

1 **Seasonal Variability of the Oxygen Minimum Zone off Peru in a**
2 **high-resolution regional coupled model**

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1 **Abstract**

2 In addition to being one of the most productive upwelling systems, the oceanic region off Peru is
3 embedded in one of the most extensive Oxygen Minimum Zones (OMZs) of the world ocean. The
4 dynamics of the OMZ off Peru remain uncertain, partly due to the scarcity of data and to the ubiquitous
5 role of mesoscale activity on the circulation and biogeochemistry. Here we use a high-resolution
6 coupled physical/biogeochemical model simulation to investigate the seasonal variability of the OMZ
7 off Peru. The focus is on characterizing the seasonal cycle in Dissolved O₂ (DO) eddy flux at the OMZ
8 boundaries, including the coastal domain, viewed here as the eastern boundary of the OMZ,
9 considering that the mean DO eddy flux in these zones has a significant contribution to the total DO
10 flux. Along the coast, despite the increased seasonal low DO water upwelling, the DO peaks
11 homogeneously over the water column and within the Peru Undercurrent (PUC) in austral winter,
12 which results from mixing associated with the increase in both the intraseasonal wind variability and
13 baroclinic instability of the PUC. The coastal ocean acts therefore as a source of DO in Austral winter
14 for the OMZ core, through eddy-induced offshore transport that is also shown to peak in Austral winter.
15 In the open ocean, the OMZ can be divided vertically into two zones: an upper zone above 400m where
16 the mean DO eddy flux is larger on average than the mean seasonal DO flux, and varies seasonally, and
17 a lower part where the mean seasonal DO flux exhibits vertical-zonal propagating features that share
18 similar characteristics than those of the energy flux associated with the annual extra-tropical Rossby
19 waves. At the OMZ meridional boundaries where the mean DO eddy flux is large, the DO eddy flux
20 has also a marked seasonal cycle that peaks in austral winter (spring) at the northern (southern)
21 boundary. In the model, the amplitude of the seasonal cycle is 70% larger at the southern boundary than
22 at the northern boundary. Our results suggest the existence of distinct seasonal regimes for the
23 ventilation of the OMZ by eddies at its boundaries. Implications for understanding the OMZ variability
24 at longer timescales are discussed.

1 **1 Introduction**

2 In addition to hosting one of the most productive upwelling systems, the South Eastern Pacific (SEP) is
3 home to one of the most extensive Oxygen Minimum Zones (OMZs) of the world ocean (Fuenzalida et
4 al., 2009; Paulmier and Ruiz-Pino, 2009). These oxygen deficient regions are key to understanding the
5 role of the ocean in the greenhouse gases budget and in climate, and in the presently unbalanced
6 nitrogen cycle (Gruber, 2008). The OMZs represent a net nitrogen loss to the atmosphere in the form of
7 N₂O (particularly the SEP OMZ: Farías et al., 2007; Arévalo-Martínez et al., 2015), in addition with
8 other toxic or climatically-active gases, such as H₂S and CH₄, respectively, in extremely low dissolved
9 oxygen (DO) concentrations (Libes, 1992; Law et al., 2013). They might even limit the ocean carbon
10 sequestration and act as CO₂ sources for the atmosphere (Paulmier et al., 2008; 2011). Furthermore, the
11 OMZs contribute to the habitat compression of marine organisms, in a zone that sustains 10% of the
12 world fish catch (Prince and Goodyear, 2006; Chavez et al., 2008). Therefore, understanding the
13 dynamics behind the OMZ becomes not just a matter of scientific interest, but also a major societal
14 concern.

15 In general, these low oxygen regions are considered to result from the interaction of biogeochemical
16 and physical processes (Karstensen et al., 2008). The SEP presents high biological productivity,
17 inducing a significant DO consumption mainly through the remineralization associated with a complex
18 nutrient cycle supported by the intense upwelling. In addition, the SEP encompasses a so-called
19 'shadow zone', a near stagnant/sluggish circulation region next to the eastern basin boundary, not
20 ventilated by the basin scale wind driven circulation (Luyten et al., 1983). Assuming a steady state,
21 lateral oxygen fluxes from subtropical water masses and diapycnal mixing are expected to balance the
22 oxygen consumption (Brandt et al., 2015). However, the diversity of environmental forcings in the SEP,
23 and the variety of timescales at which they operate (Pizarro et al., 2002; Dewitte et al., 2011; 2012) has
24 eluded a proper understanding of the processes controlling the OMZ structure and variability. On the
25 one hand, the scarcity of data and rare surveys have only permitted to document the DO temporal
26 variability at a few locations (e.g. Morales et al., 1999; Cornejo et al., 2006; Gutiérrez et al., 2008;
27 Llanillo et al., 2013). On the other hand, the highly complex interaction between physical and
28 biogeochemical mechanisms makes modeling and prediction of OMZ location, intensity and its
29 temporal variability a challenging task (Karstensen et al., 2008; Cabré et al., 2015). Low resolution
30 CMIP class coupled models still have severe biases of physical and biogeochemical origins,

1 particularly in Eastern Boundary Current systems (Richter, 2015), which has eluded the interpretation
2 of long term trends in OMZ (Stramma et al., 2008; 2012; Cabré et al., 2015). Regional coupled
3 biogeochemical modeling nonetheless has provided a complementary approach to gain insight in the
4 dynamics of OMZ and its relationship with climate (Resplandy et al., 2012; Gutknecht et al., 2013a).
5 One recent modeling effort in understanding the dynamics behind the OMZ in the Eastern Tropical
6 Pacific comes from Montes et al. (2014). This study provided a first regional simulation of the OMZ in
7 the SEP, and summarized the elements involved in maintaining the OMZ found off the coast of Peru as
8 resulting from a delicate balance between (i) the equatorial current system dynamics: the relatively
9 oxygen-rich waters carried by the Equatorial Undercurrent (EUC), the relatively oxygen poor and
10 nutrient rich waters carried by the primary and secondary Tsuchiya Jets (primary and secondary
11 Southern Subsurface Countercurrents, pSSCC and sSSCC, respectively), and (ii) the high surface
12 productivity rates induced by the coastal upwelling, which in turn triggers an intense oxygen
13 consumption in the subsurface. Their model experiments also showed that different Eddy Kinetic
14 Energy (EKE) levels, induced by different representations of the mean vertical structure of the coastal
15 current, may contribute to expand or erode the upper boundary of the OMZ.

16 The study by Montes et al. (2014) established a benchmark in terms of numerical modeling of the OMZ
17 in the SEP, focusing on its permanent regime and connection with the equatorial current dynamics. In
18 the present study, we also take advantage of the regional modeling approach in order to investigate the
19 mechanisms associated with the seasonal cycle of DO within the OMZ. The motivation for focusing on
20 seasonal variability is three-folds: 1) A better knowledge of the processes acting on the OMZ at
21 seasonal timescale is viewed as a prerequisite for interpreting longer timescales of variability (ENSO,
22 decadal); 2) the scarcity of quality long term subsurface biogeochemical data in the SEP is a limitation
23 for tackling the investigation of OMZ variability at low frequency; 3) To the authors' knowledge, this
24 issue has not been addressed in the literature for the Eastern Tropical Pacific, although it has been a
25 concern for other tropical oceans (Resplandy et al., 2012; Gutknecht et al., 2013a; Duteil et al., 2014).
26 Here, besides investigating to which extent the seasonal OMZ variability can relate to the variability of
27 the environmental forcing in the SEP (local wind, equatorial Kelvin and extra-tropical Rossby waves),
28 our interest is on examining the DO budget (i.e. the balance between oxygen sources and sinks) and
29 relating it to the physical DO flux. In particular, since the Peruvian region is the location of a relatively
30 intense eddy activity (Chaigneau et al., 2009), the question of whether or not eddy activity is involved

1 in the seasonal variability of the OMZ arises, and calls for assessing its contribution to the DO flux.
2 There is growing evidence that mesoscale activity has a key role on the biogeochemical cycles and the
3 OMZ structure in EBUS (Duteil and Oschlies, 2011, Nagai et al., 2015). Most studies addressing the
4 role of mesoscale processes in the OMZs have focused on the ventilation from the coastal domain,
5 where the primary production bloom provides nutrients and DO anomalies that are in turn transported
6 offshore (Stramma et al., 2013; Czeschel et al., 2015; Thomsen et al., 2016). Gruber et al. (2011)
7 showed that mesoscale activity is prone to reduce the biological production and offshore carbon export
8 in upwelling systems by both rectifying on the mean circulation (i.e. eddy-induced mixing tends to
9 flatten the isotherms nearshore and reduce the upwelling) and changing its nutrient transport capacity.
10 This process has been to some extent supported by observations in the Peruvian OMZ (Stramma et al.,
11 2013). In this sense, the mesoscale activity represents a ventilation pathway for the OMZ, through the
12 offshore transport of oxygen-enriched waters. The ventilation of the OMZ could also take place at its
13 meridional boundaries where strong mean DO gradients are found along with eddy activity. Recently,
14 Bettencourt et al. (2015) proposed that mesoscale eddies shape the Peruvian OMZ by controlling the
15 diffusion of DO into the OMZ at the meridional boundaries. Although it is likely that both processes
16 are important for understanding the OMZ structure, it has not been clarified to which extent the
17 variability of the OMZ could be understood in terms of the changes in the DO eddy flux into the OMZ
18 through these different pathways. The mesoscale activity also exhibits a significant meridional
19 variability off Peru (Chaigneau et al., 2009), which questions if the offshore ventilation process can
20 operate effectively for modulating the whole OMZ. Another related open question is at which
21 timescales the ventilation process through eddies-induced mixing can operate effectively. In this paper
22 we will tackle these issues from a regional modeling approach, focusing on the seasonal timescale.
23 The paper is organized as follows. After the Introduction (Section 1), we detail the observations and
24 model configuration used in the study, as well as the methodology employed in the treatment of the
25 information (Section 2). We also evaluate the realism of the simulation against the available
26 observations in reproducing the main characteristics of the OMZ. The subsequent section (Section 3)
27 characterizes the DO annual cycle inside the OMZ. Section 4 opens with the analysis of the seasonal
28 variability of the coastal OMZ, and the contribution of the DO budget terms associated with it. This
29 analysis is followed by the results on DO flux directed offshore and completed by the analysis of DO
30 flux across the OMZ meridional boundaries. Section 5 presents a discussion of the main results and

1 Section 6 presents a summary and the concluding remarks.

2

3 **2 Data description and Methods**

4 **2.1 Data**

5 **Dissolved Oxygen concentration from CARS**

6 The CSIRO Atlas of Regional Seas (CARS) is a climatological product derived from a quality-
7 controlled archive of historical subsurface ocean measurements, most of which was collected during
8 the past 50 years (Additional information might be found in the website of the project:
9 <http://www.marine.csiro.au/~dunn/cars2009/>). For the present study, we use the CARS2009 version of
10 the CARS product (Ridgway et al., 2002), which has an horizontal resolution of $0.5^{\circ} \times 0.5^{\circ}$ and 79
11 vertical levels, with a 10m resolution near the surface layer. We use CARS to assess the model skills in
12 simulating the OMZ mean state and variability. One advantage of this product is its refined
13 interpolation treatment near steep topography, in comparison to other products such as the World
14 Ocean Atlas (Dunn and Ridgway, 2002). Also, it includes the annual and semiannual oxygen cycles,
15 although the semiannual cycle is available only for the first 375 m over the region of interest due to the
16 scarcity of data.

17 **Chlorophyll-a concentration from SeaWiFS**

18 SeaWiFS 8 day composites at $0.5^{\circ} \times 0.5^{\circ}$ resolution chlorophyll product (version 4), between January
19 2000 and December 2008, is used to compute the surface chlorophyll seasonal cycle (McClain et al.,
20 1998; O'Reilly et al., 2000).

21 **Sea Surface Temperature (SST)**

22 The NOAA Optimum Interpolation SST (OISST V2) product, is contrasted against the simulation SST.
23 This product is an analysis constructed by combining observations from different platforms (satellites,
24 ships, buoys) on a regular global grid. More information about the methodology used to construct this
25 data set may be found in Reynolds et al. (2007), and the product website
26 (<https://www.ncdc.noaa.gov/oisst>). The version used in this study corresponds to daily SST maps with
27 a spatial resolution of $0.25^{\circ} \times 0.25^{\circ}$, spanning the period 2000-2008.

28 **Sea Level Height (SLH)**

29 TOPEX/JASON1 merged SLH data set, distributed by the Sea Level Research Group, University of
30 Colorado (<http://sealevel.colorado.edu/>) is used to derive the geostrophic velocity field and the mean

1 EKE field. This data set corresponds to a globally-gridded $0.25^{\circ} \times 0.25^{\circ}$ weekly product. The
2 information used corresponds to the period 1993-2008. Further details on this product may be found in
3 Nerem et al. (2010).

4 **2.2 Model simulation**

5 We use a high resolution simulation of the South Eastern Pacific, based on the hydrodynamic model
6 Regional Ocean Modeling System (ROMS) circulation model (see Shchepetkin and McWilliams,
7 2005; 2009 for a complete description of the model) coupled with a nitrogen-based biogeochemical
8 model developed for the Eastern Boundary Upwelling Systems (BioEBUS, Gutknecht et al., 2013ab),
9 hereby referred as CR BIO.

10 The model is used at an eddy-resolving resolution ($1/12^{\circ}$ at the equator) for a region extending from
11 12°N to 40°S and from the coast to 95°W -nevertheless this study only focuses on the domain spanning
12 the latitudes of Peru and Ecuador (Fig. 1)- with lateral open boundaries at its northern, southern and
13 western frontiers. The physical model resolves the hydrostatic primitive equations with a free-surface
14 explicit scheme, and a stretched terrain-following sigma coordinates on 37 vertical levels. The
15 configuration is similar to Dewitte et al. (2012), that is the open boundary conditions are provided by
16 3-daily mean oceanic outputs from SODA (Version 2.1.6) for temperature, salinity, horizontal velocity
17 and sea level for the period 1958-2008, while wind stress and speed forcing at the air/sea interface
18 come from the NCEP/NCAR reanalysis. The atmospheric fields have been statistically downscaled
19 following the method by Goubanova et al. (2011) in order to correct for the unrealistic wind stress curl
20 near the coast of the NCEP Reanalysis (see Cambon et al. (2013) for a validation of the method for
21 oceanic applications). Atmospheric fluxes were derived from the bulk formula using the temperature
22 from COADS $1^{\circ} \times 1^{\circ}$ monthly climatology (daSilva et al., 1994). Relative humidity and short wave and
23 long wave radiations are also from COADS. Bottom topography is from the GEBCO 30 arc-second
24 grid data set, interpolated to the model grid and smoothed as in Penven et al. (2005) in order to
25 minimize the pressure gradient errors and modified at the boundaries to match the SODA bottom
26 topography. This model configuration has been validated from satellite and in situ observations in
27 Dewitte et al. (2012) with a focus on mean state interannual variability. In general the model is skillful
28 in simulating the mean SST field (Fig. 1ac) as well as other main aspects of the mean circulation (e.g.
29 Peru Chile Undercurrent, EKE, see Figure 3 in Dewitte et al. (2012)), although with a slight cold bias
30 ($\sim 1^{\circ}\text{C}$), that could be partly attributed to the use of climatological heat flux forcing (Fig. 1d). The cold

1 bias observed here could be also due to a systematic warm bias in the AVHRR Pathfinder data,
2 observed in the nearshore regions where high SST gradients exist, specifically in the Humboldt region
3 (cf. Dufois et al., 2012). This product is extensively used in the construction of the OISST dataset
4 (Reynolds et al., 2007).

5 The mesoscale activity diagnosed from the mean EKE, has a comparable pattern than altimetry,
6 although with a larger amplitude (Fig. 1ef). Similar levels of mesoscale activity have been obtained by
7 previous modeling studies in the Peruvian region (e.g. Echevin et al., 2011; Colas et al., 2012).

8 The ocean model within this configuration is coupled to the BioEBUS model following similar
9 methodology than Montes et al. (2014). BioEBUS uses two compartments of phytoplankton and
10 zooplankton, small (flagellates and ciliates, respectively) and large (diatoms and copepods,
11 respectively), detritus, dissolved organic nitrogen and the inorganic nitrogen forms nitrate, nitrite and
12 ammonium, as well as nitrous oxide (see Gutknecht et al., 2013ab, for a description of the model). The
13 open-boundary conditions for the biogeochemical model are provided by the climatological CARS data
14 set (nitrate and oxygen concentrations) and by SeaWiFS archive (chlorophyll-a concentration).
15 Additional biogeochemical tracers are computed following Gutknecht et al. (2013ab). Initial
16 phytoplankton concentration is defined as a function of vertically extrapolated satellite Chl-a following
17 Morel and Berthon (1989). An offshore decreasing cross shore profile, following in situ observations, is
18 applied for zooplankton, and a vertical constant (exponential) profile is used for detritus (nitrite,
19 ammonium and dissolved organic nitrogen), respectively. In order to get a realistic solution for the
20 region, the model parameters were tuned to simultaneously fit modeled oxygen and nitrate fields to
21 observations (see Table A1 of Montes et al. (2014) for parameter values). These changes were
22 motivated by the need to adjust the microbiological rates to values observed in the SEP. Within this
23 parameter configuration, BioEBUS has been shown to be skillful for simulating the OMZ off Peru
24 (Montes et al., 2014). In particular the pattern correlations between the model and observations for both
25 the annual mean and the seasonal cycle inside the OMZ present comparable scores (>0.85 , cf. Montes
26 et al. (2014)) as well as low standard deviations (i.e. in the order of the observed values). The model
27 was run over the period 1958-2008 with a 10-year spin-up obtained by repeating the year 1958.
28 Although, after the spin-up, the simulation has reached stable conditions and the OMZ volume does not
29 drift, we focus in the present study only on the period 2000-2008.

30 The reason for focusing on the last ten years of our simulation is also motivated by the fact that the

1 atmospheric momentum forcing is close to the satellite QuickSCAT winds by construction (see
 2 Goubanova et al. (2011) for details) so that this period of the simulation is the one when the model is
 3 the most constrained by observations. Most previous modeling studies for this region (Penven, et al.
 4 2005; Montes et al., 2010, 2014; Echevin et al., 2011; Colas et al., 2012) have also used a wind forcing
 5 from the QuickSCAT scatterometer, which provides a benchmark for assessing our simulation.
 6 A monthly mean climatology is calculated for all variables over this period from the 3-day mean
 7 outputs of the model, which can be compared to the CARS data.
 8 Consistently with Montes et al. (2014), the coupled simulation is skillful in simulating the mean
 9 characteristics of the OMZ off the Peruvian coast (Figures 2 and 3). In particular the thickness and
 10 location of the model OMZ core limits are realistic, and in good agreement with previous studies (Fig.
 11 2; e.g. Paulmier et al., 2006; Cornejo and Farías, 2012; Montes et al. 2014). Note that the simulation
 12 reproduces a thinner OMZ around 10°S in comparison to CARS, which agrees with the results
 13 obtained by Montes et al. (2014), (see Fig. 2 in that study). Close to the western boundary of our model
 14 domain, the simulated OMZ also exhibits a realistic vertical structure (Fig. 3) with comparable
 15 concentration in DO than observations in the vicinity of the Equatorial Undercurrent (~100 m;
 16 Equator). Furthermore, the simulation is consistent in reproducing the oxygen consuming processes, as
 17 supported by the Apparent Oxygen Utilization (AOU; Fig. 4), also in good agreement with previous
 18 studies (cf. Figure 8 in Cabré et al. (2015)). AOU was computed as the difference between the DO
 19 concentration and the saturated oxygen (O_{2sat}) concentration ($AOU = O_{2sat} - O_2$) with O_{2sat} computed
 20 following the methodology of García and Gordon (1992). The realistic representation of the oxygen
 21 consuming processes is reflected by the Particulate Organic Carbon flux as well (Fig. 5a), whose
 22 values at 100m fall within the observed range for the region (30-60 $gC\ m^{-2}\ yr^{-1}$ in the shelf area; Dunne
 23 et al., 2005; Henson et al., 2012). In addition, the low transfer efficiency of carbon (10-15% or lower
 24 over and next to the shelf; Henson et al., 2012), from the euphotic zone to greater depths (Fig. 5b),
 25 implies that the remineralization processes take place at realistic depths, and therefore allow for a
 26 correct vertical representation of the OMZ (cf. Fig. S2 in Cabré et al. (2015) for comparison).
 27 The core of the OMZ, defined with a suboxic concentration ($[DO] < 20\ \mu M$; μM will be used to refer to
 28 $\mu mol\ L^{-1}$ in all the text and Figures), occupies nearly 23% of the domain volume (Fig. 6a), with the less
 29 oxygenated layers comprised between 5°S and 15°S, and 100 m and 600 m depth (Fig. 3). As expected,
 30 the simulation presents more details than the climatological product (Fig. 3). Moreover, we computed a

1 geographical OMZ overlapping metric following Cabré et al. (2015), which quantifies the spatial
2 agreement of the OMZ volume distribution between the simulation and CARS, varying between 0 (no
3 agreement) and 1 (perfect collocation). We obtained a value of 0.79, which is ~58% above the best
4 CMIP5 models used in Cabré et al. (2015).

5 Despite the overall good agreement between the model and observations, the modeled oxygen content
6 is however underestimated as compared to CARS in certain regions of the domain, particularly
7 southwards of 20°S (Fig. 3a) and close to the coast (Fig. 3d). In particular, the volume of the suboxic
8 DO concentration range is 6% lower in the simulation (Fig. 6a), which is comparable to the differences
9 obtained by Montes et al. (2014) in a similar analysis of the OMZ.

10 The modeled DO distribution is also characterized by finer spatial scales of variability inside the OMZ
11 compared to observations (Figures 3c and 3d). In particular, the model oxycline is shallower and with a
12 more intense DO gradient than the observations, which has been also observed in a simulation of the
13 Arabian Sea OMZ (Resplandy et al., 2012), suggesting that the CARS data set may not have a
14 sufficient vertical resolution to realistically resolve the oxycline at some locations. Also, it must be kept
15 in mind that CARS is built using all the available data from the second half of the twentieth century
16 (1940-2009), whereas we focus on the period 2000-2008 for the simulation, which is known to be a
17 colder period than the previous decades in the eastern tropical Pacific (Henley et al., 2015). Other
18 limitation for the comparison between model and data includes the errors associated with the scarcity
19 of data in some regions (Bianchi et al., 2012). Nonetheless, the simulation is in good agreement with
20 CARS in terms of mean characteristics of the OMZ, as well as the mean oxygen concentration and its
21 distribution (Figures 4, 6a).

22 In order to evaluate the realism of the seasonal cycle, we estimate the seasonal variability of the
23 volume of water within the suboxic DO concentration range 0-20 μM in both the model and data (Fig.
24 6b). The results indicate that, despite a weaker amplitude (by 15% on average), the seasonal cycle of
25 the OMZ core is relatively well simulated by the model. For hypoxic DO volume in the range 40-50
26 μM , the agreement is as good as inside the OMZ core, with a Pearson correlation value of 0.9 and a
27 volume RMS difference of 16%, between the simulation and the observations.

28 In order to summarize the model validation, we present a Taylor diagram showing the statistics of the
29 comparison between the model and observations for a depth range encompassing the OMZ (Fig. 7).
30 This analysis indicates that within the present model configuration, we reach a comparable skill than

1 the model configuration of Montes et al., (2014) (their Figure 1). The good agreement of the seasonal
 2 cycle between CARS and the simulation, in addition to the consistency of our results with those of
 3 Montes et al. (2014), provides confidence in using the model outputs for investigating the processes
 4 associated with the seasonal variability of the OMZ.

5 **2.3 Methods**

6 In this work, our approach is twofold: First, the biogeochemical processes for DO are investigated
 7 explicitly through the on-line oxygen budget (1). Although this methodology can provide a direct
 8 estimate of the seasonal variability in advection and mixing, it does not allow for a direct estimate of
 9 the eddy contribution to DO change that can also vary seasonally. The DO flux associated with
 10 different timescales of variability is therefore estimated. This consists in computing the temporal
 11 average of the cross-products between DO and velocity anomalies. Anomalies can refer either to
 12 seasonal anomalies and in that case, this provides the mean seasonal DO flux ($\langle \tilde{u} \cdot \tilde{O}_2 \rangle$, where \sim refers
 13 to the seasonal anomalies), or to the intraseasonal anomalies (calculated here as the departure from the
 14 monthly mean) and in that case, this provides an estimate of the mean DO eddy flux ($\langle u' \cdot O_2' \rangle$, where
 15 the apostrophe refers to the intraseasonal anomalies). In this paper we are also interested in the
 16 seasonality of the DO eddy flux. This is estimated from the monthly-mean seasonal cycle of the mean
 17 DO eddy flux calculated over a 3-month running window, and is now referred to as $\overline{\langle u' \cdot O_2' \rangle}$. The
 18 climatological EKE activity is estimated similarly.

19 The DO budget consists in the following Equation:

$$20 \quad \frac{\partial O_2}{\partial t} = -\vec{u} \cdot (\vec{\nabla} O_2) + K_h \nabla^2 O_2 + \frac{\partial}{\partial z} \left(K_z \frac{\partial O_2}{\partial z} \right) + SMS(O_2) \quad (1)$$

21 The first three terms on the right hand side represent the physical processes involved in the changes in
 22 oxygen concentration. The first term stands for the advection of oxygen, with \vec{u} the velocity vector
 23 (note that the model determines the vertical velocity component from the continuity equation). The
 24 second term corresponds to the horizontal subgrid-scale diffusivity (with K_h the eddy diffusion
 25 coefficient equal to $100 \text{ m}^2\text{s}^{-1}$ in this version of the model), and the third term corresponds to the
 26 vertical mixing (with turbulent diffusion coefficient K_z calculated based on the KPP mixing scheme
 27 (Large et al., 1994)). Note that the model has also numerical diffusion associated with inherent spurious
 28 diapycnal mixing of the numerical scheme, so that K_h is empirically adjusted

1 The fourth term represents the “Sources-Minus-Sinks” contribution to the oxygen changes, directly due
 2 to biogeochemical activity. Biogeochemical processes correspond to the sum of oxygen sources and
 3 sinks, namely the photosynthetic production, and the aerobic processes (oxic decomposition, excretion
 4 and nitrification). In this study, for simplicity, those will be considered as a summed-up contribution to
 5 the DO rate of change, whereas physical processes will be divided into advection and mixing terms.
 6 Each term of this oxygen budget is determined on line at each time integration. While horizontal
 7 diffusion and vertical diffusivity are explicit sources of mixing, they are not the only terms contributing
 8 to mixing. Later on in the paper, unless stated otherwise, the term mixing will refer to the integrated
 9 effect of all processes contributing mixing directly or indirectly. Besides the horizontal diffusion (
 10 $K_h \cdot \nabla^2 O_2$) and vertical mixing ($\frac{\partial}{\partial z} \left(K_z \frac{\partial O_2}{\partial z} \right)$), mixing can be also induced by non-linear advection.
 11 The latter corresponds to $(u' \partial u' / \partial x) + (v' \partial v' / \partial y) + (w' \partial w' / \partial z)$, assuming the Reynolds decomposition
 12 for the velocity field, i.e. $\vec{u} + \vec{u}'$, where \vec{u}' accounts for the intraseasonal variability (periods lower
 13 than ~3 months).
 14 In the SEP, the subthermocline seasonal variability can be interpreted as resulting from the propagation
 15 of Extra-Tropical Rossby Waves (ETRW). ETRW radiate from the coast and propagate vertically,
 16 inducing a vertical energy flux, whose trajectory follows the theoretical Wentzel-Kramers-Brillouin
 17 (WKB) ray paths (Dewitte et al., 2008; Ramos et al., 2008). The energy flux results from the phase
 18 relationship between vertical velocity associated with the vertical displacement of the isotherms, and
 19 the pressure fluctuations associated with them. In the regions sufficiently below the thermocline for
 20 DO consumption to become weak (that is DO can be considered a passive tracer), it is expected that
 21 changes in DO relate to the anomalous velocity field, and that the DO flux shares comparable
 22 characteristics than the Eliassen-Palm flux (EP flux; Eliassen and Palm, 1960). The trajectories of the
 23 WKB ray paths are a function of latitude, local stratification and the phase speed of the Rossby wave
 24 (see Ramos, et al., (2008)). The latter consists in the superposition of a certain number of baroclinic
 25 modes, in order to propagate vertically, so the phase speed can range from values between c_1 to c_n ,
 26 where c_n is the theoretical phase speed of a n^{th} baroclinic mode, obtained from the vertical mode
 27 decomposition of the local density profile.

28

29 **3 Characteristics of the DO annual cycle**

1 While the annual signal is a conspicuous feature inside the region (Fig. 6b), it could manifest
 2 differently across the OMZ. As a first step towards investigating processes driving the rate of DO
 3 change, it appears important to document the vertical structure variability of the DO annual cycle
 4 within the OMZ. The amplitude and phase of the annual harmonic of the model DO climatology is
 5 presented along a zonal section off central Peru (12°S, Fig. 8ab), where the OMZ core is extensive
 6 (Fig. 2). The DO climatology has been normalized by its RMS (Root Mean Square), in order to
 7 emphasize the regions where the amplitude in DO changes (and mean DO) is weak. The amplitude
 8 reveals a complex pattern with three regions of large relative variability: 1) near the coast (i.e. fringe of
 9 ~150 km) between the oxycline and 400 m; 2) offshore between 82°W and 84°W in the upper 400 m
 10 and 3) below 500 m. The phase lines over these three regions suggest distinct propagating
 11 characteristics: whereas in the coastal region there is no propagation, in the offshore and deep region,
 12 there is indication of a westward propagation. In the region below 500 m, the phase lines tend also to
 13 be parallel and slope downward, suggestive of westward-downward propagation (estimated phase
 14 speed of $\sim 2.5 \text{ cm s}^{-1}$). These propagating characteristics can be evidenced in the Hovmöllers of the
 15 recomposed annual cycle at the depth of 150 m (Fig. 8c) and 700 m (Fig. 8d). While at 150 m the
 16 annual signal does not clearly propagate and only shows two domains of high amplitude, separated by
 17 low amplitude values (Fig. 8c), there is a clear westward propagation of the DO anomalies at 700 m,
 18 with the phase speed increasing westward. At 400 m, the propagation is only observed west of 81°W
 19 (Fig. 8b). In addition to the large vertical structure variability of the annual cycle, the OMZ annual
 20 cycle is also characterized by a large horizontal variability in particular at its northern and southern
 21 boundaries. This is illustrated from Figure 9, that displays the amplitude of the annual cycle of the DO
 22 climatology at 400 m, and evidences amplitude peaks at the OMZ meridional boundaries (between the
 23 20 and 45 μM isopleths).

24 The annual variability pattern evidenced above results from a delicate balance between the physical
 25 processes (namely advection and mixing, cf. Eq. (1)) and the biogeochemical processes (consumption
 26 versus production). As a first step towards investigating each term of the DO budget, it is interesting to
 27 evaluate the relative contribution of the physical and biogeochemical fluxes to the DO variability at
 28 seasonal scale. The RMS of the climatological fluxes along a section at 12°S indicates that the
 29 maximum amplitude of the seasonal fluxes takes place near the oxycline and along the coast over the
 30 whole water column (Figure 10). The relative importance of the physical processes against the

1 biogeochemical processes varies across the OMZ. At the coast and near the oxycline, the annual
2 variability of the biogeochemical processes reaches values almost half those of the variability in
3 physical processes (Fig. 10c), as a consequence of the proximity to both the well lit and highly
4 productive part of the water column, and the high remineralization activity that occurs near the
5 oxycline. Towards offshore and at depth, the relative importance of the variability of the
6 biogeochemical processes reduces gradually. Near ~300 m the variability of the biogeochemical
7 processes is nearly 1/5 of the physical processes variability. Below ~300 m, and towards the lower part
8 of the OMZ core and below, the physical processes variability is one order of magnitude larger.
9 Consequently, the distribution of DO in the lower part of the OMZ is rather a function of
10 advection/diffusion than a consequence of the biogeochemical processes, although DO consumption
11 even at very low levels has the potential to generate local gradients and therefore induce advection. The
12 spatial heterogeneity in the seasonal DO changes induced by the biogeochemistry and dynamics as
13 described above, appears as an ubiquitous feature in the OMZ. To illustrate this, we estimate the
14 proportion of explained variance of the seasonal DO rate of change by the physical fluxes as:

$$15 \quad R_{phys.}^2 = (1 - RMS(Biogeochemical\ Fluxes) / RMS(Total\ Fluxes)) \cdot 100 . \quad (2)$$

16 Figure 11ab presents the results of $R_{phys.}^2$ at 100 and 450 m depth, which evidences that the relative
17 importance of the physical fluxes versus the biogeochemical fluxes in the seasonal DO variability
18 increases with depth, and is enhanced at the OMZ boundaries. On the other hand, the biogeochemical
19 fluxes explain more than 50% of the variance in seasonal DO change rate in a narrow (~ 200 km width)
20 coastal fringe that extends more offshore to the north of the domain (around 8°S; Fig. 11a) and
21 vertically down to 300 m (Fig. 11c).

22 Based on the above analysis, it is clear that the coastal region (first 200-300 km from the coast) below
23 the oxycline corresponds to a territory where the seasonal variability of biogeochemical and physical
24 fluxes have a comparable magnitude, whereas outside this region, notably in the lower part of the OMZ
25 core, the physical fluxes variability dominates over the biogeochemical fluxes variability at seasonal
26 timescale. Hereafter we examine the possibility of two distinct regimes of OMZ dynamics at seasonal
27 timescale: one associated with the upper OMZ (including coastal domain and meridional boundaries),
28 and the other one associated with the deep OMZ. In the following we investigate the processes
29 responsible for the DO flux.

1

2 **4 Seasonality of the OMZ ventilation**

3 It has been shown for the SEP that the DO content near the coast is set to a large extent from the
4 transport of oxygen deficient waters from the equatorial current system, particularly the oxygen
5 depleted sSSCC (Montes et al., 2014). Therefore, the seasonal variability of DO is likely to result in
6 part from the seasonal variability of the different branches of the EUC in the far eastern Pacific. Local
7 wind stress forcing (and its intraseasonal activity) has also a marked seasonal cycle off Peru (Dewitte et
8 al., 2011) which may impact both the upwelling dynamics -through Ekman pumping/transport- and
9 mixing. Some studies also argue that the DO exchange between the coastal domain and the OMZ takes
10 place through the offshore transport of DO poor waters by eddies (Czeschel et al., 2011), implying that
11 the variability of such processes is set up by coastal processes that determine the nature of the DO
12 source. As a first step, we investigate the mechanisms responsible for the seasonal variability in DO
13 along the coast, which can be considered as the eastern boundary of the OMZ. This is aimed at
14 providing material for the interpretation of the offshore DO flux variability.

15 **4.1 The coastal domain as the eastern boundary of the OMZ: variability and** 16 **mechanisms**

17 We analyze the seasonal variability along the coast, at a section at 12°S. Similar results are obtained for
18 latitudes between 7°S and 14°S (not shown), which corresponds to the latitude range where the PUC is
19 well defined. The results are also presented in terms of the first EOF mode, in order to ease the
20 interpretation of the variability, reduced as a spatial pattern modulated by a seasonal timeseries. It was
21 verified in particular that the consideration of the first EOF mode of each term leads to an almost
22 perfect closure of the DO budget (see below, Table 1). Figure 12 displays the first EOF mode of
23 various climatological fields in a section at 12°S near the coast and from the oxycline (45 μM isoline)
24 to the depth of 300 m. Figure 13 shows the principal components associated with the first EOF mode
25 patterns. The seasonal DO cycle is dominated by an annual component, with a peak centered in August
26 (Fig. 13a), and the largest variability at the coast below the oxycline that extends offshore and
27 downward, resulting in an elongated tongue below 100 m near ~78°W (Fig. 12a). During the first
28 quarter of the year, oxygen anomalies remain relatively stable (oxygen rate nearly zero, Fig. 13b), and
29 negative, due to a high production of organic matter in Austral summer (cf. Fig. 1c of Gutiérrez et al.,
30 2011) that stimulates a subsurface oxygen consumption associated with the degradation of this organic

1 matter. DO anomalies start to increase during the second quarter, become positive in June and reach
 2 their maximum in August (Fig. 13a). The peak anomaly in Austral winter could be understood in terms
 3 of the increased mixing (see Fig. 13a showing EKE peaking in July) associated with the increase in
 4 baroclinic instability due to the seasonal intensification of the PUC from June. Note that the pattern of
 5 the first EOF of the alongshore current coincides with the mean position of the PUC (see Fig. 12b), so
 6 that seasonal variations of the PUC can be interpreted in terms of the variations in the vertical shear of
 7 the coastal current system. Other processes that may explain the peak DO anomaly in Austral winter
 8 includes the reduced productivity and downwelling that peaks in June (Fig. 13c), associated with
 9 seasonal equatorial downwelling Kelvin wave.

10 The following investigates the tendency terms of the DO budget, in order to quantitatively interpret the
 11 DO seasonal cycle near the coast. Given that the analysis is performed inside the 45 μM isopleth, the
 12 biogeochemical flux term is largely dominated by the “Sinks” terms (aerobic processes; one order of
 13 magnitude larger than “Sources”), driven by organic matter remineralization and zooplankton
 14 respiratory metabolic terms (not shown). For clarity, the seasonal DO budget is presented synthetically,
 15 from the first EOF mode of the climatological advection, mixing (horizontal and vertical diffusion) and
 16 biogeochemical fluxes terms. Although this does not warranty a perfect closure, it eases the
 17 interpretation. Note that the residual resulting from the difference between the first EOF mode of the
 18 rate of DO changes and the summed-up contribution of all the other terms in Figure 13b is rather weak,
 19 validating to some extent our approach (see also Table 1). First of all, we find that the largest amplitude
 20 of the mode patterns is found near the coast and inside the mean PUC core (Figs. 12d to 12g). During
 21 the first part of the year (January to May), positive advection anomalies are compensated by mixing
 22 (horizontal and vertical diffusion), and maintain the rate of DO change relatively low (Fig. 13b; Table
 23 1). Biogeochemical fluxes anomalies are positive during that period, associated with a positive
 24 anomaly of primary production in the well lit surface layers, implied by the high chlorophyll-a values
 25 (Fig. 13c). A positive oxygen anomaly is sustained by the advection terms and the biogeochemical
 26 terms, and is balanced out by the mean advection of low DO waters carried by the PUC (Montes et al.
 27 2010; 2014), generating the relatively stable oxygen values (oxygen rate nearly zero).

28 From May, the rate of DO changes increases concomitantly with EKE (Fig. 13ab), followed one month
 29 later by mixing (horizontal and vertical diffusion), whereas advection and biogeochemical fluxes
 30 decrease. By June-July, the intensification in alongshore winds (Fig. 13c) starts to propel the coastal

1 upwelling, which has two compensating effects: on one hand it triggers photosynthesis in the lit surface
2 layers (DO rate turns to positive values) and on the other, it uplifts low oxygen waters from the OMZ.
3 The intraseasonal wind activity also starts to increase at that time (cf. Fig. 13c; see also Dewitte et al.,
4 2011) which favors mixing, and so the downward intrusion of positive DO anomalies (note the
5 deepening of the mixed layer in Fig. 13c). The overall effect is an increase in DO which leads to a peak
6 anomaly in August. At that time, the DO rate drops sharply due to the strong subsurface DO
7 consumption (Table 1) associated with aerobic remineralization of organic matter produced earlier in
8 the season (DO rate moves sharply to negative values) and the high mixing that brings DO depleted
9 waters from the subsurface into the deepened mixed layer. Note that this is consistent with the decrease
10 in surface chlorophyll-a (Fig. 13c) and the interpretation proposed by Echevin et al. (2008) to explain
11 the Austral summer minimum in surface chlorophyll-a observed off Central Peru.
12 This change to oxygen poor conditions combines with the natural decrease in oxygen production
13 towards the end of the upwelling season and coincides with a restratification of the water column,
14 which restricts the oxygenated waters near the surface (Echevin et al., 2008). This altogether
15 contributes to maintain a negative DO rate inside the coastal OMZ, despite the increase in anomalous
16 DO flux from the advective terms and (later on) biogeochemical processes towards the end of the year.
17 As a result, oxygen returns to low values towards the end of the year.

18 **4.2 Offshore flux**

19 While the coastal OMZ variability is heavily constrained by the environmental forcings –coastal
20 upwelling, coastal current system and local wind– due to the shallow oxycline there, the offshore
21 OMZ, as embedded in the shadow zone of the thermohaline circulation, is somewhat insensitive to
22 direct local forcing and rather experiences remote influence in the form of westward propagating
23 mesoscale eddies (Chaigneau et al., 2009) and ETRW (Ramos et al., 2008; Dewitte et al., 2008). The
24 influence of westward propagating mesoscale eddies on the OMZ translates as the transfer of coastal
25 water properties towards the open ocean (DO included), while these properties are altered during
26 transport due to physical-biogeochemical interactions (Stramma et al., 2014; Karstensen et al., 2015).
27 Towards the end of their lifetime, hydrographic and biogeochemical anomalies carried by eddies are
28 redistributed in the ocean (Brandt et al., 2015), linking the coast and the open ocean. Although most
29 eddies genesis takes place near the coast (Chaigneau et al., 2009) and seasonal ETRW have a coastally
30 forced component (Dewitte et al., 2008), we expect different characteristics of the seasonal variability

1 in DO between the coast and the open ocean, given that oxygen demand will change from one region to
2 the other. We also distinguish the mean DO flux associated with the annual component of the
3 circulation that represents the transport in DO associated with seasonal change in the large scale
4 circulation, and the annual variability of the eddy DO flux that corresponds to the annual changes in the
5 transport due to eddies. These two quantities are diagnosed at 12°S (Figs. 14 and 15). The DO has been
6 normalized by its climatological variability in order to emphasize variability patterns where DO is low.

7 **Mean seasonal flux**

8 We first document the mean DO flux associated with the annual component of the circulation. It
9 consists in the mean of the cross-product of the annual harmonics of the climatological velocity and
10 DO (Fig. 14a). The results indicate that the amplitude of the annual DO flux is maximum near the coast
11 and below ~400 m and it tends to be orientated westward-downward, following approximately the
12 trajectories of theoretical WKB paths for the annual period Rossby wave. Note that this is consistent
13 with the westward propagating pattern of DO below 400 m evidenced earlier (Fig. 8). As a consistency
14 check, we also estimated the annual energy flux vector in the (x,z) plan associated with a long extra
15 tropical Rossby wave, that is $(\langle p^{1\text{yr}} \cdot u^{1\text{yr}} \rangle, \langle p^{1\text{yr}} \cdot w^{1\text{yr}} \rangle)$ where the superscript denotes the annual
16 harmonics and the bracket the temporal average (Fig. 14b). The flux vector indicates vertical
17 propagation of energy at the annual period and the pattern of maximum flux coincides approximately
18 with the region of maximum amplitude of the mean seasonal DO flux. This suggests that the annual
19 ETRW is influential on the DO flux below ~400 m. This is interpreted as resulting from the advection
20 of DO by the ETRW since biogeochemical fluxes have much less influence on the DO rate of change
21 below 400 m (Fig. 10c) and the amplitude of the annual cycle of climatological DO eddy flux has a
22 much reduced amplitude below that depth (Fig. 15a) suggesting a reduced contribution of horizontal
23 and vertical diffusion to the DO budget. Note that the DO (Fig. 15a) was normalized prior to compute
24 the DO eddy flux in order to render both the analysis akin, and therefore contrast the flux associated
25 with the annual ETRW against the annual DO eddy flux. It was verified that the vertical structure
26 variability of the annual DO flux described above for the section of 12°S is comparable at other
27 latitudes within the OMZ. In particular the annual DO flux tends to remain homogeneous along
28 trajectories mimicking the energy paths of the ETRW at annual period which slope becomes steeper to
29 the South (not shown).

30 **Seasonal eddy flux**

1 As previously described, the annual amplitude of the climatological DO eddy flux is the largest in the
2 upper 400 m near the coast at 12°S consistently with the high EKE in this region. Since EKE is large
3 along the coast of Peru, exchange of DO induced by eddies could be expected at all latitudes, with a
4 direction that depends on the sign of the DO gradient at the coast. Figure 16 presents the annual
5 harmonic of the climatological DO eddy flux along the coast and averaged in a coastal fringe distant 1°
6 from the coast and 2° width. The maximum amplitude –reaching $\sim 1 \text{ cm s}^{-1} \mu\text{M}$ – is concentrated in the
7 upper oxycline (Fig. 16a) with a peak during Austral winter. The peak season is also confirmed by the
8 EOF analysis of the climatological DO eddy flux (not shown). Despite the relative large meridional
9 variability in the amplitude, the mean vertical structure of the DO eddy flux consists in an approximate
10 exponentially decaying profile with depth, with a decay scale of $\sim 90 \text{ m}$ (Fig. 16b) so that at 300 m the
11 seasonal DO eddy flux is only 19% of that at 100 m on average along the coast. Figure 16a also reveals
12 that the annual DO eddy flux is larger towards the northern rim of the domain and extends deeper than
13 towards the south. The high values are increasingly confined close to the surface towards the southern
14 part of the domain, in comparison to the northern part, although the vertical attenuation displays a
15 similar scale.

16 **4.3 Meridional boundaries**

17 Here, our objective is to document the seasonality of the DO eddy flux. As a first step, we estimate the
18 distribution of mean DO eddy flux, in order to identify the regions where its magnitude is large and
19 thus where it is likely to vary seasonally with a significant amplitude.

20 **Mean seasonal flux**

21 The horizontal distribution of mean DO eddy flux displays the highest values at the boundaries of the
22 OMZ core (Fig. 17), and adjacent to the $45 \mu\text{M}$ isopleth. Towards the inner OMZ, the mean DO eddy
23 flux values decrease notoriously, with a factor of nearly 10 between the interior and exterior of the $10 \mu\text{M}$
24 contour. In agreement with the observations reported in the previous section, the mean DO eddy
25 flux decreases sharply with depth (approximately one order of magnitude between 100 m and 700 m),
26 with the highest values concentrated near the oxycline as expected from the increasing oxygen
27 concentration in this part of the OMZ. In this sense, the pattern of DO eddy flux around the depth of
28 the oxycline encloses a region of high variability (not shown).

29 To gain further insight with respect to the vertical structure of the DO eddy flux and at the same time,
30 diagnose the role of the mesoscale activity at the boundaries of the OMZ, we compute the mean DO

1 eddy flux across the two sections that correspond to the northern and southern limits of the OMZ
2 (depicted in Fig. 18). These limits are defined based on Figure 17, and are located in the provinces of
3 high amplitude of the mean DO eddy flux.

4 The DO eddy flux across each of the north and south boundaries was computed by averaging the
5 product of the fluctuating velocity component normal to the boundary in the horizontal directions and
6 the fluctuating DO concentration component, thereby obtaining horizontal eddy fluxes.

7 As observed in Figure 17, the highest values for both north and south boundary sections are also
8 comprised between the oxycline and the lower OMZ core limit (Fig. 18), being almost one order of
9 magnitude smaller at greater depths (Fig. 18c). These high values, located between ~100-300 m, are
10 followed by a sharp decrease (average decrease of $1.5 \text{ cm s}^{-1} \mu\text{M}$ in 100 m). At the range of depths
11 between 100 m and 300 m, the DO eddy flux displays higher values at the southern boundary (nearly
12 twice as large) when compared with the northern boundary. This relationship is less clear when
13 analyzing the lower part of the OMZ. At both meridional boundaries, the mean DO eddy flux in the
14 upper part of the OMZ is nearly one order of magnitude larger than in the lower part.

15 **Seasonal eddy flux**

16 We now document the seasonal variability of the DO eddy flux across the OMZ boundaries analyzed
17 above (Fig. 18). An EOF analysis of the mean seasonal cycle of the DO eddy flux is performed at the
18 boundary sections previously defined. The Figure 19 presents the first EOF mode patterns along with
19 the associated timeseries. In order to estimate the uncertainty associated with the location of the OMZ
20 boundaries, we repeated this analysis for 12 nearby sections parallel to the boundaries and spaced by
21 ~20Km. This leads to an estimated error (standard deviation across the different sections) of the DO
22 eddy flux. The error is represented as a colored shading in the Figures 19bde. At both locations, the
23 first EOF accounts for a well defined seasonal cycle. At the northern boundary (Fig. 19a), the seasonal
24 cycle of the DO eddy flux peaks in Austral Winter, in phase with the DO changes along the coast (Fig.
25 16). Note that the seasonal cycle is in phase with the one of the intraseasonal activity of the horizontal
26 current normal to the section, which was estimated the same way than the climatological eddy flux (see
27 red line in Fig. 19b), supporting the idea that the climatological DO eddy flux results from anomalous
28 advection. The amplitude of the mode pattern is maximum at the oxycline with DO between 20 and 45
29 μM , and presents a sharp decrease below the OMZ core depth (Fig. 19a). This sharp decrease is
30 evidenced by the mean vertical profile of the DO eddy flux seasonal variability estimated as the RMS

1 across the section of the EOF mode pattern (Fig. 19e). The vertical structure of the DO eddy flux
2 variability indicates that there is a difference of nearly one order of magnitude between 100 and 300 m
3 depth. From that depth on, the DO eddy flux variability decreases linearly.
4 In contrast with the northern boundary, the seasonal variability at the southern boundary peaks during
5 Austral Spring (Fig. 19d), in phase with the intraseasonal activity of the horizontal currents normal to
6 the section. The amplitude of the seasonal cycle is the largest around the depth of the oxycline, and
7 remains high down to the vicinity of the OMZ core upper limit (Fig. 19c). Below the depth of the OMZ
8 core, the amplitude of the EOF mode decreases sharply (\sim one order of magnitude in 100m; Fig. 19c).
9 This is evidenced by the profile of the DO eddy flux seasonal variability, estimated in the same manner
10 as for the northern boundary (Fig. 19e). This profile shares some characteristics with its counterpart at
11 the northern boundary, meaning, a sharp decrease between the oxycline and the OMZ core depths,
12 suffering a reduction of nearly 90% (Fig. 19e). On the other hand, the variability along the southern
13 boundary is \sim 70% larger than along the northern boundary. At both boundaries, the zonal wavelength
14 of the seasonal DO eddy flux variability along the boundary is estimated to be in the order of $\sim 10^2$ km,
15 a scale that falls within the range of observed eddies diameter (Chaigneau and Pizarro, 2005), which
16 indicates that locally there can be an injection or removal of DO across the boundary on average over a
17 season. The mean DO eddy flux across the boundaries is nevertheless positive.

18

19 **5 Discussion**

20 We now discuss some limitations and implications of our results. While the model realistically
21 simulates the main characteristics of the OMZ (position, intensity, average volume and seasonal
22 variations), it still presents biases that could be influential on our results. In particular, and since the
23 coastal domain is viewed here as a boundary of the OMZ, it is important to have a realistic mean DO
24 concentration there. Compared to CARS, the simulated suboxic volume is however underestimated by
25 \sim 6%, and 85% of this error can be attributed to the coastal domain (fringe of 3° from the coast). This
26 bias could be due to several factors. Montes et al. (2014) observed variations of the suboxic volume in
27 the order of 5%, when contrasting two simulations that used different oceanic open boundary
28 conditions, which indicates a sensibility of the simulated OMZ to the physical parameters and the
29 representation of the Equatorial Current system. This bias could also be partly due to coastal sediments
30 processes (DO demanding processes) that are not represented in our simulation. Using a similar

1 configuration to the one used in the present study on the Namibian OMZ, Gutknecht et al. (2013a)
2 observed that the differences between the simulated OMZ volume and CARS increased towards the
3 shelf, which could be related to the exclusion of the DO demand from the sediments in the model. On
4 the other hand, in a study on the impact of sediment biogeochemistry upon the water column
5 biogeochemical cycles in the northern end of the California Current system, Bianucci et al. (2012)
6 argued that the sediment denitrification is balanced by a nitrification in the water column, obtaining
7 similar bottom DO concentrations between their experiments even when disregarding the DO sediment
8 demand. Nevertheless, the interaction between the sediments processes and the water column in terms
9 of DO consumption in the mid latitude upwelling systems is still unclear. Regarding this point, the
10 inclusion of a sediment module in the current model setting is planned for future work.

11 Besides other likely sources of biases related to an imperfect model setting (e.g. use of relatively low
12 resolution atmospheric forcings near the coast, absence of air-sea coupling at mesoscale, absence of
13 coupling with benthic oxygen demand or consideration of N_2 fixation), another inherent limitation of
14 our study is related with the difficulty to validate some aspects of the eddy field, in particular its
15 vertical structure. This might be overcome in the future as the Argo coverage increases (cf.
16 TPOS2020).

17 With the limitations of our regional modeling approach in mind, it is worthwhile discussing some
18 implications of our results. While previous studies have mostly focused on the role of the mean DO
19 eddy flux in shaping the OMZ boundaries (Resplandy et al., 2012; Brandt et al., 2015; Bettencourt et
20 al., 2015), we have documented here the seasonal variability in the DO eddy flux at the OMZ
21 boundaries, including the “eastern boundary” formed by the coastal system. We infer that the
22 seasonality of the DO eddy flux is controlled by different physical processes depending on the
23 boundary. At the “eastern boundary”, there is a constructive coupling between eddies resulting from the
24 instability of the PUC peaking in Austral winter, and the enhanced DO along the coast resulting from
25 an increased horizontal and vertical diffusion at the same season.

26 At the northern boundary of the OMZ, the DO eddy flux is also related to the strong EKE around $5^\circ S$
27 that peaks in Austral winter. Despite the fact that the OMZ northern boundary is embedded in the
28 equatorial wave guide, since the intraseasonal Kelvin wave activity tends to peak in Austral summer
29 (Illig et al., 2014) it can be ruled out that the seasonal cycle in DO eddy flux is strongly linked to the
30 intraseasonal long equatorial waves. The results of Echevin et al. (2011) also suggest that the enhanced

1 mesoscale activity observed near the northern OMZ boundary during winter would not be related to the
2 equatorial Kelvin wave activity, but rather to local variations of the local current system (intrinsic or
3 induced by the local wind stress). Whether or not the strong EKE found there results from the
4 instability of the coastal current system or of the EUC and the South Equatorial Current (SEC), would
5 need to be explored.

6 Regarding the southern boundary, it is interesting to note that the DO eddy flux peaks in Austral spring,
7 three months later than at the northern boundary. A possible mechanism driving the local variability
8 observed at the southern section is the generation of local baroclinic instability and vorticity input from
9 wind stress curl as observed for the California system (Kelly et al, 1998). The southern section lies
10 within the northeast rim of the Southeast Pacific Anticyclone, and the peak in the seasonal DO eddy
11 flux coincides with the reported intensity peak of the seasonal cycle of the Anticyclone, towards the
12 end of the year (Rahn et al., 2015; Ancapichún and Garcés-Vargas, 2015). In this sense, the mesoscale
13 activity in this region could be directly modulated by the winds. Dewitte et al. (2008) also report that
14 intraseasonal (internal) variability in currents can originate from the interactions between the annual
15 extra tropical Rossby wave and the mean circulation in a medium resolution oceanic regional model
16 over this region, a process also observed and documented from a high-resolution model over the North
17 Pacific (Qiu et al., 2013). The actual source of the eddy activity in this region would also deserve
18 further investigation.

19 Our study also reveals that the most prominent propagating features in DO inside the OMZ at annual
20 frequency is below ~ 300 m, where the seasonal DO flux follows approximately the theoretical WKB
21 ray paths of the annual ETRW. From that depth, the seasonal variability in physical fluxes becomes one
22 order of magnitude larger than the one of the biogeochemical fluxes (Fig. 10c). This supports the
23 observation that DO tends to behave as a passive tracer so that vertical displacements of the DO
24 isopleths mimic those of the isotherms, inducing a seasonal DO flux that resembles the energy flux
25 path of the ETRW. This mechanism adds a dimension to the understanding of the OMZ variability,
26 considering that the vertical propagation of ETRW can take place at frequencies ranging from annual
27 (Dewitte et al., 2008) to interannual (Ramos et al., 2008).

28 We now discuss some implications of our results with regards to current concerns around OMZ
29 variability at long timescales. A recent study has suggested a trend in the OMZ towards expansion and
30 intensification (Stramma et al., 2008) whose forcing mechanism remains unclear (Stramma et al.,

2010). Observations in the Pacific Ocean also suggest that the OMZ characteristics vary decadal-ly (Stramma et al., 2008; 2010). Since decadal variability can manifest as a low frequency modulation of the seasonal cycle, our study may provide guidance for investigating OMZ variability at long timescales. In particular we find that the amplitude of the seasonal cycle is nearly twice as large at the southern boundary than at the northern boundary and “coastal boundary”, which suggests a larger sensitivity of the OMZ variability at the southern boundary to the modulation of eddy activity by climate forcing. This view would preferentially link the OMZ low frequency fluctuations to mid latitudes changes in the circulation. We note however that the relative contribution of the mean DO flux and the DO eddy flux exhibits significant interannual fluctuations at the boundaries (not shown), which suggests that eddy induced DO flux may not be the only key player for understanding long term trend in the OMZ. It is interesting to note that so far, it has been difficult to reconcile the observed trend in the OMZ with the trend simulated by the current generation of coupled models (Stramma et al., 2012), which has been attributed to biases in the mean circulation and inadequate remineralization representation (Cocco et al., 2013; Cabré et al., 2015). Our results support the view that such discrepancy may partly originate from the inability of the low resolution models to account for the DO eddy flux and its modulation. Regional modeling experiments also showed that eddy activity can be modulated at ENSO and decadal timescales (Combes et al., 2015; Dewitte et al., 2012). This issue would certainly require further investigation, and could benefit from the experimentation with our coupled model platform. This is planned for future work.

Lastly, the seasonal changes in the OMZ evidenced in this work are associated with a seasonal change of the oxycline depth (and an oxycline intensity change; not shown), which can be considered a proxy for the production of greenhouse gases (CO_2 and N_2O) inside the OMZ (e.g. Paulmier et al., 2011; Kock et al., 2016). Our results suggest that the impact of the OMZ on the atmosphere through the production of climatically-active gases, such as CO_2 and N_2O , would be seasonally damped during austral winter, due to a deepening of the oxycline and a weakening of its intensity.

6 Summary and conclusions

A high resolution coupled physical/biogeochemical model experiment is used to document the seasonal variability of the OMZ off Peru. The annual harmonic of DO reveals three main regions with enhanced amplitude or specific propagation characteristics, suggesting distinct dynamical regimes: 1) The coastal

1 domain; 2) the offshore ocean below 400 m and 3) at the southern and northern boundaries. In the
2 coastal portion of the OMZ, the seasonal variability is related to the local wind forcing, and therefore
3 follows to a large extent the paradigm of upwelling triggered productivity, followed by
4 remineralization. It is shown in particular that the DO peaks in Austral winter which is associated with
5 horizontal and vertical diffusion induced by both the increase in baroclinic instability and intraseasonal
6 wind activity. This is counter intuitive with regards to the seasonality of the alongshore upwelling
7 favorable winds also peaking in Austral winter, which would tend to favor the intrusion of
8 deoxygenated waters from the open ocean OMZ to the shelf. Instead, the coastal domain can be viewed
9 as a source of DO in Austral winter for the OMZ through offshore transport. The latter is induced by
10 eddies that are triggered by the instabilities of the PUC. In the model, the offshore DO eddy flux has a
11 marked seasonal cycle that is in phase with the seasonal cycle of the DO along the coast, implying that
12 the coastal domain, viewed here as the eastern boundary of the OMZ, is a source of seasonal variability
13 for the OMZ. This appears to operate effectively in the upper 300 m. Below that depth, the DO eddy
14 flux is much reduced due to both a much weaker eddy activity, and very low DO concentration. On the
15 other hand, a mean seasonal DO flux is observed and exhibits propagating features reminiscent of the
16 vertical propagation of energy associated with the annual extra tropical Rossby wave.
17 In the upper 300 m, the OMZ seasonal variability is also associated with the DO eddy flux at the OMZ
18 meridional boundaries where it is the most intense. We find that the seasonal cycle in DO eddy flux
19 peaks in Austral winter at the northern boundary, while it peaks a season later at the southern boundary.
20 Additionally, the amplitude of the seasonal cycle in DO eddy flux is larger at the southern boundary
21 than at the northern boundary. The schematic of Figure 20 summarizes the main processes documented
22 in this paper to explain the seasonality of the OMZ.

23

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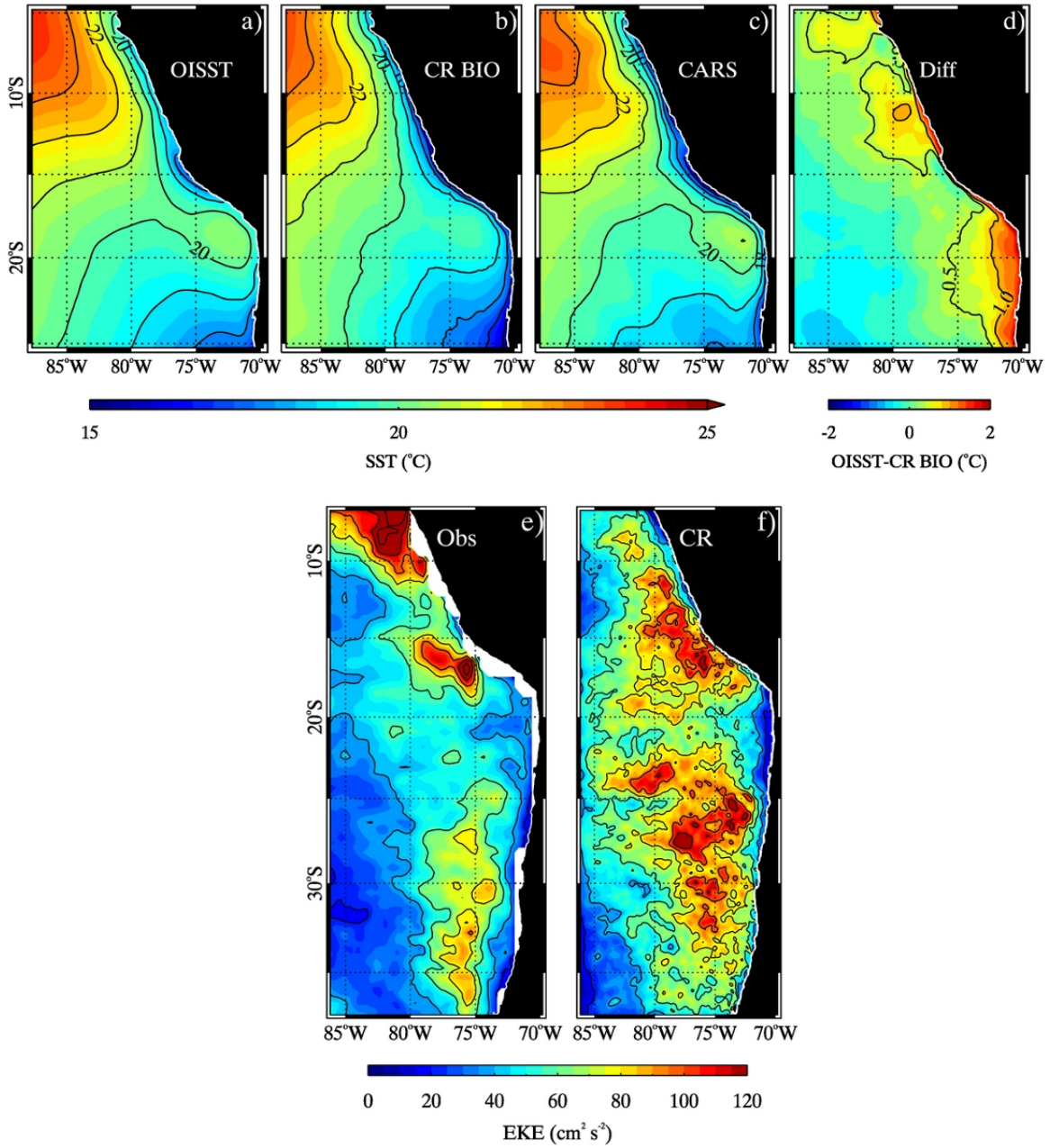
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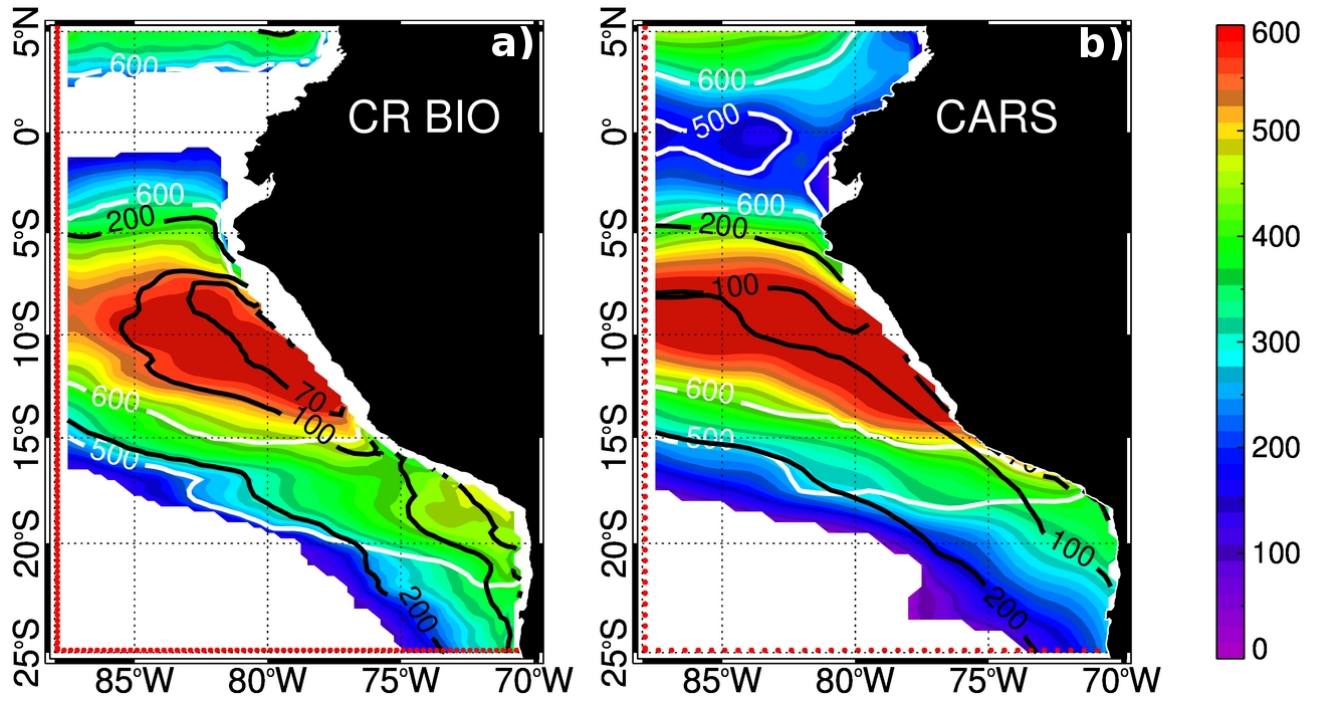
1 **Table 1.** Austral summer (DJF mean) and winter (JJA mean) seasonal anomalies of the DO budget,
2 averaged over the core of the Peru Under Current at 12°S (as depicted by the red contour in Figure 12).
3 The values for the seasonal cycle and the reconstructed first EOF mode (Figures 12 and 13) are
4 presented along with the difference Climatology-EOF. All values are in $10^{-6}\mu\text{M s}^{-1}$. Mixing here
5 consists in the summe-up contribution of horizontal diffusion and ($K_h \nabla^2 O_2$) and vertical diffusivity ($\frac{\partial}{\partial z} \left(K_z \frac{\partial O_2}{\partial z} \right)$).

7

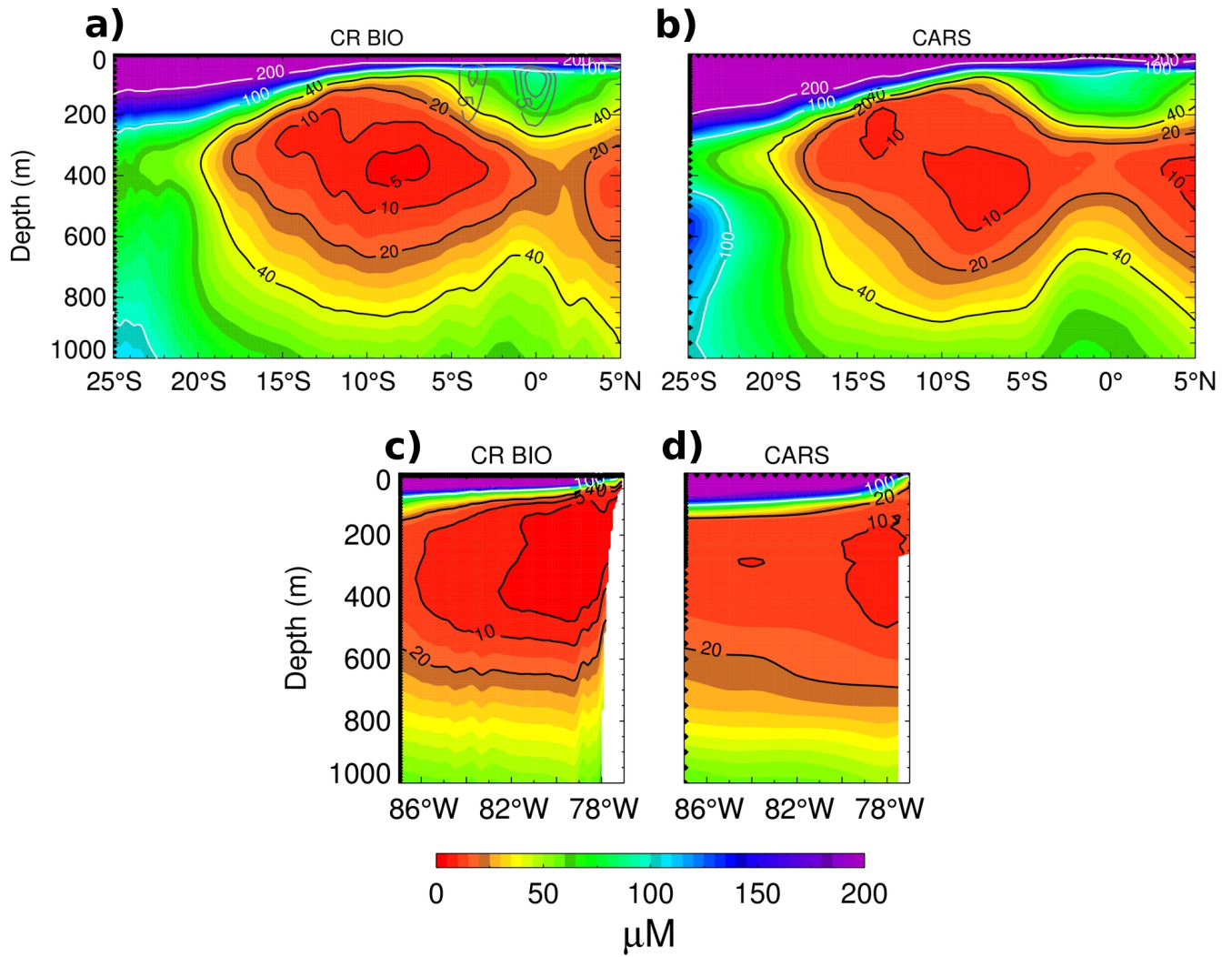
| | Climatology | | EOF | | Difference | |
|----------------------------|-------------|--------|--------|--------|------------|--------|
| | Summer | Winter | Summer | Winter | Summer | Winter |
| dO₂/dt | 1.10 | -2.74 | 1.30 | -2.67 | -0.2 | -0.07 |
| Adv | 0.61 | -9.38 | 0.85 | -9.30 | -0.24 | -0.08 |
| Mixing | -0.42 | 7.99 | -0.35 | 7.99 | -0.07 | 0.0 |
| Biogeochemical Flux | 0.91 | -1.35 | 1.00 | -1.35 | -0.09 | 0.0 |



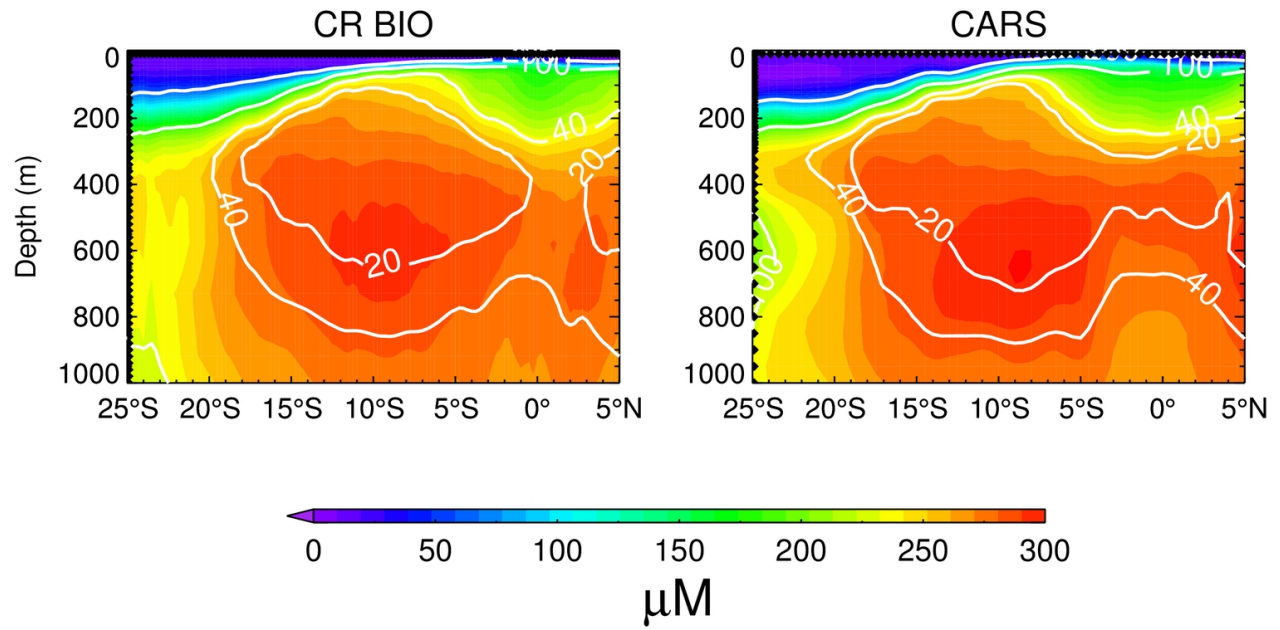
2 **Figure 1.** Mean sea surface temperature (SST) between 2000 and 2008 for (a) OISST product
3 (0.25°x0.25°), (b) the simulation (1°/12) and (c) CARS dataset (0.5°x0.5°). (d) Difference between the
4 OISST product and the simulation. Mean Eddy Kinetic Energy (EKE) between 1993 and 2008, for (e)
5 TOPEX/Poseidon Jason 1-2 merged product (0.25°x0.25°), and (f) Simulation (1°/12). EKE was
6 derived from the interannual anomalies of the geostrophic velocity field.



2 **Figure 2.** Mean Oxygen Minimum Zone core thickness (color scale in meters) for (a) the simulation
3 and (b) CARS. Depth of the lower (white) and upper (black) limits of the OMZ core are also depicted.
4 The OMZ core is defined as $[DO] < 20 \mu\text{M}$. The red dots denote the horizontal resolution of the DO
5 field.



2 **Figure 3.** Mean oxygen concentration for a meridional section at 85°W (a and b) and a cross shore
3 section at 12°S (c and d), for both the simulation and CARS. Gray contours in (a) show mean zonal
4 speed of 5,10 and 15 cm s^{-1} respectively. The black dots denote the horizontal and vertical resolution of
5 the DO field.

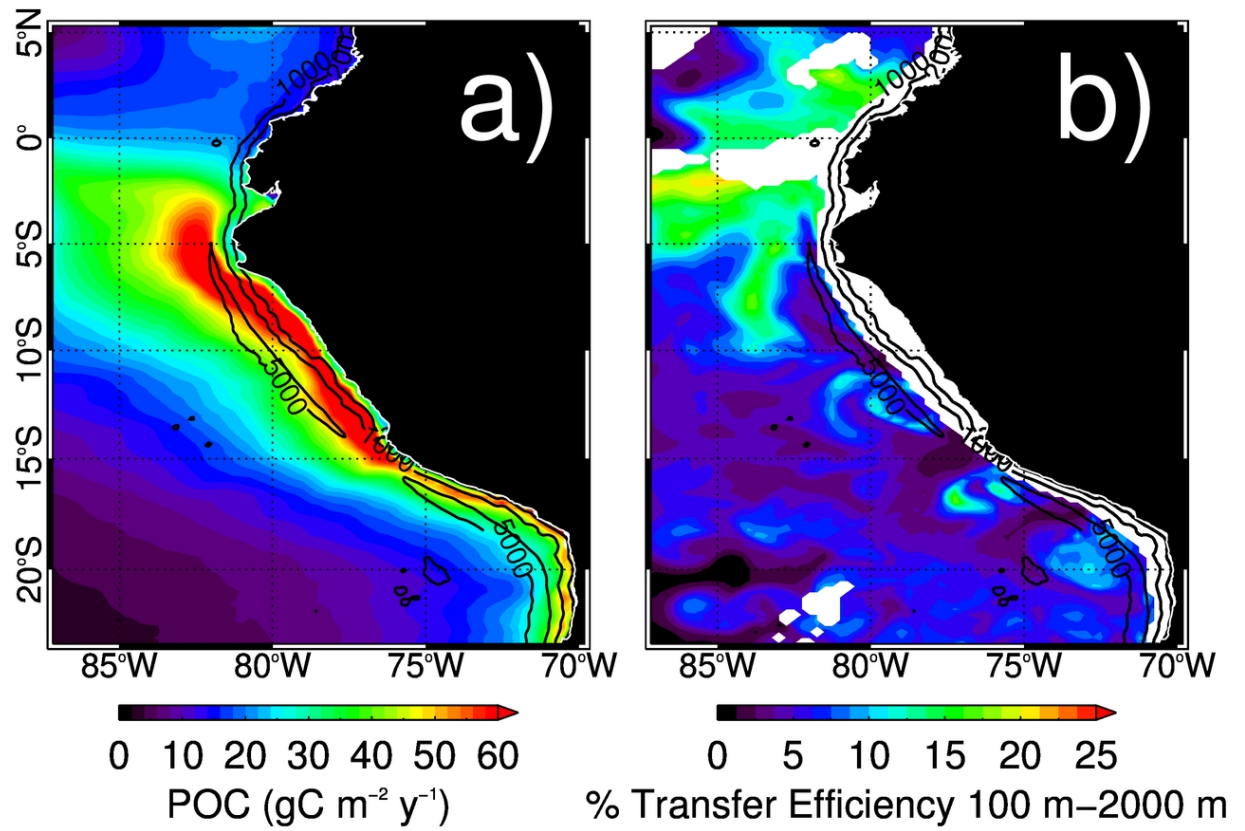


2

3 **Figure 4.** Mean Apparent Oxygen Utilization (AOU) at 85°W for both CR BIO and CARS. White
 4 contours denote the mean oxygen concentration isopleths (in μM). The black dots denote the horizontal
 5 and vertical resolution of the DO field.

6

7



2 **Figure 5.** (a) Particulate Organic Carbon (POC) flux at 100 m and (b) POC transfer efficiency between
 3 100 m and 2000 m (POC flux at 2000 m divided by POC flux at 100 m), computed from the
 4 simulation. Integrated carbon flux at the depth of 100 m: $0.8 \text{ Pg C year}^{-1}$. Black contours correspond to
 5 the 200, 1000 and 5000 meters isobaths.

6

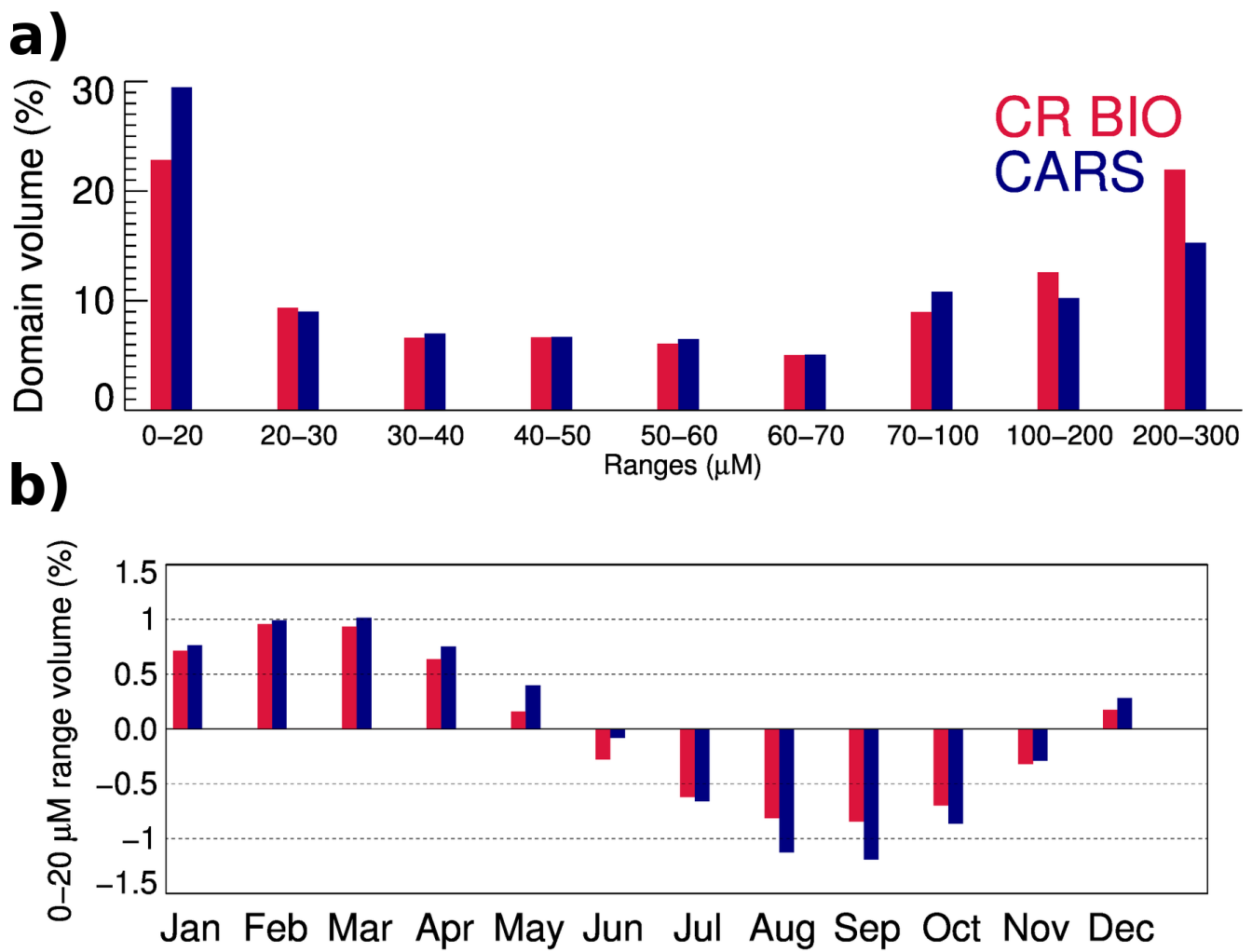


Figure 6. (a) Domain volume distribution (25°S - 5°N , 88°W - 70°W) as a function of the oxygen concentration, and (b) annual cycle, relative to the mean, of the volume distribution inside the OMZ core (DO value range corresponding to 0 - $20 \mu\text{mol L}^{-1}$), for both CARS and the simulation.

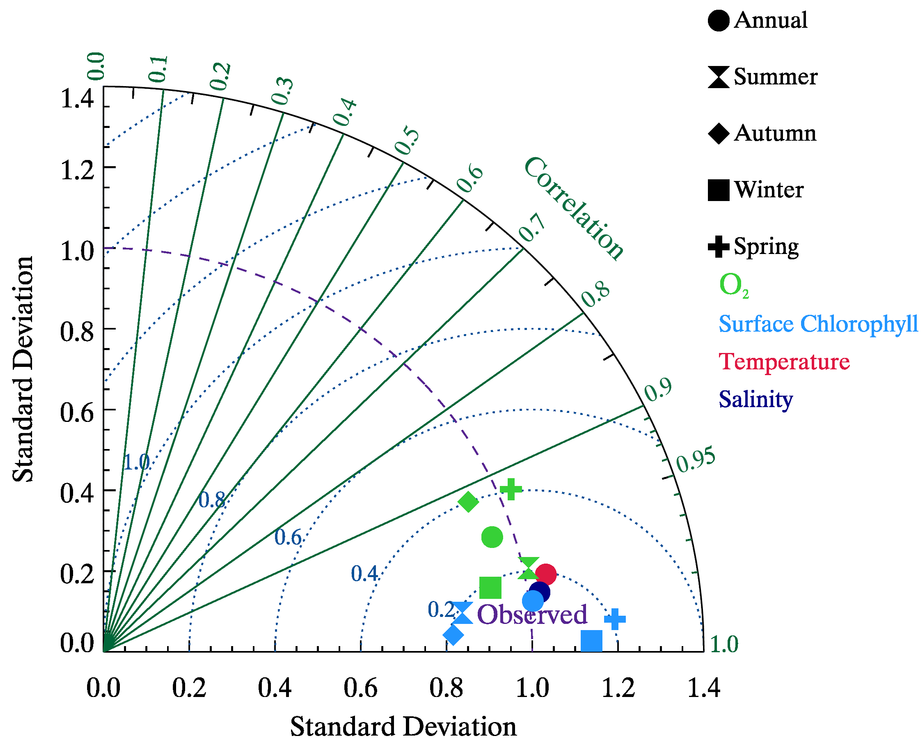


Figure 7. Taylor diagram of the seasonal mean (hourglass, diamond, square and cross) and annual mean (circle) pattern of DO and Surface Chlorophyll (25°S-5°N, 88°W-70°W). Only annual mean pattern comparisons are shown for temperature and salinity (same spatial domain). DO, temperature and salinity were vertically averaged between 100 and 600m depth (focus on the OMZ core). Only the surface chlorophyll values within 250 km next to the coast were considered. The comparisons are made between the simulation and CARS (for DO, temperature and salinity) and SeaWiFS (for surface chlorophyll). Ordinate and abscissa axes represent the standard deviation normalized by the observations standard deviation. Blue dotted radial lines indicate the RMS difference between the observations and the simulation.

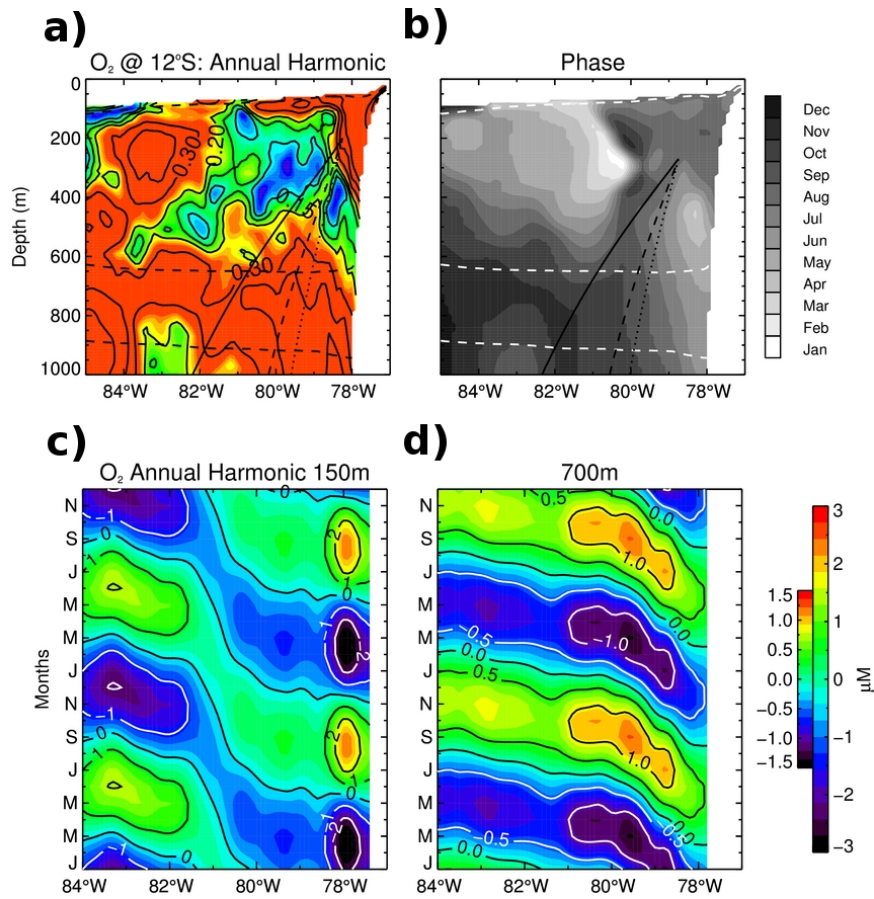
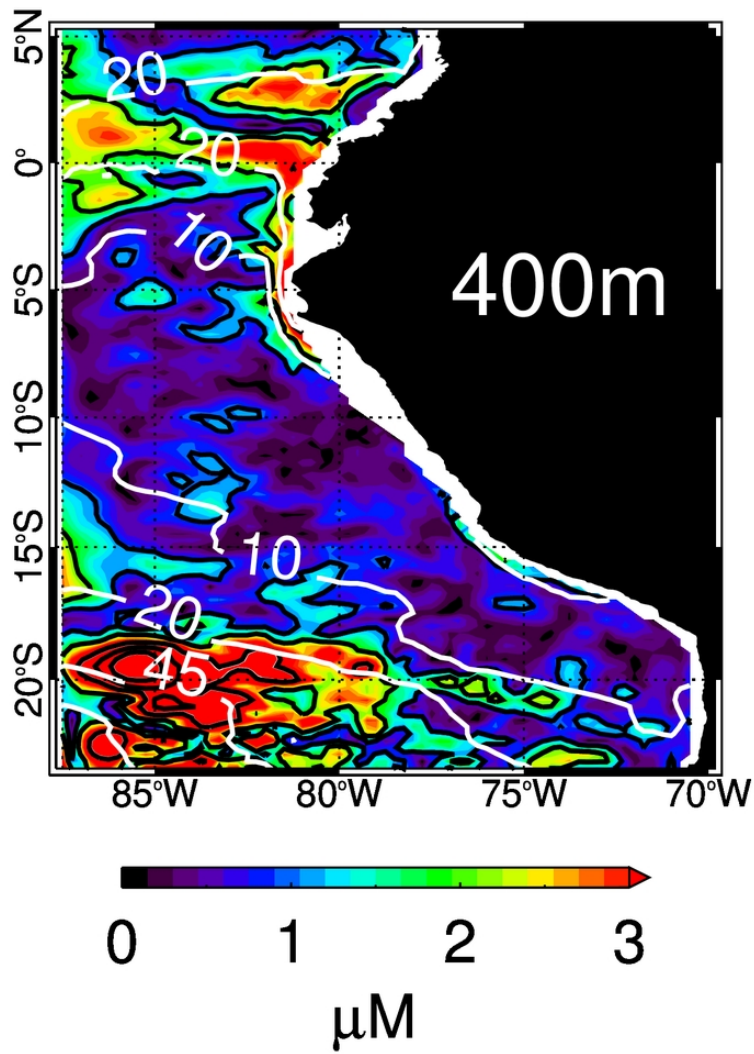
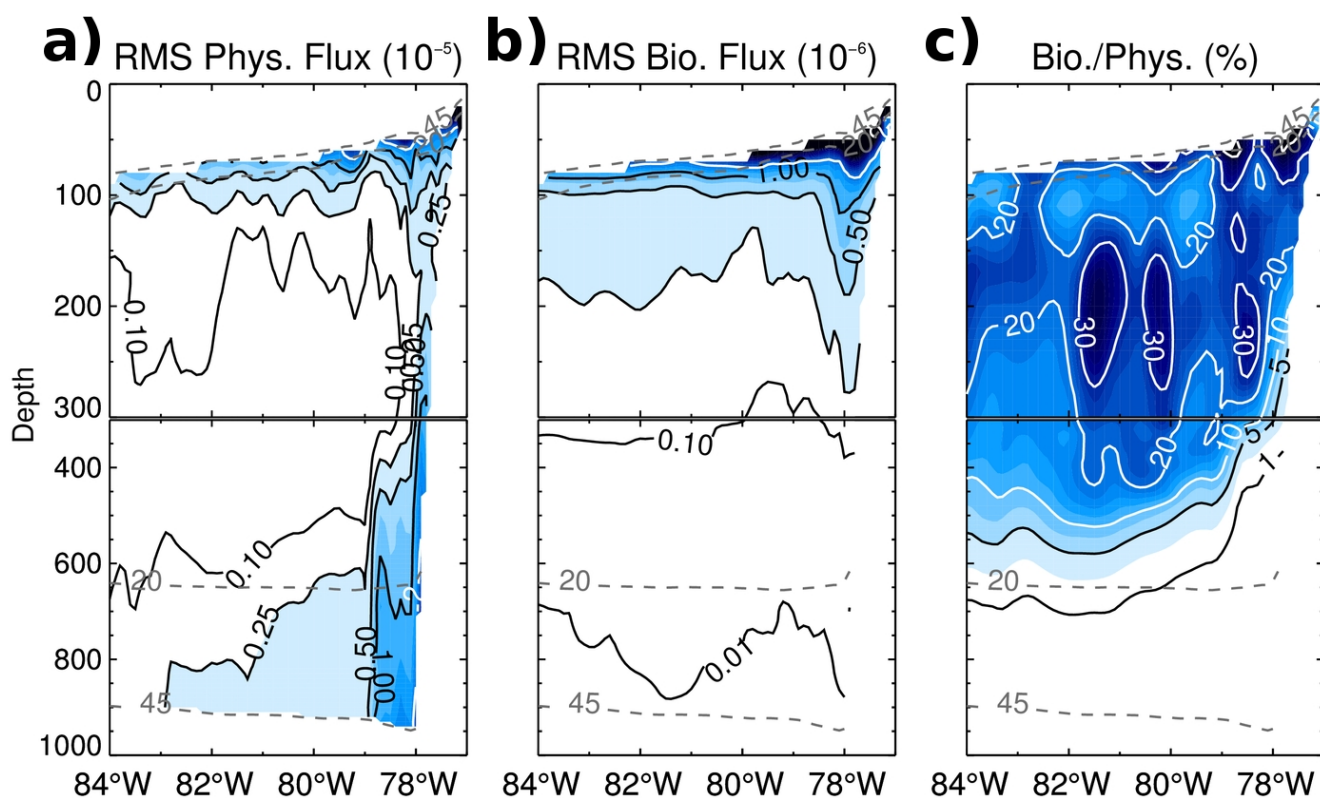


Figure 8. (a) Amplitude and (b) phase of the annual maximum (in months) of the annual harmonic of the normalized DO concentration at 12°S. The slanted vertical lines indicate the theoretical WKB ray paths at a frequency of $\omega=2\pi\cdot\text{year}^{-1}$, for different value of phase speed. The theoretical trajectories were computed using the phase speed of the first (full), second (dashed) and third (dotted) baroclinic modes of a long Rossby wave. Dashed contours in (a) and (b) depict the 45 and 20 μM mean DO values. Land and the region outside the 45 μM mean DO isopleth are masked in white. (c) Annual harmonic of the DO concentration at 12°S, at 150 m and (d) 700 m depth. Small color scale corresponds to 700 m and the large color scale denotes the levels used in (c).

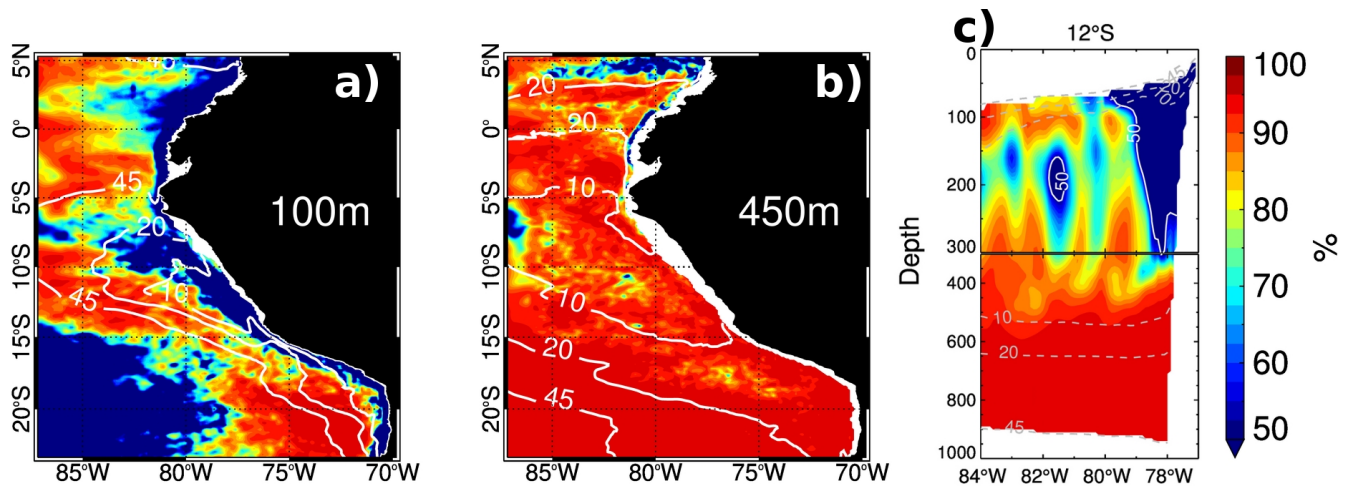


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 2
 3 **Figure 9.** Annual DO harmonic amplitude at 400 meters depth. White contours denote the 10, 20 and
 4 45 μM mean oxygen isolines. Black contours denote the 1, 2, 4, 6 and 8 μM levels.
 5



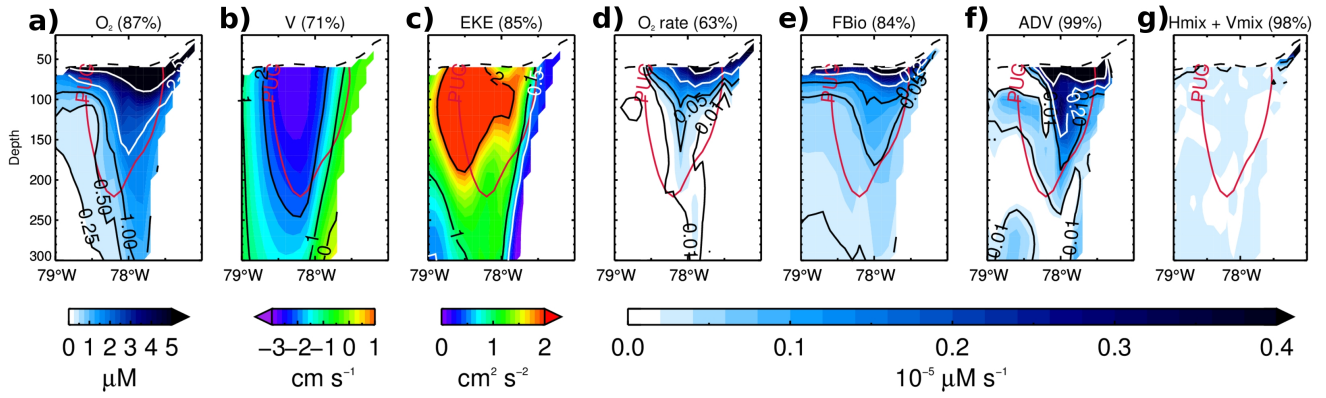
2 **Figure 10.** Root mean square of the seasonal cycle of (a) Physical and (b) Biogeochemical oxygen
3 fluxes (in $10^{-5} \mu\text{M s}^{-1}$ and $10^{-6} \mu\text{M s}^{-1}$, respectively) for CR BIO at 12°S. c) Ratio between the RMS of
4 the biogeochemical fluxes and the physical fluxes, expressed as percentage. Dashed contours depict the
5 45 and 20 μM mean oxygen values. Note the vertical scale change at 300m depth. Land and the region
6 outside the 45 μM mean DO isopleth are masked in white.

7



2 **Figure 11.** Percentage of the seasonal DO rate variance explained by the physical fluxes, at (a) 100 and
3 (b) 450 meters depth, and along a cross shore section at 12°S. Solid white lines (a,b) and dashed gray
4 lines (c) denote the 10, 20 and 45 μM mean DO isopleths. Land and the region outside the 45 μM mean
5 DO isopleth are masked in white in (c).

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1 **Figure 12.** First EOF mode pattern of (a) DO, (b) alongshore currents component, (c) Eddy Kinetic
 2 Energy, (d) oxygen rate, (e) biogeochemical flux, (f) advective terms (sum of horizontal and vertical
 3 components) and (g) mixing terms (sum of horizontal and vertical components). Percentage of
 4 explained variance by each EOF mode pattern is indicated in parentheses on top of each panel. The red
 5 contour denotes the mean position of the Peru Under Current core, defined here as alongshore
 6 southward current exceeding 4 cm s^{-1} . The black dashed contour denotes the mean DO $45 \mu\text{M}$ isopleth.
 7 Land and the region outside the $45 \mu\text{M}$ mean DO isopleth are masked in white. The EOF mode patterns
 8 were multiplied by the RMS of the PC timeseries. Multiplying the EOF pattern by the PC timeseries
 9 plotted in Figure 13 yields the contribution of the first EOF mode to the original field, in
 10 dimensionalized units (i.e. $\mu\text{M s}^{-1}$ for the tendency terms).
 11

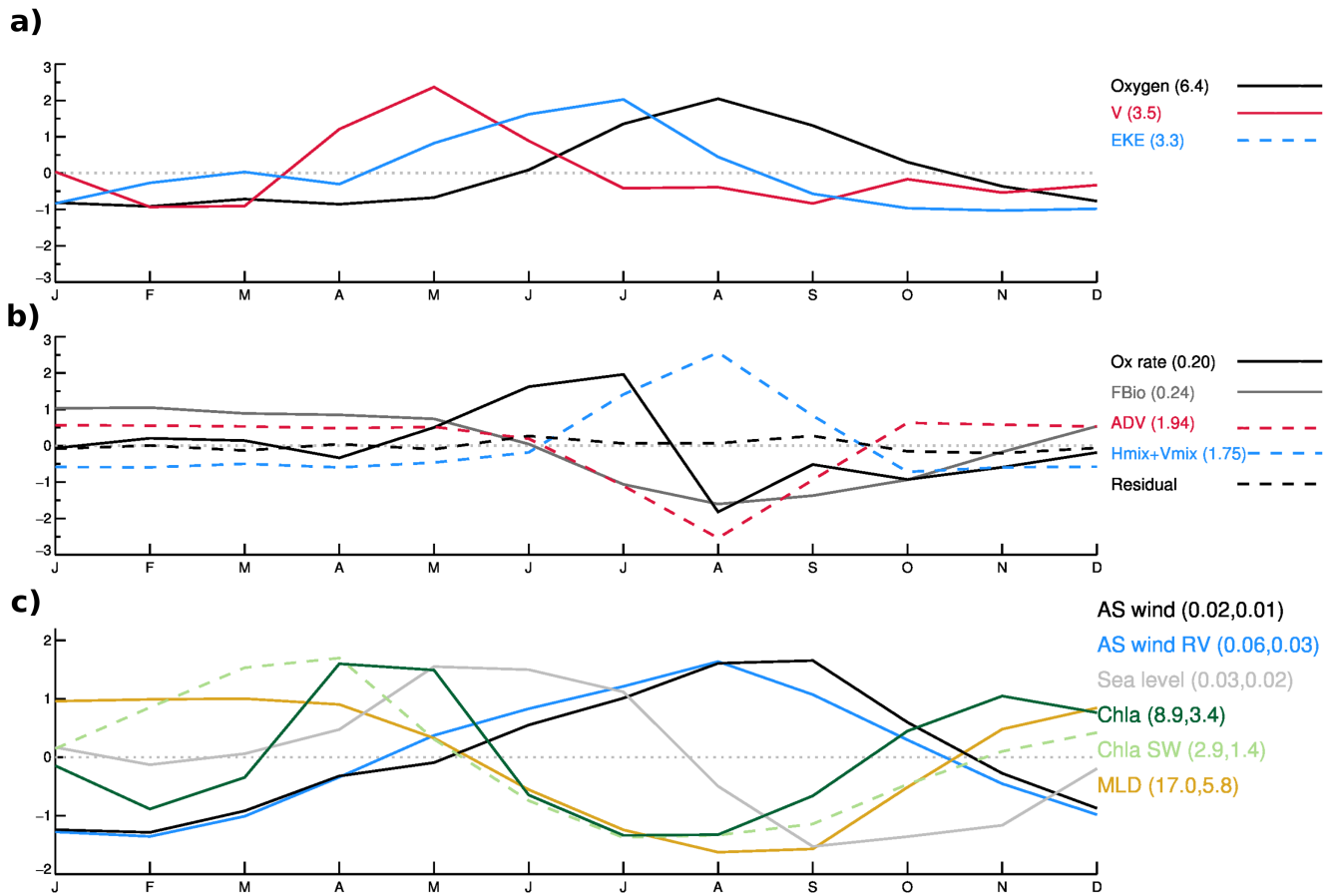


Figure 13. (a, b) Non dimensional principal components (PC) associated with the EOF patterns in Figure 12. Multiplying the principal component by the associated EOF pattern (from Fig. 12) yields a first EOF-mode reconstruction of the original field. RMS values of the principal components are indicated in parenthesis (corresponding units as in Fig. 12). The residual corresponds to the difference between the rate of DO change and the sum of all the terms of the rhs of Eq. 1 in terms of the normalized PC timeseries. The weak residual indicates that the seasonal DO budget can be interpreted from the EOF decomposition. The EOF decomposition was performed over the climatological (mean seasonal cycle) fields. (c) Normalized seasonal cycle of: coastal alongshore wind (AS wind) and coastal alongshore wind Running Variance (variance over a 30 day running window) at 12°S, sea level at the coast at 12°S, surface chlorophyll-a from CR BIO (Chla) and from SeaWiFS (Chla SW) averaged over a coastal band of 2° width at 12°S, and Mixed Layer Depth at the coast (MLD) at 12°S. Mean and RMS used to normalize each time series, are indicated in parenthesis. Original seasonal cycle is found by multiplying the normalized series by its RMS and then adding the mean. Original units are N m^{-2} , m, mg m^{-3} , and m respectively.

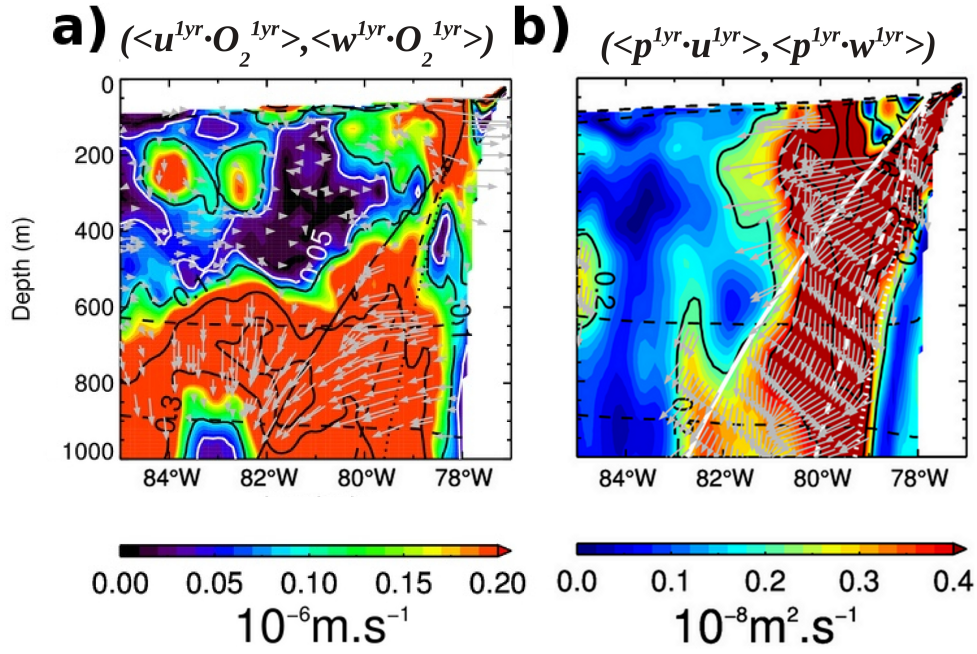


Figure 14. (a) Norm of the annual DO flux vector (i.e. $\sqrt{(\langle u^{1yr} \cdot O_2^{1yr} \rangle)^2 + (\langle w^{1yr} \cdot O_2^{1yr} \rangle)^2}$) for a cross shore section at 12°S. Arrows indicate the vector direction (i.e. $[\langle u^{1yr} \cdot O_2^{1yr} \rangle, \langle w^{1yr} \cdot O_2^{1yr} \rangle]$). The DO signal was normalized by its Root Mean Square value before computing the annual harmonic, in order to emphasize the flux patterns where DO concentration is very low. (b) Norm of the annual energy flux vector (i.e. $\sqrt{(\langle p^{1yr} \cdot u^{1yr} \rangle)^2 + (\langle p^{1yr} \cdot w^{1yr} \rangle)^2}$). Arrows inside the 0.2 contour indicate the vector direction (i.e. $[\langle p^{1yr} \cdot u^{1yr} \rangle, \langle p^{1yr} \cdot w^{1yr} \rangle]$). A range of theoretical WKB trajectories (1 year period) originating from near the coast at the surface are drawn for phase speed values of a first (full), second (dashed) and third (dotted) baroclinic modes. The range of phase speed values (modes 1-3) are obtained from a vertical mode decomposition of the mean model stratification. Dashed black contours indicate the 45 and 20 μM mean DO isopleths. Land and the region outside the 45 μM mean DO isopleth are masked in white.

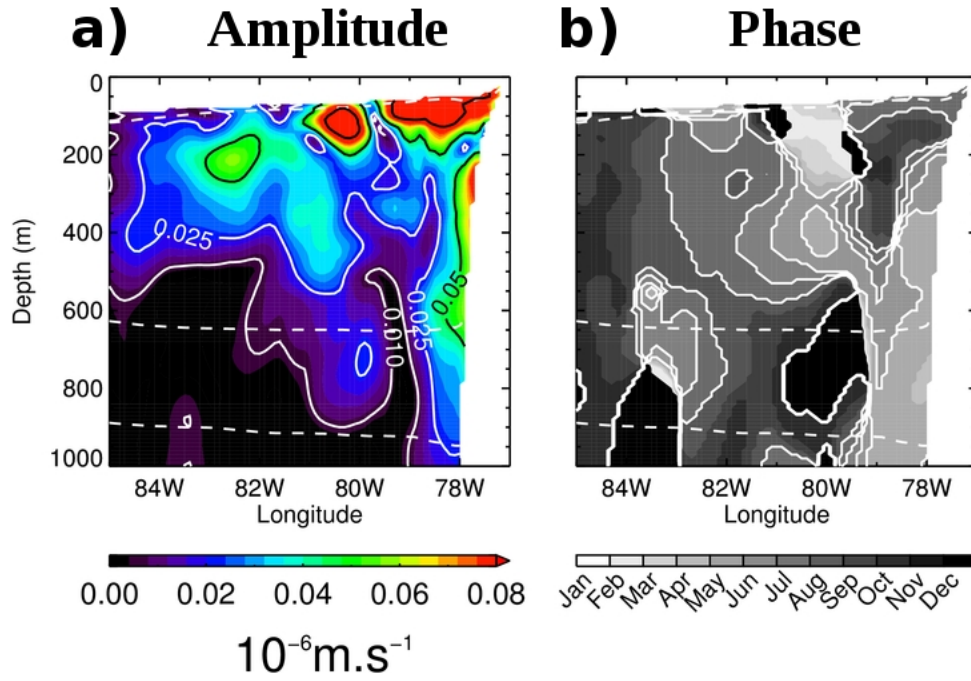
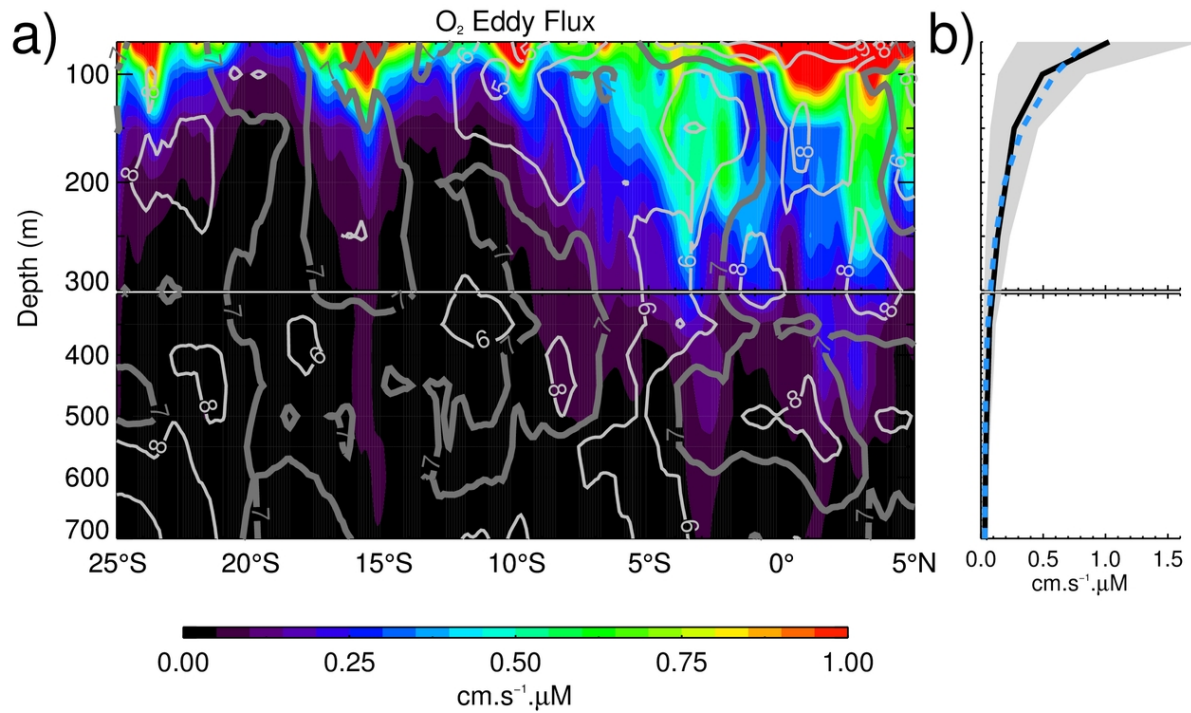
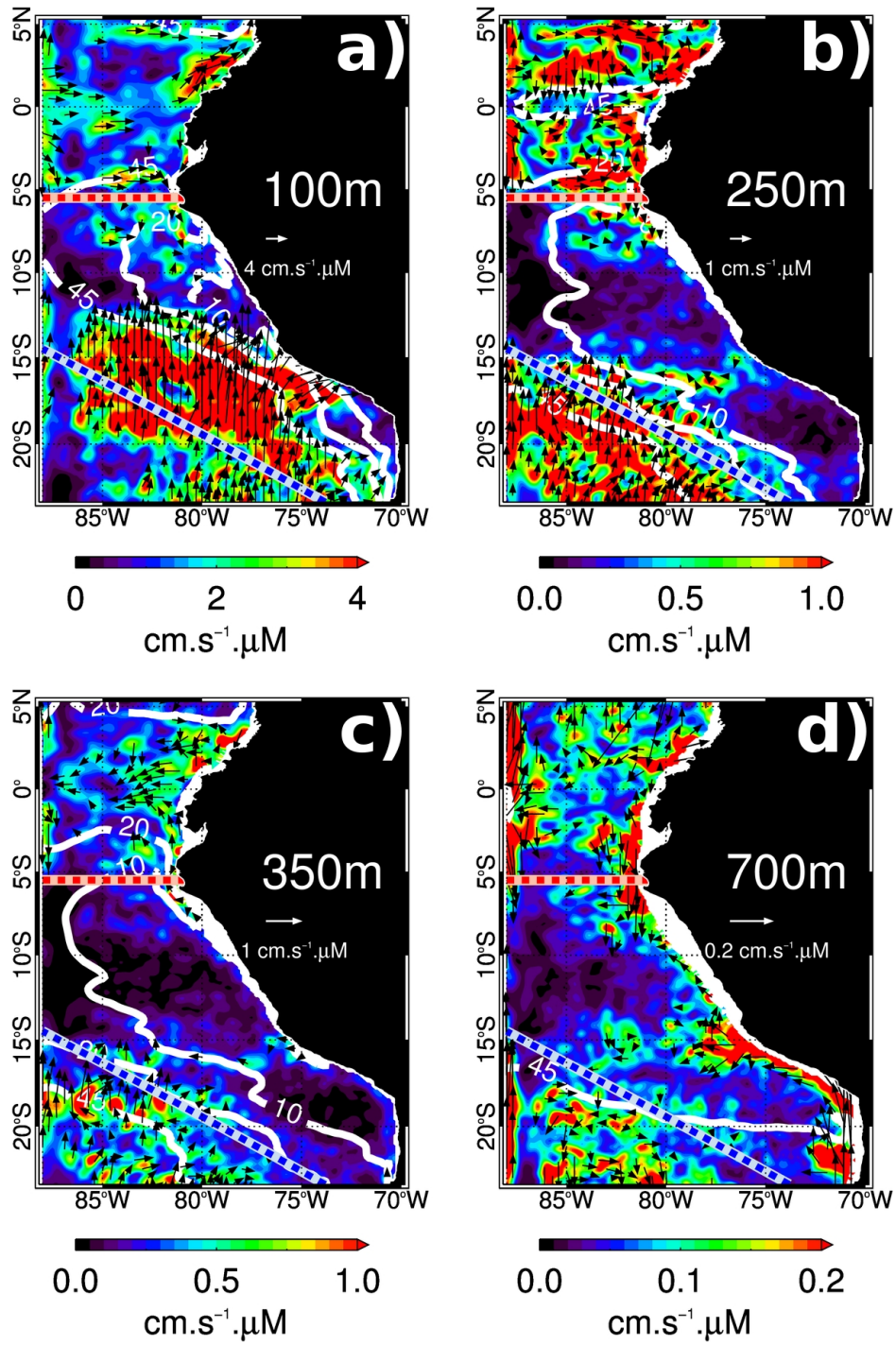


Figure 15. Zonal section of the annual harmonic of the module of the seasonal DO eddy flux vector $\left(\overline{\langle u' \cdot O_2' \rangle}, \overline{\langle w' \cdot O_2' \rangle}\right)$ at 12°S . (a) Amplitude of the harmonic and (b) phase of the annual maximum (in months). Dashed white contours indicate the 45 and 20 μM mean DO isopleths. DO was normalized by its RMS prior to carrying out analysis. Land and the region outside the 45 μM mean DO isopleth are masked in white.

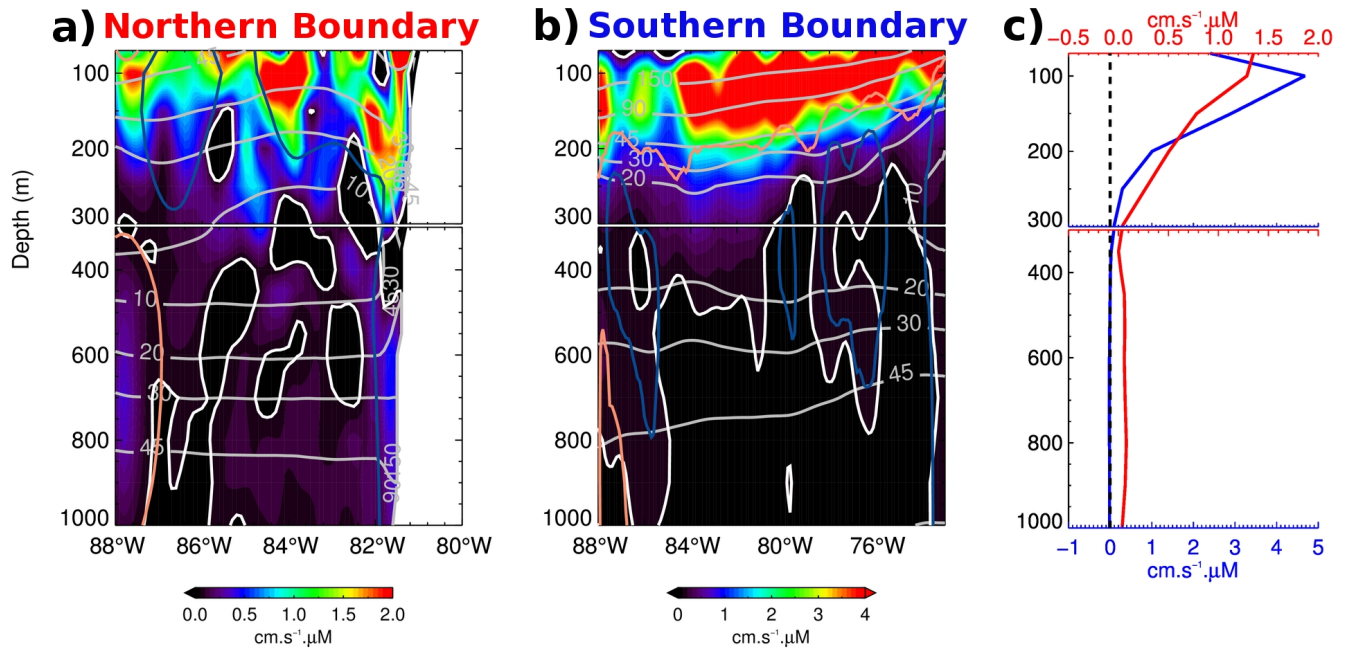


2 **Figure 16.** (a) Amplitude (color shading) and phase (months, gray contours) of the annual harmonic of
 3 the climatological DO eddy flux along the coast. The climatology of DO eddy flux was averaged over a
 4 coastal fringe of 2° width starting from 1° from the coast. (b) Meridional average vertical profile (black
 5 line), \pm RMS (gray shading). An exponential model fitted to the average vertical profile (dashed blue
 6 line) yields a vertical decay scale of ~ 90 m.



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3 **Figure 17.** Module of the mean DO eddy flux vector $(\langle u' \cdot O_2' \rangle, \langle v' \cdot O_2' \rangle)$ at (a) 100 m, (b) 250 m, (c)
4 350 m and (d) 700 m depth. Arrows -displayed only for values above the central value in each colorbar-
5 denote the vector direction and strength. White contours correspond to the 45, 20 and 10 μM mean DO
6 values. Red and blue lines denote the position of vertical sections.



1 **Figure 18.** (a) Mean DO eddy flux normal to the section denoted by the red line in Fig. 17. (b) Mean
2 DO eddy flux normal to the section denoted by the blue line in Figure 17. (c) Horizontal mean of (a)
3 and (b) (red and blue lines, respectively). Gray contours denote mean DO concentrations, and light
4 red/blue contours correspond to positive/negative values of mean currents normal to the section (1.0/-
5 1.0 cm s⁻¹ in (a) and 0.4/-0.2 cm s⁻¹ in (b)). White contour denotes the 0 value. The sign convention was
6 chosen so that a positive horizontal flux indicates transport towards the interior of the OMZ.

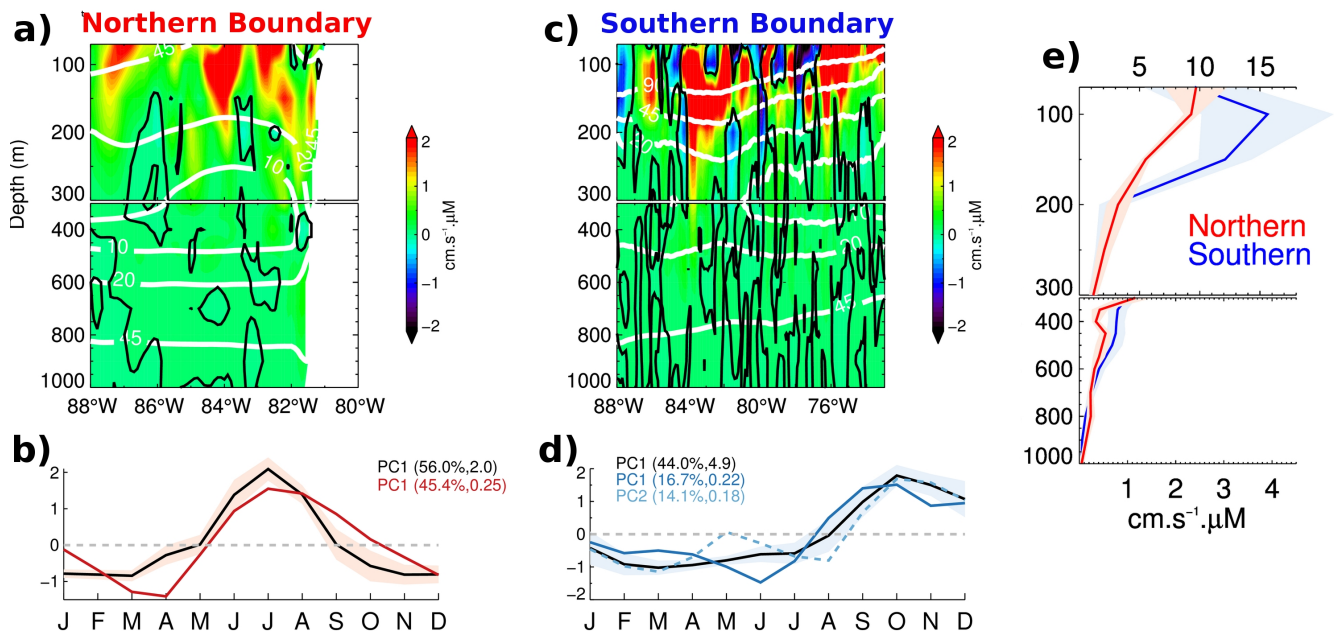
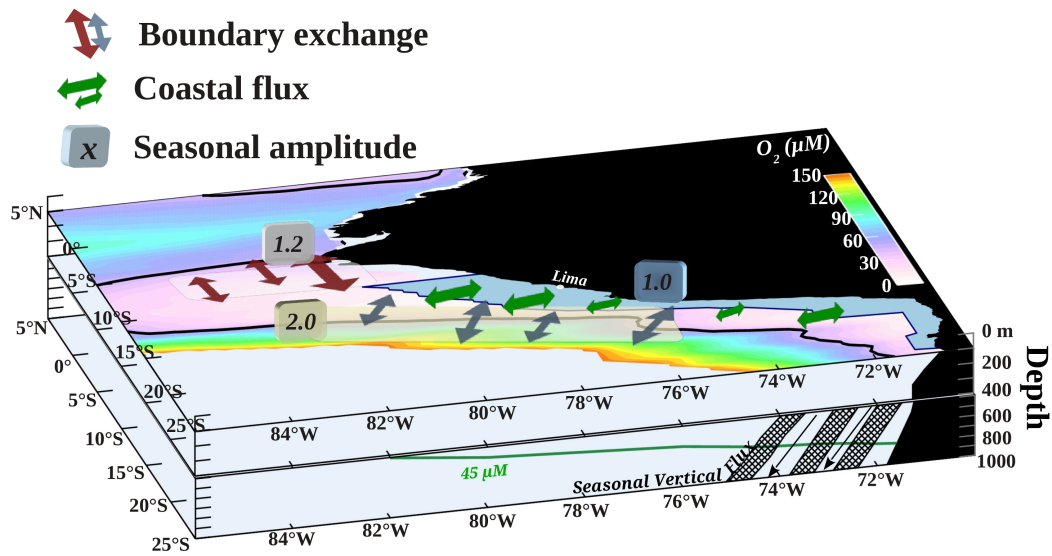


Figure 19. (a) First EOF mode of the seasonal cycle of the DO eddy flux normal to the section depicted in Fig. 17 by the dashed red line (Northern boundary). (b) Principal component (PC) timeseries associated with the first EOF mode (black line). The red line in (b) corresponds to the PC timeseries associated with the first EOF mode of the seasonal cycle of the 30-day running variance of intraseasonal currents normal to the section. (c) First EOF mode of the seasonal cycle of the DO eddy flux normal to the oblique section depicted in Fig. 17 by the dashed blue line (Southern boundary). (d) PC timeseries associated with the first EOF mode (black line). The blue curves (full and dashed lines) in (d) corresponds to the PC timeseries associated with the first and second EOF modes of the seasonal cycle of the 30-day running variance of the intraseasonal currents normal to the section (computed as in (b)). Percentage of explained variance and RMS value are indicated in parentheses in the panels (b) and (d) (in $\text{cm.s}^{-1}.\mu\text{M}$ and cm.s^{-1} , for DO eddy flux and currents respectively). White contours in (a) and (c) denote mean DO concentration values, in μM . (e) RMS of the spatial patterns (a) and (c), computed along the horizontal direction. Note the scale leap at 300 m. Red/blue shading in (b), (d) and (e) represents an estimate of the error associated with slight changes in the location of the boundaries, that is when the EOF is performed over a section that is located at a distance from the original section (cf. Figure 17) compromised between $\pm 120\text{km}$ (see text). The error corresponds to the standard deviation among 12 PC timeseries (for (b) and (d)) and EOF patterns (for (e)).

1



3 **Figure 20.** Schematic of the main processes driving the seasonal variability in the SEP OMZ: The DO
 4 eddy flux through the northern-southern boundaries and the DO flux that takes place at the coastal
 5 boundary of the OMZ. The coastal band limits are defined by the light blue shading adjacent to the
 6 coast. A scale of the seasonal amplitude of the eddy driven DO flux at each OMZ boundary is indicated
 7 (units in $\text{cm s}^{-1} \mu\text{M}$). The mean DO concentration (color shading) and the position of the $45 \mu\text{M}$
 8 isopleth (thick black contour) at 100 m depth are also represented. The vertical/offshore DO flux
 9 induced by the propagation of the annual ETRW across the $45 \mu\text{M}$ isopleth at 25°S is represented in the
 10 bottom panel.

11