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Patterns of deoxygenation – sensitivity to natural and anthropogenic drivers

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Keywords: marine oxygen, deoxygenation, climate change

Summary

Observational estimates and numerical models both indicate a significant overall decline in marine oxygen levels over the past decades. Spatial patterns of oxygen change, however, differ considerably between observed and modelled estimates. Particularly in the tropical thermocline that hosts open-ocean oxygen minimum zones, observations indicate a general oxygen decline, whereas most state-of-the-art models simulate increasing oxygen levels. Possible reasons for the apparent model-data discrepancies are examined. In order to attribute observed historical variations in oxygen levels, we here study mechanisms of changes in oxygen supply and consumption with sensitivity model simulations. Specifically, the role of equatorial jets, of lateral and diapycnal mixing processes, of changes in the wind-driven circulation and atmospheric nutrient supply, and of some poorly constrained biogeochemical processes are investigated. Predominantly wind-driven changes in the low-latitude oceanic ventilation are identified as possible factor contributing to observed oxygen changes in the low-latitude thermocline during the past decades, while the potential role of biogeochemical processes remains difficult to constrain. We discuss implications for the attribution of observed oxygen changes to anthropogenic impacts as well as research priorities that may help to improve our mechanistic understanding of oxygen changes and the quality of projections into a changing future.

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Introduction

Analyses of historical measurements of dissolved marine oxygen have revealed a systematic decline of oxygen levels in the tropical thermocline over the recent decades. Stramma et al.'s [1] compilation of oxygen measurements over the past five decades revealed the first robust evidence for an almost continuous decline of dissolved oxygen in several regions of the tropical ocean's thermocline. All but one out of the six regions chosen showed a decline in thermocline oxygen. Oxygen concentrations in the Atlantic and Pacific tropical oceans declined and no clear oxygen trend could be found for the region in the eastern part of the tropical Indian Ocean. Concentrating on the shorter period of 1970 to 1992, Helm et al. [2] could extend the analysis to zonal and vertical averages of global-ocean oxygen changes. Their results indicated that, over the time interval chosen, the volumetric oxygen decline was most pronounced at high latitudes throughout the water column and in the tropical thermocline. This was mostly confirmed by a recent full-depth linear trend analysis of oxygen change [3]. A zonally averaged oxygen increase is only found in narrow bands in the upper few hundred meters of the subtropical ocean and also well below a thousand meters in the tropical ocean (Figure 2 of Helm et al. [2]). Focussing on the upper ocean, Stramma et al. [4] presented a global map of estimated linear oxygen trends in the thermocline (at about 300 m depth) over the five decades 1960 to 2010. For this longer time interval, and consistent with other observational analyses by Helm et al. [2] and Schmidtko et al. [3], their spatially resolved analysis revealed an oxygen decline in the tropical thermocline and at high latitudes, separated by subtropical areas with positive oxygen trends (top left panel of Figure 1).

While there is some consistency among different observational estimates of recent multi-decadal oxygen trends in the upper ocean, our analysis of the results of the most recent fifth phase of the Climate Model Intercomparison Project (CMIP5, [5]) reveals a rather different picture (Fig.1). Most of the models that included oxygen in the archived output ([6, 7, 8, 9, 10, 11, 12, 13], Table 1) show increasing linear oxygen trends over the period 1960-2010 over large areas of the tropical thermocline. Trends become more negative in the subtropics, reaching largest oxygen declines at mid to high latitudes. In fact the pattern correlation between simulated oxygen trends and the observational trend estimate is very weak for all models, and even slightly negative for 6 out of 10 models (computed over the

latitude range 50° S – 50° N, Table 2). The weak pattern correlations of any model with the observational estimate, as well as among the different models themselves, is *per se* not unexpected: All CMIP5 model runs used here are forced by historical CO₂ emissions (and those of other greenhouse gases and land-use change), but have their own internal ocean-atmosphere dynamics and thus cannot be expected to simulate patterns associated with internal climate variability in phase with the observations. The UVic model of intermediate complexity [14], on the other hand, is driven by prescribed monthly varying climatological winds and thus cannot display realistic levels of internal climate variability. The common forcing of all models is greenhouse-driven global warming. The absence of statistically significant correlations between observed and simulated patterns of multi-decadal oxygen change suggests that this global warming alone cannot explain the observed patterns of multi-decadal oxygen trends. Could this mean that the observed oxygen decline in the tropical thermocline is not related to anthropogenic climate change but, for example, a result of internal natural variability in the ocean-climate system? Or does the absence of significant correlations indicate some systematic model deficiencies?

In the following we investigate the discrepancies between observed and simulated oxygen trends in more detail and aim to identify the causes of these differences. From our results we infer research priorities that may help to reduce uncertainties and yield more robust model projections. In the final section of the paper we will discuss, based on our model results, to what extent the observed changes in thermocline oxygen levels may be natural variability or anthropogenically driven change.

Possible causes of model-data mismatch

The zonally averaged pattern of the observational estimate of the 1960—2010 thermocline oxygen trend is shown in Figure 2 together with the zonal average of the mean of all CMIP5 model simulations. Also shown is the zonally averaged oxygen trend obtained by resampling and interpolating each individual CMIP5 model simulation in the same way the observations were obtained – using space and time of every individual 300 m observation. The averaged resampled model trend is almost identical to that of the full models, which gives some confidence into the robustness of the mapping algorithm applied to the still

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sparse observational data [4]. Since the computationally efficient UVic model will in the following be used for a number of sensitivity studies we show in Figure 2 along with the CMIP5 models the results of the UVic reference simulation UVic REF. Important for justifying the use of the UVic model for sensitivity experiments, the CMIP5 model mean and the UVic REF model yield very similar results for the zonally averaged 50-yr oxygen trend in the thermocline. In the tropical thermocline, only a small oxygen decline (less than $0.04 \mu\text{mol kg}^{-1} \text{yr}^{-1}$) and, in the case of the UVic model, even a slight oxygen gain is simulated in the tropical thermocline, with larger oxygen losses in the subtropics and mid latitudes. This zonally averaged pattern is essentially inverse to that of the observational trend estimate that indicates a substantial oxygen decline in the tropics (more than $0.2 \mu\text{mol kg}^{-1} \text{yr}^{-1}$) accompanied by small oxygen gains in the subtropics. Only poleward of about 55 degrees south and 40 degrees north do both simulated and observed oxygen trends show a decline and even some quantitative agreement. Despite the spatial anti-correlation of the zonally averaged trends, the globally averaged oxygen decline at 300m depth is very similar. The observational estimate ($0.06 \mu\text{mol kg}^{-1} \text{yr}^{-1}$) and the CMIP5 model mean ($0.05 \mu\text{mol kg}^{-1} \text{yr}^{-1}$) and the UVic REF model ($0.03 \mu\text{mol kg}^{-1} \text{yr}^{-1}$) closely agree with each other, despite some variability among the individual models (Table 2), see also Bopp et al. [15] and Cocco et al. [16] for inter-model variations in simulated oxygen.

The common forcing of the CMIP5 models and the UVic model are historical CO_2 emissions. However, neither the CMIP5 model mean nor the UVic model reproduces the observed oxygen decline in the tropical thermocline. To the extent that current models are dynamically realistic, this suggests that CO_2 -driven global warming alone cannot explain the negative oxygen trend in the tropical thermocline. In an attempt to understand why current models do not reproduce the observed multi-decadal oxygen decline in the tropical thermocline, in the following we investigate potential impacts of other processes on thermocline oxygen trends by running a number of sensitivity simulations with the UVic model. Restricting the sensitivity experiments to a single, relatively coarse-resolution model, can yield only first-order estimates, and more robust inferences will require follow-on studies with other earth system models. Nevertheless, the UVic results are useful for pointing to the mechanisms worthwhile exploring in more elaborate follow-on studies. For the analysis, we begin here with possible model deficiencies in physical transport and

ventilation before moving to potential impacts of poorly represented biogeochemical processes and finally speculating about the possible role of fisheries.

(i) Mixing: Previous studies have already investigated the role of mixing processes in simulating thermocline oxygen distributions. Sensitivity studies using the UVic model revealed that, for a business-as-usual CO₂ emissions scenario low oxygen regions expanded for high diapycnal mixing rates, and shrank for low diapycnal mixing rates [17]. This finding could be related to the larger decline of export production under global warming for lower diapycnal mixing [17]. The same behaviour is also found for oxygen levels in the tropical thermocline simulated over the period 1960-2010 with oxygen trends being somewhat more negative (by about 0.04 μmol kg⁻¹ yr⁻¹) for high diapycnal mixing ($k_v=0.3$ cm² s⁻¹) relative to low ($k_v=0.01$ cm² s⁻¹) diapycnal mixing (Fig. 5 of Stramma et al. [4]). The reference simulation uses a background diapycnal mixing intensity of $k_v=0.15$ cm² s⁻¹. Although systematic, this sensitivity of simulated tropical thermocline oxygen trends to diapycnal mixing remains small compared to the model data misfit displayed in Figure 2.

Sensitivity studies with different strengths of isopycnal mixing have demonstrated a significant impact on the mean state of simulated oxygen distributions [18], but have, as far as we know, not yet been analysed in terms of oxygen trends. Simulations with enhanced zonal isopycnal mixing in the tropical ocean as a pragmatic way to parameterise the effects of unresolved equatorial current bands [19] did affect the simulated evolution of eastern boundary oxygen minimum zones, but had little effect on the zonally averaged oxygen trends.

(ii) Winds: Wind changes have already been suggested by Stramma et al. [4] as possible mechanism that can substantially impact multi-decadal thermocline oxygen trends. Since the CMIP5 models are all coupled atmosphere-ocean models that simulate their own internal variability, it is plausible that atmospheric modes will be out of phase with observations. If winds play a dominant role in controlling tropical thermocline oxygen levels, this could then explain the poor agreement of coupled climate models with observed oxygen trends. For tropical oxygen, both local wind changes in the tropics and also remote wind changes at higher latitudes may be important.

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4 Changes in tropical trade winds have been observed and their possible impact on marine
5 oxygen patterns was investigated in experiments using the UVic model [20]. Results
6 indicated a see-saw like impact on oxygen levels in the tropical oceans. The compensatory
7 nature of changes in the eastern and western parts of the basins does not make tropical
8 trade wind changes a likely candidate for zonally averaged oxygen changes. Remote wind
9 changes in the Southern Ocean, on the other hand, may impact the ocean without being
10 bounded by continents and therefore have no straightforward mechanism for zonal
11 compensation of zonal wind-induced changes.
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19 Southern Ocean wind changes have been identified as one of the most dominant changes in
20 global wind patterns during the past decades [21]. We tested the potential impact of the
21 observed increase and poleward migration of the southern hemispheric westerlies by
22 applying these wind anomalies to the UVic model [22]. Starting from a spun-up pre-
23 industrial state in year 1800, the increasing westerly winds south of 40°S lead to a small
24 reduction of tropical thermocline oxygen levels (by up to 0.03 $\mu\text{mol kg}^{-1} \text{yr}^{-1}$) compared to
25 the reference experiment during the time period 1960-2010 in our model (Fig.3a). This
26 oxygen reduction is primarily related to the southward shift of the subtropical gyre
27 circulation that increases the distance between mid-latitude subduction regions and the
28 tropics and, in turn, leads to a decrease of the oxygen supply via newly ventilated oxygen-
29 rich waters of subantarctic origin [22] This effect might be underestimated in CMIP5 models
30 that do not fully reproduce the observed trends in speed and position of the southern
31 hemispheric westerlies [23], but it appears an order of magnitude too small to explain the
32 discrepancy between observed and simulated oxygen changes.
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46 Long-term wind changes have also been reported for other regions of the planet and are
47 often associated with climate modes such as the North Atlantic Oscillation (NAO) [24] or
48 Pacific Decadal Oscillation (PDO) [25]. There is robust evidence that the NAO has some
49 impacts on winter mixed-layer depths, ventilation and nutrient supply to the mid and high
50 latitudes of the North Atlantic and the upwelling region off Mauretania [26] and can
51 significantly affect mid- and high-latitude oxygen levels [27] and marine ecosystems in the
52 western and eastern North Atlantic [28]. However, the NAO seems to have little direct
53 impacts on oxygen levels in the low-latitude North Atlantic [29].
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For the Pacific Ocean, significant impacts of the PDO on tropical oxygen changes have been inferred from observations and models. PDO-driven changes in the depth of the tropical thermocline have been suggested to affect the upper-ocean supply of nutrients, associated biological production and subsequent respiratory oxygen consumption [30]. This mechanism would explain observed oxygen changes by changes in oxygen demand. Alternative mechanisms have been proposed, focusing on variations in oxygen supply. These include PDO-driven variations in the subduction-driven injection of oxygenated surface waters into the thermocline [31], the above-mentioned imprint of Southern Ocean wind changes on the source waters that eventually feed tropical oxygen minimum zones [22] and also wind-driven changes in the subtropical cells [32] and subtropical gyres [33]. The individual contributions of the different mechanisms to the observed oxygen changes are still not well understood and are a matter of ongoing research.

Forcing the UVic model with monthly mean CORE (Co-ordinated Ocean-ice Reference Experiments [34]) reanalysis winds over the period 1958-2007 (and then repeating 2007 winds for another three years until year 2010), the simulated linear 50-year oxygen trend can either be positive or negative, depending on the winds used in the model spin-up prior to the observational period (1960-2010): When the 1958-1967 mean CORE winds are used to spin up the model until year 1958 with interannually varying CORE winds applied thereafter, oxygen levels in the tropical thermocline increase in the UVic model and also in a 0.5 degree resolution forced ocean model of the NEMO (Nucleus for European Modelling of the Ocean) consortium. However, when the model is spun up with the NCEP wind climatology [35] prior to year 1958, with interannually varying CORE winds applied thereafter, oxygen levels in the tropical thermocline decrease during the period 1960-2010 (Fig. 3b). The different behavior of the two model simulations during the analysis period 1960-2010 can be related to the different model states at the end of the spin up. Although large-scale thermocline oxygen patterns at the end of the spin-ups are similar for the NCAR/NCEP and CORE wind stress forcing, some regional differences exist: At the end of the spin-up, the simulation forced with NCAR/NCEP winds shows lower thermocline oxygen concentrations (by up to 60-80 $\mu\text{mol kg}^{-1}$) along the equatorward margins of the Pacific subtropical gyres than the simulation forced with the 1958-1967 CORE winds (Figure 4a). In contrast, thermocline oxygen levels are higher (by up to 20-40 $\mu\text{mol kg}^{-1}$) at the end of the NCAR/NCEP spin-up in the western equatorial Pacific. These end-of-spin-up oxygen

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4 differences are related to weaker trade winds in the NCEP/NCAR forcing compared to the
5 1958-1968 CORE reanalysis period (Figure 4b). In year 2010, i.e. at the end of the simulation
6 forced with interannual 1958-2010 CORE forcing, the initial thermocline oxygen levels have
7 been considerably reduced and in the western tropical Pacific even reverted (Figure 4c).
8 This is consistent with the more negative oxygen trend in the tropical thermocline
9 simulated by the run spun up with NCEP/NCAR climatological winds (Fig. 4d, Fig. 3b).
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16 Our sensitivity simulations with different wind forcing fields during the spin-up reveal that
17 despite identical wind forcing during the period 1960-2010, the simulated oxygen trends
18 during this period are very sensitive to the applied wind forcing prior to the 1960s. Thus we
19 conclude that the actual oxygen trend in the tropical thermocline is influenced by oxygen
20 anomalies that may have been set during the formation of thermocline waters decades ago.
21 Given our still relatively poor knowledge of the oceanic wind forcing in the decades before
22 the 1960s, we thus cannot rule out that the failure of forced models to reproduce the
23 observed 1960-2010 oxygen trend is caused by an unrealistic representation of the legacy of
24 oxygen anomalies already 'in the pipeline' prior to the observational period. Further
25 research into the potential immediate and lagged impacts of temporally varying wind
26 patterns on tropical oxygen levels should help to clarify this issue.
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37 **(iii) Model resolution:** Current climate models and even most stand-alone global ocean
38 models have a grid resolution not fine enough to resolve the equatorial current bands that
39 are known to be important agents of lateral oxygen transport in the tropical oceans [36, 37].
40 Evidence from 0.1° tropical ocean nests embedded into global models shows considerable
41 improvement in the mean state of the simulated oxygen distributions [38]. However, effects
42 of refined grid resolution on simulated oxygen trends have not yet been investigated. As
43 found for the simulations with interannually varying CORE wind forcing, a 0.5° model
44 revealed roughly similar oxygen trends as the coarser ($1.8^\circ \times 3.6^\circ$) UVic model that only
45 includes a pragmatic parameterisation of the effects of zonal current bands via enhanced
46 zonal mixing close to the equator [19] (Fig.3b). Investigations of interannual changes in sea
47 level height have also found little impact of refined model resolution (at least not up to $\frac{1}{4}$
48 degree grid resolution) in the tropical and subtropical oceans [39]. Interannually forced
49 model simulations at much higher resolution should help to further clarify the role of small-
50 scale equatorial transport processes for the observed multi-decadal oxygen trends. It also
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remains to be investigated whether changes in eddy activity and associated transport in oxygen, nutrients or organic matter may have played a role in observed oxygen changes.

(iv) Changes in atmospheric nutrient deposition: In addition to changes in physical transport of nutrients and oxygen within the ocean, there are also changes in the supply from the atmosphere. Changes in atmospheric oxygen concentrations are dominated by the seasonal cycle of biological production and respiration and by a long-term decline due to the burning of fossil fuels [40]. These changes are, however, of the order of parts per million and thus negligibly small compared to the observed changes in tropical thermocline oxygen.

Much larger changes have occurred in atmospheric nutrient-containing aerosols and subsequent deposition, predominantly from anthropogenic sources. Deposition of nitrogen from the atmosphere to the ocean has more than tripled over the past century and is likely to increase further in the future [41]. Adding the increasing deposition of anthropogenic nitrogen to the ocean during the historical period [42] in our model simulations, we find a reduction of the simulated tropical-thermocline oxygen increase by some $0.015 \mu\text{mol kg}^{-1} \text{yr}^{-1}$ (Figure 5a). This simulated impact of nitrogen deposition on thermocline oxygen is only about half the size of that inferred from the model study by Ito et al. [43], and is also smallest of all the factors investigated in the current study. The considerable discrepancy between the two model solutions may be explained by the difference in the treatment of nitrogen cycle feedbacks, in particular via nitrogen fixation. While diazotrophs are ignored in the Ito et al. [43] study, they are included in our model simulations, where nitrogen fixation declines in response to the addition of fixed nitrogen to the surface ocean, limiting the net impact of anthropogenic nitrogen deposition on marine biogeochemistry and exerting a stabilizing negative feedback on the oceanic nitrogen inventory [44]. The importance of nitrogen fixation for the biogeochemical model response presents a serious caveat, as neither the absolute rates nor the environmental controls of marine nitrogen fixation are very well understood [e.g., 45]. Further studies are required to better constrain the potential impact of atmospheric nitrogen deposition on marine biogeochemistry, including oxygen concentrations.

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4 Another biotically important element with a substantial atmospheric supply to the ocean is
5 iron. Though mostly derived from desert soils, with some contributions from combustion
6 sources, substantial changes in iron have been reported over the past century, with
7 substantial decadal variability [46]. Recent modelling studies suggest that changes in
8 atmospheric deposition of iron may have had a larger impact on marine biogeochemistry
9 than those in nitrogen deposition [47, 43], though models differ in the treatment of the
10 marine iron cycle with still large conceptual and parametric uncertainties [48] and in the
11 simulated response of nitrogen fixation to the supply of iron. A robust estimate of the
12 possible contribution of atmospheric iron supply to the observed tropical oxygen decline
13 will require further study.
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23 **(v) Changes related to ocean acidification:** Another anthropogenic impact on ocean
24 biogeochemistry is ocean acidification [49], which may give rise to a number of direct and
25 indirect impacts on ocean biogeochemical cycles [50]. We here investigate the two main
26 potential effects on oxygen, one arising from likely acidification impacts on the production
27 of inorganic carbon [51], the other one on possible impacts on organic carbon production
28 [52].
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33 If calcium carbonate production is attenuated under ongoing acidification, there may be less
34 ballast available to efficiently sink organic particles to great depth [51]. One consequence
35 could be that remineralisation and hence oxygen consumption occurs at shallower depths,
36 possibly leading to a further expansion of low oxygen regions [53]. According to their
37 model simulations, regional oxygen changes in the upwelling area off Peru are in a range of
38 0.2-0.5 $\mu\text{mol kg}^{-1} \text{yr}^{-1}$ and thus similar to observational estimates in that region (Figure 1).
39 The zonally averaged oxygen changes in the tropical thermocline are, however, smaller
40 than 0.04 $\mu\text{mol kg}^{-1} \text{yr}^{-1}$ in the same model, and can therefore explain only a small fraction
41 (<20%) of the observational estimate of thermocline oxygen decline.
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51 Another possible effect of ocean acidification might be the enhanced fertilization of marine
52 algae by increased availability of carbon, leading to increasing carbon-to-nutrient ratios in
53 the biological uptake and possibly export as inferred from mesocosm experiments [52]. A
54 model simulation that extrapolated the local and short-term experimental results to
55 centennial time scales and the global ocean hypothesised substantial effects of increasing
56 carbon to nitrogen ratios in organic matter export on the respiratory oxygen demand and
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hence the extent of oxygen minimum zones [54]. While the volume of the tropical oxygen minimum zones increased by more than 50% in that simulation by the end of the century, changes remain relatively small over the historical time period 1960-2010 (Figure 5b). The model predicts an oxygen decline of about $0.05 \mu\text{mol kg}^{-1} \text{yr}^{-1}$ in the tropical thermocline due to the hypothetical enhancement of C:N ratios over the period 1960-2010, similar in size as the proposed effect due to a predicted reduction in calcium carbonate ballast [53] and again a small portion of the model-data discrepancy. Both acidification-related effects would not be limited to the tropical oceans and be roughly proportional to the amount of export production.

(vi) Potential impacts of fisheries: An anthropogenic forcing that has, until now, not been considered with respect to ocean biogeochemistry is that of industrial fisheries. These are known to have substantially depleted the higher trophic levels over large areas of the global ocean [55]. Whether and to what extent this drastic modification at the upper end of the food chain has had, and will have, significant impacts on marine biogeochemistry and on oxygen levels is currently unknown. General biogeochemical circulation models commonly include only lower trophic levels with zooplankton constituting the highest trophic level. Thus the explicit interaction with fish is generally neglected, but still implicitly included by the closure terms for energy and mass. These are typically prescribed by globally constant linear or quadratic zooplankton mortalities [e.g., 14].

It is not immediately clear how zooplankton biomass will change in response to fisheries. Fishing pressure on large predators can lead to a decrease on the grazing pressure on smaller (planktivorous) fish, which in turn could lead to an increase of the grazing pressure on zooplankton, whereas fishing pressure on smaller fish can lead to a direct decrease of the grazing pressure on zooplankton. Any changes in zooplankton and phytoplankton biomass will affect export production and eventually oxygen utilisation. It is therefore possible that the expansion of fisheries over the historical period has left an imprint on changing oxygen concentrations, the possible extent of which has not been estimated yet.

In a heavily idealised sensitivity study we varied the parameter that determines the strength of the quadratic zooplankton mortality in the UVic model such that changes in zooplankton loss amount to about 10 times the estimated yield of global fisheries (2.3-3.4

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4 TgN/yr [56, 57]), reflecting an assumed 10% transfer efficiency between trophic levels [58].
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6 This was obtained for alterations of the quadratic zooplankton mortality parameter from its
7 standard value of $0.06 \text{ (mmol N/m}^3\text{)}^{-1}\text{day}^{-1}$ to $0.09 \text{ (mmol N/m}^3\text{)}^{-1}\text{day}^{-1}$ (simulation HIGH)
8 and $0.03 \text{ (mmol N/m}^3\text{)}^{-1}\text{day}^{-1}$ (simulation LOW), respectively (Figure 5b). Results of these
9 idealised models show that local changes in export production can have some impact on
10 thermocline oxygen levels. Under historical climate change and an assumed linear increase
11 of industrial fisheries between 1900 and 2000 (staying constant afterwards), the simulated
12 impact on the zonally averaged oxygen changes is of similar magnitude than the impacts
13 found for potential CO₂-induced changes in elemental stoichiometry. A robust estimate
14 about the detailed impact on regional oxygen distributions or even the sign of oxygen
15 change would require a more elaborate food-web model [e.g. 59] and will require further
16 study.
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27 Discussion and implications for future research

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30 We have identified a systematic discrepancy between the observational estimate of tropical
31 thermocline deoxygenation during the past 50 years and the oxygen trend simulated by
32 hindcasts of current state-of-the-art biogeochemical climate models. In general, models
33 driven by anthropogenic CO₂ emissions do not reproduce the observed significant decline
34 in tropical thermocline oxygen levels. Based on our current state-of-the-art models, we
35 cannot, at present, mechanistically link CO₂-induced global warming and tropical upper-
36 ocean deoxygenation. This is consistent with results of a recent fingerprint analysis [60].
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44 A number of sensitivity experiments performed with the UVic model and employing
45 different wind fields as well as parameterisations of anthropogenic perturbations other than
46 CO₂-induced global warming did find systematic impacts of increasing atmospheric
47 nutrient deposition, possible direct CO₂ impacts on marine biogeochemistry, and also
48 possible fisheries impacts on oxygen levels in the tropical thermocline. While the individual
49 effects are, with the possible exception of changes in applied wind-stress fields (Fig.3b),
50 small compared to the model-data misfit (Fig. 5), this reveals that anthropogenic impacts on
51 marine oxygen levels may not be restricted to CO₂-induced climate warming, but that
52 atmospheric nutrient deposition, changes linked to ocean acidification, and impacts of
53 global ocean fisheries may be factors that need to be considered in order to explain
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observed oxygen changes. Other mechanisms not investigated here may also have the potential to impact thermocline oxygen levels. These include potential changes in plankton size [61] or stoichiometry [62], ecological changes with possible impacts on biogeochemistry [63], and also changes in physical transport processes, e.g. via the tropical zonal current bands [64] that are not yet resolved by current models. Table 3 is an attempt to list the different causes and mechanisms of thermocline oxygen changes as well as our current level of understanding.

While the current study relies heavily on sensitivity experiments with a single, relatively coarse-resolution earth system model of intermediate complexity, future research will have to address and hopefully reduce the large uncertainties that remain about the degree of realism in the model description of these individual mechanisms that may have been responsible for observed changes in marine oxygen concentrations. The process that was found to generate largest changes in simulated oxygen trends was variability in the wind forcing, including forcing prior to the observational period. Better observational constraints on winds before the 1960s will help to yield more robust simulations of oxygen changes during the 50 years between 1960 and 2010. Information contained, for example, in ship logbooks, reanalyses combining historical pressure data and dynamical models, and proxy information may all help to improve the historical wind products. The fact that current models do not seem to be able to reproduce the patterns of thermocline oxygen changes that have been identified by decades of high-quality oxygen observations is a scientific challenge that has to be resolved in order to raise confidence in our models' reliability when it comes to simulating the response, and possible feedbacks, of biogeochemical cycles to expected future environmental change.

Additional Information

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6 AL) is acknowledged.
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10 11 **Data Accessibility**

12 CMIP5 model data have been obtained from the Earth System Grid Federation (ESGF) CMIP5
13 repository (e.g. cmip-pcmdi.llnl.gov/cmip5/data_portal.html) and have been accessed in February
14 2014. All own model output reported in this this article is available via
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16 <http://thredds.geomar.de>
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20 21 **Competing Interests**

22 The authors have no competing interests.
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27 **Authors' Contributions**

28 AO conceived the study and drafted the manuscript. SS analysed the historical oxygen data, WK analysed and resampled
29 the CMIP5 modelled oxygen fields, JG and AL ran UVic model simulations, OD ran NEMO simulations. All authors
30 contributed to the analysis, discussion and writing of the paper.
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Tables

Table 1: Model configurations used in Figure 1.

Tab. 1: Brief description of the CMIP5 [Taylor et al., 2012]) and UVic models used in this study. Column (1) indicates the Earth System Model (first line) and the respective biogeochemical ocean module (second line) used. Column (2) indicates vertical and horizontal resolution of the ocean model. For details see the references given in columns (3) and references therein. For our analysis we combined model output from emission driven Earth System Model runs from the historical period (1960-2005; esmHistorical) with runs from the RCP85 scenario (2006-2010; esmrcp85). CMIP5 model data are available from the Earth System Grid Federation (ESGF) CMIP5 repository (e.g. cmip-pcmdi.llnl.gov/cmip5/data_portal.html) and have been accessed in February 2014.

Model	Ocean	Reference
IPSL-CM5A-LR NEMO-PISCES	31 lev, 0.5 - 2°	Dufresne et al., 2013
MPI-ESM-LR HAMOCC 5.2	40 lev, 1.5°	Giorgetta et al., 2013
GFDL-ESM2M TOPAZ2	50 lev, 0.3° - 1°	Dunne et al., 2012
GFDL-ESM2G TOPAZ2	63 lev, 0.3° - 1°	Dunne et al., 2012
CESM-BGC BEC	60 lev, 1.125° × 0.27° - 0.53°	Gent et al., 2011
NorESM1-ME HOMOCC5.1	53 lev, 1.125° x 1.125°	Bentsen et al., 2013
MRI-ESM1 MRI.COMu	50 lev, 1° × 0.5°	Yukimoto et al., 2011
CMCC-CESM Pelagos	31 lev, 2° × 0.5-2°	Vichi et al., 2011
HadGEM2-ES diatHadOCC	40 lev, 1° × 0.3° - 1°	Collins et al., 2011
UVic 2.9 Kiel marine BGC	19 lev, 3.6° x 1.8°	Keller et al., 2012

Table 2: Global mean trend and pattern correlations between observational estimate and simulated thermocline oxygen trends over the period 1960 to 2010.

Data set	Global mean O ₂ trend at 300 m 1960–2010 (μmol kg ⁻¹ yr ⁻¹)	Pattern correlation with observed O ₂ trend at z=300m (50°S–50°N)
Observational estimate (Stramma et al., 2012)	-0.064	
Mean model	-0.045	0.04
UVic	-0.028	-0.32
CESM1	-0.073	-0.03
CMCC	-0.098	0.16
GFDL-ESM2G	-0.0004	-0.05
GFDL-ESM2M	-0.040	0.14
HadGEM2-ES	-0.060	0.03
IPSL-CM5A-LR	-0.055	-0.15
MPI-ESM-LR	-0.024	-0.08
MRI-ESM1	-0.008	-0.09
NorESM1-ME	-0.031	0.02

Table 3: Summary of causes, mechanisms and likely sign of tropical thermocline oxygen changes.

Cause	Mechanism	Sign of O ₂ change	Magnitude	Level of understanding
Warming	Decrease in solubility	-	medium	high
	Enhanced stratification: reduced vertical mixing, less O ₂ ventilation (less O ₂), lower nutrient supply (more O ₂)	+/-	small	low for net effect
Acidification	Reduced CaCO ₃ ballasting: shallower respiration	-	small	medium
	CO ₂ fertilisation	-	small	low
Atmospheric nutrient supply	N fertilisation	-	small	medium
	Fe fertilisation	-	small	medium
Wind-driven circulation changes	Strengthening and poleward displacement of Southern Ocean winds	-	small	medium
	Changes in tropical trade winds	+/-	small	medium
	Changes in low & mid-latitude winds, including wind changes prior to O ₂ changes	+/-	large (?)	low
Fisheries	Impact on pelagic foodweb and carbon export	?	small	very low
Ecological changes	Size, export & remineralisation;	?	?	very low
	Stoichiometry	?	?	very low

Figure and table captions

Figure 1. Linear trend of dissolved oxygen at 300 m depth computed over the 50-year period 1961 to 2010. The top left panel shows the observational estimate of Stramma et al. [4], all other panels show results of numerical model simulations with trends computed from the simulated annual means. Model names indicated at each panel are resolved in Table 1. The “mean model” is the trend averaged over all shown model trends. UVic is the reference simulation by the Kiel version of the University of Victoria model used by Getzlaff et al. [22], all other model data are taken from the Coupled Model Intercomparison Project CMIP5. Units are $\mu\text{mol kg}^{-1} \text{yr}^{-1}$.

Figure 2. Zonally averaged linear trend of thermocline ($z=300\text{m}$) oxygen change over the 1960-2010 period as derived from observations (black) [4], the CMIP5 model mean (red dashed) and the reference UVic model (red solid) [22]. The blue dashed curve shows the zonally averaged trends resulting from the data mapping algorithm applied to the resampled model, with a resampling scheme that corresponds to the time and location of the oxygen observations.

Figure 3. Zonally averaged linear trend of thermocline oxygen ($z=300\text{m}$) as derived from observations (black) and the reference UVic model solution (red), together with (a) oxygen trends simulated by a UVic model run applying observed increasing westerly winds in the Southern Ocean (blue dashed) [22], and (b) results of simulations with CORE wind forcing over the period 1958-2010, and with either CORE wind climatology (cyan for UVic, grey for the somewhat higher resolution 0.5 degree NEMO model) or NCEP wind climatology (dashed blue) used in the model spin-up prior to year 1958.

Figure 4. (a) Annual mean oxygen difference at $z=300\text{m}$ in year 1960 between the runs with NCEP/NCAR wind stress applied during spin up minus the run with CORE wind stress applied. (b) Difference in wind stress applied during the spin up: NCEP/NCAR climatology (1957-1996 period) minus 10-year CORE climatology (1958-1967 period). Length of the reference vector corresponds to 0.1 N m^{-2} . (c) Annual mean oxygen difference between the two runs as in (a) at $z=300\text{m}$ in year 2010 after interannual CORE wind stresses have been applied for the period 1958 to 2010. (d) Difference in the linear oxygen trend over the 50-year period 1960-2010 simulated by the run with NCEP/NCAR climatology spin up minus the run with CORE climatology spin up. Units are $\mu\text{mol kg}^{-1}$ in (a) and (b), and $\mu\text{mol kg}^{-1} \text{yr}^{-1}$ in (c).

Figure 5. Zonally averaged linear trend of thermocline oxygen ($z=300\text{m}$) as derived from observations (black) and the reference UVic model solution (red), together with (a) results of UVic model simulations applying an observational estimate of atmospheric nitrogen deposition (dashed blue) and CO_2 dependent variable C:N stoichiometry (green) [54], and (b) results of idealised sensitivity simulations with increased (dashed blue) and decreased (dashed red) zooplankton mortality (see text).

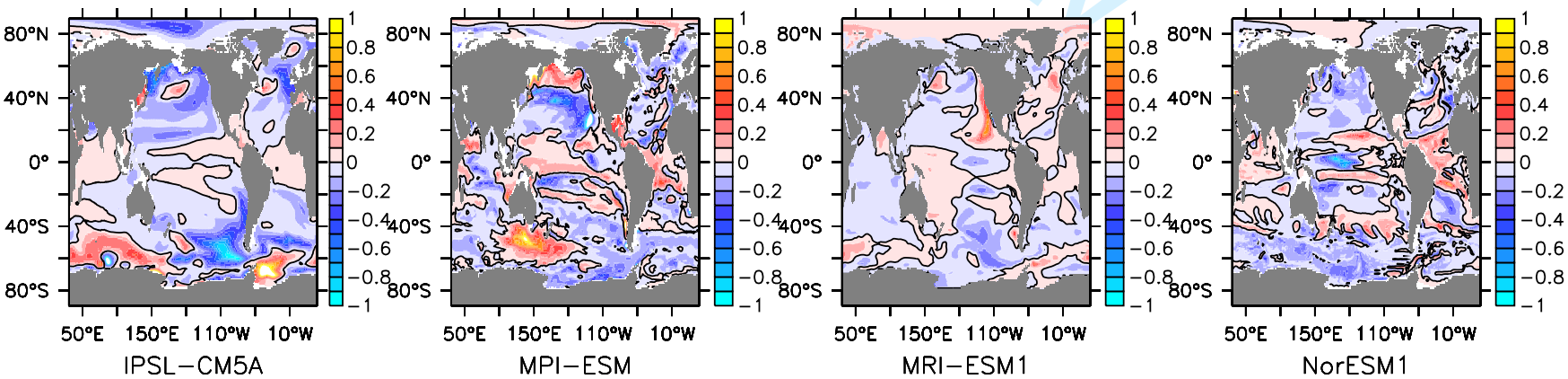
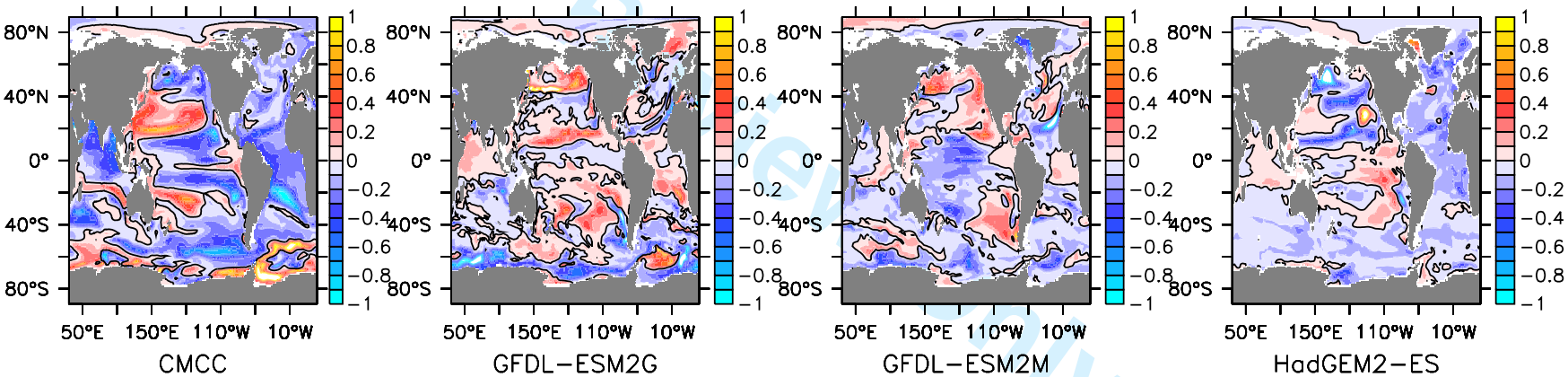
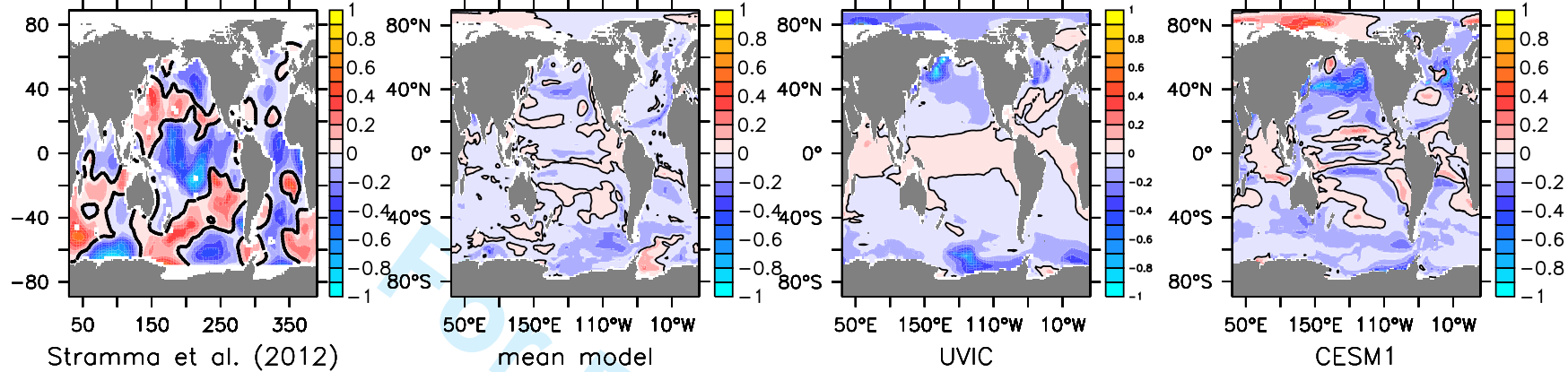
Table 1. List of the models used.

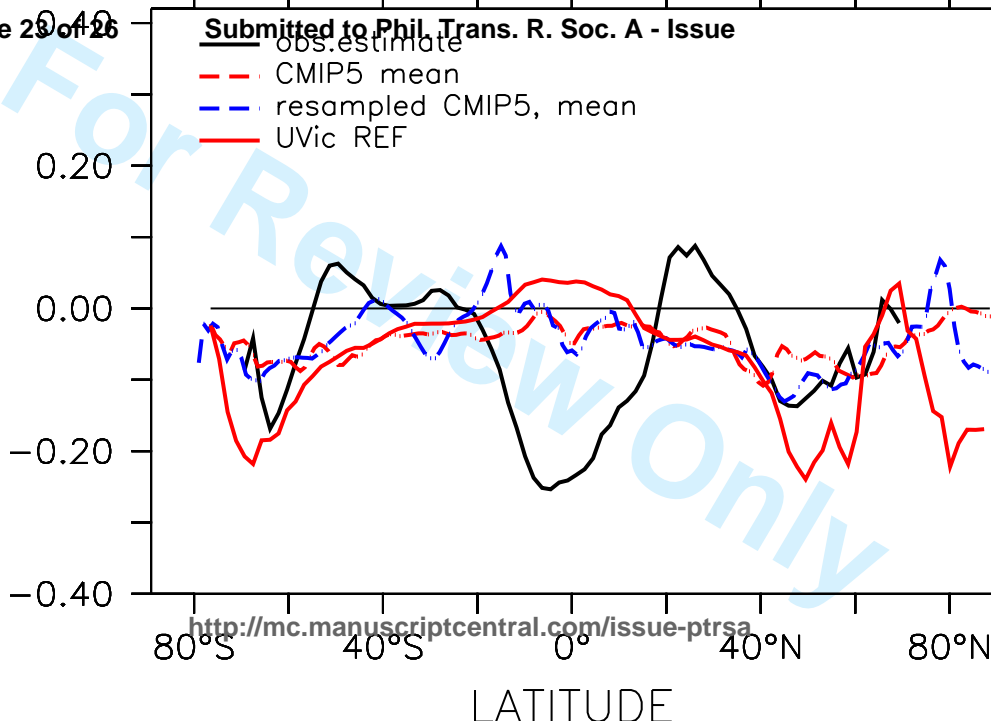
Table 2. Mean trend and pattern correlations between observational estimate and simulated thermocline oxygen trends over the period 1960 to 2010.

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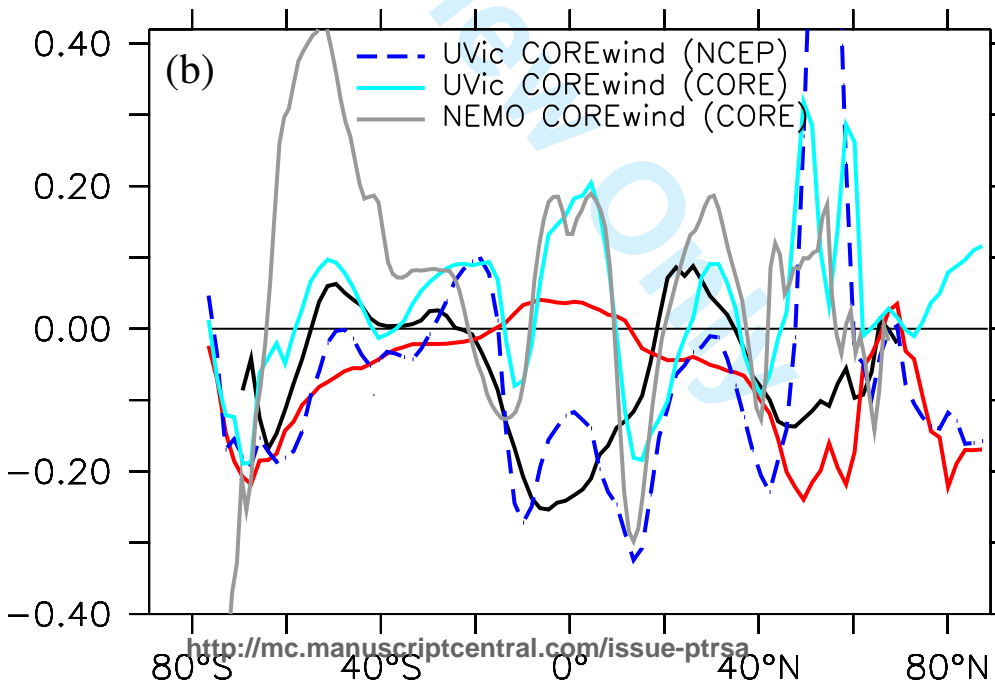
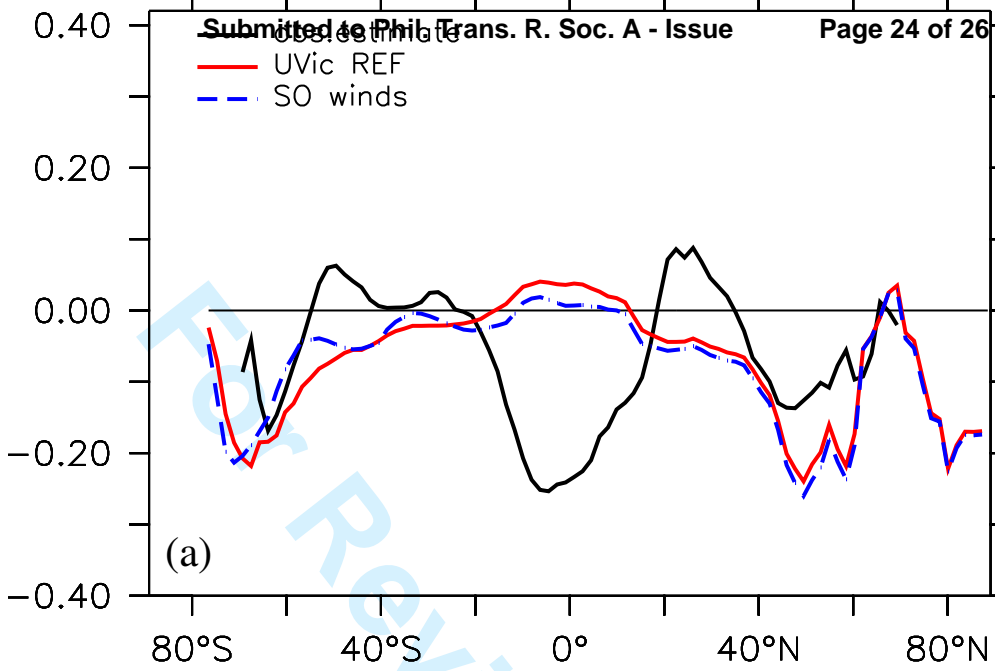
Table 3: Summary of causes, mechanisms and likely sign of tropical thermocline oxygen changes.

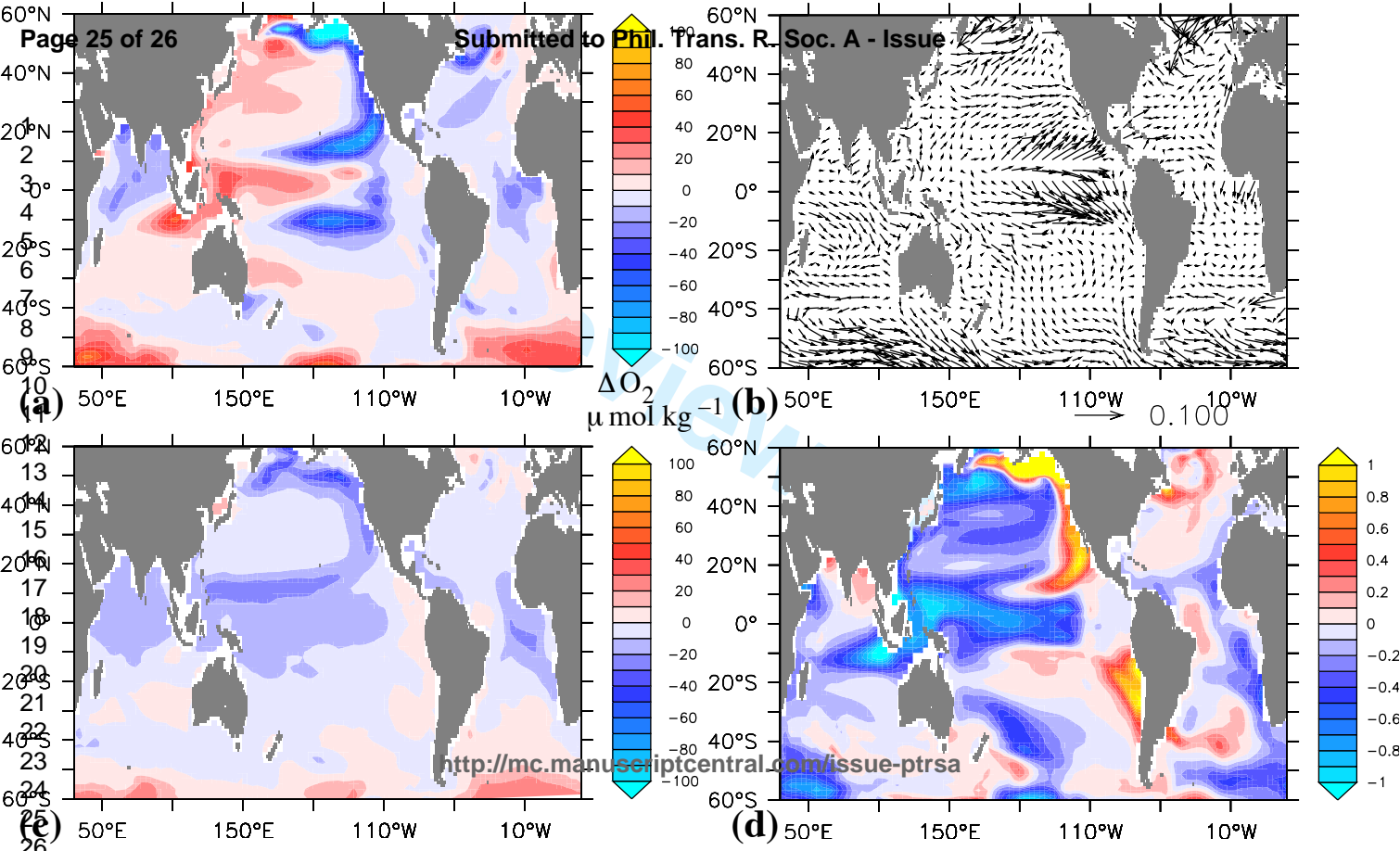
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