the mesoscale currents is responsible for an eddy damping that occurs  
 444 during the entire eddy lifetime through an enhanced Ekman pumping and a negative wind stress work. The  
 445  $F_e K_{eg}$  composites (Fig. 10) show indeed that the wind work due to the current-stress interaction is negative  
 446 everywhere above the eddies (unlike what is stated in Renault et al, 2016). It seems quite logical since the

447 current-induced wind stress anomalies are always opposed to the current anomalies on both sides of the eddies  
448 (see schemes in Figs.10c,d).

449

### 450 5.3 Implications for the South East Pacific

451 In the South-East Pacific, mesoscale eddies are responsible for a significant fraction of the heat advection (Colas  
452 et al, 2012). Thus the mesoscale activity modulation induced by the current-stress interaction may affect the  
453 regional heat balance. Following Chaigneau et al (2011) and using the temperature anomaly composites pre-  
454 sented in Fig.6, we estimate the available heat anomaly transported by the composite coherent eddies. We found  
455 heat anomalies of  $-3.1 \cdot 10^{18}$  J ( $-1.5 \cdot 10^{18}$  J) for cyclones and  $2.5 \cdot 10^{18}$  J ( $1.6 \cdot 10^{18}$  J) for anticyclones in Abs (Rel,  
456 respectively). Based on the eddy census in Abs and Rel, we can crudely approximate that the contribution of  
457 coherent eddies to the heat transport is reduced by at least 50 % by the current-stress interaction. Note that an  
458 accurate estimate of the coherent eddies contribution to the total eddy heat transport ( $\sim 20\text{-}35\%$ , Colas et al,  
459 2012) is difficult to achieve. A dedicated study would be required to quantify precisely the overall effect of the  
460 current-stress interaction on the regional heat balance.

461

462 This effect on the mesoscale dynamics could also modulate the intense biogeochemical activity in the South  
463 East Pacific EBUS. McGillicuddy et al (2007); Eden and Dietze (2009) and Anderson et al (2011) find significant  
464 biological impacts of the eddy-induced Ekman pumping. However, none of these studies focused on EBUS where  
465 the effect of the mesoscale eddies actively subduct nutrients and phytoplankton (Lathuilière et al, 2010; Gruber  
466 et al, 2011), in contrast with offshore, more oligotrophic regions. Consequently, the mesoscale air/sea coupling  
467 could certainly modify the horizontal and vertical transport of biogeochemicals tracers and further impact the  
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477 Jet Propulsion Laboratory, Pasadena, CA, [https://podaac.jpl.nasa.gov/dataset/OSCAR\\_L4\\_OC\\_third-deg](https://podaac.jpl.nasa.gov/dataset/OSCAR_L4_OC_third-deg). Numerical data were  
478 obtained by model experiments described in Sect. 2.



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## Tables

|  |                             |                             |
|--|-----------------------------|-----------------------------|
|  | Abs                         | Rel                         |
| a) Number of cyclones detected                             | 2082                        | 1955                        |
| Mean radius  | 46 km                       | 44 km                       |
| Mean lifetime  | 41 days                     | 35 days                     |
| Number of cyclones with a lifetime longer than 40 days     | 578<br>(28 % of the total)  | 492<br>(25 % of the total)  |
|  | Abs                         | Rel                         |
| b) Number of anticyclones detected                         | 2299                        | 2022                        |
| Mean radius  | 47 km                       | 45 km                       |
| Mean lifetime  | 34 days                     | 30 days                     |
| Number of anticyclones with a lifetime longer than 40 days | 520<br>(23 % of the total)  | 383<br>(19 % of the total)  |
| Number of subsurface anticyclones                          | 1133<br>(49 % of the total) | 1015<br>(50 % of the total) |

Table 1: Characteristics of the (a) cyclones and (b) anticyclones detected in the two simulations

## Figures

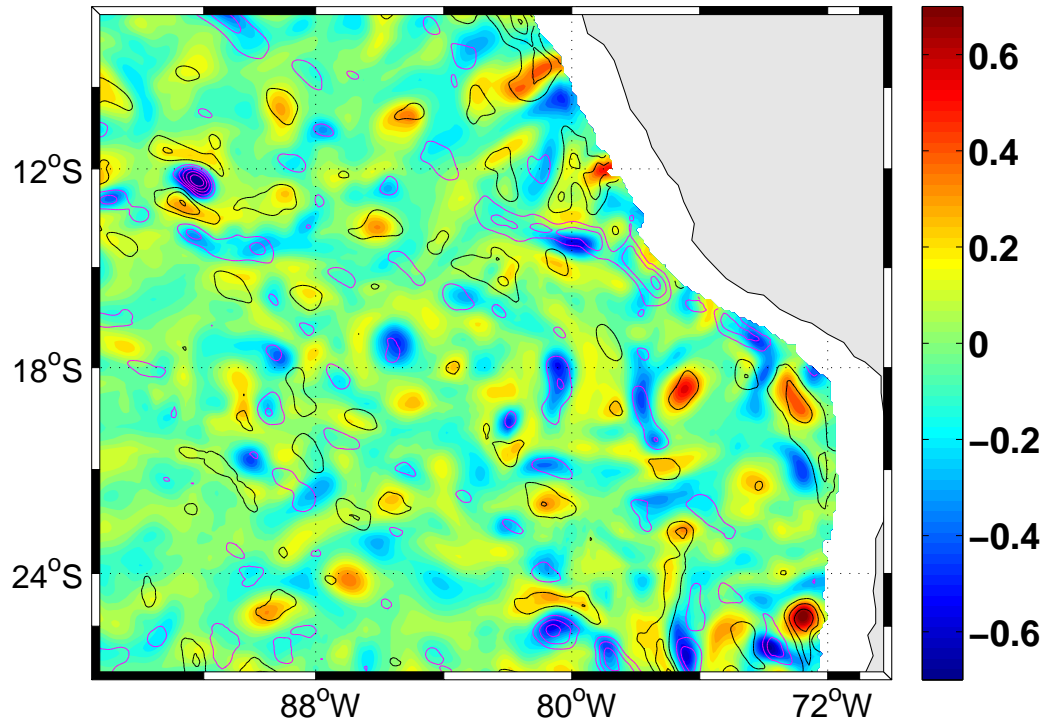


Fig. 1: Mesoscale anomalies of the surface current vorticity (colors, in  $10^{-5} \text{ s}^{-1}$ ) and  $\text{curl}(\vec{\tau}_s)_{res}$  (contours, with a  $3 \times 10^{-8} \text{ N m}^{-3}$  interval, black line is for negative values and magenta line for positive values), monthly-mean for July 2004 in Rel coupled simulation. Mesoscale anomalies are computed using two Gaussian spatial filters with 30 km and 150 km standard deviations to remove the large and small scale structures.



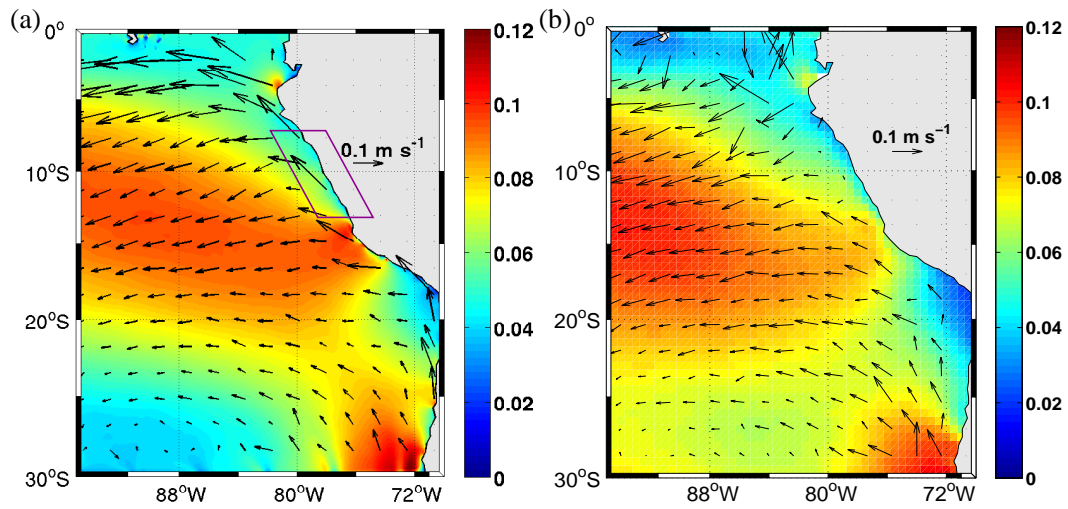


Fig. 2: 2003-2008 mean wind stress intensity (colors, in  $\text{N m}^{-2}$ ) and surface current (arrows) for (a) Rel simulation and (b) observations from QSCAT wind data (Dunbar et al, 2006) and Ocean Surface Current Analysis–Real Time (OSCAR) product (Bonjean and Lagerloef, 2002). A  $0.015 \text{ N m}^{-2}$  constant bias has been added to the model wind stress intensity. The  $7^\circ\text{S} - 13^\circ\text{S}$  coastal region is indicated by the magenta box

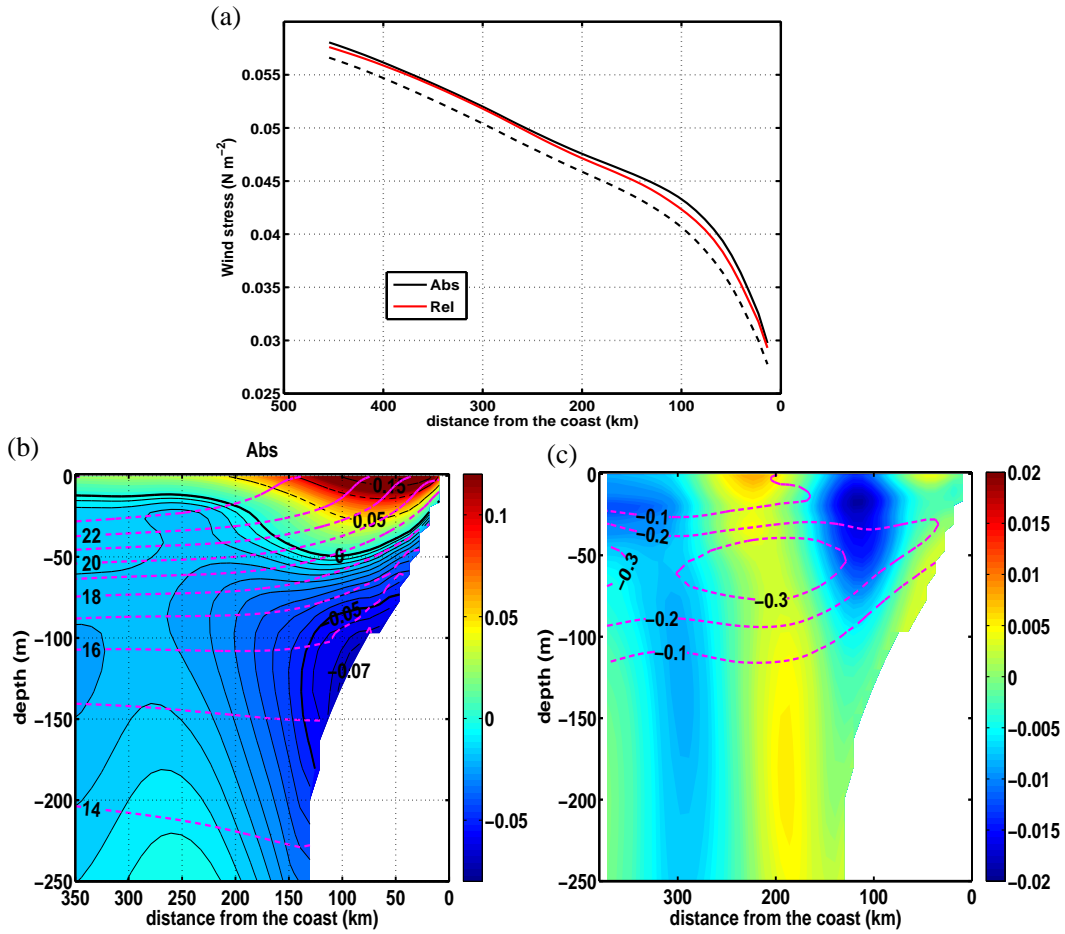


Fig. 3: (a) Mean wind stress intensity ( $\text{N m}^{-2}$ ) with respect to the distance from the coast (in km) for Abs (black plain line) and Rel (red plain line) simulations. The black dotted line represents a wind stress computed offline following equation (2) and using  $v_s^a$  and  $v_1^o$  from Abs. (b) Mean alongshore current velocities (shading, in  $\text{m s}^{-1}$ ) and temperature (dashed magenta line, in  $^\circ\text{C}$ , with a  $1^\circ\text{C}$  interval) for Abs. Thin plain black velocity contour interval is  $0.5 \text{ cm s}^{-1}$  and black bold contours marks 0 and  $-5 \text{ cm s}^{-1}$ . Thin dotted black velocity contour interval is  $5 \text{ cm s}^{-1}$  (c) Difference between the mean alongshore current velocities (shading, in  $\text{m s}^{-1}$ ) in Rel and in Abs. Contours marks the temperature differences (in  $^\circ\text{C}$ ) with a  $0.1^\circ\text{C}$  interval. All fields are averaged over the 2003-2008 period and from  $7^\circ\text{S}$  to  $13^\circ\text{S}$ .

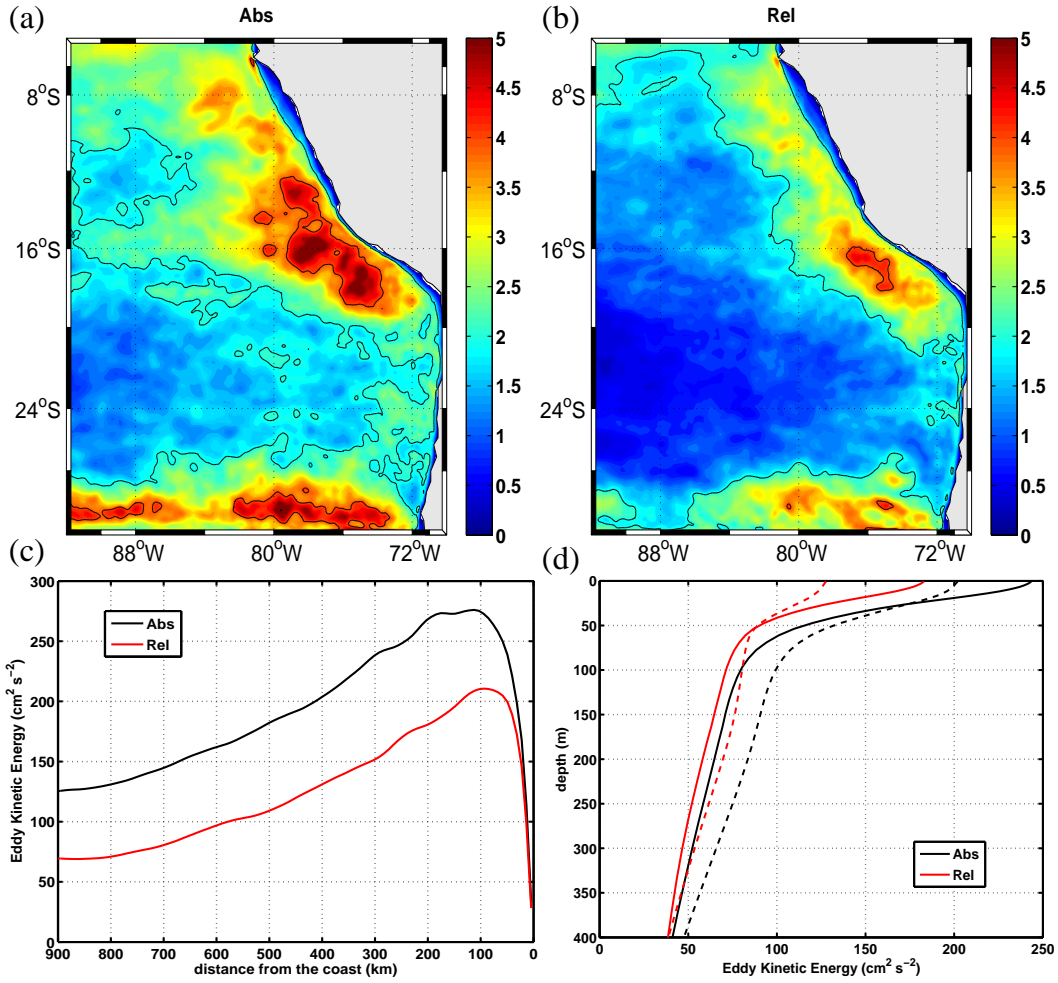


Fig. 4: (a)-(b) 2003-2008 mean eddy kinetic energy (EKE) integrated from the surface to 400 m depth (in  $\text{m}^3 \text{s}^{-2}$ ), in Abs and Rel simulations. (c) 2003-2008 mean surface EKE ( $\text{cm}^2 \text{s}^{-2}$ ), averaged between  $7^\circ \text{S}$  and  $13^\circ \text{S}$ , with respect to the distance from the coast (in km). (d) EKE vertical profile ( $\text{cm}^2 \text{s}^{-2}$ ), averaged over the 2003-2008 period, from  $7^\circ \text{S}$  to  $13^\circ \text{S}$ , and from 0 to 200 km offshore (plain line) or from 300 to 450 km offshore (dotted line). Red (black) lines are for Rel (Abs, respectively).

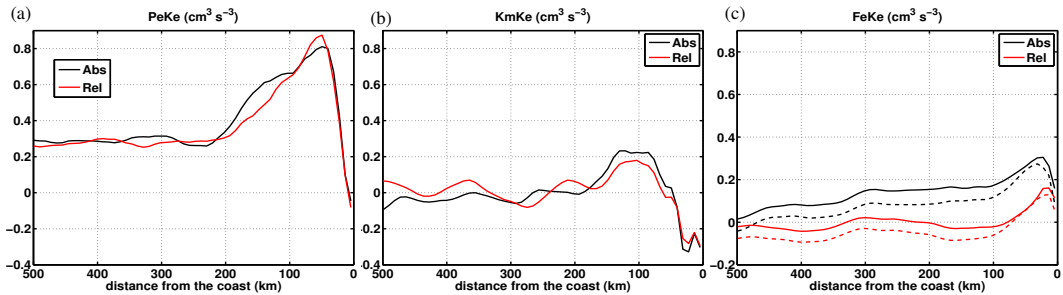


Fig. 5: EKE generation terms (in  $\text{cm}^3 \text{s}^{-3}$ ) with respect to the distance from the coast (in km), averaged over the 2003-2008 period, from  $7^\circ \text{S}$  to  $27^\circ \text{S}$  for Abs (black) and Rel (red). (a)  $P_e K_e$  term, (b)  $K_m K_e$  term (c)  $F_e K_e$  (plain line) and  $F_e K_{eg}$  (dotted line) term.

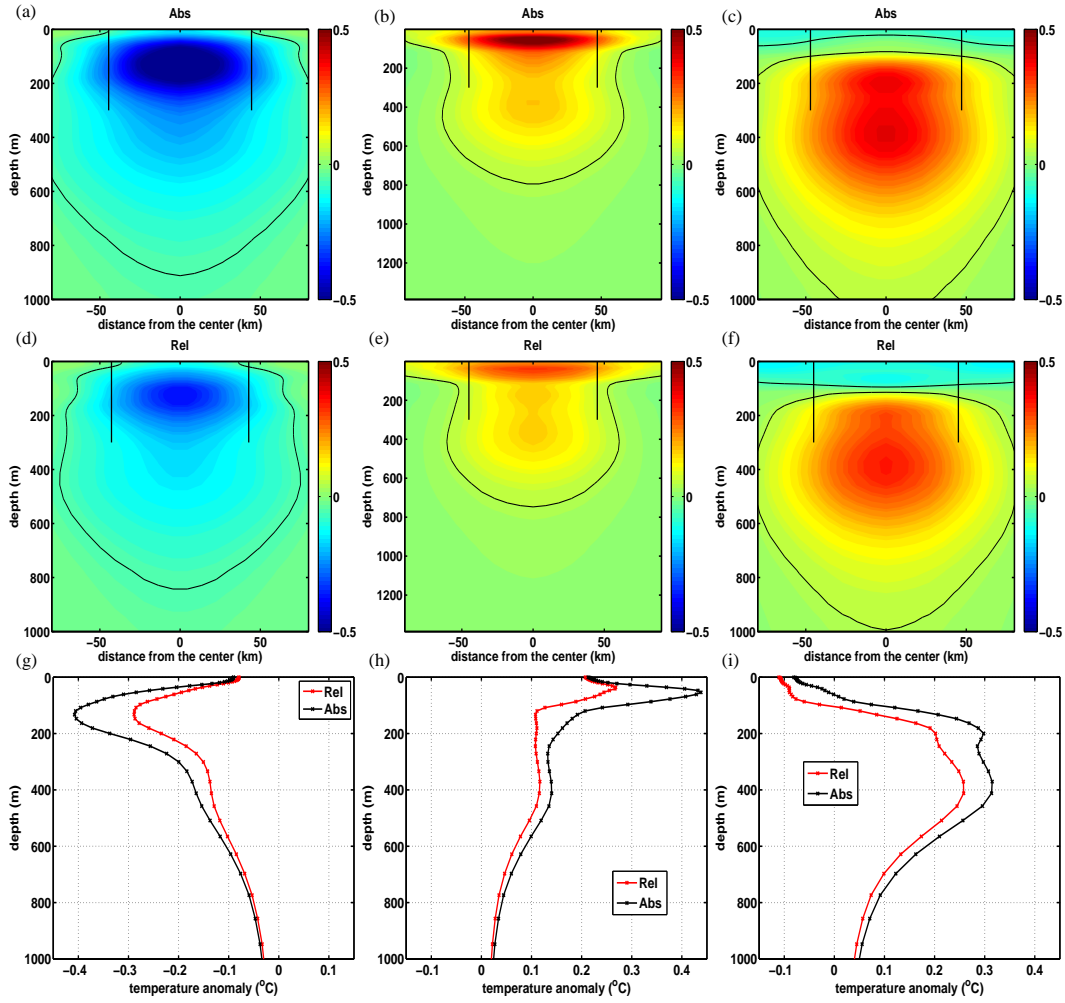


Fig. 6: Vertical structure of the temperature anomaly (in  $^{\circ}\text{C}$ ) composite in (a) the cyclones detected in Abs, (b) the surface and (c) subsurface anticyclones detected in Abs, (d) the cyclones detected in Rel, (e) the surface and (f) subsurface anticyclones detected in Rel. Thin line marks the  $0.05^{\circ}\text{C}$  contour and the bold vertical lines marks the mean eddy diameter. A radial average has been performed on the 3D composites to obtain the mean vertical structure. (g)-(i) : mean vertical profile of the temperature anomaly (in  $^{\circ}\text{C}$ ) in (g) the cyclones and (h) the surface and (i) subsurface anticyclones. Red (black) lines are for Rel (Abs, respectively).

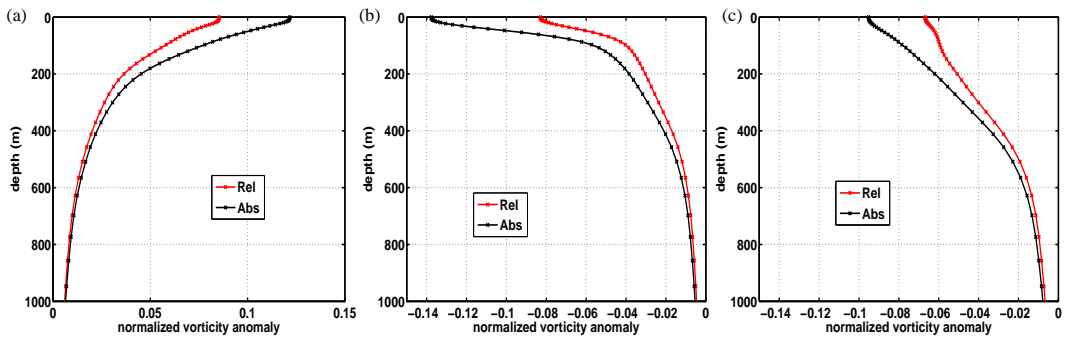


Fig. 7: Mean vertical profile of  $\frac{\zeta}{f}$  anomaly in (a) the cyclones and (b) the surface and (c) subsurface anticyclones. Red (black) lines are for Rel (Abs, respectively).

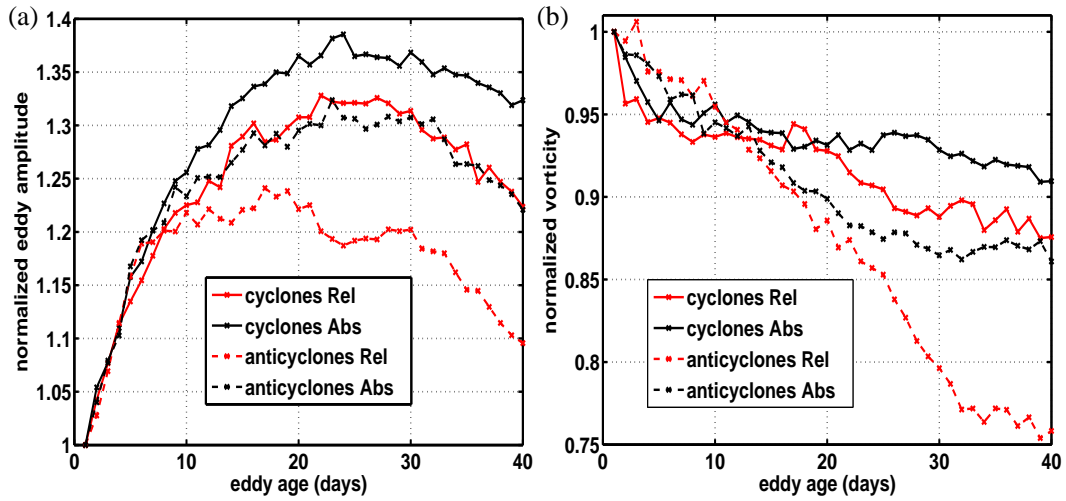


Fig. 8: Evolution of the (a) eddy composite amplitude and (b) the mean vorticity in Abs (black lines) and Rel (red lines) for the cyclones (plain lines) and anticyclones (dotted lines). For each day after its detection, the eddy amplitude is computed from the daily SSH anomaly composite as the difference between the SSH extremum and the mean SSH anomaly on the circle fitting the eddy (see Sec. 2.2). The amplitude and vorticity are normalized by the values at the first detection day.

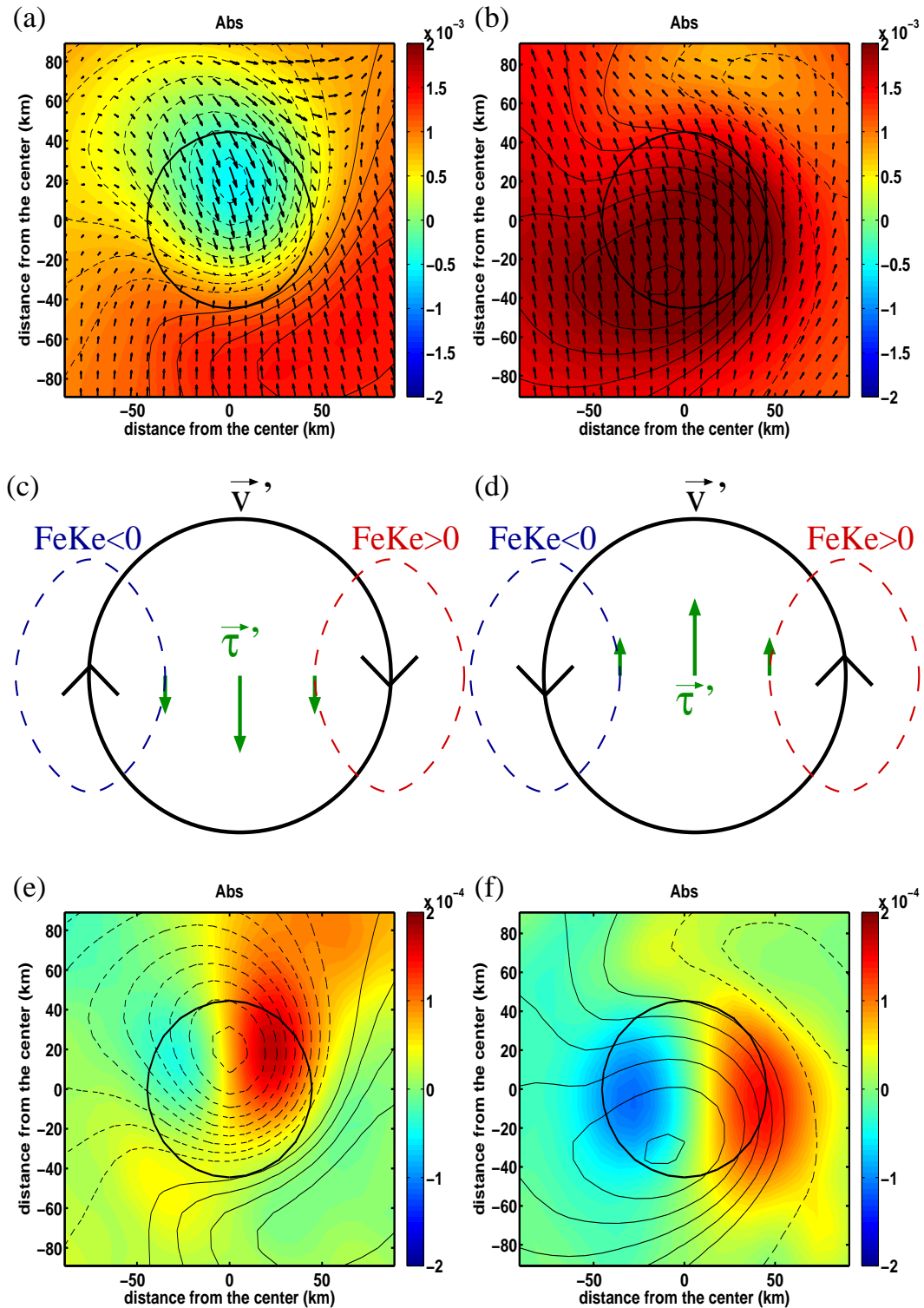


Fig. 9: Composites computed using Abs simulation. (a)-(b) Composites of the wind stress intensity anomaly (colors, in  $\text{N m}^{-3}$ ) above the (a) cyclones and (b) anticyclones. Composites of the wind stress anomaly are represented by the arrows and the thin black contours represent the composites of the SST anomaly with a  $0.02 \text{ }^\circ\text{C}$  interval, plain (dotted) lines are for positive (negative, respectively) values. (c)-(d) Scheme of the wind work associated to the SST-induced wind stress anomaly on a (c) cyclone and (d) an anticyclone. The current anomaly is represented by the black arrows, the wind stress anomaly is represented by the green arrows and the regions with positive (negative) wind work are marked with red (blue, respectively) dotted lines. (e)-(f) Composites of the wind work ( $\text{m}^3 \text{s}^{-3}$ ) on the (e) cyclones and (f) anticyclones in colors. The thin black contours represent the composites of the SST anomaly with a  $0.02 \text{ }^\circ\text{C}$  interval, plain (dotted) lines are for positive (negative, respectively) values. Black circles represent the mean eddy radius.

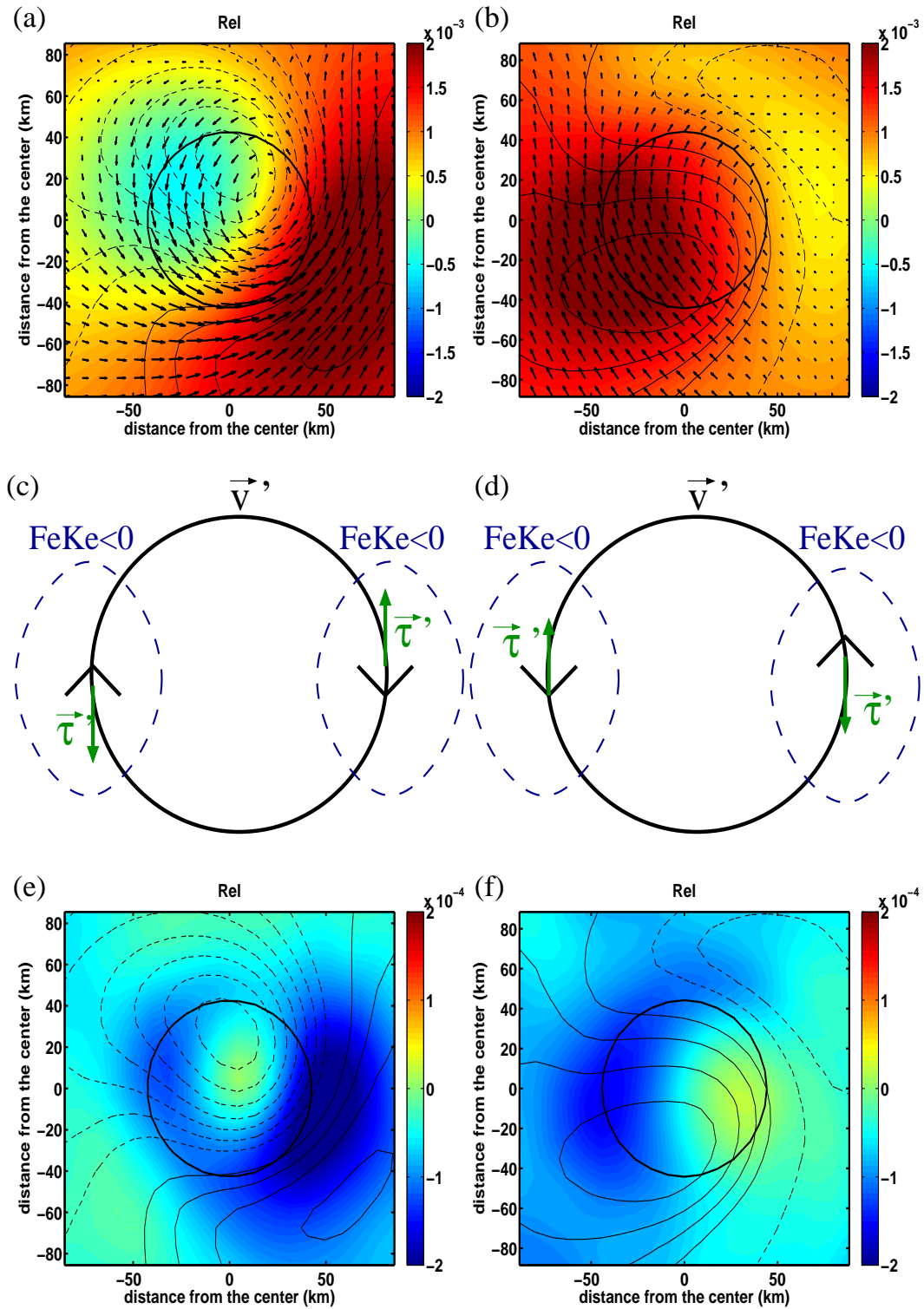


Fig. 10: Same as Fig. 9 but for Rel.

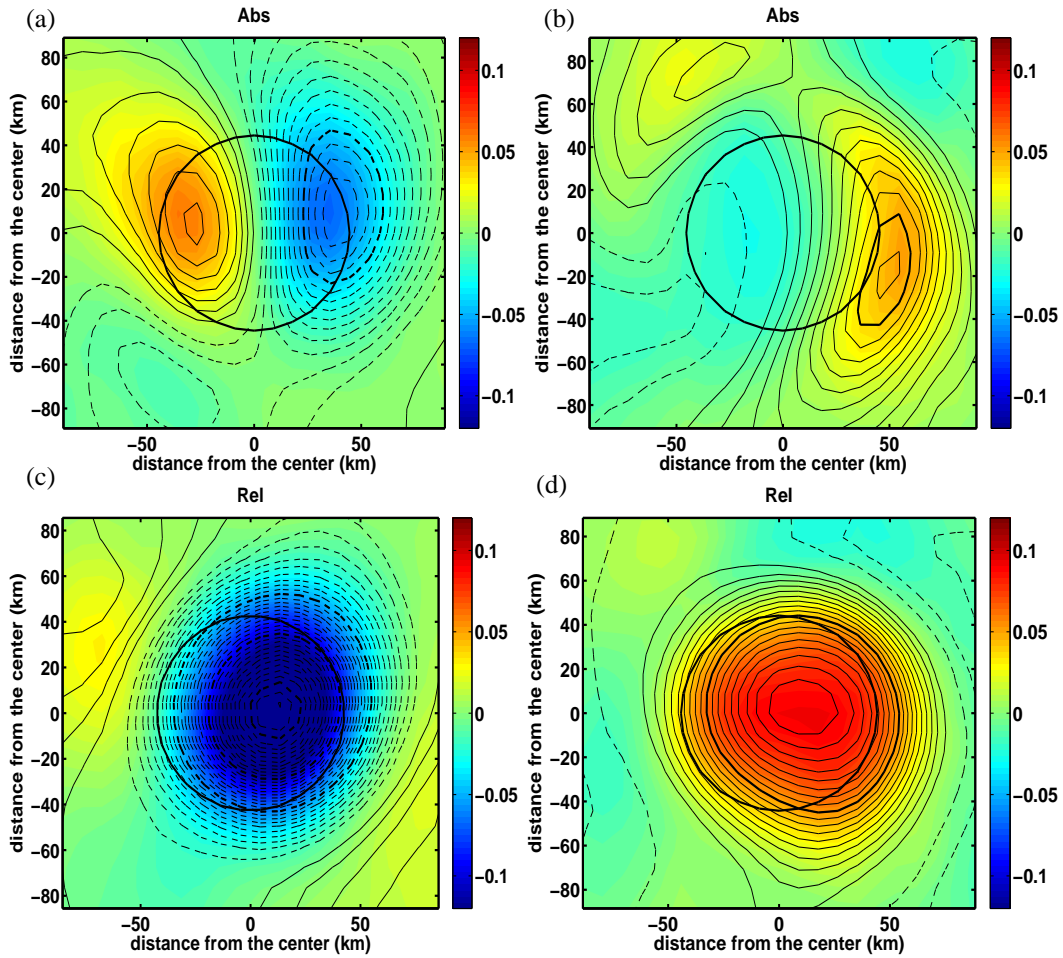


Fig. 11: Ekman pumping  $W_{Ek}$  (in  $\text{m day}^{-1}$ ) composite in (a) the cyclones detected in Abs, (b) the anticyclones detected in Abs, (c) the cyclones detected in Rel, and (d) the anticyclones detected in Rel. Black thin (bold) contours represents  $s_1 \text{grad}_{cr}(SST)$  in (a), (b) and  $s_1 \text{grad}_{cr}(SST) + s_2 \text{curl}(\vec{v}_1^{\delta'})$  in (c), (d), with a 0.005 (0.05, respectively)  $\text{m day}^{-1}$  interval. Dotted (plain) contours are for negative (positive, respectively) values. Black circles represent the mean eddy radius.



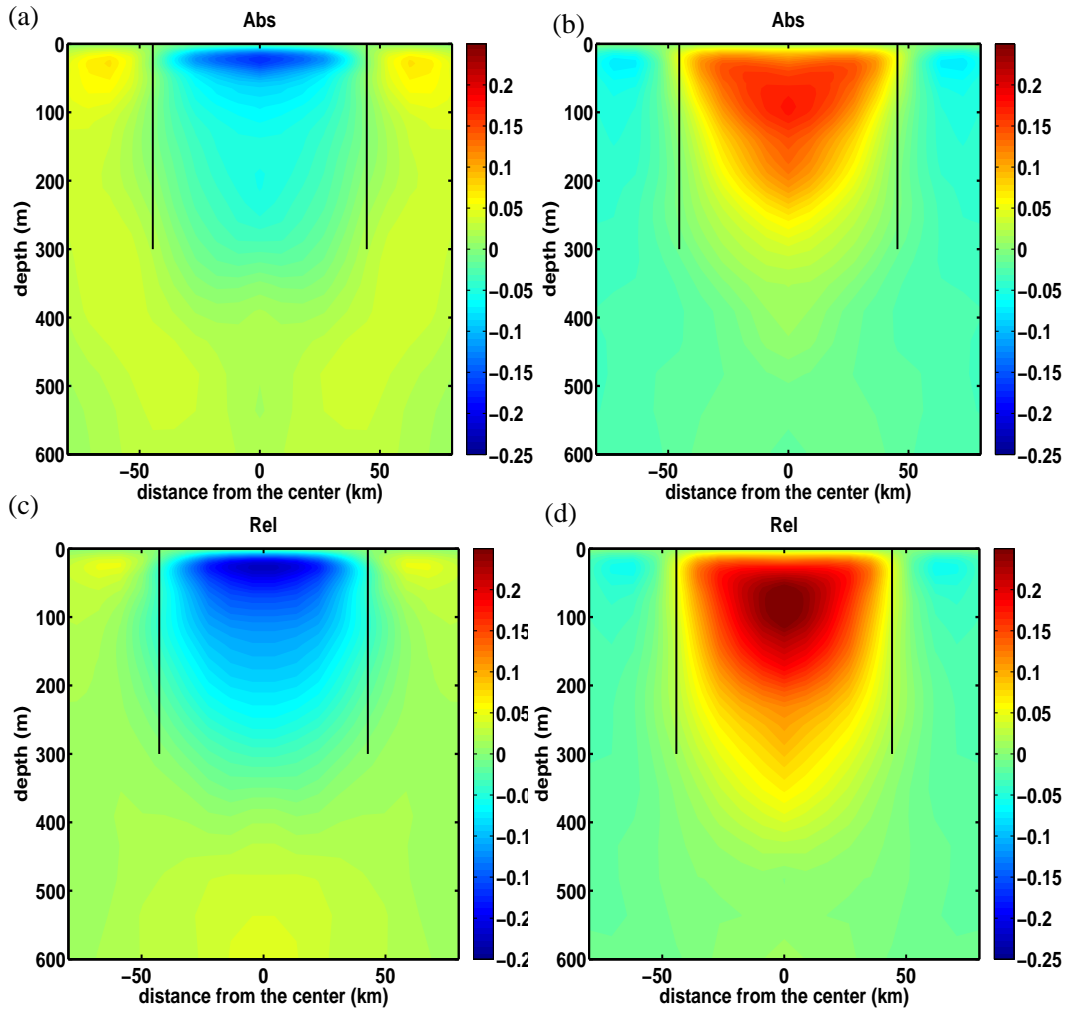


Fig. 12: Vertical structure of the vertical velocity anomaly (in  $\text{m day}^{-1}$ ) composite in (a) the cyclones detected in Abs, (b) the anticyclones detected in Abs, (c) the cyclones detected in Rel, and (d) the anticyclones detected in Rel. The bold vertical lines marks the mean eddy diameter. A radial average has been performed on the 3D composites to obtain the mean vertical structure.