

**“Modulation of the North Atlantic Deoxygenation by The Slowdown of the
Nutrient Stream”**

by F. Tagklis, T. Ito, A. Bracco

*In black the original comments by the editor, followed by our responses in blue.

Associate Editor Decision: Reconsider after major revisions (17 Jul 2019) by Andreas Oschlies:

Comments to the Author:

The manuscript is a nice overview over northern hemisphere upper ocean deoxygenation as simulated by a few CMIP5 models. It concentrates on the different declines in the upper North Atlantic (small) and North Pacific (large). The slowdown of the North Atlantic nutrient stream is suggested to explain the different deoxygenation rates in the two basins. This provides an interesting story, which I find plausible and worthwhile of eventual publication. However, all arguments are based on relatively vague qualitative descriptions of basin scale patterns of simulated changes rather than being mechanistically 'confirmed' (line 199) or quantitatively estimated as providing 'most' of the simulated deoxygenation (line 222). There is no mentioning of the possible role of other processes (e.g. changes in up- or downwelling, subpolar gyre strength, fresh water supply and stratification, carbon export by vertical mixing (including deep convection in the N.Atlantic) of particulate or organic matter - are these included in the diagnosed export production of Fig. 9?).

Two expert reviewers have provided useful comments and thoughtful criticism, which I find valid. Unfortunately, the current responses of the authors fail short of satisfactorily addressing those points. Saying that the required model fields are unfortunately not available is not an appropriate scientific reasoning (it would only say that the study was ill-designed - which, I think, is not the case). One could, for example, use one model for which the required temporal resolution output is available and estimate the errors made when considering annual mean tracer fields only. I therefore would like to ask the authors to carefully address all the points raised by the reviewers and also discuss possible other mechanisms as well. I expect that this will become a scientifically very useful contribution and look forward to seeing a revised version.

Many thanks and best regards, -Andreas

Response to Editor's Comments

We much appreciate your time assessing the manuscript and the review comments. Most of the issues raised by the two expert reviewers were addressed in the revised manuscript and the following response letter.

We want to briefly outline the main changes:

- Three more models were added in the analysis.
- Figures and table were updated accordingly
- Table 1: We changed the northern boundary from 50°N to 48°N, thus some minor changes in the numbers.
- We provide additional analysis with a more quantitative character on the mechanisms in the last section of the manuscript, leading to the likely causes of the reduction in AOU of the subtropical North Atlantic. More details are given in the Response to Rev. 2.
- We provide additional analysis and discussion on the possible role of other processes in controlling the AOU of the Atlantic Basin. More details are provided in the Response to Rev. 2

Please see below a more detailed response to reviewer's comments.

Best regards,

On behalf of co-authors,

Filippos Tagklis

Interactive comment on:

**“Modulation of the North Atlantic Deoxygenation by The Slowdown of the
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Anonymous Referee #1:

This study presents an analysis of output of dissolved O₂, PO₄, temperature, advective velocities, and sinking POC flux from four ESMs in the CMIP5 ensemble to investigate the drivers of upper ocean deoxygenation, comparing the North Atlantic to the North Pacific. The compensating effects of the temperature-controlled decrease in O₂ solubility, ocean circulation/ventilation effects, and changes in biological O₂ utilization are examined and attributed to identified trends in basin O₂ content in the ESM outputs comparing the 1970-2000 period with the predicted 2070-2100 period under RCP 8.5 forcing. A contrasting pattern between the North Atlantic and North Pacific is identified, with deoxygenation proceeding more rapidly in the Pacific despite a smaller temperature increase in that basin. Solubility driven deoxygenation in the North Atlantic is revealed to be compensated for by a mechanism rooted in the slowdown of the AMOC, which slows the important Gulf Stream nutrient stream, decreasing lateral injection of nutrients into the subtropical North Atlantic, decreasing export production and concomitant biological O₂ utilization in the subsurface. Deoxygenation in the North Pacific is revealed to proceed via dual solubility and change in ventilation controls. This is a well-written, clear, and concise manuscript detailing the authors study. I only have some comments regarding the Methods.

We appreciate the positive comments on this manuscript. We have addressed your questions related to the Methods below.

Line 81-83: What method was used to interpolate the model output to the common WOA grid?

The method used was bilinear interpolation using Climate Data Operators. This information has been added to the text.

Line 152-160: Why not perform a similar t-test as performed for O₂ to test for statistical significance of the temperature increases identified?

Following reviewer's suggestion, we performed a similar t-test for temperature and apparent oxygen utilization in figures 4 and 6 in the main text.

Why were these 4 models chosen and not others from the CMIP5 ensemble? The given reason is that these 4 provided the variables of interest. Surely more than 4 CMIP5 models provide output of O₂, PO₄, temperature, advective velocities, and POC sinking flux for the historical and RCP8.5 cases?

We added three more models in the analysis. Now, the subset consists of 7 models in total. We looked into two more models, IPSL-CM5B-LR which appears to have significant errors, and HADGEM-CC which during the evaluation appeared to have very low values of nutrient in the Atlantic basin, but most importantly the monthly maximum ocean mixed layer depth variable ‘omlmax’ was not available in any of the ESGF-CoG nodes. This variable was used in the additional analysis. We decided to exclude both.

Table 1: Why not include a column for the multi-model mean to be congruent with that which is provided in Figures 1-8?

Following the reviewer’s suggestion, we included the suggested column to the table to be congruent with the rest of the presentation in the manuscript. The table has been updated with the additional models.

Line 178: “The rate of solubility change ranges from -12.07 μM to -14.81 μM ” A rate implies a change with respect to another quantity or dimension, in this case time. I recommend switching these numbers to -0.12 $\mu\text{M}/\text{yr}$ or mention the rate in parenthesis, etc.

We want to thank the reviewer for pointing this out. Indeed, the rate refers to a centennial time scale. We clarified in the revision. Switching these numbers to an annual rate would imply a linear trend, which is not justified in the current analysis. We mention in all figures and table that changes are calculated as differences between two 30 years-periods 1970-2000 and 2070-2100.

Interactive comment on:

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Anonymous Referee #2:

The study presents an analysis of phosphate, dissolved oxygen, temperature, apparent oxygen and velocity changes in the upper ocean for the northern hemisphere from 4 different CMIP5 Earth system model projections. A plausible narrative is presented as to how changes in the dissolved oxygen in the northern basins are controlled in terms of overturning and temperature changes. There is a clear asymmetry between the North Atlantic and North Pacific with a weakening of the meridional overturning in the former basin. Taking that view forward, the narrative is engaging in terms of discussing the likely changes in terms of a weakening of the western boundary currents and the associated nutrient stream. However, the study is lacking in providing supporting quantitative analysis to endorse the above interpretations as to how the dissolved oxygen and nutrient distributions are controlled. There are no estimates of the nitrate flux carried by the western boundary currents and no estimates of nitrate transports along sections running across the basin. There is no real proof that either the nutrient transport change or the temperature changes are controlling the dissolved oxygen changes. Thus, a very plausible set of interpretations are presented that need to be made more quantitative and so become more robust and convincing.

I recommend that the authors address the following points:

- 1. What is the northward transport of nutrients in the nutrient stream and by how much has that weakened?**
- 2. The nutrient stream is providing a redistribution of nutrients, but it is unclear whether the changes in this redistribution is confined within the subtropical gyre/subpolar gyre of the North Atlantic?**
- 3. Alternatively, the changes in the overturning drives a weakening in the change in nutrient transport across the equator from the South to the North Atlantic.**

For points 2 & 3, see Palter and Lozier (2008) supporting a more confined basin view and Sarmiento et al. (2004) and Williams et al. (2006) supporting a cross basin view. Both processes could be occurring to different extents.

While I agree with the view that the horizontal transport and redistribution of nutrients is probably key to the response, there needs to be mention of the implied changes in the vertical transfer of nutrients and oxygen within the basins.

A minor point, there are a lot of abbreviations that could be removed to make the text more readable.

In summary, the manuscript presents an engaging view of how the nutrients and dissolved oxygen distributions are controlled in the Earth system model projections. However, the authors need to provide more quantitative evidence to support their interpretations, rather than rely on changes in property maps.

We appreciate reviewer's comments. We have tried to address and clarify the suggested points as follows.

Regarding points 1, 2, and 3:

We estimated changes of the northward supply of phosphate at 10°N along with nutrient inventory of the subtropical gyre. In the new Figure 10 of the revised manuscript, time series represent the zonally and vertically integrated northward transport of phosphate ($\overline{vPO_4}$) at 10°N over the 0-700 meters depth range, decomposed in the overturning ($MO = \overline{vPO_4}$) and gyre ($GY = \overline{v'PO_4}$) components, along with the nutrient inventory (NI) zonally, meridionally and depth-integrated over 10°N-48°N and 0-700 meters. The overbar indicates the zonal mean, and the primes indicate the departure from the zonal mean. For better comparison, we applied a low pass filter of 10-years, and then we normalised the time series by subtracting their mean and dividing by their standard deviation. The coloured values [Figure 10] represent the per cent centennial change of each transport component and nutrient inventory. The subtropical gyre nutrient inventory closely follows the declining trajectory of the overturning component of the northward nutrient transport at 10°N for all models but IPSL-CM5A-LR.

Regarding the vertical transport of nutrients in the euphotic layer, we additionally explored the changes in the vertical entrainment of thermocline nutrients. Results suggest that different process are at play in the subpolar versus subtropical the gyre, which we describe in the manuscript (line 243, Figure 10).

Regarding the comment “There is no real proof that either the nutrient transport change or the temperature changes are controlling the dissolved oxygen changes.”. To convince the reviewer about the role of temperature on oxygen changes we provide here the centennial changes of the O₂ saturation ($\Delta O_{2,sol}$) which is inversely proportional to the temperature changes, but we decided to exclude this figure as repetitive. Indeed, it does not provide more information than the ΔT maps as shown below in figure S1 to be compared to figure 4 in the main text.

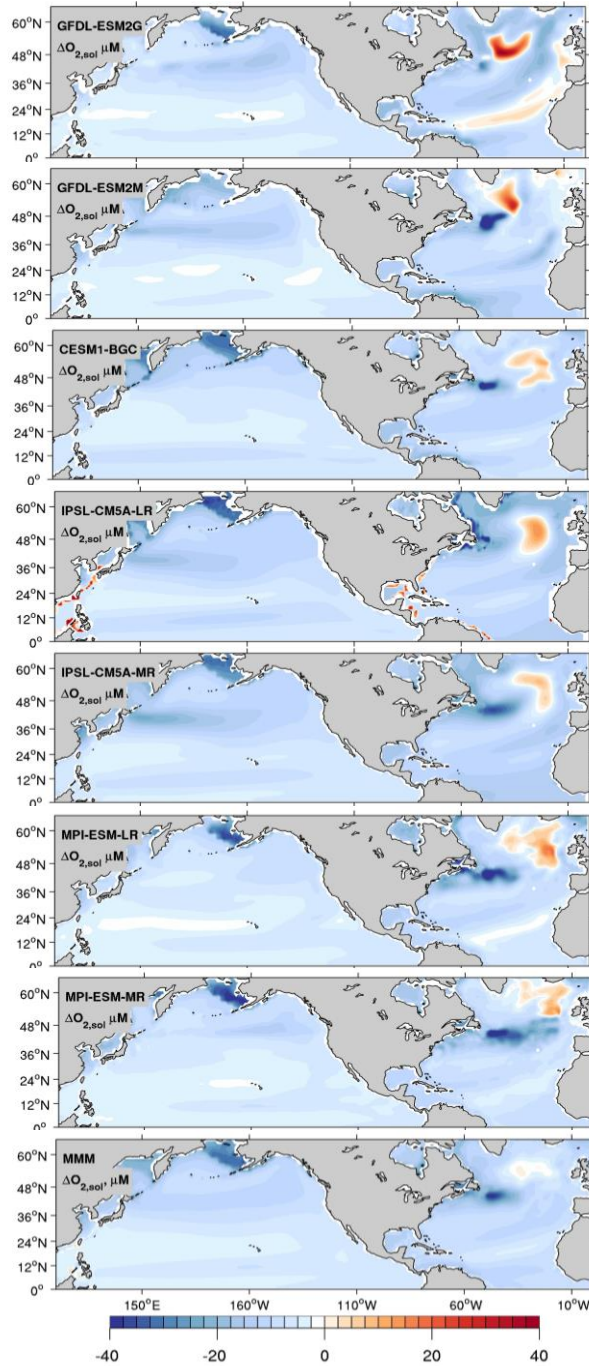


Figure S1 Centennial change of oxygen saturation calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages.

Modulation of the North Atlantic Deoxygenation by The Slowdown of the Nutrient Stream

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Abstract. Western boundary currents act as transport pathways for nutrient-rich waters from low to high latitudes (nutrient streams) and are responsible for maintaining mid- and high-latitude productivity in the North Atlantic and North Pacific. This study investigates the centennial oxygen (O₂) and nutrient changes over the Northern Hemisphere in the context of the projected warming and general weakening of the Atlantic Meridional Overturning Circulation (AMOC) in a subset of Earth System Models included in the CMIP5 catalogue. In all models examined, the Atlantic warms faster than the Pacific Ocean, resulting in a greater basin-scale solubility decrease. However, this thermodynamic tendency is compensated by changes in the biologically-driven O₂ consumption which dominates the overall O₂ budget. These changes are linked to the slow-down of the nutrient stream in this basin, in response to the AMOC weakening. The North Atlantic resists the warming-induced deoxygenation due to the weakened biological carbon export and remineralization, leading to higher O₂ levels. On the contrary, the projected nutrient stream and macro-nutrient inventory in the North Pacific remain nearly unchanged.

20 **Introduction**

Deoxygenation of the oceans is potentially one of the most severe ecosystem stressors resulting from global warming given the high sensitivity of dissolved oxygen to ocean temperatures. Unrestrained anthropogenic CO₂ emissions and consequent warming are likely to disrupt marine habitats and influence the cycles of many biogeochemically essential elements (Gruber, 2011). Global-scale deoxygenation has taken place during the second half of the 20th century (Stramma et al., 2008), and a widespread recognisable signal of O₂ decline is emerging beyond the envelope of natural variability (Schmidtko et al., 2017; Ito et al., 2017). The Earth Systems Models (EaSMs) included in the CMIP5 (Coupled Model

Intercomparison Project – Phase 5) catalog project a robust (across models) decline in dissolved O₂ inventory for the 21st century despite the differences in models’ complexity, biogeochemical parameterizations and warming responses. Under the “business as usual” scenario, all models predict enhanced hypoxic conditions and dissolved oxygen loss (Bopp et al., 2013;Cocco et al., 2013).

The dissolved oxygen is controlled by air-sea exchange, circulation, and biology, and the dissolved oxygen concentrations in the interior ocean reflect a balance between ventilation, circulation and biological consumption. Warming climate can cause shifts in this balance. The solubility of dissolved oxygen is inversely proportional to seawater temperature, and air-sea O₂ exchange is a relatively fast process in the ice-free open ocean, of the order of $O(20$ days) (Broecker and Peng, 1974;Wanninkhof, 1992). All else unchanged, in a warming climate there would be a corresponding O₂ decline closely following the temperature-solubility relationship of seawater (Najjar and Keeling, 1997). However, changes in ocean stratification, ventilation and biological productivity can further change dissolved oxygen. During the transient trajectory of the climate system as it adjusts to anthropogenic forcing, near-surface waters warm faster than deeper waters, leading to an increase in ocean stratification. In a more stratified ocean, the ventilation of sub-surface waters diminishes, reducing the O₂ supply to the ocean interior (Bopp et al., 2002;Frölicher et al., 2009). Furthermore, increased stratification is expected to weaken the meridional overturning circulation and therefore the ventilation of the waters deeper than 1000 m (Meehl and Stocker, 2007). At the same time, the weakening of the overturning circulation may decrease the overall vertical mixing and therefore the supply of nutrient-rich waters to the euphotic layer, thus causing a reduction in biological productivity and carbon export. As upwelling becomes less effective in uplifting nutrient-rich waters, export production of organic material and oxygen consumption through respiration also diminishes, but as water parcels spend more time in the ocean interior, the oxygen consumption integrated over time may increase (Rykaczewski and Dunne, 2010).

Western Boundary Currents (WBCs) plays an essential role in biogeochemical cycling. In the northern hemisphere, WBCs represent an advection pathway for nutrients from the ocean boundaries into the open waters. They are known as “nutrient streams” and are responsible for maintaining basin-scale high productivity in the mid- and high-latitudes over interannual and longer timescales (Letscher et al.,

55 2016;Palter et al., 2005;Williams et al., 2011;Williams et al., 2006). High nutrient concentrations extend from tropical coastal areas into the interior of the Pacific and Atlantic Oceans, following the Kuroshio Current and the Gulf Stream (Pelegrí and Csanady, 1991). From a dynamical perspective, recent studies have shown that the nutrient supply due to the lateral transport in the subtropical euphotic zone dominates over the vertical transport (Letscher et al., 2016), with mean and eddy horizontal cross-boundary nutrient
60 transport accounting for ~75% of the total nutrient supply into the subtropical gyres (Yamamoto et al., 2018). Therefore, changes in this horizontal nutrient transport, through changes in the WBC characteristics, can have a profound influence on the basin-scale biogeochemical cycling.

The primary objective of this study is to investigate how and why the dissolved oxygen content of the North Atlantic and the North Pacific basins is projected to change in the 21st century using a suite of
65 EaSM integrations. In particular, we aim at understanding and quantifying the role of the nutrient streams in the centennial scale deoxygenation and nutrient loading of these two basins. We first verify the EaSMs' skill in reproducing the mean state of relevant biogeochemical variables and then analyze the model projections to the end of the 21st century.

Data and Methods

70 For this study, we analyse seven CMIP5 EaSMs for which the variables of interest are available. The suite includes two versions of the Geophysical Fluid Dynamics Laboratory (GFDL) Earth System Model, GFDL-ESM2G and GFDL-EASM-2M (Dunne et al., 2013;Dunne et al., 2012), the Community Earth System Model, CESM1-BGC (Long et al., 2013;Moore et al., 2013a, b), two of the Institute Pierre Simon Laplace model, IPSL-CM5A-LR and IPSL-CM5A-MR (Dufresne et al., 2013), and two of the
75 Max Plank Institute model, MPI-ESM-LR and MPI-ESM-MR(Giorgetta et al., 2013a;Giorgetta et al., 2013b). The EaSMs vary regarding the parameterisations of the ocean circulation and biogeochemical modules, but the biogeochemical component in all cases is formulated as Nutrient-Phytoplankton-Zooplankton-Detritus (NPZD) type. For each member, we examine the last 30 years (1970-2000) of the twentieth century in the historical simulations and the last 30 years (2070-2100) of the twenty-first century

80 under the future projections based on the Representative Concentration Pathway 8.5 scenario or “rcp8.5”
(Riahi et al., 2011a; Riahi et al., 2011b; Taylor et al., 2012).

All the variables used in the CMIP5 analysis are three-dimensional and annually averaged fields interpolated onto a common $1^\circ \times 1^\circ$ longitude-latitude grid domain and 33 depth levels, consistent with the World Ocean Atlas. The interpolation was done with a bilinear interpolation using Climate Data
85 Operators. The variables of interest are dissolved oxygen (O_2), temperature (T), phosphate (PO_4), particulate organic carbon export at 100m depth (EP) and current speed ($CS = V_{CS} = (\sqrt{u^2 + v^2})$) in units of meters per second. Oxygen solubility ($O_{2,sat}$) is calculated from potential temperature and salinity following Garcia and Gordon (1992). Apparent oxygen utilisation (AOU) is then determined as the difference between the $O_{2,sat}$ and O_2 ($AOU = O_{2,sat} - O_2$). AOU changes quantify contributions from
90 processes other than warming, such as remineralisation of organic matter and/or the rate of transport and mixing of water mass (Sarmiento and Gruber, 2006). The separation of oxygen changes ΔO_2 into a biologically/transport-driven component, $\Delta(AOU)$, and a thermodynamically-driven component, $\Delta O_{2,sat}$, is based on the assumption that the surface oxygen is always in equilibrium with the overlying atmosphere. However, intense air-sea interactions during wintertime at the high latitudes often cause
95 under-saturated surface O_2 , leading to a non-negligible preformed AOU (Ito et al., 2004). Unfortunately, stored variables in the model outputs do not allow a more precise estimation.

It has been shown that in the CMIP5-EaSMs the biogeochemical tracers are not always equilibrated with respect to the ocean circulation. To account for the magnitude and sign of this model drift, in all analyses we used the pre-Industrial Control simulations (piControl) and removed the drift by
100 defining, for example, $O_{2,trend} = \{O_2^{rcp8.5(B)} - O_2^{hist(A)}\} - \{O_2^{piControl(B)} - O_2^{piControl(A)}\}$ where A and B indicate the periods 1970-2000 and 2070-2100.

Results

Model Evaluation

We first evaluate the model representation of the distributions of key biogeochemical variables, including
105 PO_4 , O_2 and AOU. We focus on the Northern Hemisphere ($10^\circ N$ - $65^\circ N$) and concentrate on the upper layer

of the ocean (depth range 0-700 m). The CMIP5 climatological values are calculated over the period 1970-2000 in the “esmHistorical” experiments. Annual mean climatologies from the World Ocean Atlas 2009 (WOA09) (Locarnini et al., 2010; Garcia et al., 2010; Antonov et al., 2010) are used as an observational reference. Note that in Figures 1-3 the Pacific and Atlantic basins are plotted in separate panels with different color scales because of the large differences in their mean values.

The observed PO_4 concentrations [Figure 1] range from $\sim 0.8\mu\text{M}$ in the subtropical North Pacific (STNP) gyre to values greater than $2.7\mu\text{M}$ in the subpolar North Pacific (SPNP) gyre and the eastern boundary and equatorial upwelling region at the lower latitudes. The EaSM are broadly in agreement over the North Pacific regarding the PO_4 spatial gradients, with the exception of CESM1-BGC that underestimates the latitudinal differences with higher nutrient levels overall. In all models, there is a slight underestimation of PO_4 in the subpolar region, that is reflected in the multi-model mean (MMM) where values are about $\sim 0.3\mu\text{M}$ smaller than in the WOA09. In the North Atlantic, the observed concentrations range from $\sim 0.2\mu\text{M}$ in the subtropical (STNA) gyre to $\sim 1.15\mu\text{M}$ in the subpolar (SPNA) gyre. In contrast to the Pacific ocean, there are significant model-to-model differences in the PO_4 spatial pattern. All models but IPSL-CM5A-LR overestimate the concentrations of PO_4 , with CESM1-BGC displaying the largest bias, followed by GFDL-ESM2M.

The simulated pattern of dissolved oxygen is better captured than PO_4 by each model individually and therefore by the MMM, especially in the Atlantic basin. In the Pacific ocean, the observed dissolved oxygen concentrations range from $\sim 160\mu\text{M}$ in the STNP gyre to $\sim 50\mu\text{M}$ in the SPNP gyre. GFDL-ESM2M, IPSL-CM5A-LR and IPSL-CM5A-MR overestimate dissolved oxygen in the STNP by $\sim 35\mu\text{M}$ and CESM1-BGC and MPI-ESM-MR underestimate oxygen concentration in the same area. The end result is a MMM that compares relatively well to WOA09 due to the compensating biases. In the North Atlantic the concentrations of dissolved oxygen range from $\sim 180\mu\text{M}$ in the STNA gyre to $\sim 340\mu\text{M}$ in the western SPNA and ventilation sites. The latitudinal gradient reflects both the temperature gradient and the presence of well-mixed and ventilated cold subpolar waters.

In terms of AOU, the CMIP5-ESMs integrations capture the observed climatological distribution with more robust (across models) patterns in the Atlantic region [Figure 3]. In the Pacific Ocean, the

AOU concentrations range from ~30μM in the STNP gyre to ~250μM in the SPNP gyre. The overall higher values of AOU in the Pacific compared to the Atlantic basin, are due to the older age of the waters and the limited physical O₂ supply to intermediate and deep waters. In the Atlantic Ocean, low AOU values are found in the SPNA as convection and deep water formation decrease the AOU in this region. The narrow band of higher AOU values around ~60μM that extends from the tropics to the east into the basin following the Gulf Stream and the North Atlantic Current (NAC) pathway is captured by all models with different intensity, and is present in the MMM, even if slightly weaker than observed due to biases in the representation of the Gulf Stream separation and NAC location. **The skill of each model in capturing the mean nutrient concentration [Figure 1] is also reflected in the intensity of AOU [Figure 3]. For example, CESM1-BGC as the one extreme in the North Atlantic, overestimating (>1.2μM) the nutrient concentration, shows the highest (>70μM) AOU values, while both IPSL versions underestimate the nutrient concentrations and show low AOU values (<70μM).**

145 Centennial Changes

We next examine hemispheric centennial changes of the physical and biogeochemical variables in the North Pacific and North Atlantic oceans. Changes are calculated as the differences between the 30-year period 2070-2100 in the rcp8.5 scenario and 1970-2000 in the historical simulations. We choose to use 30-year periods to ensure that year to year changes are mostly averaged out. **For O₂, T and AOU we also verify the statistical significance of the drift-corrected trends by testing if the average concentrations during 2070-2100 under the rcp8.5 scenario are significantly lower than those during 1970-2000 period relative to the interannual variability within each 30-year period.** We did so using a t-test and evaluating

$$t = \frac{-\{(\overline{x}_{rcp8.5} - \overline{x}_{his}) - \Delta x_{piControl}\}}{\sigma \sqrt{\frac{1}{N_1} + \frac{1}{N_2}}} \quad \text{where } \sigma \text{ is defined as } \sqrt{\frac{N_1 s_1^2 + N_2 s_2^2}{N_1 + N_2 - 2}}, \text{ and the degree of freedom is}$$

d.f.=N₁+N₂-2. In our case, the number of records in each sample set is the same N=N₁=N₂=30 and s₁, s₂ the corresponding sample variance. Preindustrial control simulations are used to correct for the model drift as mentioned earlier.

Under the rcp8.5 scenario, both basins warm by 0.5 - 4°C [Figure 4], and the warming is generally stronger in the Atlantic than in the Pacific. A localised patch of cooling stands out in the SPNA in all

models but in different locations. This patch is known as “warming hole” (Drijfhout et al., 2012; 160 Rahmstorf et al., 2015a,b) and is a response to the reduced poleward transport of heat due to the AMOC slowdown, which is common to all models (Tagklis et al., 2017). The location of the warming hole depends on each model representation of the NAC pathway. Despite the presence of this cold patch, basin-scale averages between 10°N-48°N, shown in Table 1, reveal that the North Atlantic takes up more heat than the Pacific, and warms on average $\Delta T \sim 1^\circ\text{C}$ more than the Pacific. This mean difference is consistent 165 across the models. Additionally, in the Atlantic the pattern of the warming is consistent among the models, with stronger warming at the gyre boundaries, both at the tropical-subtropical and subtropical-subpolar boundaries.

Even though the Atlantic ocean is warming faster than the Pacific, the centennial changes of O_2 in Figure 5 reveal a more moderate deoxygenation rate in the Atlantic compared to the Pacific. The trends 170 shown in the figure are statistically significant nearly everywhere, according to a t-test at the 99% confidence level. The oxygen trend in the Atlantic is “patchy” with the subtropics resisting to deoxygenation especially in correspondence of the Gulf Stream/NAC paths (Tagklis et al., 2017). The subpolar regions offshore Newfoundland and Labrador, on the other hand, lose the most oxygen in this basin, in correspondence with the largest warming signal. The basin scale averages in Table 1 confirm 175 that the Atlantic Ocean is losing oxygen at a lower rate than the Pacific in all seven models. The modelled basin averaged O_2 changes are in the range between -3.4 and -12 μM but mostly in the -6 μM range in the Atlantic and between -10 to -18.1 μM in the Pacific. This corresponds to O_2 decline of 3% in the Atlantic and 10% in the Pacific compared to their 1970-2000 mean state.

The inverse proportionality of the solubility of oxygen to seawater temperature implies that 180 negative/positive changes in temperature are reflected as positive/negative changes in oxygen solubility $\Delta\text{O}_{2,\text{sat}}$. In thermocline waters, a temperature change by 1°C causes a solubility decrease of about $7\mu\text{M}$. Given the modeled warming trends, oxygen solubility decreases in both basins for all seven models, except for the warming holes in the SPNA. The rate of solubility change in the Atlantic Ocean ranges from -8.4 μM for GFDL-ESM2G to -12.2 μM for IPSL-CM5A-MR; in the Pacific Ocean ranges from -

185 5.1 μM for MPI-ESM-LR to -8.6 μM for IPSL-CM5A-LR (see Table 1). The solubility decline is more pronounced in the subpolar Atlantic as expected, but this is in contrast to the net O_2 change in all models.

The AOU signal explains the different O_2 trend [Figure 6]. In the subtropical regions, the AOU decreases in all models in the North Atlantic, but increases overall in the Pacific, even if with inter-model regional differences. As the ocean's surface warms and becomes more stratified, AOU generally increases
190 due to weakened ventilation and sustained biological O_2 consumption which dominates over the physical supply. The effect of respiration is accumulated as water spends more time in the ocean interior, leading to a decline of O_2 . This is verified in most of the North Pacific and in the subpolar North Atlantic. In the subtropical North Atlantic, however, AOU and stratification decouple due to changes in lateral transport and biological oxygen utilization as shown next. Basin-scale averages of ΔAOU in Table 1 are in the
195 range -1.5 μM for GFDL-ESM2G to -6.6 μM for MPI-ESM-2M in the Atlantic and in the range +4.6 μM for MPI-ESM-LR to +10.65 μM for CESM1-BGC in the Pacific. The question that naturally follows is: how could the subtropical North Atlantic have a significant decrease in AOU under the increasing stratification? It is unlikely that the thermocline ventilation increases under this condition. Also, the mechanism at work must be specific to the North Atlantic Ocean.

200 In all EaSMs examined the speed of the Gulf Stream and NAC extension decreases; in contrast, the speed of the Kuroshio Current does not change noticeably [Figure 7]. Consequently, the “nutrient stream” in the North Atlantic loses part of its strength. Since it is a major supply pathway of macronutrients for the North Atlantic, the nutrient inventory and the biological productivity decline in the subtropical gyre. This mechanism is confirmed by the significant decline of the PO_4 inventory projected
205 in the North Atlantic [Figure 8], and by the weakening in carbon export in all models [Figure 9].

The weakened remineralization results in the regional AOU decline, which can compete against the effect of weakened ventilation. In the North Pacific, on the other hand, the PO_4 inventory displays a moderate increase, again following the currents' behaviour. Basin-scale averages of ΔPO_4 in Table 1,
210 range from -0.006 μM for IPSL-CM5A-LR to -0.16 μM for CESM1-BG. In the North Pacific, the nutrient decline is close to zero. Further support for this proposed mechanism can be found in the North Atlantic in the IPSL-CM5A-LR model where the weakest current speed decline [Figure 7] is associated with the

weakest PO_4 decline ($-0.006 \mu\text{M}$), the strongest warming and stronger deoxygenation ($-12 \mu\text{M}$; Table 1) among the models.

It is important to note there is no overall agreement in the patterns or signs of centennial changes in export production, ΔEPC_{100} , among the models. Also, the pattern of the carbon (C) export does not necessarily correspond to the changes in AOU, which instead follow the concomitant changes in ventilation. AOU reflects the integrated respiration rates over the ventilation pathways, so it is not surprising that the patterns look different between ΔAOU and ΔEPC_{100} . Having said this, it is expected that basin-scale decrease in carbon export and respiration are likely to cause a decrease in AOU. The C export decreases globally, but the magnitude of the decline is particularly strong in the North Atlantic [Figure 9]. It generally decreases under increasing stratification because of the reduced upwelling and entrainment of subsurface macro-nutrients, which partially compensates the deoxygenation due to the reduced ventilation. The net effect on the AOU is dominated by ventilation in the North Pacific and the subpolar North Atlantic. However, this is not the case in the subtropical North Atlantic. The decline of the C export is much stronger due to the compounding impacts of the increased stratification and the weakened North Atlantic nutrient stream, as evidenced by the decline in the phosphorus inventory [Figure 8]. This is consistent with the decline in nutrient supply in the North Atlantic and the resultant decrease in AOU. On the contrary, the AOU in the subtropical gyre of the North Pacific increases, despite the weakened C export, suggesting that the weakened ventilation in this region contributes the most to deoxygenation.

In figure 10, we further analyze the mechanisms at play in the Atlantic Basin. To investigate more in depth the nutrient inventory changes in the North Atlantic, we estimate changes of the northward supply of phosphate at 10°N along with the nutrient inventory of the subtropical gyre. Figure 10 time series represent the zonally and vertically integrated northward transport of phosphate ($\overline{v\text{PO}_4}$) at 10°N over the 0-700 meters depth range, decomposed in the overturning ($MO = \overline{v\text{PO}_4}$) and gyre ($GY = \overline{v'\text{PO}'_4}$) components, along with the nutrient inventory (NI) zonally, meridionally and depth-integrated over 10°N - 48°N and 0-700 meters. The overbar indicates the zonal mean, and the primes indicate the departure from the zonal mean. For better comparison, we apply a low pass filter of 10-years, and then we normalise the

240 time series by subtracting their mean and divide by their standard deviation. The coloured values represent the per cent centennial change of each transport component and nutrient inventory. The subtropical gyre nutrient inventory closely follows the declining trajectory of the overturning component of the northward nutrient transport at 10°N for all models but IPSL-CM5A-LR.

To better understand the reduction in carbon export, we explore the nutrient supply to the surface euphotic layer through the vertical entrainment of thermocline nutrients. If averaged over a broad area, the downward export of organic matter is mostly replenished by the upwelling and vertical mixing of nutrients during cool seasons. Over the subtropical oceans, wind-driven Ekman downwelling dominates the mean large-scale circulation, so the winter-time deepening of the mixed layer can be the primary pathway for the vertical nutrient supply. The entrainment flux of nutrient (P) can be represented as, $E_{flux} = \Lambda (P_{th} - P_m) \frac{\partial h}{\partial t}$, with the operator $\Lambda = 1$ when the mixed layer thickness increases $\frac{\partial h}{\partial t} > 0$. $(P_{th} - P_m)$ is the vertical difference in nutrient concentration between the thermocline and mixed layer. Integrating over one year, we approximate the annual entrainment as $E_{flux,ann} \sim H * dPz$, where H is the difference between the yearly maximum and minimum mixed layer depth, and dPz is the vertical nutrient difference between the surface and the 300m. The centennial changes of those two terms are presented in Figure 10 as percent changes along with the changes of the ΔEPC_{100} . The change in sign of both terms ΔH and $\Delta(dPz)$ across SPNA and STNA are in response to different processes. We direct the reader's attention to the lower panels of Figure 10 and the multi-model mean behavior. In the subpolar regions, the maximum mixed layer depth is significantly reduced ($\Delta H < 0$) with the suppression of convective mixing, while the vertical gradient of the nutrient increases ($\Delta dPz > 0$). This indicates that the reduction in export production is primarily caused by the increased stratification and weakened vertical mixing of nutrients into the surface euphotic layer. The increased vertical nutrient gradient cannot cause the weakened export production. On the contrary, in the subtropical Atlantic region, the maximum mixed layer depth deepens with time ($\Delta H > 0$) along the WBC. Positive $\Delta H > 0$ tends to increase the entrainment of nutrient in the euphotic layer but the vertical gradient of the nutrient decreases ($\Delta dPz < 0$), as a result of the total nutrient inventory decline. The reduction in export production in the subtropics is likely caused by the weakened vertical gradient of nutrient. Increased seasonality of the mixed layer depth cannot

explain the reduction in export production. The close relationship between the reduction in basin-scale nutrient inventory and the zonal mean (meridional overturning) nutrient transport indicates that the weakened nutrient stream is causing the weakened export production in the subtropical North Atlantic.

Conclusions

270 We analyzed a subset of seven EaSMs included in the CMIP5 catalogue to understand current and future state of oxygen distribution in the upper 700 m of the water column in the northern hemisphere. During the historical period 1970-2000, models reproduce the observed mean state of dissolved oxygen concentration, capturing spatial variations in apparent oxygen utilisation and, most importantly, reproduce the “nutrient stream”. By the end of this century, the upper water column in the business as usual scenario
275 is projected to warm more in the North Atlantic compared to the North Pacific. Despite this tendency, the subtropical North Atlantic resists to deoxygenation. As the ocean warms, O₂ saturation decreases globally, with the exception of the warming holes in the North Atlantic, but the two basins differ especially in the AOU. In the subtropical North Atlantic, the basin-mean AOU decreases and is decoupled from the stratification-induced reduction in ventilation. In all models but one (IPSL-CM5A-LR), the
280 AMOC weakening is associated with a decline in the current speed of the Gulf Stream and its extension and, in turn, to a decline in the nutrient stream. Lateral nutrient supply, quantified by the reduction in phosphate inventory, decreases, and so does biological productivity, as confirmed by the negative trend in ΔEPC100 . The decline in biological productivity and consequent retention of O₂ (by weakened biological consumption) in the subtropical North Atlantic are sizable enough to compensate the O₂
285 solubility trend. The decline in the nutrient stream is not detected in the North Pacific, where biological productivity does not change as dramatically as in the Atlantic, and the solubility trend dominates.

Our results imply that the ocean deoxygenation progresses more intensely in the North Pacific Ocean even though its heat uptake is moderate compared to its neighbour ocean. This faster and stronger decline appears to be supported by the relatively stable P inventory of the North Pacific. The macro-
290 nutrient inventory of the North Pacific is “charged up” with the higher concentrations of nutrients in comparison to the North Atlantic due to the old age of the Pacific waters. In contrast, the North Atlantic nutrient inventory is more dynamic given that nutrient streams critically depends on the AMOC and its

feedbacks on the western boundary current system. This difference has significant consequences given that the background, climatological O₂ levels are much lower in the Pacific basin, again due to the older age. The Pacific Ocean indeed hosts already two of the four most voluminous oxygen minimum zones. Higher rate of O₂ loss can potentially lead to more frequent and intense hypoxic events, with devastating impacts for the marine ecosystem (Penn et al., 2018). The length of the EaSM integrations does not allow to verify if the reduction in the biological activity of the subtropical North Atlantic is only transient, and if a rebound may take place once a new climate equilibrium is achieved (Moore et al., 2018). Further investigations and higher resolution model outputs are also needed to better constrain the regional patterns of biological productivity and oxygen changes.

Acknowledgments

We acknowledge the World Climate Research Programme's Working Group on Coupled Modeling, which is responsible for CMIP, and the U.S. Department of Energy's Program for Climate Model Diagnosis and Intercomparison for providing support and software infrastructure in partnership with the Global Organization for Earth System Science Portals. The CMIP data were obtained from the CMIP5 Data Access Portal (http://cmip-pcmdi.llnl.gov/cmip5/data_portal.html). The World Ocean Atlas Data (WOA09) were obtained from the National Center for Environmental Information (www.nodc.noaa.gov/OC5/SELECT/woaselect/woaselect.html). We thank two reviewers and the Editor for their useful comments.

We acknowledge the support by the NOAA Climate Program Office, Climate Variability and Predictability Program, through grant NA16OAR4310173. In its initial stages this work was supported by a grant from the National Science Foundation (NSF-OCE 097394084).

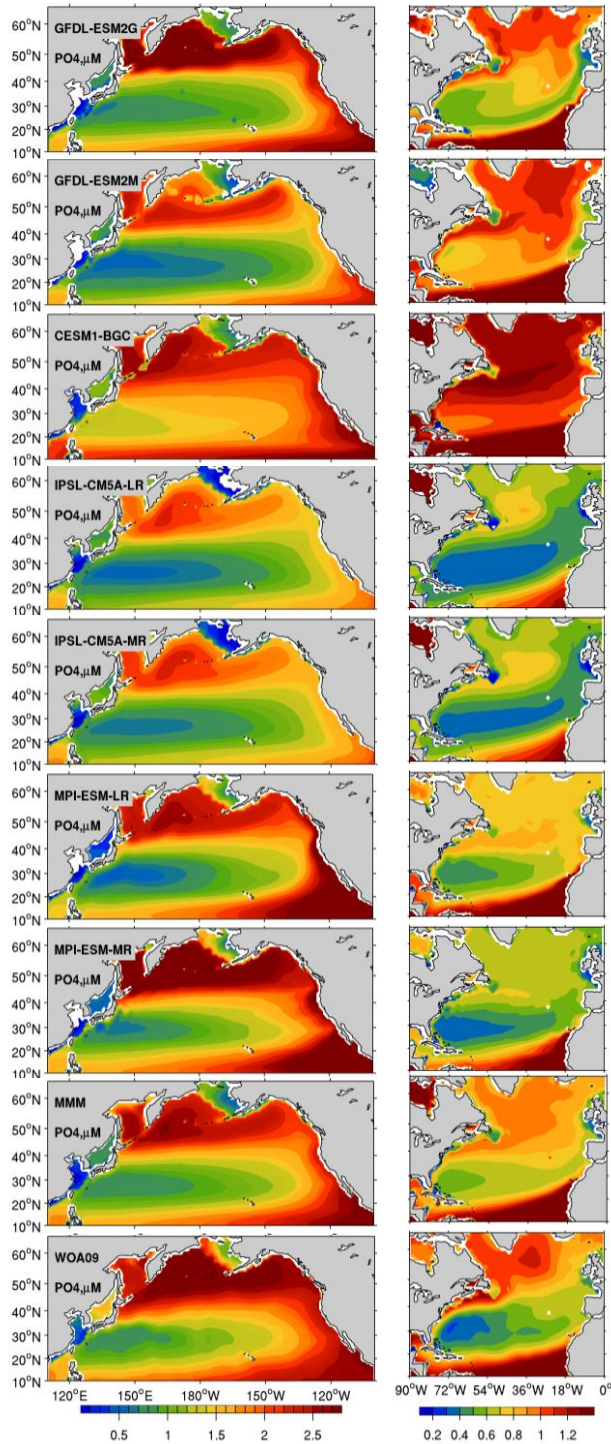


Figure 1 : Upper ocean (0-700 m) concentration of phosphate (PO₄), for the period 1970-2000 in a subset of the CMIP5 models (esmHistorical), Multi-Model-Mean (MMM), and World Ocean Atlas 2009 (WOA09). The North Pacific and Atlantic basins are plotted with different colour ranges to better highlight the spatial patterns in models and observations.

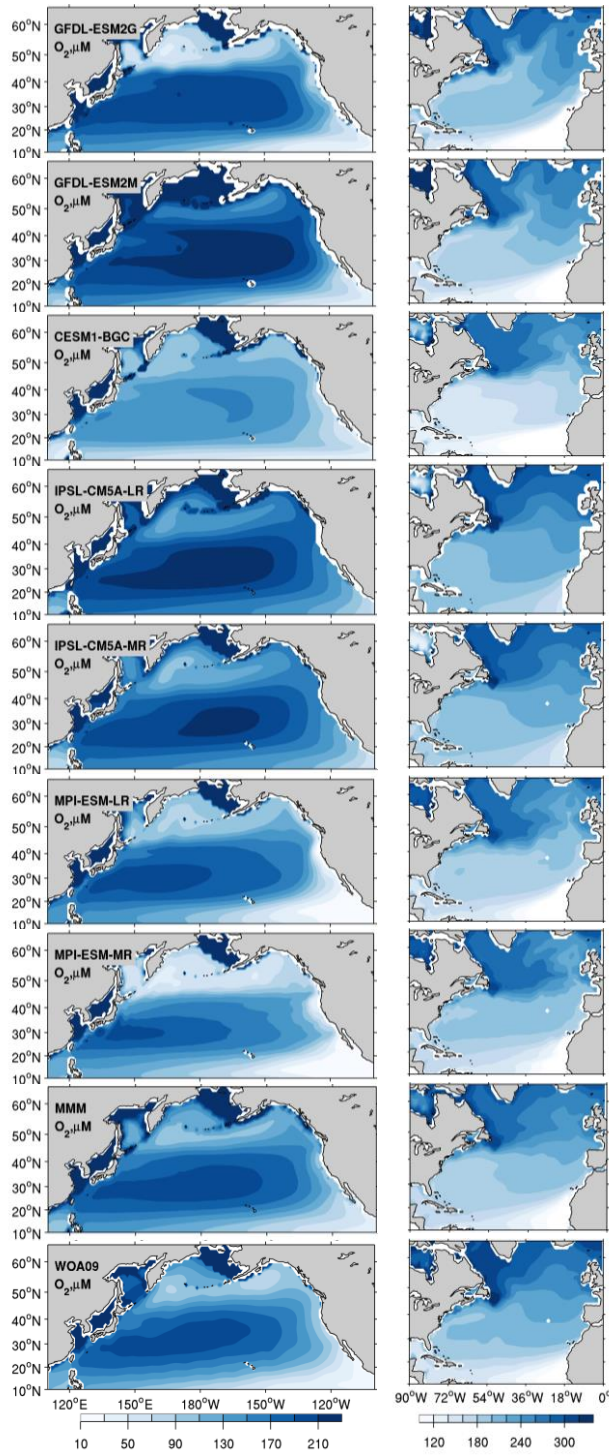


Figure 2 Upper ocean (0-700 m) concentration of dissolved oxygen (O_2), for the period 1970-2000 in a subset of the CMIP5 models (esmHistorical), Multi-Model-Mean (MMM), and World Ocean Atlas 2009 (WOA09). The North Pacific and Atlantic basins are plotted with different colour ranges to better highlight the spatial patterns in models and observations.

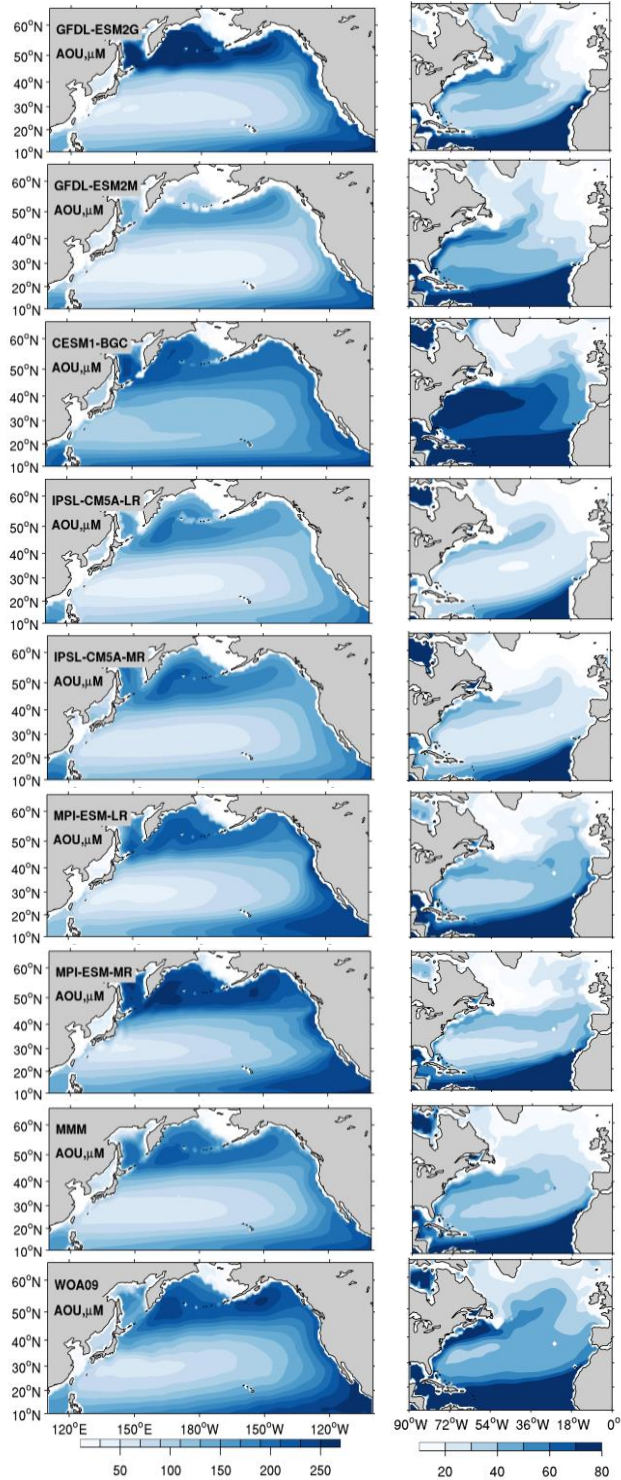


Figure 3 Upper ocean (0-700 m) concentration of apparent oxygen utilization (AOU), for the period 1970-2000 in a subset of the CMIP5 models (esmHistorical), Multi-Model-Mean (MMM), and World Ocean Atlas 2009 (WOA09). The North Pacific and Atlantic basins are plotted with different colour ranges to better highlight the spatial patterns in models and observations.

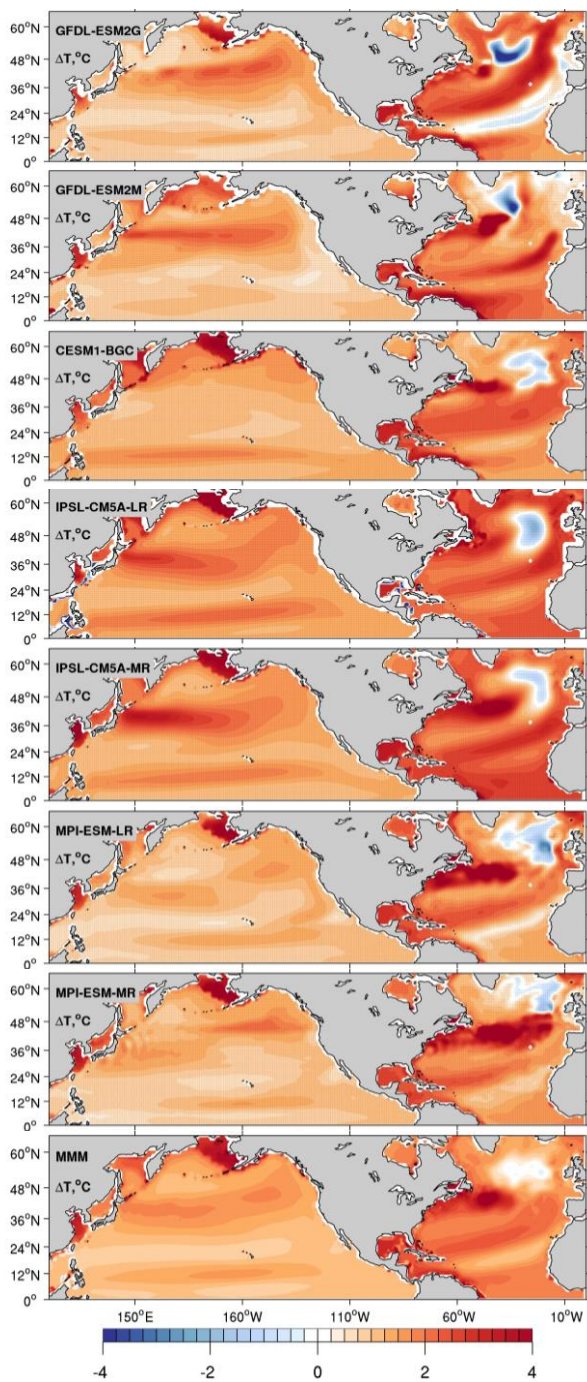


Figure 4 Centennial change of T calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages. Black dots indicate areas where the results are statistically significant at the 99% confidence level according to a t-test.

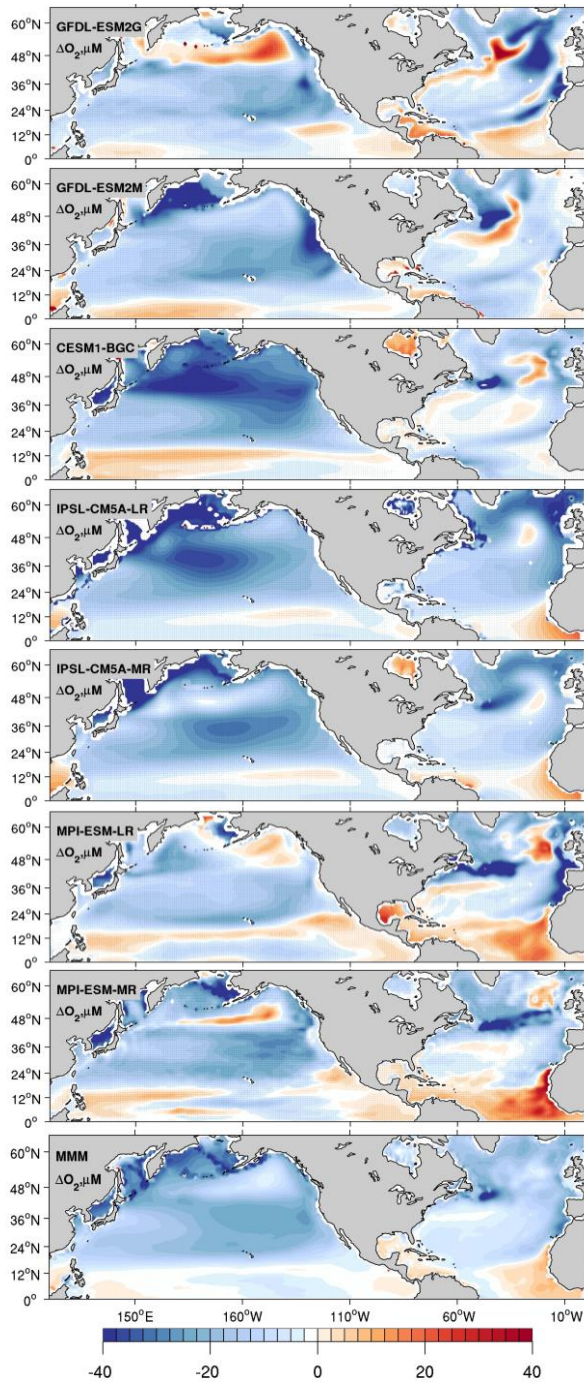


Figure 5 Centennial change of dissolved oxygen calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages. Drift is removed from the piControl simulation. Black dots indicate areas where the results are statistically significant at the 99% confidence level according to a t-test.

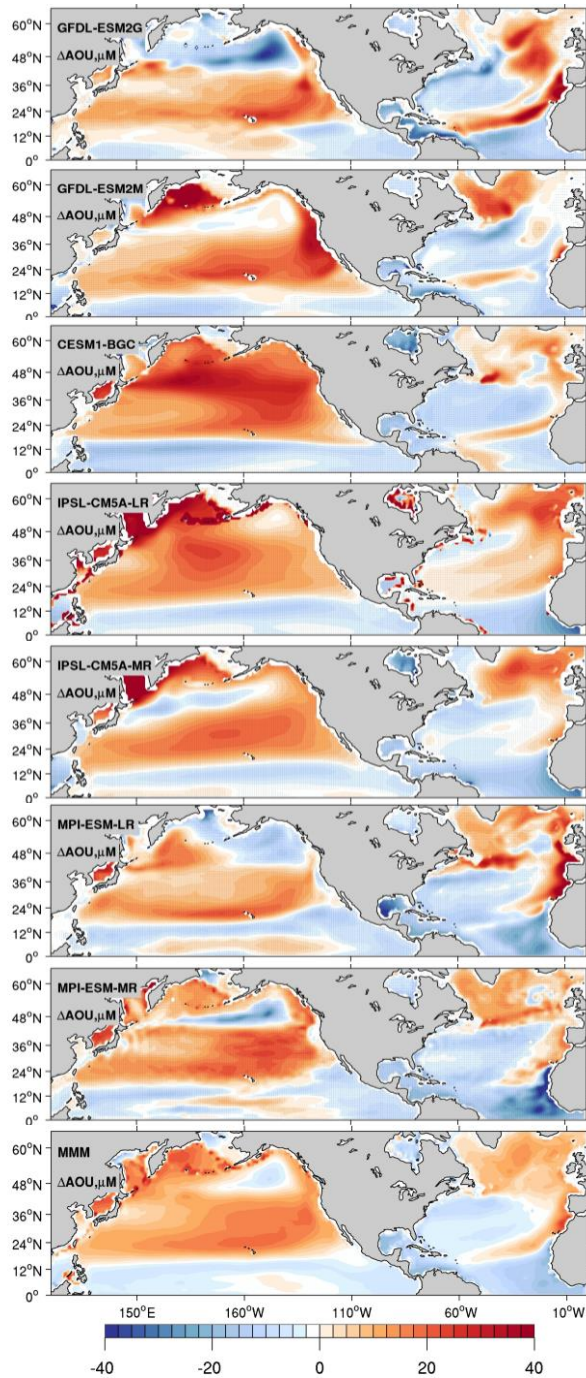


Figure 6 Centennial change of apparent oxygen utilization calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages. Drift is removed from the piControl simulation. Black dots indicate areas where the results are statistically significant at the 99% confidence level according to a t-test.

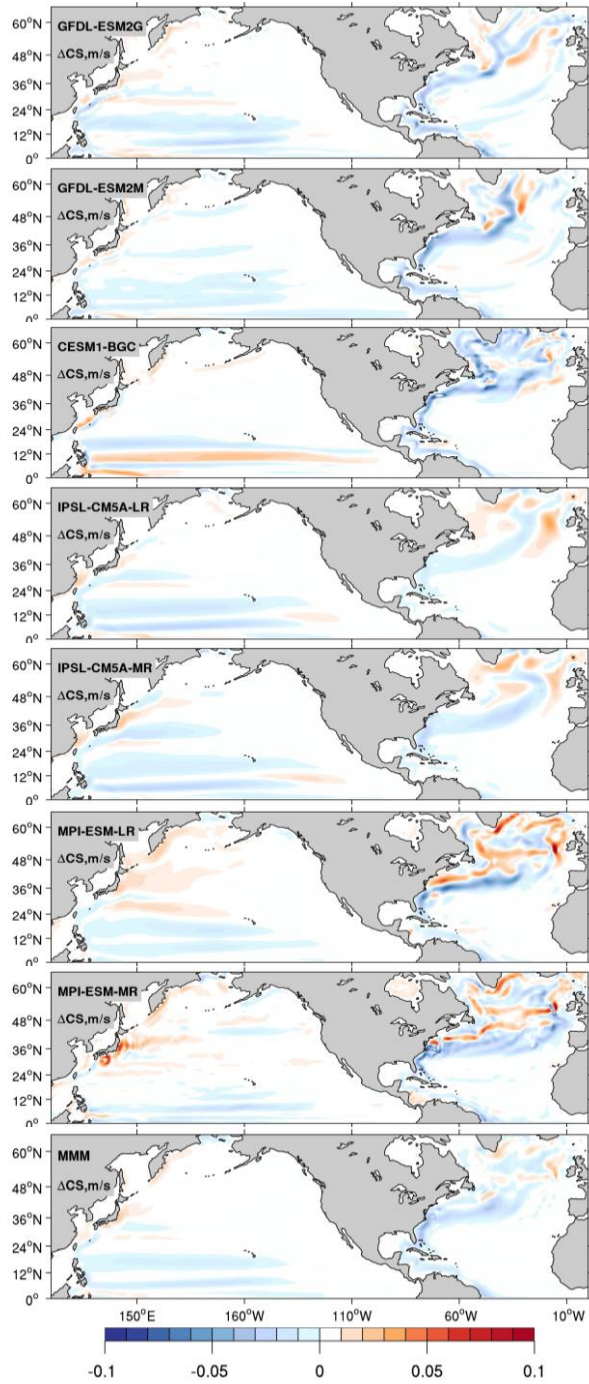


Figure 7 Centennial change of current speed calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages.

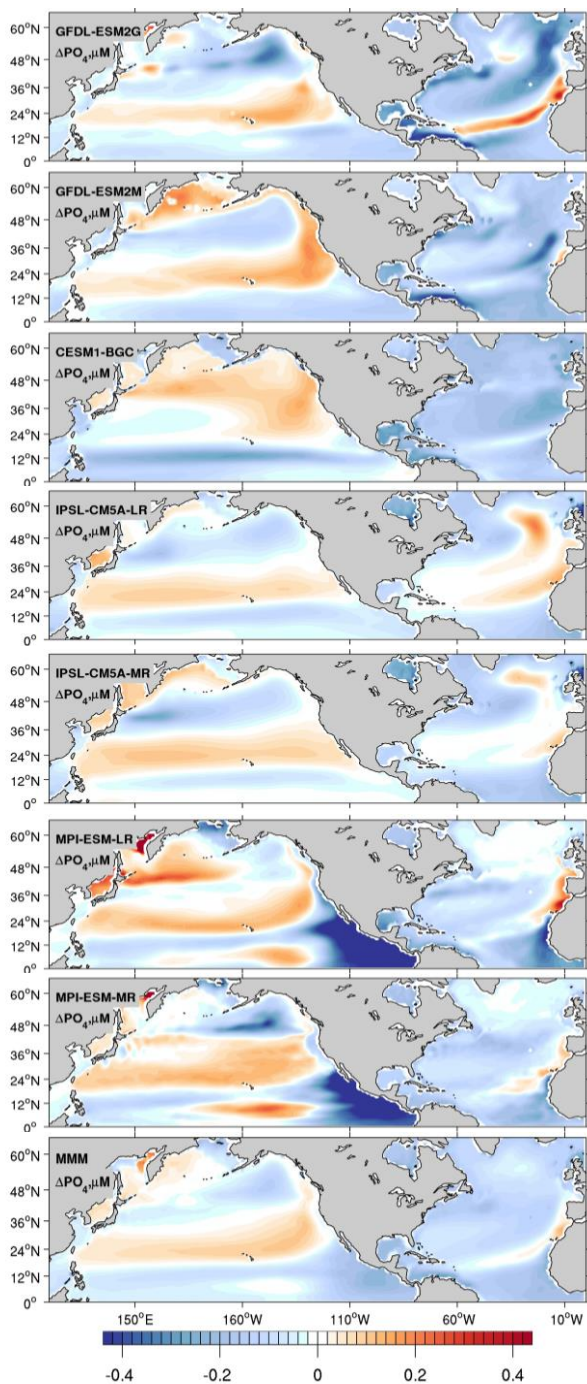


Figure 8 Centennial change of PO_4 calculated as the difference in 30-year averages between (2070-2100) and (1970-2000). All plotted values are 0-700 m depth averages.

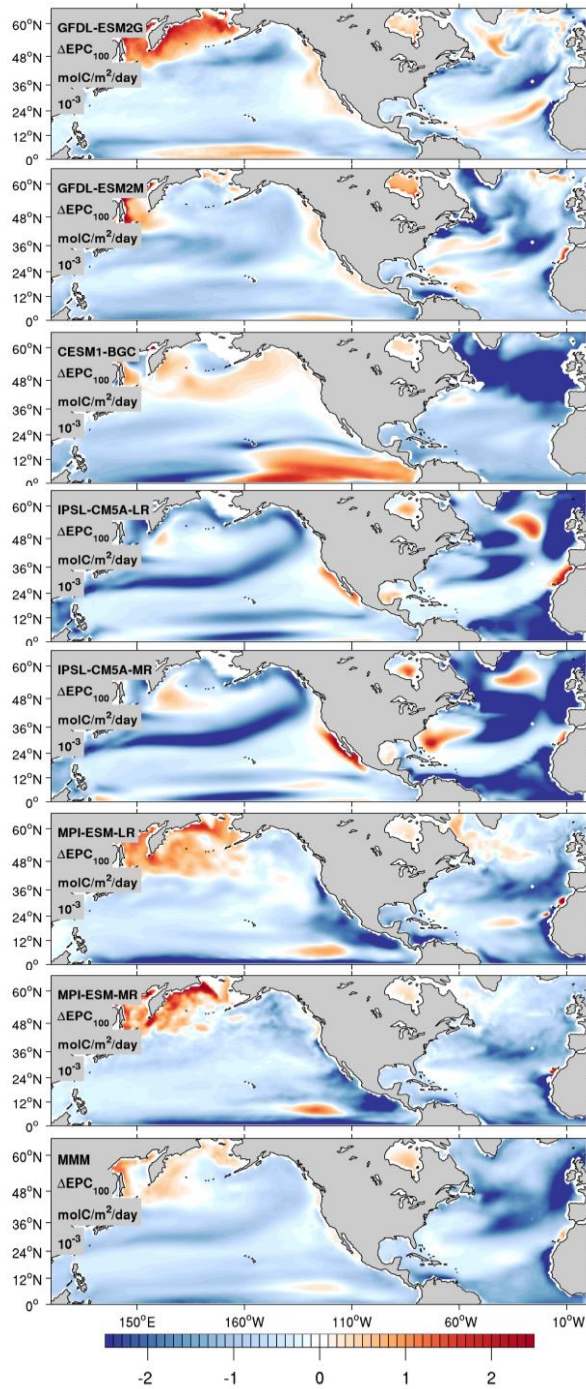


Figure 9 Centennial change of export production calculated as the difference in 30-year averages between (2070-2100) and (1970-2000).

Table 1: Averaged changes of temperature (ΔT), dissolved oxygen (ΔO_2), oxygen solubility ($\Delta(O_{2,sat})$), apparent oxygen utilisation ($\Delta(AOU)$), and nutrient ($\Delta(PO_4)$) between 10°N-48°N for Pacific and Atlantic basins averaged over the upper 0-700 m. The changes are calculated as the differences between the 30-year period 2070-2100 in the rcp8.5 scenario and 1970-2000 in the historical simulations.

	ΔT		ΔO_2		$\Delta O_{2,sat}$		ΔAOU		ΔPO_4	
	Pac	Atl	Pac	Atl	Pac	Atl	Pac	Atl	Pac	Atl
GFDL-ESM2G	1.14	1.69	-12.4	-6.9	-6.1	-8.4	6.3	-1.5	-0.007	-0.13
GFDL-ESM2M	1.14	2.21	-16.1	-6.1	-5.8	-11.0	10.1	-5	0.001	-0.15
CESM1-BGC	1.20	2.00	-16.8	-6	-6.2	-10.3	10.7	-4.2	-0.020	-0.16
IPSL-CM5A-LR	1.60	2.03	-18.1	-12	-8.6	-10.0	9.5	-1.7	0.000	-0.006
IPSL-CM5A-MR	1.60	2.45	-16.0	-7	-8.5	-12.2	7.5	-4.8	-0.005	-0.04
MPI-ESM-LR	1.07	1.90	-10.0	-6	-5.1	-9.5	4.6	-3.6	-0.001	-0.10
MPI-ESM-MR	1.12	2.00	-12.2	-3.4	-5.4	-10.0	6.7	-6.6	0.001	-0.10
MMM	1.26	2.04	-14.5	-6.7	-6.5	-10.2	7.9	-3.9	-0.004	-0.10

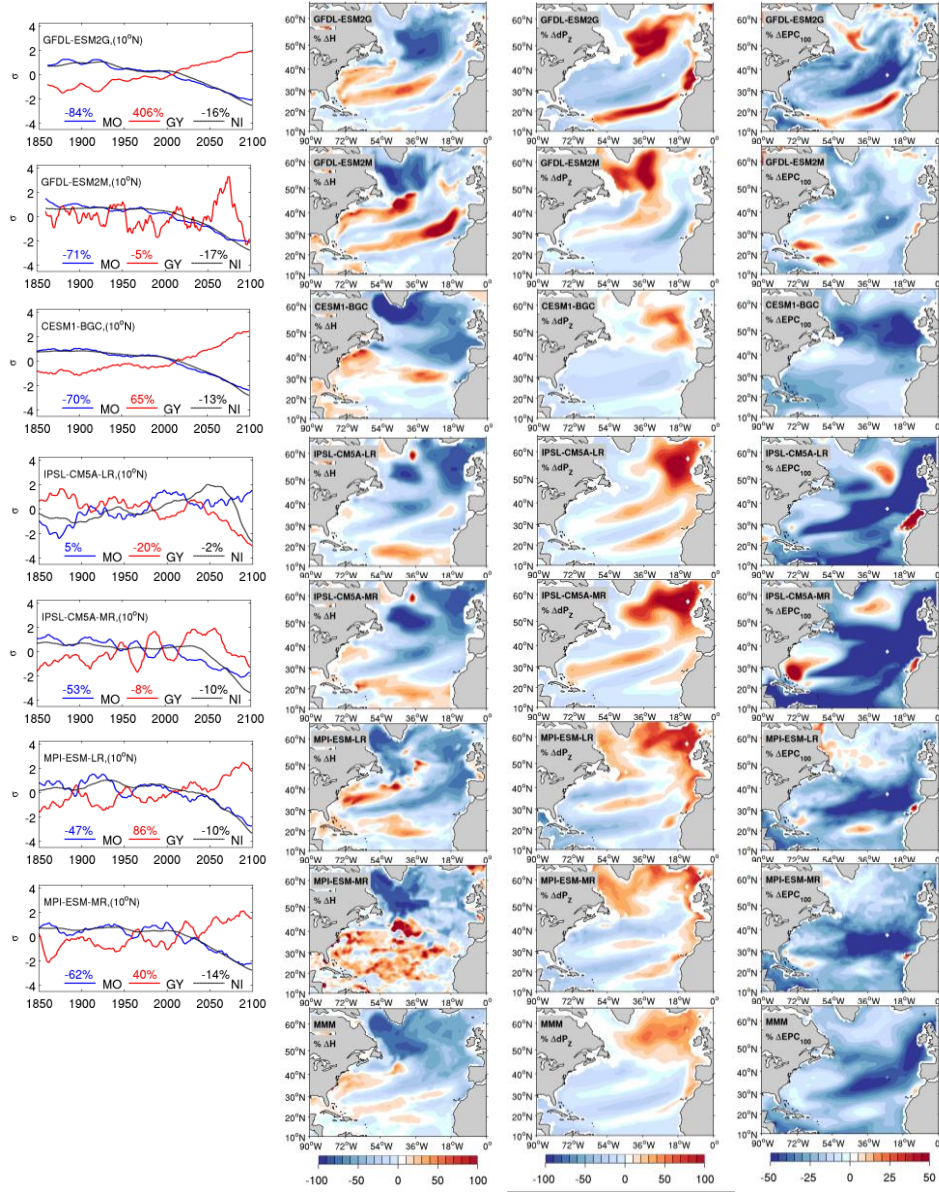


Figure 10 (Left column) Normalized timeseries of zonally (70°W-17°W) and depth (0-700m) integrated northward nutrient transport components, ($MO = \overline{v'PO_4}$) and gyre ($GY = \overline{v'PO_4}$), at 10°N and nutrient inventory (NI) of the subtropical gyre (10°N-48°N, 0-700 meters). Coloured values represent the percent centennial change of each variable. Panels represent the percent centennial change of H (year maximum seasonal change of mixed layer depth), dPz (vertical gradient of PO_4), EPC_{100} export production, all calculated as the difference in 30-year averages between (2070-2100) and (1970-2000).

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