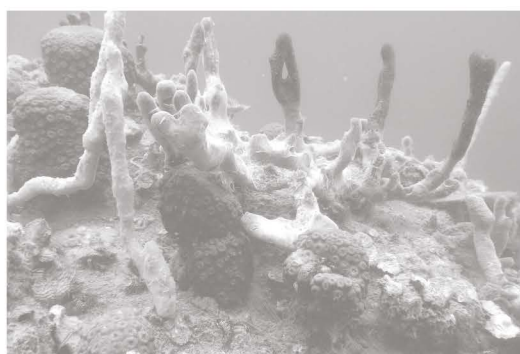# IUCN

Ocean deoxygenation: Everyone's problem

Causes, impacts, consequences and solutions

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3.1 Ocean deoxygenation from climate change

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Summary

- According to the most recent observational estimate, the ocean lost 2% of its oxygen inventory between 1960 and 2010.
- About 15% (range 10-30%) of the oxygen loss is attributed to warming-induced decline in solubility (high confidence).
- Less than 15% of the oxygen decline can be attributed to warming-induced changes in respiration of particulate and dissolved organic matter. Enhanced respiration will tend to generate oxygen deficits close to the sea surface. The increased near-surface vertical oxygen gradient may even increase ocean uptake of oxygen from the atmosphere (low confidence).
- The majority of oxygen loss has been caused by changes in ocean circulation and associated ventilation with oxygen from the ocean surface (medium confidence).
- Current state-of-the-art models simulate deoxygenation rates more than 2 times smaller than the most recent data-based global estimate.

Processes	Likely impacts on deoxygenation
Increasing temperatures reduce the solubility of oxygen (and other gases) in water.	• Explains about 50% of oxygen loss in the upper 1000 m, corresponding to a solubility-driven oxygen loss of about 0.013 Pmol O_2 yr^{-1}). • Until now the solubility-driven contribution to oxygen loss below 1000 m depth amounts to about 2% (about 0.001 Pmol O_2 yr^{-1}). • According to the most recent estimate of global deoxygenation (0.096 ±0.042 Pmol O_2 yr^{-1}), solubility changes account for 15% (range 10-30%) of the total oxygen loss during 1960-2010.
Warming-driven increase in respiration rates of particulate organic matter.	• Metabolic rates increase with temperature, so will respiration rates. Remineralization of sinking organic particles will therefore tend to occur at shallower depths, generating oxygen deficits closer to the air-sea interface and tending to increase the oxygen flux from the atmosphere into the ocean. • Overall effect computed from model simulations expected to be a small (< 0.005 Pmol O_2 yr^{-1}) oxygen gain.
Reduced burial in the sediments.	• By enhancing respiration rates in the water column and in the sediment, ocean warming is expected to reduce the amount of organic matter buried in the sediments. Less burial means that more organic matter is respired, generating additional respiratory oxygen loss in the water column. • An upper bound is a complete cessation of burial, which would correspond to a pelagic oxygen sink of about 0.002 Pmol O_2 yr^{-1}, i.e. 2% of the total oxygen loss estimated for 1960-2010.
Warming-driven increase in the respiration of dissolved organic matter.	• Respiration rates of dissolved organic matter (DOM) are expected to increase with temperature. Estimates are difficult because of the poorly known composition and biological accessibility of DOM. Incubation experiments indicate some warming-induced acceleration of respiration, which might explain up to 10% or 0.01 Pmol O_2 yr^{-1} of the observed oxygen loss. A corresponding decline of the ocean's DOM inventory has not yet been observed.
Respiration of enhanced release of methane from melting gas hydrates.	• Warming of bottom waters may result in enhanced destabilization of methane gas hydrates, leading to enhanced release of methane from sediments and subsequent aerobic respiration of methane to CO_2. • There is little observational evidence for a warming-induced acceleration of methane release taking place already.
Circulation changes in response to global warming.	• As the ocean warms from the surface, stratification is expected to increase, with a tendency for a slowing down of the ocean circulation. A slowed-down circulation is expected to account for up to 50% of the observed deoxygenation in the upper 1000 m, and for up to 98% in the deep ocean (> 1000 m). Spatial patterns and individual mechanisms are not yet well understood.

3.1.1 Introduction

Ocean deoxygenation is caused by an imbalance of sources and sinks of oxygen dissolved in sea water. This section addresses how individual oxygen sources and sinks can vary in response to climate change. Based on our still immature quantitative understanding of the climate sensitivity of the various processes at play, estimates will be provided, wherever possible, to what extent past and likely future changes in marine dissolved oxygen can be attributed to individual mechanisms and, ultimately, to anthropogenic climate change.

Waters of the surface ocean are essentially in equilibrium with the oxygen partial pressure of the atmosphere that contains more than 99% of the molecular oxygen available on Earth. Burning fossil fuels consumes oxygen, but because of the large atmospheric oxygen inventory, the associated relative decline of atmospheric oxygen (0.001% per year) is about two orders of magnitude smaller than the current rate of ocean deoxygenation. Air-sea gas exchange typically equilibrates tens of metres of surface waters within days (Kihm & Körtzinger, 2010), with warmer waters having a lower solubility and therefore, at the same saturation levels, holding less oxygen than colder waters. Oxygen concentrations in the surface ocean are therefore highest in cold high-latitude waters (Figure 3.1.1) where deep waters form and thereby transport oxygen into the deep ocean interior to depths well exceeding 1000 metres. Waters above the deep waters of polar origin and below the surface mixed layer belong to the stratified thermocline

that is ventilated via a mostly wind-driven circulation that moves water equatorward and westward along isopycnals in large anticyclonic gyres in the subtropical and tropical oceans (Luyten et al., 1993).

Photosynthesis by marine phytoplankton is an oxygen source within the ocean's light-lit surface waters in close contact with the atmosphere. On regional and seasonal scales, photosynthesis typically makes up for a small fraction of the total air-sea oxygen exchange (Garcia & Keeling, 2001). There are no significant oxygen sources in the dark ocean interior. Here, oxygen can be supplied only via physical processes that transport oxygen-rich waters from the surface to depth (Figure 3.1.2).

The main marine oxygen sink is the respiration of organic matter by microbes and multicellular organisms. The total amount of respiration is controlled by the availability of organic matter, which ultimately depends on primary production in the surface ocean, often limited by nutrients, light or both (Arteaga et al., 2014). Changes in surface-ocean environmental conditions can thus cause changes in oxygen consumption in the ocean interior. By forming particles that sink through the water instead of moving with it, the marine biology can lead to relatively fast changes in respiration in ocean-interior water parcels that may have been supplied with oxygen via direct contact with the atmosphere the last time hundreds of years ago. Changes in particle properties and in respiration rates may also affect the amount of organic matter that is buried in the sediments. Since photosynthetic production of organic
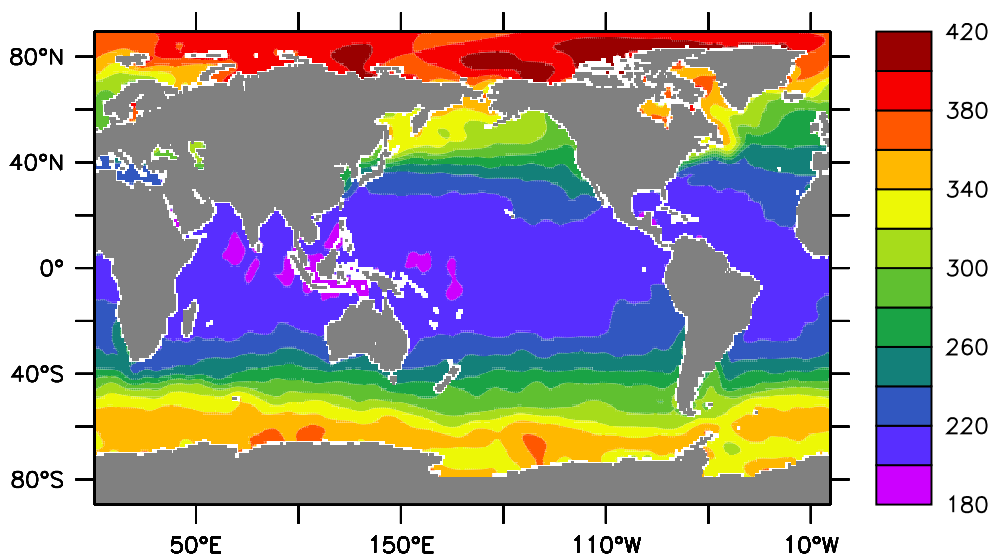

Figure 3.1.1 Global distribution of annual-mean oxygen concentrations in the surface waters compiled from the World Ocean Atlas 2009 (Garcia et al., 2010). Units are µmol kg^{-1}.
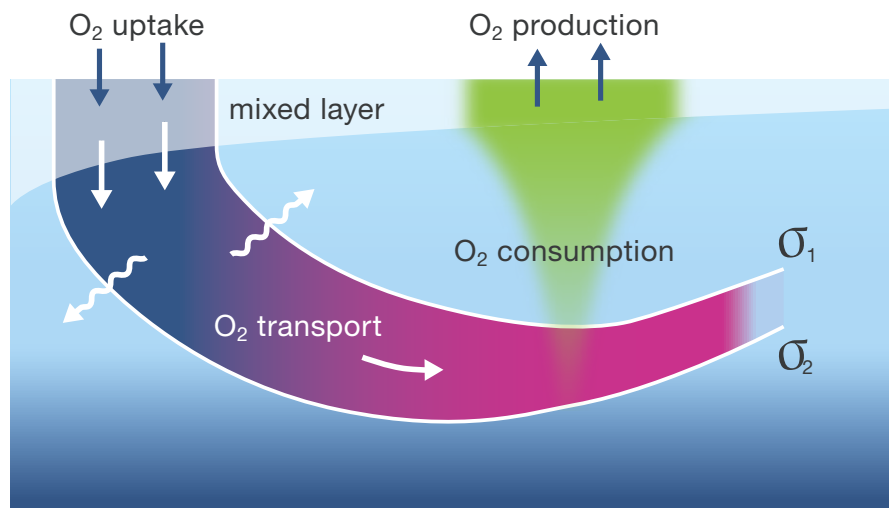
Figure 3.1.2 Schematic representation of oxygen supply to the ocean interior via physical transport that occurs predominantly along isopcynal surfaces (σ_i) and oxygen consumption by respiration of organic matter exported downward out of the surface ocean.

matter generates oxygen, the subsequent burial and the associated avoidance of respiration constitutes a net oxygen source. Thus, a reduction in burial rates or increase in remobilization from the sediments may contribute to marine deoxygenation. The same holds for a hypothetical reduction in anaerobic remineralization, e.g. by denitrification, at the expense of more aerobic remineralization. However, as denitrification occurs in low-oxygen environments, it is currently estimated to increase rather than decrease. Another biological process that may reduce marine oxygen levels is the microbial oxidation of methane released from gas hydrates found in the sediments at water depths of a few hundred metres and that may destabilize under increasing bottom water temperatures in response to global warming.

The ocean oxygen inventory can thus change via three main pathways:

i. Changes in initial oxygen concentrations in surface waters in immediate contact with the atmosphere.
ii. Changes in respiratory oxygen consumption in the ocean interior.
iii. Changes in ventilation via mixing and circulation that affect the time span, during which respiratory oxygen losses can accumulate in ocean-interior waters since their last contact with the atmosphere.

Since the oxygen content of sea water has been measured accurately for over one hundred years, changes can be derived from compilations of such observations. Schmidtko et al. (2017) estimate a decline

in the ocean's oxygen inventory by about 2% between 1960 and 2010. Current climate-biogeochemistry models, however, do not adequately reproduce observed patterns of oxygen changes and, overall, tend to underestimate oxygen variability and trends (Oschlies et al., 2018). In particular, models simulate a decline in the global ocean's oxygen inventory by typically less than 1% over the past 50 years (Table 3.1.1; Bopp et al., 2013). On average, simulated deoxygenation is more than two times slower than the most recent observational estimate. This discrepancy indicates that an accurate quantitative understanding of ocean deoxygenation is still lacking, and that current estimates of contributions from individual processes and even causes must be regarded with some degree of caution. The following presents observational and theoretical evidence regarding climate-related mechanisms of oxygen changes and discusses the extent to which ocean deoxygenation can be linked to anthropogenic climate change.

3.1.2 Temperature effects on solubility

The solubility of oxygen in sea water determines the concentration of dissolved oxygen in the surface water that is in equilibrium with the overlying atmosphere. It thereby controls the supply of oxygen to the ocean – point (i) above. Oxygen solubility is, like for any gas, a function of temperature and salinity, and decreases with increasing temperature and salinity. For example, oxygen saturation decreases by about 5% for a 2 °C warming from 4 °C to 6 °C. For current climate change, effects of changes in salinity are generally many times

Table 3.1.1 Global ocean oxygen inventory (Pmol), trend (Tmol/decade) and percentage change over the past 50 years as derived from observations (Schmidtko et al., 2017), as simulated by the nine models of the 5th phase of the Climate Model Intercomparison Project (CMIP5) that contain oxygen as tracer (more detailed information about the models is provided in Table 1 of Oschlies et al., 2017), and three simulations of the University of Victoria Earth system climate model of intermediate complexity (UVic 2.9) also described in more detail in Oschlies et al. (2017). Standard deviations as given by Schmidtko et al. (2017) for the observational estimate, and the variance among the 9 CMIP5 models, respectively.

Model	Inventory Pmol	Tmol/decade	% change over 50 yr
CESM1	260,2	-420,6	-0,81
CMCC CESM	271,8	-893,6	-1,64
GFDL ESM2G	275,7	-215,5	-0,39
GFDL ESM2M	244,8	-356,1	-0,73
HadGEM2 ES	192,3	-229,4	-0,60
IPSL CM5A LR	199,3	-529,4	-1,33
MPI ESM LR	255,5	-214,5	-0,42
MRI ESM1	320,4	-30,9	-0,05
NorESM1 ME	342,7	52,4	0,08
UVic 2.9	225,6	-463,2	-1,03
model average	258,8	-330,1	-0,64
Obs. estimate	227,4	-960	-2,11

smaller than the temperature-induced changes (Weiss, 1970). By analysing temporal changes in temperature, one can directly compute the solubility-driven component of oxygen changes. This approach has been used, among others, by Helm et al. (2011) and Schmidtko et al. (2017), who inferred that about half of the oxygen decline measured in the upper 1000m over the past decades could be attributed to temperature-driven changes in solubility (Table 3.1.2). This explains part of the tight relationship between decreasing O_2 inventories and increasing ocean heat content that has been described by Ito et al. (2017) for the upper ocean. Integrated over the entire water column, Schmidtko et al. (2017), however, attribute only 15% of their estimate of oceanic oxygen loss to direct solubility effects. This means that oxygen changes in the deep ocean are, until now, almost independent of solubility changes. This is consistent with the fact that solubility effects originate at the sea surface and have to be transported physically with the circulation into the ocean interior. Water ages in the deep ocean are too old to be substantially affected by anthropogenically driven solubility changes, yet.

3.1.3 Temperature effects on metabolic rates

Chemical rates generally increase with increasing temperature (Arrhenius, 1889). This applies to both primary production (Eppley, 1972) and respiration

(Pomeroy & Wiebe, 2001). Primary production is, in most regions of the world ocean, limited by resources (i.e. nutrients or light) and not temperature (Maranon et al., 2014). Similarly, the total amount of marine respiration is primarily limited by the availability of substrate, i.e. organic matter. In the current ocean, almost all organic matter exported from the ocean's surface layer is respired, except for a very small portion that is buried in the sediments that corresponds to a net oxygen production of 0.002 Pmol O_2 yr^{-1} globally by avoided aerobic respiration (Wallmann, 2010), a rate that is 50 times smaller than the total ocean oxygen loss of 0.096 Pmol O_2 yr^{-1} estimated by Schmidtko et al. (2017).

A likely impact of warming-induced acceleration of metabolic rates is a systematic change in the spatial patterns of respiration and associated oxygen consumption in the water column. The decrease of respiration with depth, z, was first described empirically by an exponential function in the form of $R=R_0e^{-a/z}$ (Wyrtki, 1962) and later replaced by a power law function $R=R_0(z/z_0)^{b-1}$ (Martin et al., 1987) that was found more appropriate to describe measurements of the downward flux of particulate organic matter over the entire water column. As noted by Brewer and Peltzer (2017), such temperature-independent empirical descriptions, that are still used in a number of conceptual and numerical models of marine biogeochemical

Table 3.1.2 Approximate contribution of individual mechanisms to the observed ocean deoxygenation between 1960 and 2010, referenced to the recent deoxygenation estimate by Schmidtko et al. (2017). Observational evidence is very strong for the solubility driven contribution (+++), positive for ventilation changes (+), plausible for temperature effects on oxidation of dissolved organic matter and methane (o) and absent for temperature effects on remineralization profiles and burial of organic matter (-).

Process	% upper 1000m (0.026 Pmol O_2 yr^{-1})	% below 1000m (0.070 Pmol O_2 yr^{-1})	% total (0.096 Pmol O_2 yr^{-1})	Observational evidence
Solubility	50%	2%	15%	+++
Temperature effects on remineralization profile and air-sea exchange of O_2			up to -5%	-
Potential warming-induced cessation of burial	< 6%	< 0.3%	< 2%	-
Temperature effects on remineralization of dissolved organic matter			up to 10%	0
Temperature effects on methane oxidation	0	0	0	0
Circulation changes	up to 50%	up to 98%	up to 85%	+

cycles, lead to difficulties in accounting for impacts of warming on respiration and associated effects on tracer distributions. Differences in the consideration of temperature effects can have substantial impacts on projected changes in marine biogeochemical cycles under global warming (Taucher & Oschlies, 2011). In a model study, Segschneider and Bendtsen (2013) applied a temperature perturbation, corresponding to business-as-usual global warming at the end of the 21st century, to remineralization rates. They found that accelerated remineralization of sinking organic matter at higher temperatures made respiration happen at shallower depths. In their model, oxygen concentrations were reduced by a few mM in the upper few hundred metres in the simulation using projected end-of-21st century temperatures based on the RCP 8.5 emissions scenario compared to pre-industrial temperatures. Because organic matter was respired at shallower depths under elevated temperatures, less substrate was exported to greater depth where respiration was reduced and oxygen concentrations were elevated. Assuming that primary production is resource-limited and not temperature-limited and therefore unaffected by warming, Segschneider and Bendtsen (2013) showed that metabolic effects of global warming lead to elevated CO_2 and reduced O_2 concentrations in the near-surface waters resulting from shallower respiration. This leads to a net outgassing of about 0.2 Pg C yr^{-1}, corresponding to a net oceanic oxygen flux into the ocean of 0.02 Pmol O_2 yr^{-1} in their model study simulating end of 21st century ocean warming. The direct effect of warming on the cycling of freshly produced organic matter is thus expected to be a net oceanic oxygen gain rather than a loss. Given that less than one quarter of the oceanic heat uptake simulated by the year 2100 has happened before the year 2010, this would compensate for only a small fraction (less than ~0.005 Pmol O_2 yr^{-1}, i.e. about 5%, Table 3.1.2) of the observed estimate of total ocean deoxygenation over the past 50 years.

Since respiration will tend to be faster in a warmer ocean, less organic matter is left over to reach deeper waters and the sea floor. This might affect burial and thereby have a small net effect on total oxygen consumption. However, even if burial of organic matter would cease completely and all organic matter would instead be respired in the water column, this could account for only 2% of the deoxygenation observed over the past 50 years. As only about 10% of the burial occurs in deep waters (Wallmann, 2010), cessation of burial might have largest impact on upper ocean deoxygenation (Table 3.1.2).

Substrate limitation therefore rules out major effects of temperature-driven enhancements of metabolic rates on globally integrated decomposition, and associated oxygen consumption, of freshly produced organic matter. However, enhanced respiration of old and more refractory dissolved organic carbon (DOC), that constitutes a significant oceanic carbon pool of more than 600 Pg C (Hansell et al., 2009), under increased temperatures may lead to a net oxygen decline. While the oceanic DOC pool is composed of a wide spectrum of largely uncharacterized organic compounds, it is still

unclear what limits its consumption and to what extent warming may accelerate the respiration of DOC. If the deoxygenation signal observed between 1960 and 2010 were to be explained solely by respiration of DOC, about 10% of the oceanic DOC pool would have been lost already, for which there is no observational evidence.

From incubation experiments, Bendtsen et al. (2015) inferred that the respiration of the more recalcitrant components of DOC (lifetime of months) increased about twice as fast with increasing temperatures (Q_{10} > 5) than respiration of the more labile DOC pool (e-folding lifetime of days, Q_{10} ~ 2). Possible effects of a warming-induced acceleration of DOC remineralization in the open ocean are still uncertain. For coastal waters in the Baltic Sea – North Sea transition zone, Bendtsen and Hansen (2015) simulated a decline in oxygen concentrations by 30-40 µM for an assumed 3 °C temperature rise. Assuming this as a globally representative value and relating to the heat required to warm a litre of sea water by 3 °C (11.550 J) yields a deoxygenation rate by enhanced respiration of DOC of about 0.4-0.5 nmol O_2 J^{-1}, which is an order of magnitude smaller than the observed relationship between deoxygenation and oceanic heat uptake (2-10 nmol O_2 J^{-1}) for an older deoxygenation estimate (Keeling & Garcia, 2002) that approximately doubles for the most recent deoxygenation estimate of Schmidtko et al. (2017) and is consistent with model-derived estimates of 3.7 ± 0.8 nmol O_2 J^{-1} reported by Resplandy et al. (2018)). If extrapolated to the global ocean, current estimates of local enhancement of DOC respiration by warming may thus explain as much as 10% of the observed ocean deoxygenation. More studies are required to investigate this further.

3.1.4 Oxygen demand of methane oxidation

Oxidation of methane released from sediments at depths of a few hundred metres presents another oxygen sink that may increase when bottom waters warm and thereby destabilize some of the methane hydrates, leading to outgassing of methane from the sea floor, of which most is expected to be oxidized in the water column to CO_2 (Kessler et al., 2011; Mau et al., 2017). Methane hydrate destabilization is expected to begin this century (Biastoch et al., 2011), but there is no observational evidence for warming having caused, until now, enhanced outgassing of methane from the sea floor (Mau et al., 2017). It is therefore unlikely that enhanced methane release from the sea floor and subsequent oxidation has significantly contributed to the ocean deoxygenation observed until now. Assuming sustained future warming in 2 x CO_2 and 4 x CO_2 scenarios, however, model simulations predict that warming-induced release of methane and its subsequent oxidation in the water column may lead to substantial oxygen losses, that can amount to 10-30% of solubility driven oxygen losses (see above) on millennial timescales (Yamamoto et al., 2014).

3.1.5 Effects of stratification and circulation changes

As well as the direct effects of warming on marine oxygen sources and sinks, there are indirect effects of warming via changes in transport brought about by changes in ocean stratification and circulation. Climate model simulations show a close correlation between declining oxygen levels and increasing water ages on isopycnal surfaces in the thermocline (Long et al., 2016). Because of the scarcity of abiotic transient tracer observations required to derive water mass ages, this inference is, until now, only based on model simulations. It does, however, suggest that reduced ventilation – point (iii) above – is a dominant driver of upper-ocean deoxygenation. A quantitative and mechanistic understanding of the individual processes is still lacking and attribution to different forcing agents such as anthropogenic warming, internal climate variability or a combination of both (i.e. changing internal variability in a warming world) remains challenging (Andrews et al., 2013).

Ocean ventilation can change as a result of buoyancy forcing in the form of changes in heat and water cycles, and as a result of changes in wind patterns. Changes occur on a multitude of space and time scales and may be linked to anthropogenic climate change and/or natural variability of the climate system, e.g. related to the Pacific Decadal Oscillation (PDO), El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO) or Atlantic Multidecadal Oscillation (AMO). Up to now, results of numerical models do not generally agree well with observed patterns of oxygen change in the upper ocean (Oschlies et al., 2017), suggesting systematic deficiencies in current models and, presumably, our understanding. The following is an attempt to review what is known about the contribution of individual ventilation mechanisms to ocean deoxygenation.

As the ocean warms from above, stratification tends to increase. At high latitudes, particularly in the northern

Figure 3.1.3 Warming induced glacial melt © Dan Laffoley.

hemisphere, stratification increases also as a result of enhanced freshwater discharge due to warming-induced glacial melting (Figure 3.1.3). Overall, a stronger stratification suppresses downward transport of oxygen-saturated surface waters and upward transport of oxygen-deficient deeper waters to the sea surface. This leads to a net reduction in the air-sea flux of oxygen (Figure 3.1.4) and the transport of newly ventilated oxygen-rich waters into the ocean interior. This is, however, partly compensated by reduced upward transport of nutrients and an associated decline in biological production and the export of organic matter from the productive surface zone to depth (Bopp et al., 2013) and its subsequent remineralization in the ocean interior. Because respiratory oxygen demand depends on the stoichiometric composition of organic matter, as yet uncertain effects of elevated CO_2 concentrations on carbon-to-nutrient ratios in the exported organic matter may have considerable effects on oxygen consumption (Oschlies et al., 2008). As such stoichiometric effects would be related to elevated CO_2 and not to climate change per se, they will not be considered further here.

Besides increasing upper-ocean stratification, surface warming has also been linked to a poleward migration of isopycnal outcrops (Durack & Wijffels, 2010), which tends to increase the pathways and transit times of waters in the ventilated thermocline. With buoyancy fluxes setting the transformation of water masses in the surface mixed layer, the circulation in the ocean's thermocline is predominantly wind-driven (Luyten et al., 1983). Of prime importance for the ventilation of the thermocline are changes in wind patterns. These are often related to modes of natural climate variability (e.g. PDO, ENSO, NAO, AMO) and the separation from wind changes due to global warming is challenging against the large internal variability of the climate system. A warming-related strengthening has been reported for the Pacific trade winds since the early 1990s (England et al., 2014) and the Southern Ocean westerlies since the 1960s (Marshall, 2003; Thompson & Solomon, 2002).

Regarding the recent strengthening of the Pacific trade winds, Ridder and England (2014) inferred from a model study that the oceanic oxygen content decreases with increasing tropical winds, as a result of enhanced upwelling, enhanced biological production, and enhanced respiration. This interpretation focuses on wind-driven changes in oxygen consumption, in line with a regional study of the subtropical North Pacific
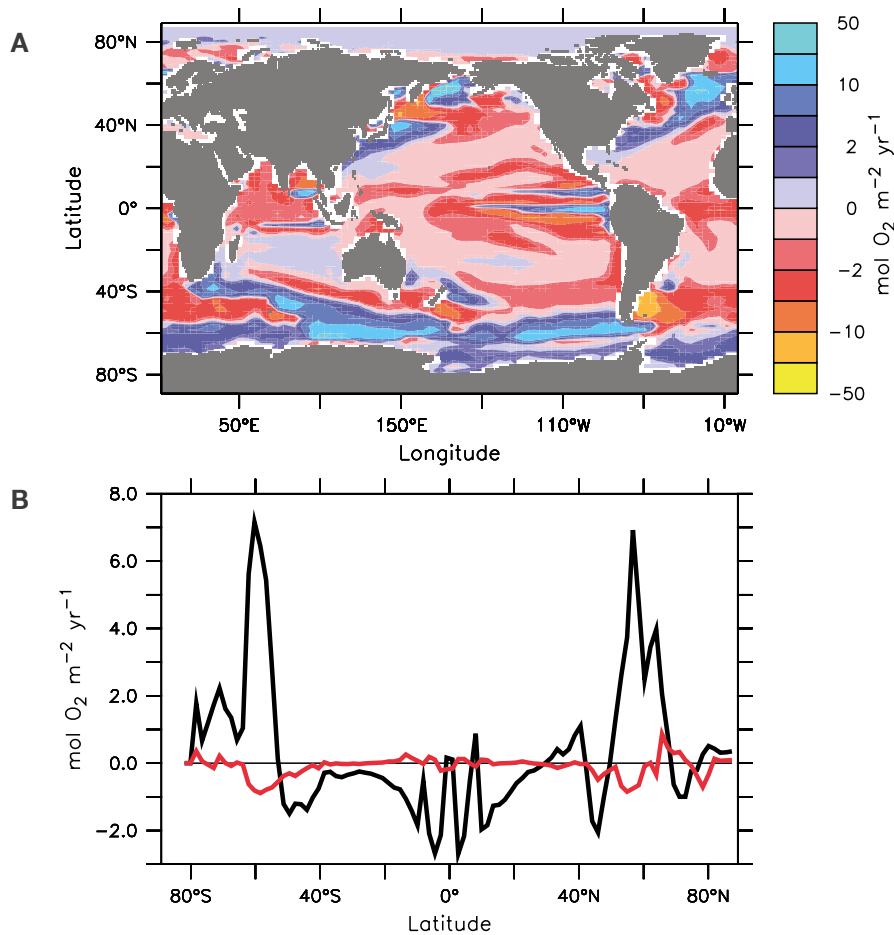
Figure 3.1.4 (A) Annual-mean air-sea oxygen fluxes averaged over the 50-year period 1960-2010 as simulated by the UVic model (reference simulation of Oschlies et al. (2017) using climatological wind forcing). (B) The same air-sea oxygen flux shown in zonally averaged form (black) together with the zonally averaged simulated linear trend in the air-sea oxygen flux integrated over the period 1960-2010 (red).

by Deutsch et al. (2014). Based on observations of sedimentary nitrogen isotopes and a biogeochemical ocean model forced with atmospheric reanalysis fields, they concluded that the volume of low-oxygen waters in the eastern subtropical North Pacific contracted during much of the 20th century and only expanded after about 1990 under strengthening trade winds (with no net change over the past 150 years). In their analysis, the volume of low-oxygen waters was inversely correlated with the depth of the thermocline, with greater depths and smaller low-oxygen volume during warm phases under more sluggish trade winds, and with shallower depths and larger low-oxygen volume during colder phases under stronger trade winds. This interpretation, based solely on changes in oxygen demand, has been challenged by a model study of Duteil et al. (2014) who stressed the importance of changes in the oxygen supply. Besides intensifying upwelling, biological production and respiratory oxygen consumption, stronger trade winds also lead to an intensified wind-driven shallow overturning circulation

and enhanced supply of waters that are relatively rich in oxygen to the tropical thermocline. Changes in wind-driven tropical ocean oxygen supply and demand thus tend to counteract. Yang et al. (2017) demonstrated in their model analysis that impacts of changes in local respiration and oxygen supply via mixing and ventilation tend to have opposing effects on oxygen levels in the eastern tropical Pacific thermocline and also on shorter variations associated with El Nino events.

Regarding the recent and possible future enhancement of Southern Ocean westerlies, a model study by Getzlaff et al. (2016) identified an enhanced ventilation of the thermocline with oxygen-rich southern hemispheric mode waters, leading to a small increase in the global ocean's oxygen content that counteracts up to 10% of the overall warming-induced deoxygenation. The same study also revealed that the observed poleward displacement of the southern hemispheric wind system leads to longer ventilation pathways and regionally lower

oxygen concentrations in the waters arriving from higher southern latitudes in the tropical ocean's thermocline.

According to climate models, enhanced freshwater discharge into the polar oceans results in reduction in the deep ocean overturning circulation (Stouffer et al., 2006). Recent observations indicate some decline of the Atlantic meridional overturning circulation against a background of large intra- and inter-annual variability (Srokosz & Bryden, 2015). The meridional overturning circulation associated with the formation of Antarctic Bottom Water is globally of similar strength compared to the overturning circulation of North Atlantic Deep Water. During recent decades a general warming of these bottom water masses spreading northward into the Pacific, Atlantic and Indian Oceans together with a general reduction in overturning strength has been observed (Johnson et al., 2008; Purkey & Johnson, 2010). This is mirrored by an oxygen decrease by more than 0.012 Pmol yr^{-1} in the Southern Ocean below 1200 m, i.e. more than 10% of the global ocean oxygen loss (Schmidtko et al., 2017). As deduced from a modelling study by Matear et al. (2000), oxygen changes in the Southern Ocean are predominantly caused by circulation changes, with changes in export production and respiration playing only a minor role. Interestingly, the deoxygenation signal is not apparent in an analysis of oxygen concentrations of retreating Antarctic Bottom Water (Andrie et al., 2003; van Wijk & Rintoul, 2014), limiting the decrease to the deep-water masses above the newly formed bottom water. Both, the replacement of well oxygenated bottom waters by older water masses as well as the accumulating oxygen consumption in these ageing water masses lead to the oxygen decrease observed in the deep waters. According to Schmidtko et al. (2017) almost three quarters of the total oceanic oxygen loss occur in waters deeper than 1200m. With negligible solubility effects at these depths and little direct observational evidence for biological effects, these deep-ocean oxygen changes appear essentially driven by changes in the overturning circulation (Table 3.1.2).

3.1.6 Conclusions

Recent observational estimates indicate that the marine oxygen inventory has decreased by about 2% over the past 50 years, with three quarters of the oxygen loss appearing to be located at depths below 1200 m (Schmidtko et al., 2017). Warming is considered to be a major driver, although direct solubility effects explain only about 15% of the total changes (50% in the upper 1000 m, Helm et al., 2011; Schmidtko et al., 2017). The remaining part is thought to result from indirect warming effects on stratification and circulation, possibly respiration, changes in marine biology and biogeochemical feedbacks. It is noteworthy that current state-of-the-art models simulate an ocean-wide oxygen decline of only about one third of the observed rate (Table 3.1.1). Agreement is better in the upper ocean with models reaching, on average, 70% of the observed oxygen decline at 300 m depth even though the spatial patterns of oxygen change are not reproduced well (Oschlies et al., 2017). Models also tend to underestimate the interannual to decadal variability of oxygen concentrations at subtropical time-series sites in the Pacific and Atlantic (Long et al., 2016) and of air-sea fluxes inferred from time-series observations in the equatorial Pacific (Eddebbar et al., 2017).

The systematic underestimation of temporal variability and trends in marine oxygen levels by current models is a strong indication of gaps in our understanding of the mechanisms of marine deoxygenation. There may be systematic deficiencies in the models, e.g. too coarse resolution and too diffusive physics, or unaccounted variations in respiratory oxygen demand. Additionally, amplifying feedback processes may not be properly accounted for. These include biogeochemical feedbacks via the release of nutrients, in particular phosphate and iron, from sediments turning anoxic (Niemeyer et al., 2017). Other, impacts not directly related to climate change, such as eutrophication by atmospheric pollutants (Ito et al., 2016), possible stoichiometric effects of elevated CO_2, or even impacts of industrial fisheries (Getzlaff & Oschlies, 2017) may contribute to ocean deoxygenation. According to current models, we can, until now, attribute at least one third of the observed oxygen loss of about 0.096 Pmol yr^{-1} to anthropogenic climate change but cannot rule out that climate change explains essentially all of the observed oxygen loss.

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