**Reply to comments of BG Discussion**
**Reviewer #1**

We thank the reviewer for his/her constructive comments. Due to a comment from Reviewer#2 arguing that the results from the regional simulations forced by ESM oxygen trends (labelled RCM' in the previous version of the manuscript) should be presented before the simulations forced by climatological oxygen boundary conditions, the paper has been thoroughly reorganized. The figures and tables presenting the biogeochemical trends have been modified. This does not change the general message of our paper, but large portions of the text have been modified.

**Comments:**
*C1) I think that a more quantitative and critical evaluation of the regional model perfor
mance after the spinup is missing. A clear discussion of how well the "baseline", or
present day is being represented in the regional simulation is necessary, in particular
for sensitive parameters such as thermocline and oxycline depth.*

R: We have added new figures showing cross-shore sections of mean state and bias of temperature and dissolved oxygen (DO) for the three regional simulations (period 2006-2015). The model is compared to CARS climatology (interpolated on the model's grid). These figures, included in the supplementary material, are briefly described in the text (lines 254-256 for temperature and lines 366-371 for DO)

*C2) Some measure of uncertainty in the percentage of change by the end of the century
for each variable is needed. This percentage values form most of the base of the whole
discussion and are calculated on the basis of linear trends. A quantitative estimate of
how well a linear model fits the timeseries examined, or perhaps an estimate of the
actual temporal variability around the trend could make the interpretation of the long-
term changes more robust.*

R: We have estimated the trend uncertainty based on a bootstrap method. We construct 10 000 synthetic time series by randomly removing data points in the annual series. We converted the trend uncertainty into a percentage uncertainty, now reported in Tables 3,4,5. We also have computed the $R^2$ from the least square estimation in the tables. Most of the trends are significant at the 10% level. The significant trends are reported in bold font in the tables. We now explain how the uncertainty is computed in the methodology section 2.8.

**Specific comments**

*C: Section 2.3: To choose the global model for regional downscaling, the authors use
averaged vertical profiles of a meridional section and compare the bias with an
observation-based gridded product (World Ocean Atlas 2009). It is not clear to me
if the model was sampled to represent the time period of WOA09, which years is the
WOA09 climatology representing?*

R: The temperature and salinity from the CMIP5 historical simulations were averaged between 1950 and 2005 to compare with the WOA2009 climatology, which includes observations mainly collected between 1950s and 2009. The nutrient and oxygen profiles from the CMIP5 historical simulations were averaged between 1980 and 2005. They are compared to the WOA2009 which includes biogeochemical observations mostly in recent decades (i.e. after 1980) in the equatorial pacific.

C: *Some ideas in section 2.3 need to be more quantitative. E.g. phrases like "too low", "realistic enough" are somewhat subjective. The authors mention that the temperature and salinity biases are weak, but what does weak mean? How do we compare the weak salinity and temperature biases to the biogeochemical biases?*

R: The reviewer is right. First we added temperature and salinity profiles in Fig.1 to allow for visual comparison between the ESMs. Second, we computed a normalized bias, defined as:
  $NB(z)= |X_{model}(z)-X_{obs}(z)|/X_{obs}(z) \times 100$  for each variable $X(=T,S$, nutrients, $O_2$). This allows to quantify the amplitude of the normalized bias between the ESMs and compare the normalized bias of different variables. The depth-averaged values of the normalized bias are reported in Table 1. We find that the normalized bias for temperature and salinity are weaker than those for nutrients and oxygen. We corrected the text to avoid vague terms and be more quantitative (see section 2.3).

C: *Lines 293-298: The authors describe a shoaling of the mixed layer depth in all simula tions and the agreement or disagreement with a gridded product. I find this confusing since this idea comes after they mention that the "thickness of the surface layer more than doubles" (line 287).*

R: By surface layer we did not mean the mixed layer in this paragraph, but the surface layer with waters warmer than 20°C. We defined D20 in lines 303-304 and rephrased the sentence (line 312)

C: *Also, they note that the mixed layer is calculated differently in the model and in the gridded product. How is the mixed layer calculated in the model then?*

R: The model surface boundary layer is computed from the value a critical Richardson number computed using the KPP formulation, whereas the observed mixed layer depth was computed from individual temperature profiles. However previous modelling work show that the surface boundary layer thickness is very close to the model mixed layer (Liu and Fox-Kemper, 2017). We added this information and this reference (lines 319-321)

C: *Line 279: The term thermocline depth needs to be clearly defined as the isotherm of 20C, as is indicated in figure 6 and as was done with the oxycline (line 341) or nitracline.*

R: We agree with the reviewer that this is unclear. As noticed by another reviewer, D20 and thermocline may be located at different depths. We now no longer refer to the thermocline, simply D20.

C: *Lines 333-339: In the text, they mention that figure 10 shows the evolution of nearshore DO concentration, but the trends in this figure are calculated over a region that differs from the coastal box used through the analysis. There is no mention or explanation of why these trends were calculated in an oceanic box that differs in size and distance from the coast than the rest of the analysis.*

R: We agree with the reviewer that some clarification is needed here. In this section we compare the nearshore DO content in the RCMs and ESMs between 100 and 200m depth. However, the coarse resolution and topography of the ESM implies that few grid points are present in this depth range in the 100 km band (in particular in GFDL). We believe that the comparison is thus more accurate in the 150km-300 km offshore band. In the same way as the oxycline is quite deep in R-GFDL we had to extend the width of the box to 200 km. This was not the case for the nitracline (depth of nitrate isosurface 21 umol) which was shallower and could thus be computed in the 0-100 km band (Fig.11d).

We added explanation in lines 366-367.

C: *Line 382: Positive trends in surface biomass were found in R-GFDL and R-IPSL, but the nitracline only deepens in R-IPSL, in R-GFDL the nitracline gets shallower. The increase in surface biomass would be surprising only in R-IPSL.*

R: The reviewer is right. We corrected the text accordingly.

**Typos and minor issues**

*Line 21: The resolution of the model is not consistent through the text, In the abstract is 10 km, but in the description of the model (line 100) is ∼ 12 km.*

R: The resolution is 12 km. We corrected the error in the abstract.

*Line 31: "small pelagic fisheries"*
R: Corrected.

*Line 50: IPCC is not defined*
R: We replaced by CMIP5 and wrote out the meaning of the acronym line 51.

*Line 52: "Oyarzún"*
R: Corrected.

*Line 58: AR is not defined*
R: Corrected.

*Line 76: change 2017 for 2018.*
R: Corrected.

*Lines 82-83: The phrase "most recent climate scenarios" is not clear to me. Do you imply that the RCP's are recently developed scenarios? that we are following these scenarios? please clarify.*
R: We modified the sentence: "..under climate scenarios taking into account economic and population growth assumptions (e.g. RCP8.5) and over longer time periods (e.g. 100 years)." (lines 85-86)

*Lines 117-118: Is it possible to fix the exponential with the symbol and superscript?*
R: Corrected.

*Line 124: CMIP5 is not defined*
R: It is now defined line 51.

*Line 141: Needs a comma after "However"*
R: Corrected.

*Line 171: There is no entry on the reference list for Echevin et al., 2010.*
R: Thank you for noticing this error, we added the correct reference (Echevin et al. 2012).

*Line 200: Section 2.7 is missing*
R: Corrected.

*Line 216: The number of the figures they are referring to is missing.*
R: Corrected.

*Lines 255-262: This section is described as if the trends where those of the ESMs, when figure 4 shows the change in the RCMs. Also, there is no consistency with the use of "R+model" to indicate the downscaled simulation.*

R: We corrected the text and figure title to clarify what comes form the ESMs (downward longwave flux, net downward shortwave flux) and what results from the RCM bulk formulae computation (net longwave, wind stress) (lines 275-281; Fig.4).

*Line 316: Another example of a subjective phrase "weak dissolved O2 concentrations".*
R: We modified the sentence, cited a value for the oxygen concentration, and added a reference (line 347).

*Line 318: There is no entry for Espinoza et al., 2019 in the reference list.*
R: We modified the reference (Espinoza-Morriberón et al., 2019).

*Line 321: You mean the RCM eastward surface flow?*
R:  No, this is actually the ESM eastward subsurface flow, as 95°W is the location of the RCM western boundary. We modified the sentence as follows: "we first evaluate the ESM eastward subsurface flow (which enters the western boundary of the RCM) at 95°W" (line 353).

*Line 328: "The trend is relatively weak. . ."*
R: This sentence has been changed due to changes in the figure (see our general comment above).

*Line 330: I find that the use of parentheses to indicate the opposite of an idea in a paragraph is confusing and inefficient. I invite the authors to use parentheses for clarification and citations only and not to save space. See Robock, A. 2010. Parentheses are (are not) for references and clarification (savings space). Eos, Trans. Amer. Geophys. Union, 91(45): 419).*
R:  The sentence has been modified due to changes in the figures (lines 360-365).

Table 1. Needs a better description of terms. What does 10 m mean? 10 m wind?
R: 10 m indicates the thickness of the ESM ocean surface layer.  The legend of the table (now Table 2) has been modified .

*C: Fig. 1. For clarity, I would suggest to make the vertical axis of each subplot equal and visualizing the extent of the influence of the OMZ on nitrate is not evident.*

R: Fig1. Vertical axis is now 0-500 m for oxygen panel in Fig.1 Note that it is 0-250 m for temperature and salinity to better highlight differences of the thermocline and subsurface salinity maximum structures.

*C: Also, the thickness of the lines representing the selected ESM's is not really different from the rest. Perhaps the legend should refer to these as "solid colored lines" instead of "thick colored lines."*

R: We have increased the thickness of the lines in Figure 1.

*C: Fig. 2. The description of the legend is not consistent with what is being showed and what is described on the text. i.e., b) and d) should be output from the RCM (downscaled).*
R: Corrected.

*Fig. 3. The word "value" is missing in the legend just before (c).*
R: Corrected.

*Fig. 4. In the legend (c) is missing.*
R: Corrected.

*Fig. 11. In the legend, fix the superscript in μmol L-1.*
R: Corrected.

*Fig. 16. The legend is wrong, there are no figures 16d-f.*
R: Corrected.

*Fig. 17. In a) the title of the figure is wrong. These should be the trends of the ESMs not RCM as mentioned in the legend and in the text (line 508). It should be indicated somewhere in the legend that the trends in b) and c) correspond to the R-GCM' sensitivity experiments.*

R: Figure 17 has been modified. The results from the simulations forced by dissolved oxygen climatological boundary conditions. Thus they correspond to the RCM' values and not to ESM values. The legend and the text have been modified accordingly.

**Reviewer #2:**

**General Comments**

*The authors explore the projected physical and biogeochemical state of the Northern Humboldt Current System (NHCS) under future climate change using a regional circulation model (RCM) forced by three global earth system models (ESMs). They describe changes in a range of ocean properties from temperature to zooplankton biomass, focusing on trends relative to historical conditions as well as the differences among the different ESM and RCM projections.*

*Future conditions in eastern boundary upwelling systems like the NHCS are of considerable interest due to the biogeochemical, ecological, and socioeconomic importance of these regions. It's also well known that fine scale dynamics in these regions are important and are not well captured by coarse resolution global models, so there is interest in the potential added value provided by dynamical downscaling. Therefore, this is valuable work and is at the cutting edge of regional ocean projection. The inclusion of biogeochemistry, the use of multiple ESMs to force the regional model, and the bias correction of the forcing are all notable and positive elements of the research.*

R: We thank the reviewer for his encouraging and constructive comments.

*The manuscript is mostly descriptive; the authors note that further mechanistic analysis is left to future research. In my view, the most important results are the comparisons of projected changes between the global and regional models. We know global models have biases, but it's when the projected change is altered by downscaling that a stronger case is made for the need to downscale. The authors find that this is the case for biogeochemical, but not physical, variables. I have a number of specific comments below, but my main concerns are with several choices in the methods, detailed below.*

**Specific Comments**

*I have three main concerns on the methods:*

*1. (Section 2.3). The choice of which ESMs to use has been justified based on historical comparisons with observations. However, there is a growing body of research arguing against this method, since these historical model evaluations do not necessarily correspond to how well a model captures the response to future climate forcing. "Emergent constraints" have been offered as a more relevant method for evaluating climate models (Hall et al. 2019). In the absence compelling reasons why a model is unrealistic for the future change, the default should be to pick a suite of models that capture the range of potential futures.*

R: We agree with the reviewer that the method of "emergent constraints" is a relevant method for selecting ESMs in order to project the impact of climate change on particular variables, and that even ESMs with strong biases can be used in that method. Indeed it is possible that a model may represent a correct relation between present state and future conditions even with an important bias in the present state. If our study were to be done again today, we would probably investigate this approach and the method of emergent constraints would be a good candidate. However, in the present study we are interested in the projections of several parameters (stratification, upwelling, OMZ, productivity), thus

we would have had to find different "emergent constraints" for each of these variables, which may be intricate, and moreover, may lead to select different models for each constraint. Also, we have to admit that we were not aware of this approach at the beginning of our study, which has mainly been used in basic climate studies and not for regional downscaling (to our knowledge). Therefore, we consider that it is beyond the scope of the present work to select ESMs based on an emergent constraint which remains to be identified for the region of study, but we agree that it would be interesting to investigate further such an approach. We added a short paragraph to discuss this aspect at the beginning of the discussion (discussion section 4.1, lines 509-519).

*2. (Section 2.4). As I understand it, this method produces forcing with no high-resolution (sub monthly) variability. High frequency wind variability can be very im portant especially to the BGC in Eastern Boundary systems. For example, Gruber et al (2006) attribute model chlorophyll biases to the use of monthly forcing. For future projections, one can add representative high frequency variability (e.g., from historical reanalysis) as a third term on the right hand side of equation (1). Similar has been done for historical sensitivity analyses (Frischknecht et al. 2015, Jacox et al. 2015).*

R: We fully agree with the reviewer on that point: high frequency wind variability can be important in EBUS, in areas where upwelling tends to occur episodically, as stated in Gruber et al. (2006). Off central Peru, upwelling favorable winds are persistent over longer periods than for example off Central Chile or Northern California. So, we may expect a relatively moderate effect of high frequency wind variability off Peru. A previous work by Echevin et al. (2014) showed indeed that its impact in the Peru upwelling system on some of the key biogeochemical fields is not strong. In this study, they performed sensitivity experiments on the boundary and atmospheric forcing. Two simulations were compared, one with daily wind stress (named REF, see table  2  in Echevin et al., 2014)  and one with monthly climatological wind stress (named CLIM). The mean state  computed over 7 years of simulation (Fig.9 in Echevin et al., 2019) displays very little change between the REF and CLIM simulations for cross-shore profiles of chlorophyll,  nitrate, phosphate, silicate and  iron. Thus we believe that the impact of the wind sub-monthly variability may not play an important role in this system and would not strongly impact the low frequency variability we focussed on. Note also that we had no choice but to use the monthly forcing as daily wind forcing was not available for all the ESMs we selected at the time we started our study.

The text has been modified as follows: "Note that submonthly wind variability may impact significantly surface chlorophyll in northern California (e.g. Gruber et al., 2006). However, previous regional modeling experiments in the NHCS showed a weak impact (less than 10% difference) of daily wind stress with respect to monthly wind stress on 7-year-averaged biogeochemical variables (Echevin et al., 2014). This suggests that using monthly winds may not impact significantly the climate trends reported in this study." (lines 179-184)

*3. (Section 2.5): First, it's unclear why one would not bias-correct the physical ocean boundary conditions. For consistency they should be treated like the surface and ocean BGC fields. Second, oxygen should be treated the same as the other biogeochemi cal variables. While I understand the concern about unrealistic oxygen values, the oxygenation trend is inextricably linked to the trend in nitrate concentration (Fig. 11) and presumably other nutrients, and in turn with trends in productivity. It doesn't make sense to deem the oxygen trend unrealistic and the others realistic. Furthermore, since oxygen and nitrate variability are closely coupled, imposing the ESM change in one but not the other introduces biogeochemical inconsistencies that may compromise the RCM findings. The analysis of oxygen using climatological boundary conditions is still*

*interesting as it allows one to separate different contributions to the regional change, but it's not consistent with the rest of the analysis. Therefore, the main text should include the GCM change, with the context that you are trying to bound the range of possible futures, not to predict exactly what happens in the future. The oxygen analy sis using climatological boundary conditions can move to discussion.*

R: As suggested by Reviewer#1, we now include in Figure 1 the ESMs temperature and salinity profiles in the equatorial region. We also compute a normalized bias defined as follows:
 NB= (z)= |X_model(z)-X_obs(z)|/X_obs(z) x 100.
This bias has been computed for each variable X=(T,S, nutrients, O2) in a new Table 1. This allows comparing the amplitude of the normalized bias between the ESMs. We also find that the normalized biases for temperature and salinity are weaker than those for nutrients and oxygen, which justified our approach. In other words, we estimate objectively that biogeochemical variables are less well represented than physical ones, which led us to not correct the bias of physical variables. Another difficulty would be to correct the bias of equatorial currents close to the equator where geostrophy is not valid. An interesting alternative could be to use a reanalysis (e.g. SODA) as a climatological present state and add ESM anomalies to this climatological state. This approach is however beyond the scope of the present study.
We also agree with the reviewer that separating oxygen from the other biogeochemical variables is not consistent as oxygen values can feedback on nitrate concentration. Thus we now present in the results section the RCM solutions obtained with the ESM oxygen trends as boundary conditions, and move results from the RCM simulations with WOA climatological oxygen conditions to the discussion section (lines 562-570). This led to many modifications in the text and figures.

**Detailed Comments:**
*L20, 428, 547: Suggest removing "business as usual": See Hausfather and Peters (2020).*
R: We now use the term "worst case scenario".

L87: Unclear what "in the following paragraphs" refers to. The whole rest of the paper?
R: We rephrased this sentence.

L115-121: The differences are described and are stated to be important, but it's not clear what is the motivation for these changes.
R:  The changes are due to the fact that the PISCES model described in Aumont et al. (2015) is a more recent version of the PISCES model (PISCES version 2). We had to use an earlier version (PISCES version 0) coupled to ROMS at the time of the study. The PISCESv2 had not been coupled to ROMS at the beginning of our study.
We modified the text as follows: " Detailed parameterizations of PISCES (version 2) are reported in Aumont et al. (2015). Note that we used an earlier version of the model (PISCESv0) in this study, as PISCESv2 had not been coupled to ROMS yet at the beginning of our study. Here we describe the following parameterizations of PISCESv0:…" (lines 119-121).

Section 2.6: The temporal coverage of these data sets is quite short for evaluating historical model performance, given that the decadal variability in the ESMs should not be expected to align with nature. Something like 30 years would be more appropriate, but in any case the authors should be wary of caveats associated with using short observational records.
R: We agree with the reviewer that a longer regional simulation over the historical time period would be appropriate to filter decadal variability. However, to avoid performing regional simulations over 150

years (1950-2100), we chose to limit our historical simulations to the period 1997-2015. Following Reviewer#1's comment, we added a supplementary figure displaying the cross-shore sections of mean temperature and dissolved oxygen over 2006-2015 and a comparison with climatological observations. After a spin-up phase of 8 years (1997-2005), the simulations are equilibrated and the biases are reasonable.

L205: This is probably fine as a proxy, but it's worth noting that it doesn't explicitly represent upwelling, including the curl-driven component. If so inclined, one could get a more accurate upwelling metric by integrating the Ekman and geostrophic components over the region of interest or by using the vertical velocity at the based of the Ekman layer (Jacox et al. 2018). It would also be helpful here to describe the calculation of the cross-shore geostrophic transport.

R: There seems to be a misunderstanding. We compute the upwelling index by integrating over the Ekman layer depth and over the region of interest the cross-shore current, which is almost exactly equal to the Ekman current and the geostrophic current (see Fig. 10 in Oerder et al., 2015). The same calculation is done by Jacox et al. (2018), who showed that the horizontal transport computed in this manner is very close to the upwelling computed using model vertical velocities. We added a reference to Jacox et al. (2018) in the manuscript (see lines 217-224).

L221: Thanks to its high spatial resolution and the bias correction of the forcing.
R: We agree. Corrected (line 240).

*Fig. 3: I think it would be more appropriate to show the bias corrected ESM change (i.e., remove the mean ESM SST bias so that they all start from the same place). I also don't think a % change is best for SST, at least if the units are Celsius. In Fig. 3 the ESM % changes are lower because they are starting from a warm-biased state. But the magnitudes of projected temperature changes are as large as or larger than the RCM. Lastly, throughout the manuscript some indication of significance should be added to the trends.*

R: We performed the suggested changes. Fig.3b now shows the ESMs SST evolutions starting from the same value in 2006. We agree that this shows better that the RCMs SST follows the ESMs SST when the initial bias is corrected. We also added the SST change values (instead of percentages) in the figure. Last, we have estimated the trend uncertainty using a bootstrap method based on a 10 000 member synthetic distribution derived by randomly removing data from the annual series. The method is described in the methodology section 2.8. (lines 233-236). We converted the trend uncertainty into a percentage uncertainty, now reported in Tables 3, 4, 5. We also have computed the $R^2$ from the least square estimation in Tables 3,4,5. Most of the trends are significant at the 10% level. The significant trends are reported in bold font in the tables.

*L248-249: Did Bakun actually project cooling, or just intensified upwelling? There could be intensified upwelling but still warming due to dominance of the surface heating.*

R: The reviewer is right. Bakun (1990) projected intensified wind-driven upwelling and suggested that it would induce a cooler surface ocean and more foggy coastal regions. Bakun et al. (2010) only projected intensified wind-driven upwelling. We suppressed the sentence.

*L252-253: It's not clear to me the evidence that this pattern results from the upwelling and subsequent lateral transport/damping of subsurface anomalies.*
R: We agree with the comment. We suppressed the statement.

*Fig. 4: Would be informative to see the net longwave and shortwave radiation (not just downwelling).*
R: We added the net longwave radiation (now Fig. 4d). The shortwave radiation presented here is the net shortwave radiation coming from the ESM. We modified the titles of Fig.4 to clarify which forcing comes from the ESM (net shortwave and downward longwave radiation) and which results from the RCMs air-sea interactions (net longwave radiation and wind stress).

*L266: Initially it seems strange that the offshore transport trend is 2x greater than the wind stress trend, since the transport is linearly related to the wind stress. But, it does make sense because when you calculate the Ekman transport (i.e., Fig. 5a minus Fig. 5b), the change is ~10%, consistent with the winds. This should be explained in the text, and I suggest adding the Ekman transport as a third panel to Fig. 5.*
R: We added the Ekman transport in Fig.5 (now Fig.5c). It shows clearly that the net offshore transport is equal to the sum of the Ekman transport and geostrophic transport. We modified the text accordingly (lines 287-288).

*L269-270: It's also interesting that since there's a long-term trend in Ekman transport but not in geostrophic transport, the relative contribution of the geostrophic transport increases over time.*
R: We agree. We added this comment in the text (line 291).

*L275: Do you mean they are locally influenced by the passage of waves? Or are you suggesting the waves actually propagate (advect) the anomalies somehow?*
R: We rephrased the beginning of the paragraph as follows: "Nearshore subsurface temperature anomalies are impacted by equatorial subsurface temperature anomalies in two ways: thermocline anomalies may propagate along the equatorial and coastal wave guide (e.g. Echevin et al., 2012, 2014; Espinoza-Morriberón et al., 2017, 2018), and temperature anomalies may be transported eastward and poleward by the near-equatorial subsurface jets (Fig.2a; Montes et al., 2010, 2011). The latter is particularly strong during eastern Pacific El Nino events (*e.g.* Colas et al. 2008 for the 1997-1998 event)." (lines 298-302).

*L279: I would be careful about equating the depth of an isotherm (D20) with the depth of the thermocline (i.e., the depth of maximum temperature gradient). Temperature biases (or changes) will alter D20 but not necessarily the thermocline depth.*
R: We no longer equate D20 with thermocline. We changed the title of the figures and modified the text as follows: "The thermal structure of the upper layer is strongly impacted by climate change in the eastern equatorial Pacific. The depth of the 20°C isotherm (hereafter D20) is used to characterize the thickness of the warm surface layer. "(lines 303-304).

*L297: How is MLD calculated?*
R: The model surface boundary layer (hbl) is computed from the value a critical Richardson number computed using the KPP formulation. Previous modelling work show that the surface boundary layer thickness is very close to the model mixed layer (Liu and Fox-Kemper, 2017). We added this information and this reference (lines 319-321).

*Figure 8: I would like to see the ESM tendencies on here as well*
R: We added the ESM tendencies in Fig.8. In the figure the anomalies at 95°W are from the ESMs only as 95°W is the location of the RCM western open boundary. We modified the text accordingly (lines 333-335).

*L308-309: I don't understand why this statement is here. I would delete it*
R: Agreed. We suppressed the statement.

*L351 and elsewhere: "deemed realistic enough" isn't very convincing. I don't think you have to argue for the realism of the ESM changes, rather you are looking at the regional impact of the ESM changes as one potential future scenario.*
R: Agreed. We suppressed the statement.

*L352: Since the 95W location is discussed a number of times, it would be helpful to show it on a map (e.g., Fig. 2a along with the coastal region)*
R: Agreed. We added a red line in Figure 2b and added information in the legend.

*L360-365: It's hard to compare a concentration in one place (Fig. 11a) with a depth level in another place (Fig. 11c), especially when arguing that one is the driver of the other. Can these be presented in a more consistent way?*
R: We modified the figure and now plot the depth of the 21 μmol L$^{-1}$ nitrate iso-surface at 95°W in the RCM boundary condition (Fig.11a). It is now directly comparable to Figs.11c,d. We modified the text accordingly (lines 391-393).

*L389-397: I must say I'm surprised to see trends of opposite sign in the upper 10m. Surely one can't have opposite trends at different depths within the mixed layer. Per haps this is a seasonal signature, e.g., in summer the mixed layer is very shallow (~5m) and increased chlorophyll in the seasonal mixed layer is driving the overall trend. But the authors should look at this in more detail to explain how chlorophyll at 2m can have an opposite trend from chlorophyll at 7m.*

R: We agree with the reviewer's comment. Following his comment, we modified the figure and now plot vertical sections of the seasonal trends computed for summer and winter (Fig.13). This is quite interesting as we see clearly (1) the deeper mixed layer depth in winter than in summer, (2) a different behavior for R-CNRM due to a stronger nitrate limitation. We added the following paragraph in the text:
" The seasonal trends in R-GFDL and R-IPSL are consistent with a shoaling of the mixed layer (Fig.7), which reduces light limitation of phytoplankton growth (e.g. Echevin et al., 2008; Espinoza-Morriberón et al., 2017) and increases surface primary productivity in summer and winter. In contrast, the R-CNRM trend in the mixed layer is negative in summer. This is likely caused by the strong deepening of the nitracline in R-CNRM (Fig.11c) and the seasonality of the wind-driven upwelling. As the upward flow is weaker in summer, the upwelling of less rich waters into the mixed layer may trigger a nutrient limitation of phytoplankton growth. On the other hand, as the upward flow remains strong during winter, nutrient limitation does not occur. Light limitation of phytoplankton growth reduces because of the shoaling of the mixed layer, enhancing phytoplankton growth (as in the two other RCMs). Moreover, visual correlation between decadal variability of the chlorophyll content and nitracline depth

in R-CNRM (e.g. the oscillations in 2070-2100 in Fig.11c and Fig.12c) also suggests that nitrate limitation of phytoplankton growth may play a role." (lines 334-346).

*L430: This may be true, but without a heat budget it's speculative. There will be other contributions as well (e.g., local surface fluxes). The text at L456-460 is good.*
R: We modified the sentence.

*L547-552: There is also a summary statement like this in the previous section (L427-429). One of them should be cut – probably the earlier one.*
R: We left the earlier statement as it is placed in the summary. We believe that repeating this statement in a similar manner in the conclusion section does not burden the manuscript.

L561-565: I think this is all speculation, so should be presented as hypotheses rather than fact (unless there is evidence to support it)
R: We agree. We have modified the text accordingly: " We **can speculate** that this happens for two reasons: the enhanced thermal stratification due to the warming **may** alleviate light limitation and vertical dilution, and the reduction of wind-driven offshore transport **may** allow plankton to accumulate near the coast. These processes **could** partly compensate the reduction of primary productivity due to a deeper nitracline and reduced wind-driven coastal upwelling." (lines 616-619).

**Technical Corrections**
L117-118: Does the period in a.T indicate multiplication?
R: Yes. We modified the text. It is now written as $e^{a.T}$

L368-369: Quasi-absent doesn't make sense. Maybe negligible? Insignificant?
R: We modified the text.

Table 1: Suggest including in the caption the meaning of abbreviations (mainly Pg and Zg) and the meaning of (10m) in the number of vertical levels column. Also I don't think the full references are needed in the table, they can be in the reference list.
R: We modified the caption and the references in the table accordingly.

Figure 4: in caption (d) should be (c)
R: We modified the caption accordingly.

Figure 6, 7, 13, 16: Values should be positive for depth
R: We modified the figures accordingly.

Figure 17: Top panel should be ESM?
R: Fig.17 has been modified as it now shows the RCM results forced by the climatological oxygen boundary conditions.

# Physical and biogeochemical impacts of RCP8.5 scenario in the Peru upwelling system

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**Abstract:**

The northern Humboldt current system (NHCS or Peru upwelling system) sustains the world's largest small pelagic fishery. While a nearshore surface cooling has been observed off southern Peru in recent decades, there is still considerable debate on the impact of climate change on the regional ecosystem. This calls for more accurate regional climate projections of the 21st century, using adapted tools such as regional eddy-resolving coupled biophysical models. In this study three coarse-grid Earth System Models (ESMs) from the Coupled Model Intercomparison Project (CMIP5) are selected based on their biogeochemical biases upstream of the NHCS and simulations for the RCP8.5 climate scenario are dynamically downscaled at ~12 km resolution in the NHCS. The impact of regional climate change on temperature, coastal upwelling, nutrient content, deoxygenation and the planktonic ecosystem is documented. We find that the downscaling approach allows to correct major physical and biogeochemical biases of the ESMs. All regional simulations display a surface warming regardless of the coastal upwelling trends. Contrasted evolutions of the NHCS oxygen minimum zone and enhanced stratification of phytoplankton are found in the coastal region. Whereas trends of downscaled physical parameters are consistent with ESM trends, downscaled biogeochemical trends differ markedly. These results suggest that more realism of the ESMs **circulation, nutrient and dissolved oxygen fields** is needed in the eastern equatorial Pacific to gain robustness in the projection of regional trends in the NHCS.

**1 Introduction**

Eastern Boundary Upwelling Systems (EBUS) are oceanic systems where alongshore winds generate the upwelling of deep, cold and nutrient-replete waters. This drives a high biological productivity and thriving small pelagic fisheries which are major sources of income for the adjacent countries. In particular, the Peruvian Upwelling System (also known as the

Northern Humboldt Current System, NHCS in the following), located in the South Eastern Tropical Pacific, is the most productive EBUS in terms of fish catch (Chavez et al., 2008), due to its rich anchovy fishery. Moreover, the subsurface water masses in the NHCS are located in the poorly ventilated so-called "shadow zone" of the south eastern Pacific (Luyten et al., 1983). This low ventilation creates a subsurface water body with very low oxygen concentration, the oxygen minimum zone (OMZ). The OMZ results from a balance between oxygen consumption by respiration of large amounts of organic matter exported from the highly productive surface layer, and ventilation by the equatorial current system composed of eastward jets transporting relatively oxygenated waters (Czeschel et al., 2011; Montes et al., 2014). A particular aspect of the NHCS OMZ is its very low oxygen concentration (anoxia) at relatively shallow depths, which impacts the local marine ecosystem (Stramma et al., 2010; Bertrand et al., 2011).

In the recent decades, public concern has risen about the impact of climate change on EBUS. Using ship wind observations, Bakun (1990) showed that upwelling-favorable winds increased over recent decades (1950-1990) in several EBUS. He proposed that nearshore winds would continue to intensify due to an enhanced differential heating between land and sea, driven by a stronger greenhouse effect over land. However, this hypothesis has been challenged in the NHCS because of observation bias (*e.g.* Tokinaga and Xie, 2011) and poleward displacement of the South Pacific Anticyclone (Belmadani et al., 2013; Rykaczewski et al ., 2015). Nevertheless, in situ and satellite Sea Surface Temperatures (SST) show a conspicuous surface coastal cooling off southern Peru (15°S) since the 1950s. This cooling, consistent with a wind increase found in ERA40 reanalysis, suggests a possible intensification of the wind-driven upwelling (Gutierrez et al., 2011).

Recent analysis of the 5th Coupled Model Intercomparison Protocol (CMIP5) global circulation models (GCMs) reported that the intensification of nearshore winds under scenarios of carbone dioxyde concentration increase is mainly confined to the poleward portions of EBUS (Wang et al., 2015; Rykaczewski et al., 2015, Oyarzún and Brierley, 2019). However, the evolution of winds in the NHCS remains unclear (note that the NHCS stricto sensu was not included in these studies). Furthermore, the realism of IPCC GCMs is hampered by the coarse resolution of the model grids (~100-200 km), that does not allow to represent the details of coastal orography and coastline that influence the coastal wind structure.

A few downscaling studies focusing on regional wind changes in the NHCS have provided invaluable information. NHCS upwelling-favorable winds may weaken in the future, mainly during the productive austral summer season (Goubanova et al., 2011; Belmadani et al., 2014). However, only idealized extreme scenarios (preindustrial, doubling (2xCO2) and quadrupling (4xCO2) of carbon dioxyde concentration) from a single GCM (IPSL-CM4, Marti et al., 2010) of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) were downscaled in these studies. In line with these studies, Echevin et al. (2012) used a regional ocean circulation model (RCM) forced by statistically downscaled atmospheric winds from Goubanova et al. (2011) to downscale the NHCS ocean temperature and circulation changes under 2xCO2 and 4xCO2 scenarios. They found a strong warming in the surface layer, of up to ~+5°C nearshore in the 4xCO2 scenario with respect to preindustrial conditions, and an upwelling decrease during austral summer. Following the same regional modeling approach and using the downscaled winds from Belmadani et al. (2014), Oerder et al. (2015) found a year-round reduction in upwelling intensity, mitigated by an onshore geostrophic flow. The shoaling of upwelling source waters in the 2 scenarios suggests that upwelled waters could become less nutrient-rich, and thereby reduce nearshore primary productivity (Brochier et al., 2013).

The impact of climate change on the NHCS productivity, oxygenation and acidification has been even less investigated. Assuming Bakun's (1990) hypothesis of increasing coastal winds, Mogollón and Calil (2018) found a moderate increase (5%) of the NHCS productivity using a RCM. However, they did not take into account the large-scale stratification changes driven by climate change that may significantly contribute to nearshore stratification and mitigate the upwelling (Echevin et al., 2012; Oerder et al., 2015). Following a similar approach, Franco et al. (2018) found a sustained acidification of NHCS shelf and slope waters under the Representative Concentration Pathway 8.5 scenario (RCP8.5, the so-called worst case AR5 climate scenario corresponding to a 8.5 Wm$^{-2}$ heat flux driven by the greenhouse effect, *e.g.* van Vuuren et al., 2011), driven by changes in surface fluxes of atmospheric $CO_2$ concentration and subsurface dissolved inorganic carbon concentrations. However, as in Mogollón and Calil (2018),  the impact of climate change on NHCS surface winds, circulation and stratification was unaccounted for in Franco et al. (2018).

In brief, previous regional modelling experiments were either obtained from (i) the downscaling of one single GCM or Earth System Model (a GCM including a biogeochemical model, hereafter ESM), (ii) the analysis of relatively short time periods (e.g 30 years in the stabilized phase of the 2xCO2 and 4xCO2 scenarios in Echevin et al., 2012; Oerder et al., 2015; Brochier et al., 2013), or (iii) simplified approaches that did not account for all physical forcings (*e.g.* Mogollón and Calil, 2018; Franco et al., 2018). More work is thus needed to evaluate the robustness of these findings under climate scenarios taking into account economic and population growth assumptions (*e.g.* RCP8.5) and over longer time periods (*e.g.* 100 years).

In the present work, 3 different ESMs are dynamically downscaled in the NHCS using a regional coupled dynamical-biogeochemical model. The studied time period is 2005-2100 under the RCP8.5 scenario. The regional trends from RCMs are compared to illustrate the diversity of climate change regional impacts. RCM trends are also contrasted with those of the ESMs in order to highlight the impact of the downscaling process. In the next section (section 2) the regional model, the selection process of ESMs and the downscaling methodology are described. Results are presented in section 3: we describe the trends of key physical and biogeochemical parameters such as temperature, coastal upwelling, thermocline depth, oxygenation, nitrate and productivity. The approach and implications of our work are discussed in section 4. The conclusions and perspectives are drawn in section 5.

## 2 Methodology

### 2.1 Ocean model

The Regional Ocean Modeling System (ROMS) was used to simulate the ocean dynamics. The ROMS AGRIF (version v3.1.1 is used in this study) resolves the Primitive Equations, which are based on the Boussinesq approximation and hydrostatic vertical momentum balance (Penven et al., 2006; Shchepetkin and McWilliams, 2009). A fourth-order centered advection scheme allows the generation of steep tracer and velocity gradients (Shchepetkin and McWilliams, 1998). For a complete description of the model numerical schemes, the reader can refer to Shchepetkin and McWilliams (2005).

The model domain spans over the coasts of south Ecuador and Peru from 5°N to 22°S and from 95°W to 69°W. It is close to the one used in Penven et al. (2005). The horizontal resolution is 1/9°, corresponding to ∼12 km. The bottom topography from **STRM30 (Becker et al., 2009)** is interpolated on the grid and smoothed in order to reduce potential errors in the

horizontal pressure gradient. The vertical grid has 32 sigma levels.

 Wind speed, air temperature and humidity, and ROMS SST are used to compute latent and sensible heat flux online using a bulk parameterization (Liu et al., 1979).

**2.2 Biogeochemical model**

 ROMS is coupled to the Pelagic Interaction Scheme for Carbon and Ecosystem Studies (PISCES) biogeochemical model. PISCES simulates the marine biological productivity and the biogeochemical cycles of carbon and main nutrients (P, N, Si, Fe; Aumont et al., 2015) as well as dissolved oxygen (DO) (*e.g.*, Resplandy et al., 2012, Espinoza-Morriberón et al., 2019). It has three non-living compartments which are the semi-labile dissolved organic matter, small sinking particles and large sinking particles, and four living compartments represented by two size classes of phytoplankton (nanophytoplankton and diatoms) and two size classes of zooplankton (microzooplankton and mesozooplankton). The ROMS-PISCES coupled model has been used to study the climatological (Echevin et al., 2008), intraseasonal (Echevin et al., 2014), and interannual variability of the surface productivity (Espinoza-Morriberón et al., 2017) and oxygenation (Espinoza-Morriberón et al., 2019) in the NHCS. Detailed parameterizations of PISCES (version 2) are reported in Aumont et al. (2015). **Note that we used an earlier version of the model (PISCESv0) in this study, as PISCESv2 had not been coupled to ROMS yet at the beginning of our study.** Here we describe the following parameterizations of PISCESv0: i) diatoms and nanophytoplankton growth, microzooplankton grazing and mortality, mesozooplankton mortality depend on temperature (T) and are proportional to $e^{a.T}$ with a=0.064 °C$^{-1}$; ii) mesozooplankton grazing on nanophytoplankton and diatoms is proportional to $e^{b.T}$ with b=0.076 °C$^{-1}$. These differences, in particular the larger temperature-enhanced mesozooplankton grazing with respect to phytoplankton growth, can play an important role in the context of surface warming in the NHCS. Boyd et al. (1981) measured grazing of Peruvian copepods, however further laboratory experiments are needed at different temperatures to calibrate these rates.

**2.3 Selection of the Earth System Models**

        Three CMIP5 ESMs are selected for the regional downscaling. The selection process is based on the nutrients simulated by the ESMs and on the evaluation of biogeochemical bias. Only five ESMs (CNRM, GFDL, IPSL, CESM and Nor-ESM) represent the four nutrients (silicate, phosphate, nitrate and iron) and DO required by PISCES. As different ESM versions were available, a total of eight ESMs (CNRM-CM5, GFDL-ESM2M, GFDL-ESM2G, IPSL-CM5A-MR, IPSL-CM5A-LR, IPSL-CM5B-LR, CESM1, Nor-ESM1-ME) were compared to observations from the World Ocean Atlas (WOA2009, Fig.1). Following Cabré et al. (2015), the ESM **DO, nutrients, temperature and salinity** were averaged at 100°W between 5°N and 10°S, near the location of the western open boundary of the RCM, **for the period 1980-2005 (1950-2005 for T  and S)**. This meridional section intersects eastward jets: the Equatorial Undercurrent (EUC) at 0°S and the off-equatorial Southern Subsurface Countercurrents (SSCCs) at ~ 4°S and ~ 8°S (Montes et al., 2010). These jets transport physical and biogeochemical properties to the Peru upwelling region (Montes et al., 2010, 2014; Oerder et al., 2015; Espinoza-Morriberón et al., 2017, 2019).

        **Visual examination of the ESMs temperature and salinity profiles (Fig. 1) suggests that the corresponding biases are weak in comparison with other variables. The comparison between the biases of different variables can be**

**quantified by computing a bias normalized by the mean state, averaged between 0 and 500 m depth (0-250m for temperature and salinity,  see Table 1). The ESMs normalized temperature bias is weaker than the biogeochemical biases (Table 1).**

145    All ESMs simulate an oxygen decrease with depth (Fig.1a), but oxygen values are too low (i.e. <10 μmol $l^{-1}$) in CESM1-BGC, GFDL-ESM2M, GFDL-ESM2G and NorESM1-ME. Slightly negative values are attained below 300 m depth for GFDL-ESM2G. CNRM-CM5. In contrast, the 3 IPSL model versions, which all include PISCES as biogeochemical component, overestimate the oxygen content above ~600 m depth. Note that only CESM1-BGC is able to reproduce the observed oxygen increase below 400 m depth, which corresponds to the lower limit of the OMZ.

150    In terms of nitrate concentration, the most realistic models in the upper 300 meters are GFDL-ESM2G, GFDL-ESM2M and CESM1-BGC (Fig.1b). However, the model biases become negative and increase strongly at depths greater than 300 m. A negative bias found in the three IPSL ESMs (~ 3-4 μmol $l^{-1}$ for IPSL-CM5A-MR and IPSL-CM5A-LR and ~6-8 μmol $l^{-1}$ for IPSL-CM5B-LR) is roughly constant over depth. CESM1-BGC, GFDL-ESM2G and NorESM1-ME display too low nitrate concentrations below 250 m depth, possibly due to denitrification in the anoxic OMZ (Fig.1a).

155    The GFDL-ESM2M phosphate profile is very close to the observations (Fig.1c, Table 1), whereas the three IPSL ESMs and CNRM-CM5 underestimate phosphate concentrations with a roughly constant bias over depth (negative bias of ~0.5-1 μmol $l^{-1}$). In contrast, NorESM1-ME, GFDL-ESM2G and CESM1-BGC overestimate the phosphate concentrations.

The IPSL and CESM1-BGC silicate profiles are close to observations above ~250 m depth, whereas the positive bias in GFDL-ESM2M and NorESM1-ME increases below 200 m depth.  The CNRM-CM5 negative bias is moderate between 50
160    and 300 m depth (Fig.1d).

To conclude, as the three IPSL ESMs and the CNRM-CM5 include the PISCES biogeochemical model also used in the regional simulations and provide reasonable nutrient bias with respect to the other ESMs (Table 1), IPSL-CM5A-MR (which nitrate and phosphate bias is weaker than the two others IPSL ESMs, Table 1) and CNRM-CM5 are selected. We also select GFDL-ESM2M, which represents well the nitrate and phosphate profiles in the upper layers, and whose bias did not increase
165    at depth as in GFDL-EMS2G. CESM1-BGC also has weak biases with respect to the latter ESMs (Table 1), but some variables were not available from the archive (*e.g.* 10 m wind) at the beginning of this study. We thus restrict our study to the downscaling of three ESMs. The main characteristics of the selected ESM ocean models (grid spacing and biogeochemical structure) are summarized in Table 2. We refer to the ESMs as CNRM, IPSL and GFDL in the following sections and figures.


170    **2.4 Atmospheric forcing methodology:**

A bias correction is used to construct monthly forcing files (*e.g.* Oerder et al., 2015; note that daily files were not available for all ESMs). For each forcing variable X (i.e. X=wind velocity, air temperature, ...), the bias-corrected variable X' is computed as follows:

$$X' = X_{OBSclim} + (X_{ESM-RCP8.5} - X_{ESM-hist-clim})    (1)$$

175    $X_{OBSclim}$ corresponds to a monthly climatology of observed values, $X_{ESM-RCP8.5}$ corresponds to the coarse-grid ESM values for each month, and $X_{ESM-hist-clim}$ to a monthly climatology of the coarse-grid ESM values during the historical period (2000-2010). This allows substracting the ESM mean bias, assuming that it remains identical over the historical period and over 2000-

2100. This method has been used in several papers (Cambon et al. 2013; Echevin et al. 2012; Oerder et al., 2015). The SCOW (Risien and Chelton, 2008) surface wind and COADS (Da Silva et al., 1994) downward shortwave and longwave fluxes and air parameters (temperature and specific humidity) climatologies were used for $X_{OBSclim}$. **Note that submonthly wind variability may impact significantly surface chlorophyll in other EBUS, such as off northern California where the wind variability is much stronger than off Peru (*e.g.* Gruber et al., 2006). Indeed, a previous regional modelling study in the NHCS showed a weak impact (less than 10% difference) of daily wind stress with respect to monthly wind stress on 7-year-averaged biogeochemical variables (Echevin et al., 2014). This suggests that using monthly winds may not impact significantly the climate trends reported in this study.**

### 2.5 Open boundary and initial conditions for physics and for biogeochemistry

As in Echevin et al. (2012) and Oerder et al. (2015), the ESM monthly sea level, temperature, salinity, horizontal velocity at the locations of the RCM open boundaries are directly interpolated on the model grid without bias correction. Given the important bias of the ESM mean biogeochemical state (*e.g.* Bopp et al., 2013; Cabré et al., 2015), we apply the bias correction described in Eq.(1): we add the WOA2009 (1°x1°) monthly climatology of the biogeochemical variables (nitrate, silicate, phosphate, iron, DIC, DOC, alkalinity, oxygen) and the annual mean anomalies (see Eq.(1)). The 3D fields were interpolated on the ROMS grid using the ROMSTOOLS package (Penven et al., 2008).

The three simulations are initialized as follows. Initial conditions from the ESM physical parameters of the historical simulation (2000-2010 January average) and WOA biogeochemical values (January) constitute the initial state. A 9-year spin up simulation from 1997 to 2005 is then performed to reach equilibrium. The runs are then forced by RCP8.5 conditions until 2100. State variables and biogeochemical rates (*e.g.* primary production) are stored every 5 days. The regional simulations are named R-IPSL, R-CNRM, R-GFDL in the following.

### 2.6 Additional data sets

Two ocean reanalysis products are used to evaluate the ESM equatorial circulation and thermocline in present conditions. The SODA 2.3.4 reanalysis (Carton and Giese, 2008) over the period 1992–2000 assimilates observational data in a general circulation model with an average horizontal resolution of 0.25°. The recently available GLORYS12V1 reanalysis over the period 1993-2017 is also used (Ferry et al., 2012). Altimeter data, in situ temperature and salinity vertical profiles and satellite SST were jointly assimilated in GLORYS12V1 (Lellouche et al., 2018). This product is freely distributed by the Copernicus Marine Environment Monitoring Service.

Several sets of observations are used to evaluate the realism of ESMs and RCMs in present conditions (i.e. 2006-2015 period). In situ data include CARS2009 gridded fields of temperature, nitrate and oxygen (0.5°x0.5°, Rigway et al., 2002), high resolution (0.1°x0.1°) regional monthly climatologies of temperature (Grados et al., 2018) and oxygen (Graco et al., 2018) including measurements collected during IMARPE (The Sea Institute of Peru) cruises. AVHRR satellite SST (2006-2015) is used to assess the RCM SST. Surface chlorophyll-a monthly climatologies from SeaWIFs (1997-2010) and MODIS (2002-2015) are used to evaluate the RCM surface chlorophyll.

**2.7 Coastal indices**

215     Time series of coastal indices characterizing the variability over the central Peru shelf for specific variables are computed. The variables are averaged in a coastal band extending from the coastline to 100 km offshore and between 7°S and 13°S.

        An index of coastal upwelling, the cross-shore transport in a coastal band, is computed from the model output (Colas et al., 2008; Oerder et al., 2015; **Jacox et al., 2018**). The mean horizontal transport is computed each month in a coastal strip

220     extending from 7°S to 13°S and from the coast to 100 km offshore. The transport is integrated vertically over the Ekman layer depth. **The latter is diagnosed as follows: we compute the surface geostrophic current using model sea surface height, and integrate the thermal wind relationship from the surface to the depth (equal to Ekman layer depth) at which the cross-shore current and the cross-shore geostrophic current differ by less than 10% (see Oerder et al., 2015 for more details). The computation of this index is more straightforward than one based on model vertical velocity (Jacox et al.,**

225     **2018) and leads to similar values (*e.g.* see Fig.4 in Jacox et al., 2018).** In contrast with coastal upwelling indices based on Ekman transport only, this index takes into account the role of the cross-shore geostrophic current which can modulate the coastal upwelling (*e.g.* during El Niño events, Colas et al., 2008; Espinoza-Morriberón et al., 2017).

**2.8 Statistical methods**

230     Only time scales longer **than 5-7 years (*e.g.* El Niño time scales**) are studied in this work. Therefore the time series are low-pass filtered using a ten-year moving average. This allows to filter the ENSO variability which is very strong in the NHCS, but not the focus of the present study. Linear trends of the time series are computed using a least squares method. The percentage of change between 2006 and 2100 associated with the linear trends are listed in 3 Tables (**Table** 3 for physical variables, **Table 4** for oxygen and nitrate, and **Table 5** for chlorophyll and zooplankton). **Statistical significance is presented**

235     **as a 90% confidence interval, based on a bootstrap method: we compute a 10 000 member synthetic distribution derived by randomly removing data in the annual series. The confidence limits of the trends are converted into confidence limits for the percentages reported in the tables.**

**3 Results**

240     In the following sections we show that the RCM is able to represent the main characteristics of the NHCS coastal upwelling system thanks to its high spatial resolution (relatively to the ESMs) and to the bias correction of the forcing. We then describe the long-term trends over the period 2006-2100 under the RCP8.5 scenario for key downscaled physical (surface and subsurface temperature, heat and momentum fluxes, upwelling) and biogeochemical parameters (oxygen and

245     nutrient content, primary productivity, planktonic biomass) in the upwelling system but also in the equatorial band offshore of the NHCS. For selected variables we also compare the downscaled simulations and the coarse-grid ESMs. In the next section, we first characterize the downscaled physical fields.

**3.1 Physical mean state and variability**

7

**Sea Surface Temperature spatial patterns**

We first contrast the Sea Surface Temperature (SST) patterns of the ESMs and RCMs to highlight the efficiency of the dynamical downscaling. The actual observed SST displays the cold water tongue along the coast and associated cross-shore SST gradient, characteristic of coastal upwelling (Fig.2a). The RCM simulates correctly these upwelling features (Fig.2b). The fine representation of the coastline, shelf and slope topography and **bias-corrected** alongshore winds (see section 2.4) all play a role in the correct representation of the upwelling structure. The upwelling vertical structure is also well reproduced in the RCMs. Mean cross-shore temperature profiles (within 500 km from the coast and between 7°S and 13°S) display the typical nearshore isotherms shoaling in the 0-100m layer and deepening below, in good agreement with the CARS climatology (Fig.S1a-d).

In contrast, the ESM SST (CNRM is shown here as an example, similar results are found for IPSL and GFDL) in present conditions (2006-2015) displays a warm bias of 2-4°C typical of ESMs (Flato et al., 2013) and no clear sign of coastal upwelling (Fig.2c). In 2091-2100, the RCM displays a coastal upwelling of waters ~2-3°C warmer than in 2006-2015 (Fig.2d). Again the ESM SST spatial pattern in 2091-2100 (Fig.2e) resembles that of 2006-2015. Coastal upwelling is not present and a warming of ~2-3 °C is found over the main part of the domain.

**Trends of nearshore SST**

A steady warming of the surface coastal ocean is found in the three regional simulations (Fig 3a). SST increases rapidly in R-IPSL since the 2020s, reaching +4.5°C in 2100, whereas it increases since the 2030s in the other simulations, reaching +3.5°C and +2°C in R-CNRM and R-GFDL respectively. Interestingly, decadal variability can produce decades during which the SST increase is stalled (a.k.a. "warming hiatus"), *e.g.* in 2035-2045 in R-CNRM and in 2040-2060 in R-GFDL. The ESM linear trends are very similar to the RCM nearshore warming trends (Fig.3b, Table 3). Here the offset between the three ESM SST evolutions due to the different SST bias in 2005 (between 4-6°C among the ESMs) has been corrected in order to better compare RCM and ESM trends. As an example, the spatial structures of the R-CNRM and CNRM SST anomalies are compared (Figs.3c,d). The similarity between the two anomaly patterns is striking. Both display a maximum warming near the coasts and west of the Galapagos where upwelling occurs.

**Temporal variability of heat and momentum fluxes**

As expected from greenhouse effect, downward longwave radiation from the ESMs increases steadily over the 21st century under RCP8.5 (Fig.4a). The increase is stronger in IPSL (+10%, see Table 3) and CNRM (10%) than in GFDL (7%). This induces a decrease of the surface ocean cooling associated to net longwave radiation in the RCMs (Fig.1d). Contrasted net downward shortwave radiation trends are simulated by the ESMs (Fig.4b). Insolation decreases quasi-linearly in CNRM (-7%) and in GFDL (-4%), however it is modulated by decadal variability in GFDL (note the slight insolation increase in 2090-2100). On the other hand, IPSL displays no trend (0%). Furthermore, alongshore wind stress, the main driver of coastal upwelling, decreases in R-CNRM (-11%) and R-IPSL (-9%) in contrast with R-GFDL (+2%) (Fig 4c). The wind stress decrease found in R-IPSL and R-CNRM is consistent with that found in CMIP3 simulations (Goubanova et al.2010; Belmadani et al. 2013).

**Coastal upwelling**

Coastal upwelling (measured as the net offshore flux, see section 2.7) decreases strongly in R-IPSL (-23%) and R-CNRM (-25%) (Fig.5a, Table 3). **These downtrends are consistent with the wind stress downtrends (Fig.4c) and mainly due to the Ekman transport contribution (Fig.5c).** In contrast, coastal upwelling remains stable in R-GFDL. The upwelling is modulated by decadal variability, whose amplitude can reach 5-10% of the mean value. Decadal variability may generate decades of upwelling increase (*e.g.* 2090-2100 in R-CNRM) masking the long term decrease. Upwelling decadal variability is mainly forced by variations of the onshore geostrophic transport, **which on average compensates ~50% of the Ekman transport. As Ekman transport decreases over time in R-IPSL and R-CNRM, the relative contribution of the geostrophic transport increases over time.** This onshore current is driven by the higher sea level in the equatorial portion of the upwelling system than in its poleward portion (Colas et al. 2008; Oerder et al., 2015). This flow is occasionally remarkably strong (*e.g.* in 2090 in R-CNRM, 2035-2040 and 2065 in R-GFDL), whereas the trends are weak.

**Subsurface temperature anomalies**

Nearshore subsurface temperature anomalies are impacted by equatorial subsurface temperature anomalies in two ways: thermocline anomalies may propagate along the equatorial and coastal wave guide (*e.g.* Echevin et al., 2011, 2014; Espinoza-Morriberón et al., 2017, 2018), and temperature anomalies may be transported eastward and poleward by the near-equatorial subsurface jets (Fig.2a; Montes et al., 2010, 2011). The latter is particularly strong during eastern Pacific El Nino events (*e.g.* Colas et al., 2008 for the 1997-1998 event). The thermal structure of the upper layer is strongly impacted by climate change in the eastern equatorial Pacific. The depth of the 20°C isotherm (hereafter D20) is used to characterize the thickness of the warm surface layer. It increases in all ESMs, at different rates (Fig.6a). The deepening is roughly linear in GFDL (+5%, Table 3) and CNRM (+26%). In contrast, it increases non-linearly in IPSL, first by ~1.5 m/decade between 2005 and 2065 , and then by ~5 m/decade between 2065 and 2100. Note that D20 is shallower in the ESMs (~30-40 m) than in observations (~52 m in WOA) and in two ocean reanalysis (~56 m in GLORYS2V1 and -58 m in SODA). A shallow thermocline is likely to be more impacted by greenhouse-induced surface warming in the model simulations than in the real ocean.

D20 coastal trends in the RCMs (Fig.6b) are roughly similar to the offshore ESM equatorial trends. The coastal deepening is moderate in R-GFDL (+12%, Table 3). In contrast, a strong linear deepening is found in R-CNRM (+101%). As in the equatorial region, the D20 deepening is non linear in R-IPSL and the thickness of the warm surface layer more than doubles (+207%). The RCM D20 values at the beginning of the century are within the range of estimated values from observations and reanalyses whereas D20 is slightly too deep in the ESMs (Fig.6c), which highlights the dynamical downscaling ability to reduce part of this systematic bias. The RCM trends are roughly in line with the ESM coastal trends. D20 deepening can be amplified (*e.g.* 207% in R-IPSL vs 126% in IPSL) or mitigated (12% in R-GFDL vs ~21% in GFDL, Fig.6c) depending on the model. Decadal variability from the equatorial region propagates to the coastal regions with little change.

We now investigate the evolution of the RCMs **mixed layer. The RCM surface boundary layer thickness (hbl),**

**determined by comparing a bulk Richardson number to a critical value (KPP parameterization; Large et al., 1994), is a good proxy of the model mixed layer (*e.g.* Li and Fox-Kemper, 2017). The R-GFDL mixed layer in 2006-2015 is in fairly good agreement with the mixed layer depth (computed from temperature profiles) from the coarse 2°x2° gridded climatology of de Boyer Montegut et al. (2004), whereas R-IPSL and R-CNRM values are ~ 3 m shallower.**

A shoaling of the mixed layer is found in all simulations (Fig.7), in line with the surface heating (Fig.4a, b) and reduced wind-driven mixing (Fig.4c). The shoaling is slightly stronger in R-IPSL and R-GFDL than in R-CNRM, possibly due to the stronger surface warming in R-IPSL (Table 3).

The near-equatorial subsurface, coastal subsurface and surface temperature linear trends of the RCMs **and ESMs** are compared in Figure 8. Near-equatorial subsurface trends are weakest in GFDL and strongest in IPSL, which is consistent with the stronger D20 deepening in IPSL (Fig.6a). A similar ranking from weakest (R-GFDL) to strongest warming (R-IPSL) is found for the coastal subsurface warming and coastal surface warming. The equatorial water masses are transported towards the coasts (Montes et al., 2010; Oerder et al., 2015) and the subsurface layer trends increase by 6% in R-GFDL, 23% in R-CNRM and 10% in R-IPSL with respect to the near-equatorial trends. **The ESM trends are close to the RCM trends, which suggests that the nearshore subsurface warming is dominated by the eastward transport of warm near-equatorial subsurface waters both in the ESMs and RCMs.** In the coastal region, the upper part of the 50-200m subsurface water volume is upwelled into the mixed layer where additional heat is deposited by the local atmospheric fluxes (Figs. 4a,b).The coastal SST trends increase with respect to the coastal subsurface anomalies (+17% in R-GFDL, +37% in R-CNRM, +44% in R-IPSL), underlining the impact of different local heat fluxes. The amplitude of the ESM SST trend is very close (<10% change) to that of the RCM for R-IPSL and R-CNRM, which is consistent with the spatial patterns of SST change shown in Figs.3c,d. **Interestingly, the R-GFDL SST increase is ~20% weaker than that of GFDL.**

**3.2 Biogeochemical response of the NHCS under RCP8.5 scenario**

We now investigate the impacts of regional climate change on the main biogeochemical characteristics of the NHCS, namely oxygenation, nutrients and productivity.

**OMZ trends in response to the equatorial circulation**

The suboxic (O$_2$< 5 µmol L$^{-1}$, Karstensen et al., 2008) subsurface waters found in the NHCS result from a subtle balance between the eastward and poleward transport of relatively oxygenated waters from the equatorial region into the upwelling region, the ventilation due to mesoscale circulation (Thomsen et al., 2016; Espinoza-Morriberón et al., 2019) and the local oxygen consumption due to the respiration of sinking organic matter. The eastward currents in the offshore equatorial region thus play an important role in the ventilation of the OMZ (Stramma et al., 2008; Montes et al., 2014; Cabré et al., 2015; Shigemistu et al., 2017; Espinoza-Morriberón et al., 2019). Following Cabré et al. (2015) we first evaluate the ESM eastward subsurface flow (which enters the western boundary of the RCM) at 95°W (Fig.9a). As estimates of mean velocity from ocean reanalysis range between 0.05 m s$^{-1}$ (GLORYS12V1) and 0.09 m s$^{-1}$ (SODA), the uncertainty of the eastward flow is very high. The eastward flow in R-GFDL (in 2005-2010) is ~10% weaker than in SODA. In contrast, the

eastward flow is underestimated by ~50% in R-CNRM and R-IPSL with respect to SODA, probably because of a weak EUC and/or weak SSCCs in these coarse-grid ESMs (Cabré et al., 2015). Over 2006-2100, the eastward velocity is stable (<1%, Fig.9a, Table 3) in R-CNRM and decreases weakly in R-IPSL (-9%) and in R-GFDL (-14%).

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The evolution of the eastward dissolved oxygen (DO) flux at 95°W (Fig.9b) follows approximately that of the mass flux. Due to a strong increase in equatorial DO (not shown), the DO flux uptrend is strong in R-CNRM (33%, Table 4). This contrasts with the moderate decrease of the DO flux (~ -5%) in the other two simulations. Note that the DO eastward flux is ~25-30% stronger in R-IPSL than in R-CNRM at the beginning of the century. As the eastward flow in the 2°S-10°S equatorial band is stronger in R-IPSL than in R-CNRM (not shown) and the water is more oxygenated in this latitudinal band

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than within 2°S-2°N (*e.g.* Figure 4 in Cabré et al., 2015), this results in a stronger DO eastward flux in R-IPSL than in R-CNRM.

We now investigate the nearshore subsurface DO concentration in a box located between 150 km and 300 km offshore, in order to take into account a sufficient number of coarse ESM grid points in the 100-200m depth range. The RCM is able to

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represent the cross-shore structure of the OMZ with a fair degree of realism (Figs.S1-2). The OMZ bias are weak (< 10 µmol L$^{-1}$, Fig.S2) below ~100m and increase near ~50-100 m, in the depth range of the oxycline/thermocline. The nearshore DO concentration in the upper part of the OMZ (between 100 and 200m, Fig.10a) in 2006-2015 is slightly higher in R-GFDL (~+20 µmol L$^{-1}$) than in the observations (~15-18 µmol L$^{-1}$) and lower in R-IPSL (~10 µmol L$^{-1}$) and R-CNRM (~-5 µmol L$^{-1}$, see also Fig.S1).

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In contrast, the ESMs strongly overestimate DO in the OMZ (Fig.10b). The eastward flux at 95°W supplies DO to the nearshore OMZ in greater proportions in R-GFDL than in R-IPSL and R-CNRM (Fig.9b), partly explaining the discrepancies at the beginning of the century.

The nearshore trends are very different in the three regional simulations. The DO content is virtually unchanged in R-GFDL (-3%, Table 4) and decreases slowly (-21%) in R-IPSL, whereas it increases strongly in R-CNRM (+483% ~ 30 µmol

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L$^{-1}$ increase). R-GFDL is also marked by a stronger multidecadal variation than the other RCMs. The trends have the same sign as those of the ESMs (Fig.10b), but DO changes are reduced by half in the RCMs (*e.g.* ~+60 µmol L$^{-1}$ in CNRM versus ~+30 µmol L$^{-1}$ in R-CNRM, ~-6 µmol L$^{-1}$ in IPSL versus ~-2.5 µmol L$^{-1}$ in R-IPSL).

The depth of the 0.5 mL L$^{-1}$ (22 µmol L$^{-1}$) DO iso-surface (hereafter named "oxycline") is often used as a proxy for the OMZ upper limit (*e.g.* Espinoza-Morriberón et al., 2019) characterizing the vertical extent of the habitat of many living

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species of the coastal ecosystem (**Bertrand et al., 2010, 2014**). As R-CNRM oxycline is quite deep (Fig.S2), we averaged its values over a wider coastal box (0-200 km) in Figure 10c. The oxycline at the beginning of the century is well positioned in R-GFDL, and slightly shallower than the observed oxycline in R-IPSL and R-CNRM (Fig.10c). Between 2006 and 2100, the oxycline shoals slightly (less than 10 m) in R-GFDL and R-IPSL whereas it deepens of more than 100 m in R-CNRM. Similar trends are found for the "upper oxycline" defined by the 1 mL L$^{-1}$ isoline (not shown, see Table 4).

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**Nitrate trends**

We now investigate the evolution of subsurface nitrate concentrations at 95°W, the western boundary of the RCM (see red line in Fig.2b). A decrease is found in all simulations. This is illustrated by the shoaling of the 21 µmol L$^{-1}$ nitrate iso-

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surface (Fig. 11a). The downtrends vary between strong (78% in R-CNRM) and moderate deepening (24% in R-IPSL and 26% in R-GFDL, Table 4). Nitrate depletion was also found in IPSL CMIP3 4xCO$_2$ scenario (Brochier et al., 2013). It is likely caused by a reduced nutrient delivery from the deep ocean to the upper layers of the ocean associated to enhanced thermal stratification, reduced vertical mixing and overall slowdown of the ocean circulation (*e.g.* Frölicher et al, 2010). Due to the stronger eastward flow in R-GFLD (Fig.9a), the associated nitrate eastward flux is ~50% stronger than in R-IPSL and R-CNRM (Fig. 11b). The fluxes decrease in all simulations (-27% in R-CNRM, -20% in R-IPSL, -18% in R-GFDL, Table 4, Fig.11b).

Following Espinoza-Morriberón et al. (2017), the depth of the 21 µmol L$^{-1}$ nitrate iso-surface (hereafter D21) in the coastal region is chosen as a proxy of the nearshore nitracline depth (Fig. 11c). In spite of **the offshore nitracline deepening (Fig.11a) and** decreasing nitrate flux (Fig.11b), the nearshore nitracline shoals in R-GFDL (-25%). In contrast, it deepens in R-IPSL (+32%) and in R-CNRM (+82%). This shows that the equatorial forcing is not always the main forcing of the evolution of the nearshore nitracline depth: whereas it seems to drive nitrate depletion in R-CNRM and R-IPSL, the maintained coastal upwelling in R-GFDL  (Fig.5a) may partly compensate this effect. It is also notable that the nitracline may shoal even though coastal upwelling does not increase (*e.g.* in R-GFDL, Fig. 5a). This points to potential changes in nitrate vertical distribution, possibly due to a reduction of nitrate assimilation driven by biomass variations (see section 3.3). The **ESMs and RCMs nearshore nitracline trends are consistent for CNRM and IPSL: nitracline deepens by 97% (34%) in CNRM (IPSL) and by 82% (32%) in R-CNRM (R-IPSL). In contrast, nitracline shoaling is strong in R-GFDL (-25%) and negligible in GFDL (+2%).** However, note that D21 is too shallow in RCMs (~20-35 m over 2006-2015) **with respect to observations**  (~100 m in CARS) due to an overly high nitrate concentration in subsurface layers (figure not shown). This bias was also found in previous ROMS-PISCES regional simulations of the NHCS (e.g. see also Fig.3 in Espinoza-Morriberón et al., 2017) possibly due to a lack of denitritification.

**Chlorophyll and  primary productivity annual variations**

Regional downscaling has a strong impact on the nearshore planktonic biomass. Chlorophyll is used in the following as a proxy of total phytoplankton biomass. The surface chlorophyll concentration at the beginning of the century (Fig.12a) agrees relatively well with MODIS mean chlorophyll  (~4.25 mg Chl m$^{-3}$) in R-IPSL (~4.2 mg Chl m$^{-3}$) and R-GFDL (~4.5 mg Chl m$^{-3}$) whereas it is ~30% higher in R-CNRM (~5.5 mg Chl m$^{-3}$). Note that MODIS and SeaWIFs satellite observations differ by ~1 mgChl m$^{-3}$ due to different algorithms (O'Reilly et al., 1998; Letelier and Abbott, 1996) and different time periods (cf section 2.6).  Moderate uptrends are found in R-GFDL (+12%) and R-IPSL (+17%, Table 5). The latter seems at odds with the weak nitracline deepening (<10 m between 2006 and 2100) in R-IPSL (Fig.11c). Strong multidecadal variability with almost no trend (2%) is found in R-CNRM, in spite of the marked nutricline deepening (~20 m, Fig.11c).

**RCMs are able to correct the ESM inability to represent nearshore surface chlorophyll concentration (Fig.12b). Indeed, ESM surface chlorophyll range between ~0.6-0.7 mgChl m$^{-3}$ (GFDL)  and ~0.01-0.1 mgChl m$^{-3}$ (CNRM), almost an order of magnitude smaller than observed values.** The ESM trends display very contrasted patterns (Fig.12b). Surface chlorophyll concentration  decreases in all cases, with negative trends between -11% and -104%,  a behavior not simulated in the RCMs.

430    The total chlorophyll content, depth-integrated over 0-500m (which includes the euphotic layer) (Fig.12c), displays weak uptrends in R-IPSL (+2%, Table 5) and R-GFDL (+3%) and a moderate decrease in R-CNRM (-5%). **Note also the very marked multidecadal variability in R-CNRM**. In contrast, weak downtrends (-3%) are found in two of the ESMs (IPSL and GFDL, Fig.12d). Note that the R-CNRM downtrend (-5%) is weaker (-8%) with respect to CNRM (-32%).

    The different evolution of the RCM surface and total chlorophyll content implies that the vertical distribution of
435    phytoplankton biomass is modified in the long term. The vertical and cross-shore structure of seasonal chlorophyll trends indicates that both R-GFDL and R-IPSL simulate a chlorophyll increase in the mixed layer near the coast, and a decrease below (Figs.13a-c). Interestingly, this suggest that total biomass changes cannot be monitored using satellite measurements, as the subsurface plankton depletion cannot be observed. **The seasonal trends in R-GFDL and R-IPSL are consistent with a shoaling of the mixed layer (Fig.7), which reduces light limitation of phytoplankton growth (*e.g.* Echevin et al., 2008;**
440    **Espinoza-Morriberón et al., 2017) and increases surface primary productivity in summer and winter. In contrast, the R-CNRM trend in the mixed layer is negative in summer. This is likely caused by the strong deepening of the nitracline in R-CNRM (Fig.11c) and the seasonality of the wind-driven upwelling. As the upward flow is weaker in summer, the upwelling of less rich waters into the mixed layer may trigger a nutrient limitation of phytoplankton growth. On the other hand, as the upward flow remains strong during winter, nutrient limitation does not occur. Light**
445    **limitation of phytoplankton growth reduces because of the shoaling of the mixed layer, enhancing phytoplankton growth (as in the two other RCMs). Moreover, visual correlation between decadal variability of the chlorophyll content and nitracline depth in R-CNRM (*e.g.* the oscillations in 2070-2100 in Fig.11c and Fig.12c) also suggests that nitrate limitation of phytoplankton growth may play a role.**

    To further investigate the drivers of the surface chlorophyll trends, RCM and ESM primary productivity (PP) trends are
450    shown in Fig.14. RCM PP surface trends are weak (between -2% and +7%). In particular, the weak trend in R-IPSL (-2%) is at odds with the surface chlorophyll increase (+17%, Fig.12a). In all RCMs, PP is strongly impacted by decadal variability as a consequence of upwelling (Fig.5a) and nitracline depth variability (Fig.11c). These surface trends contrast with the more pronounced ESM PP trends, in particular for IPSL  (-25%) and CNRM (-113%). However, one may question the meaning of the ESM PP trends associated with very weak (an irrealistic) ESM chlorophyll concentrations (Figs.12b,d). The RCM depth-
455    integrated PP trends are consistent with to those of surface PP but differ from the ESMs, especially for R-CNRM (-7%) and CNRM (-66%).

    Overall, the contrasted trends found in the RCMs and ESMs, even when a similar biogeochemical model is used (*e.g.* PISCES in IPSL and CNRM), illustrate the necessity to regionally downscale ESM variability to reduce systematic bias and better represent local processes impacting on productivity.

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**Zooplankton biomass variations**

    The two zooplankton groups represented by RCMs are aggregated in a single group to allow a comparison with the ESMs. In contrast with surface phytoplankton, the order of magnitude of surface zooplankton biomass is comparable in EMSs and RCMs, with the exception of CNRM in which zooplankton concentrations are very weak. Besides, RCM surface
465    zooplankton also displays a different evolution than RCM phytoplankton. First, multidecadal variability is quite strong and

trends are weak. Zooplankton slightly accumulates in R-GFDL (+4%, Fig.15a, Table 5), in line with phytoplankton (+12%, Fig.12a), suggesting the possibility of a grazing increase. In contrast, surface zooplankton displays no trend in R-IPSL in spite of a marked surface phytoplankton increase (+17%). These weak surface zooplankton trends contrast with the stronger ESM downtrends (from -15% (GFDL) to -98% (CNRM), Fig.15b).

470       Depth-integrated zooplankton biomass decreases moderately in all RCMs, from -5% (R-GFDL) to -15% (R-IPSL) (Fig.15c). The GFDL and IPSL depth-integrated zooplankton downtrends are relatively close to the RCM downtrends. CNRM stands out as atypical with a decrease of half of its zooplankton biomass, while the decrease in R-CNRM is moderate (-11%). The spatial structure of the trends varies significantly over the vertical and in the cross-shore direction (Figs.16a-c). The accumulation of zooplankton in R-IPSL and R-CNRM near the coast is consistent with a reduction of the offshore

475 advection due to Ekman transport (Fig.5c). As for chlorophyll (Fig.13), the zooplankton decrease below 10 m depth suggests that monitoring of zooplankton must be carried out in the surface layer and below to measure long-term trends.


## 4 Summary and discussion

### 4.1 Summary of the main results

480       The dynamical downscaling of the ocean circulation and ecosystem functioning for three ESMs is performed in the NHCS for the strongly warming, so-called "worst-case" RCP8.5 climate scenario. The RCM simulations all show an intense warming of the surface layer within 100 km from the Peruvian coasts, reaching between +2°C and +4.5°C in 2100. We can speculate that the nearshore surface warming is closely associated with a subsurface warming in the near-equatorial region (95°W, 2°N-10°S) which propagates into the NHCS. The coastal warming is weakest when the wind-driven upwelling is

485 maintained (*e.g.* in R-GFDL), and strongest when it is reduced (*e.g.* in R-IPSL and R-CNRM, see also Echevin et al., 2012; Oerder et al., 2015). The coastal warming found in the RCMs is close to that found in the ESMs, but surface and subsurface temperature mean biases (for the period 2006-2015) are greatly reduced in the RCMs.

      Biogeochemical trends from the RCMs and ESMs are compared. Two of the three RCMs display a weak decrease of the near-equatorial (95°W, 2°N-10°S) eastward oxygen flux into the NHCS, associated with a moderate slowdown of the

490 eastward equatorial circulation and weak changes in oxygen concentrations in the equatorial region. Consequently, a relatively weak deoxygenation occurs in the nearshore region. This contrasts with the third RCM, in which the near-equatorial region becomes very oxygenated, which triggers a strong oxygenation of the OMZ.

      Nutrient supply from the near-equatorial region to the NHCS decreases in all RCMs due to progressive nitrate depletion of equatorial waters and to decreasing eastward flux. This drives a deepening of the nearshore nitracline in two of

495 the RCMs, and a shoaling in the third RCM in which wind-driven coastal upwelling is maintained.

      Chlorophyll concentration displays contrasted coastal trends. First, in all RCMs, surface chlorophyll does not decrease, in contrast with ESM downtrends (from -11% to -104%). Surface chlorophyll increases (>10%) in two RCMs, while the total chlorophyll biomass remains stable, indicating an enhanced vertical stratification of phytoplankton in the surface layer in 2100. Total phytoplanktonic biomass (*i.e* integrated over the water column) in the coastal zone remains

500 relatively stable in spite of a slightly decreasing primary productivity driven by a weakening upwelling (in two RCMs) and a deepening nutricline (in two RCMs). This counterintuitive evolution of surface phytoplankton could be partly driven by the

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reduced offshore transport (related to coastal upwelling) which allows floating organisms to accumulate in the coastal band. Reduced offshore transport may also induce a greater residence time of phytoplankton in the coastal area hence a stronger prey availability favoring grazing and a larger zooplankton biomass. However, the total zooplankton biomass tends to decrease in all RCMs, which shows that complex nonlinear effects (*e.g.* temperature and predator-prey relations) drive plankton trends. Note that RCM zooplankton downtrends can be weaker than the ESM downtrends used to drive fish global models (*e.g.* Tittensor et al. , 2018). In the following subsections we discuss in more details the surface temperature trends, the near-equatorial conditions impacting the NHCS and the impact of the downscaling on the plankton trends.

**4.1 Selection of the ESMs**

**The choice of which ESMs to downscale has been justified on the basis of the comparison of the ESMs historical simulations to climatological observations. We are aware that these evaluations do not necessarily correspond to how well a model may capture the response to future climate forcing. The "Emergent Constraints" approach has been offered as a relevant method for evaluating climate models (*e.g.* Hall et al. 2019). In this approach, a statistical relation (F) between a present state variable (X) and a future state variable (Y) is derived (Y=F(X)) using an ESM ensemble, regardless of ESM bias. The relation is then used to derive a future response using the best knowledge of the present state (X_obs) using Y=F(X_obs). Following such an approach would have been useful to select the ESM models that fit best with the relation F. However, as we are interested in several variables (thermal stratification, upwelling, productivity, OMZ), this would necessitate finding distinct "emergent constraints" for these variables, and thus possibly selecting different ESMs for each constraint, which may be intricate. Such an approach is however promising and should be envisaged in future work.**

**4.2 SST warming**

Enhanced surface heat fluxes and coastal upwelling of offshore-warmed source waters appear to be the main drivers of the nearshore SST evolution. The strongest nearshore warming (+4.5°C in 2100) found in R-IPSL likely results from the superposition of four effects: (i) a stronger warming of subsurface waters in the near-equatorial region subsequently transported towards the coastal region, (ii) a reduced cooling due to a decreasing coastal upwelling driven by the wind relaxation, (iii) a stable shortwave flux and (iv) an increasing downward longwave flux due to the greenhouse effect. Moreover, IPSL-CM5 ranks among the high-sensitivity climate models of CMIP5 due to a large positive low-level clouds feedback (Brient and Bony, 2013). The weaker surface warming in R-CNRM (+3.5°C in 2100) may be mitigated by the weaker insolation. Last, the weakest warming in R-GFDL (+2°C in 2100) can be explained by (i) the weakest offshore subsurface temperature anomalies, (ii) the strongest wind-driven coastal upwelling (which brings deeper colder waters to the surface layer) and (iii) the weakest greenhouse forcing. As upwelling-favorable winds are more likely to decrease than to increase in low-latitude EBUS such as the Peruvian system (Goubanova et al., 2011; Belmadani et al., 2014; Rykacsewski et al., 2015), an upwelling reduction and strong SST warming appears to be the most robust projection. However, a rigorous estimate of the forcing terms in the nearshore heat budget would necessitate the online computation of each term (*e.g.* Echevin et al., 2018).

Warmer surface waters may have severe consequences on the functioning of the Humboldt current ecosystem as a whole (Doney, 2006; Doney et al., 2012). For instance, in spite of the broad temperature range of small pelagic fish species (*e.g.* anchovy, sardine or jack mackerel) habitat (*e.g.* Gutierrez et al., 2008), the temperature anomalies associated with El Niño events may drive the NHCS into conditions detrimental for pelagic recruitment. Moreover, previous modelling studies based on the RCP8.5 scenario suggest that Peruvian fisheries will be impacted by the poleward migration of exploited species to encounter cooler waters (*e.g.* Cheung et al., 2018).

**4.3 Near-equatorial eastward flow and OMZ variability**

Eastward EUC and SSCCs are supposed to be strong drivers of OMZ variability as they transport relatively oxygenated equatorial waters into the OMZ (Cabré et al., 2015; Shigemitsu et al., 2017; Montes et al., 2014; Espinoza-Morriberón, 2019; Busecke et al., 2019). This is in line with our results: in all RCMs, the DO trend in the OMZ is consistent with the trend of the offshore DO eastward flux. The EUC is supposed to be mainly forced by the zonal pressure gradient across the equatorial Pacific, associated to the trade winds and the Walker circulation (hereafter WC; Stommel, 1960). However, most of the CMIP5 climate models fail to reproduce the WC intensification observed in the recent period (1980-2010) (*e.g.* Kociuba and Power, 2015). Furthermore, the EUC decrease in the eastern equatorial Pacific in GFDL and in IPSL (respectively -26% and -22% decrease between 2005 and 2100 for the mean velocity between 2°N and 2°S, 95°W, 50-200 m depth, Figure not shown) is not consistent with the WC trends reported in Kociuba and Power (2015). Note also that EUC trends vary significantly across the equatorial Pacific (Drenkard and Karnauskas, 2014). EUC dynamics are also likely sensitive to stratification changes in the equatorial thermocline (McCreary, 1981). In brief, to our knowledge, the mechanisms driving EUC long-term variability in the eastern equatorial Pacific remain to be investigated.

SSCCs long-term variability, which contributes to the NHCS trends (*e.g.* Montes et al., 2014), is also unknown. At basin scale, the primary SSCC (near 4°-6°S at 90°W) is supposed to be forced partly by trade winds and alongshore winds in the NHCS, by mass exchange between the Pacific basin and the Indian ocean, and by surface heating in the tropics (McCreary et al., 2002; Furue et al., 2007). The problem is that SSCCs are not resolved in CMIP5 models due to coarse resolution (*e.g.* see Fig.4 in Cabré et al., 2015). Last, the observed deoxygenation of water masses in equatorial regions (Stramma et al;, 2008) is underestimated in global models (Oschlies et al., 2018). These uncertainties imply that the ventilation of the NHCS OMZ by the eastward jets may be difficult to project using CMIP5 ESMs.

In order to investigate further the impact of the ESM oxygen conditions on the RCM results, we conducted a series of sensitivity simulations (called R-GCM') using climatological seasonally-varying WOA DO concentrations at the regional model open boundaries. Boundary conditions for all the other biogeochemical variables are unchanged with respect to the reference simulations (RCM) (note that we are aware that this simplification introduces inconsistencies in the biogeochemical properties of the water masses. However the results are worth reporting). As expected, the DO eastward flux (Fig.17a) now follows roughly the mass flux evolution (Fig.9a) and decreases weakly in each simulation. The huge nearshore DO trend previously found in R-CNRM (+483%, Fig.10b) is now much weaker in R-CNRM' (+36%) and of a comparable order of magnitude as the other RCM's (Fig.17b). Furthermore, the marked decrease of the eastward DO flux in R-GFDL' appears to drive a strong nearshore DO decrease. This confirms that strong changes in the near-equatorial eastward ventilation flux

impact the OMZ, in line with previous studies (*e.g.* Shigemitsu et al., 2017). However, ventilation of the OMZ by this
mechanism is not the only driver of oxygen variability. Indeed, nearshore deoxygenation can vary (it is slightly more intense
in R-IPSL than in R-GFDL, Fig.10a) in spite of rather similar decrease of the near-equatorial DO eastward fluxes, possibly
owing to different local physical and biogeochemical processes (and thresholds). Computing a rigorous DO budget in the
coastal region is needed to investigate in more details the local processes at stake.

**4.4 Plankton trends**

A stable and, in one case, increasing concentration of chlorophyll are found in the surface layer (0-5m), in spite of
primary production decrease (*e.g.* in R-CNRM and R-IPSL, Fig.14). Several mechanisms could contribute to partly
compensate the PP decrease.

The shoaling of the mixed layer may constrain phytoplankton vertically and increase surface concentration. The
increased temperature in the near-surface layer (0-50 m depth) induces a faster growth rate of phytoplankton cells (Eppley,
1972). Furthermore, the decrease of upwelling and offshore export (Fig.5) may concentrate more biomass in the coastal
region and contribute to the phytoplankton persistence in R-IPSL and R-CNRM. However, performing a budget of
phytoplankton in the model would be needed to estimate precisely the relative contribution of each process , but this is
beyond the scope of the present study.

Examination of RCM zooplankton biomass shows weak trends (0-4%) in the surface layer and weak downtrends
(between -5% and -15%) for total biomass (Fig.15). R-IPSL and R-GFDL zooplankton biomass decrease faster than
phytoplankton, which corresponds to a trophic attenuation of the transfer of biomass to upper levels. A similar attenuation has
been found in regional simulations of the Benguela upwelling system under the IPCC-AR4 A1B scenario (corresponding to
the more moderate RCP6.0 scenario; Chust et al., 2014). The RCM zooplankton trends also contrast with the ESM
downtrends. These discrepancies can be attributed to local physical processes (transport and mixing associated to the
mesoscale) not represented in the ESMs, but also partly to the use of an earlier version of the ecosystem model (PISCES) run
with a set of biogeochemical parameters adapted for the NHCS (see Table 1 in Echevin et al., 2014). The stronger total
zooplankton biomass downtrends in R-CNRM and R-IPSL suggest a strong impact of the temperature increase, possibly due
to the higher zooplankton mortality in a warmer environment. However, the model's microzooplankton and mesozooplankton
result from a nonlinear interplay of temperature and predation/mortality effects. Further interpretation of these trends would
require dedicated sensitivity experiments and performing a zooplankton budget. This is beyond the scope of the present study
which aims to present an overview of the main low trophic level trends.

*5* **Conclusions and perspectives**

Regional downscaling of three coarse-grid ESMs is performed in the NHCS over the 21$^{st}$ century so-called "worst-
case" RCP8.5 climate scenario using a high-resolution regional coupled biodynamical model. The **downscaling procedure
allows to correct ESM bias**. All regional simulations reproduce an intense warming (2-4.5°C) of the surface layer within 100
km from the Peru coasts. **The surface warming is strongest when the subsurface equatorial warming is strong and the
wind-driven coastal upwelling weakens in the future**. Downscaled  trends are consistent with those obtained from the

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610 ESMs.

The biogeochemical impacts of climate change are more contrasted among **RCMs and ESMs**. A slowdown of the eastward near-equatorial circulation may reduce the ventilation of the NHCS and induce a nearshore deoxygenation trend. However the long-term variability of oxygen content of equatorial water masses also impacts the nearshore oxygen trends. As observed deoxygenation trends in the eastern equatorial Pacific are not well reproduced by ESMs (Stramma et al., 2008, 615 2012) and CMIP5 ESM systematic biases are strong in this region (Cabré et al., 2015; Oschlies et al., 2018), these shortcomings limit the predictability of downscaled oxygen trends in the NHCS. One important conclusion of our study is that reducing the biases in oxygen concentration and zonal circulation trends in the eastern Equatorial Pacific ocean is crucial to project the future evolution of the NHCS oxygen minimum zone.

Downscaled surface chlorophyll in the coastal region does not decrease, in contrast with the signal projected by the 620 ESMs. In two RCMs, the surface chlorophyll remains high in the coastal region. We can speculate that this happens for two reasons: the enhanced thermal stratification due to the warming may alleviate light limitation and vertical dilution, and the reduction of wind-driven offshore transport may allow plankton to accumulate near the coast. These processes could partly compensate the reduction of primary productivity due to a deeper nitracline and reduced wind-driven coastal upwelling. Downscaled zooplankton downtrends are also relatively weak (between -5% and -15%) but appear to strengthen when the 625 warming is stronger. In all RCMs, downscaled plankton trends differ markedly from those simulated by ESMs, in particular in the surface layer  (0-5m), which illustrates the strong impact of the regional dynamical downscaling. This also underlines the necessity to interpret ESM biomass-based regional projections of fisheries (*e.g.* FISHMIP, Tittensor et al., 2018) with great caution.

As previous works point to a relaxation of upwelling-favorable wind conditions in the NHCS (*e.g.* Belmadani et al., 630 2014), dynamically downscaled wind projections as well as more realistic large scale dynamical and biogeochemical conditions in the near-equatorial regions are needed to improve the robustness of our results in future studies. Furthermore, many aspects of the regional impact of climate change have not been explored, such as for example interannual variability associated with ENSO in a warmer NHCS or the acidification of coastal waters. These impacts will be addressed in future studies.

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