



# Ocean deoxygenation: Everyone's problem

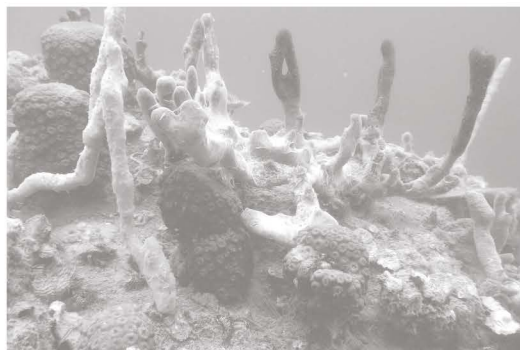
Causes, impacts, consequences and solutions

Edited by D. Laffoley and J.M. Baxter



## 2.1 Global evidence of ocean deoxygenation

Lothar Stramma and Sunke Schmidt



IUCN GLOBAL MARINE AND POLAR PROGRAMME





## 2.1 Global evidence of ocean deoxygenation

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### Summary

- The global oxygen inventory has decreased by ~2% over the period 1960 to 2010, this finding is supported by regional time series data that indicate a continuous decrease in oceanic dissolved oxygen.
- Ocean model simulations predict a decline in the dissolved oxygen inventory of the global ocean of 1 to 7% by the year 2100, caused by a combination of a warming-induced decline in oxygen solubility and reduced ventilation of the deep ocean.
- Open-ocean deoxygenation is resulting mainly from a warming ocean, increased stratification and changing circulation which interact with eutrophication-induced hypoxia (oxygen concentration below ~60 to 120  $\mu\text{mol O}_2 \text{ kg}^{-1}$ ) and biological activity in shelf regions.
- Climate change related longer-term oxygen trends are masked by oxygen variability on a range of different spatial and temporal scales.
- The decline in the oceanic oxygen content can affect ocean nutrient cycles and the marine habitat, with potentially detrimental consequences for fisheries, ecosystems and coastal economies.
- Oxygen loss is closely related to ocean warming and acidification caused by  $\text{CO}_2$  increase driven by  $\text{CO}_2$  emissions as well as biogeochemical consequences related to anthropogenic fertilization of the ocean; hence a combined effort investigating the different stressors will be most beneficial to understand future ocean changes.

Global deoxygenation effects	Consequences
Continuous loss of global oceanic dissolved oxygen content.	<ul style="list-style-type: none"> <li>• First ecosystem changes can be observed, habitat compression occurs and large-scale changes in ecosystems are expected for the future.</li> </ul>
Increasing deoxygenation and anoxia occurs in large areas of the continental shelves.	<ul style="list-style-type: none"> <li>• Reduced fish catches with decreasing economic profit in coastal states are expected.</li> </ul>
Deoxygenation is overlain by climate variability, interactions and understanding of impacts is limited.	<ul style="list-style-type: none"> <li>• Periods with enhanced, reduced or even increased deoxygenation might appear, and predictions of the future deoxygenation are insecure.</li> </ul>
Open ocean oxygen loss is closely connected to ocean warming and acidification caused by CO <sub>2</sub> increase driven by CO <sub>2</sub> emissions.	<ul style="list-style-type: none"> <li>• Continuous CO<sub>2</sub> emissions will increase deoxygenation, warming, and acidification of the ocean.</li> </ul>
Specific ecosystem services can be negatively affected by combined deoxygenation, pollution and ocean acidification.	<ul style="list-style-type: none"> <li>• Negative impacts are expected on biological regulation, nutrient cycling and fertility, food, ornamental resources (like corals, pearls, shell material), tourism and recreation.</li> </ul>

### 2.1.1 Introduction

Despite its far-reaching consequences for humanity, the focus on climate change impacts on the ocean lags behind the concern for impacts on the atmosphere and land (Allison & Bassett, 2015). In the ocean a reduction in ventilation and decrease of oceanic dissolved oxygen are two of the less obvious but important expected indirect consequences of climate change on the ocean (Shepherd et al., 2017). Deoxygenation of the open ocean is one of the major manifestations of global change. It accompanies ocean warming and ocean acidification as one of the three major oceanic consequences of rising atmospheric CO<sub>2</sub> levels (Levin & Breitburg, 2015). In addition, agricultural runoff and human waste cause decreasing oxygen concentrations in estuaries and coastal areas as well as the adjacent open ocean.

Oxygen is distributed in the ocean through an intimate interplay of physics and biology. Oxygen is transported by ocean circulation and mixing processes from the well-ventilated surface into the depths of the ocean interior. The ocean surface layers are saturated with oxygen due to atmospheric exchange and photosynthesis, primarily by phytoplankton. Oxygen is consumed by all animals throughout the ocean, but the majority of consumption occurs by bacterial respiration of organic matter. Both the supply and consumption of oxygen are severely modulated by climate change in ways that are not yet fully understood (Levin, 2018). On the continental shelf, nutrient input and eutrophication-induced hypoxia influences the oxygen concentration and hence the coastal ecosystem. Variations in marine oxygen concentrations can induce major changes to remineralization processes and associated sources and

sinks of important nutrient elements such as nitrogen, phosphorus and iron in the water column and underlying sediments (Oschlies et al., 2018). For instance, radically different microbial and chemical processes operate at low oxygen levels, such as anaerobic remineralization of organic matter by denitrification, leading to a net loss of fixed nitrogen from the ocean. Therefore, oxygen levels can act as a 'switch' for nutrient cycling and availability.

For the past several hundred million years, oxygen concentrations in Earth's atmosphere have been comparatively high (e.g. Mills et al., 2016). However, parts of the ocean seem never to have been far from anoxia (oxygen depleted) and have occasionally suffered major oceanic anoxic events (OAEs). In the past OAEs seem to have been promoted by warm climates, and some have been associated with major environmental crises and global-scale disturbances in the carbon cycle (Watson, 2016). Palaeo-records from the late Permian and Cretaceous provide evidence for past periods of dramatically reduced oceanic oxygen levels, and these had major consequences for marine ecosystems leading to mass extinctions. Major low-oxygen events in the past were primarily associated with warmer climates and higher atmospheric CO<sub>2</sub> levels. Hence deoxygenation, accompanying the present changes in climate, has to be understood in detail to be able to predict future ocean changes and propose societal actions that can help to avoid or minimize future anthropogenic inputs to oceanic deoxygenation events.

### 2.1.2 Geographic definition

Until recently knowledge of ocean deoxygenation was based on local time series of individual measurements

or from geographically restricted areas of oxygen data collections. In recent years the various data for the entire world ocean from the surface to the ocean bottom were used to investigate the trends in oxygen from both measurements as well as numerical model runs. The computed oxygen inventory for the whole world ocean from the sea surface to the ocean bottom varies for the different ocean basins and reflects oceanic volume, bathymetry and deep-water ventilation rates (Schmidtko et al., 2017). The largest oxygen inventories exist in the well-ventilated deep Atlantic Ocean (Figure 2.1.1A).

In the eastern Pacific and Atlantic Oceans as well as in the northern Indian Ocean the oxygen concentration is lower at mid ocean depths due to weak water renewal by sluggish ocean circulation. These regions are often called oxygen minimum zones (OMZs), e.g. marked in Figure 2.1.1 by the  $80 \mu\text{mol O}_2 \text{ kg}^{-1}$  contour. In the eastern portions of ocean basins, the sluggish circulation is caused by the asymmetric shape of the subtropical gyres, which lead to weak ventilated regions at the eastern sides equatorward of the subtropical gyres, sometimes referred to as shadow zones (see Box 2.1.1). Formation of well oxygenated deep water can be

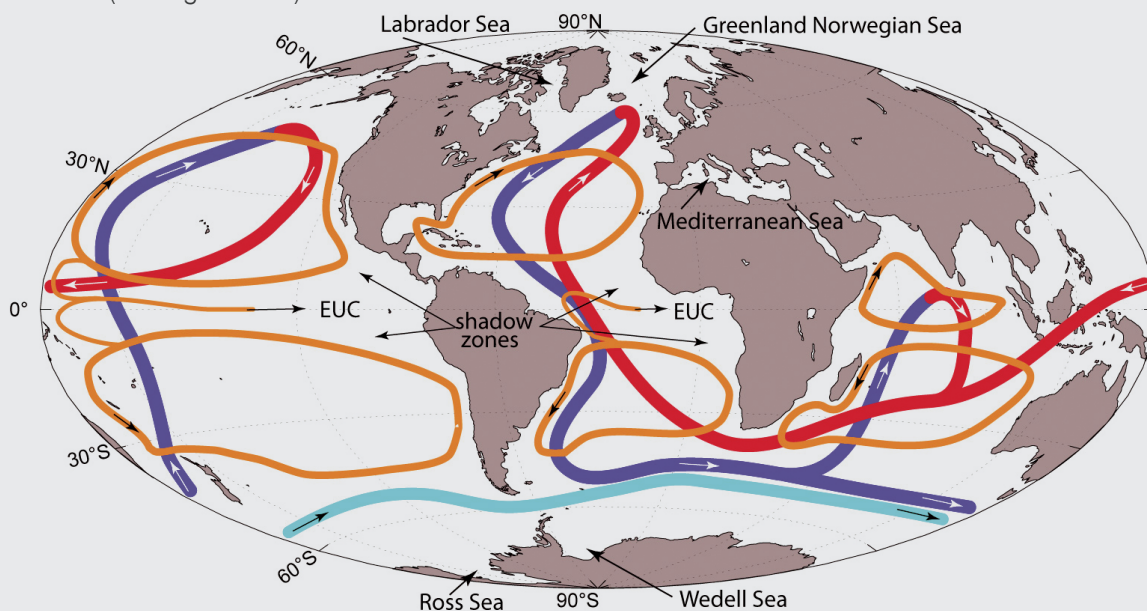
reduced by ocean surface warming in two ways: by a reduction in ventilation volume, and by a temperature increase of the water formed. Formation of this deep water that drives thermohaline ocean circulation on a global scale takes place in a few areas: the Greenland-Norwegian Sea, the Labrador Sea, the Mediterranean Sea, the Weddell Sea and the Ross Sea (see Box 2.1.1). These are the source regions ventilating the deep ocean basins via deep western boundary currents. These deep currents later upwell and form near-surface currents that complete the thermohaline circulation pattern.

### 2.1.3 Trends and impacts

Ocean models predict a decline of total dissolved oxygen in the global ocean in the order of 1 to 7% by the year 2100, caused by a combination of warming-induced decrease in oxygen solubility and reduced ventilation of the deep ocean (e.g. Keeling et al., 2010; Oschlies et al., 2018; Schmidtko et al., 2017). Oxygen trends, variability and impacts based on measurements are often analysed locally or across small geographical regions and should be used to verify the model results.

#### Box 2.1.1 Major ocean circulation features

The distribution and changes in oxygen are related to the large-scale wind driven and thermohaline ocean circulation (Box Figure 2.1.1)



Box Figure 2.1.1 Schematic distribution of the wind driven subtropical gyres (orange) and their contribution to the Equatorial Undercurrent (EUC) and the warm (red) and cold (blue) components of the thermohaline overturning circulation. The cool Antarctic Circumpolar Current (light blue). The shadow zones of the Pacific and Atlantic oceans are marked by arrows as well as the Labrador Sea, the Greenland-Norwegian Sea, the Mediterranean Sea, the Ross Sea, and the Weddell Sea.

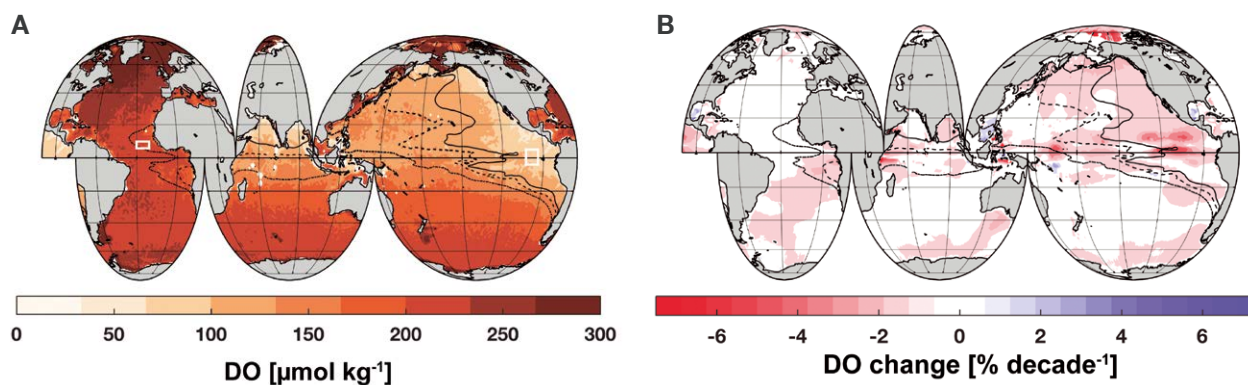


Figure 2.1.1 A): Global ocean top to bottom oxygen inventory (colour coded) with lines indicating boundaries of oxygen-minimum zones (OMZs), dash-dotted regions with less than  $80 \mu\text{mol kg}^{-1}$  oxygen anywhere within the water column; dashed lines and solid lines similarly represent regions with less than  $40 \mu\text{mol kg}^{-1}$  oxygen and  $20 \mu\text{mol kg}^{-1}$  oxygen respectively. B): Change in dissolved oxygen per decade in % for the time period 1960 to 2010, based on Schmidtko et al. (2017). Areas  $5^{\circ}\text{N}$ - $5^{\circ}\text{S}$ ,  $105^{\circ}$ - $115^{\circ}\text{W}$  and  $10^{\circ}$ - $14^{\circ}\text{N}$ ,  $20^{\circ}$ - $30^{\circ}\text{W}$  used here are marked in white in A).

Despite data limitations, it is feasible to detect significant oxygen changes over time in some regions. These studies cover different time frames and regions (Keeling et al., 2010). Regional observational data (Stramma et al., 2008) indicate a continuous decrease in oceanic dissolved oxygen content in most regions of the global ocean. Despite the already low oxygen content in the tropical ocean, a decrease of oxygen was observed for selected regions in all three tropical ocean basins for the period 1960 to 2008 (Stramma et al., 2008). These time series reveal vertical and horizontal expansion of the intermediate-depth low-oxygen zones and reduced oxygen content in the OMZ cores, where the oxygen concentration is lowest.

Despite limited data some large-scale analyses have been undertaken based on measurements in the literature. A recent and comprehensive analysis of observed oxygen changes indicates a 2% (i.e.  $96 \text{ Tmol yr}^{-1}$  ( $\text{Tmol} = 10^{12} \text{ mol}$ )) decline of total global ocean oxygen content over the 50 year period 1960 to 2010 (Schmidtko et al., 2017). Five distinct regions with significant oxygen loss stand out that cannot be directly attributed to solubility changes alone. These are the tropical regions of all basins, which contain most of the upper-ocean OMZ, the North Pacific, the South Atlantic, the Southern Ocean and the Arctic Ocean (see Figure 2.1.1B). Different drivers for the oxygen decrease in these regions have been proposed (Schmidtko et al., 2017). For the upper (0 - 1200 m depth) North Pacific a decline in North Pacific Intermediate Water formation is thought to be the main cause of decreasing oxygen content, whereas for the deep (1200 m to ocean bottom) South Atlantic a reduced or changing ventilation process as well as multi-decadal variability seem to be responsible

(Schmidtko et al., 2017). The amount of oxygen lost globally in the upper ocean has been estimated to be about  $26 \text{ Tmol O}_2 \text{ yr}^{-1}$  (0 to 1200 m layer for 1960 to 2010; Schmidtko et al., 2017) similar to the  $24 \text{ Tmol O}_2 \text{ yr}^{-1}$  (0 to 1000 m for 1958 to 2015) estimated by Ito et al. (2017). A larger decrease reported for the 100 – 1000 m layer of  $55 \text{ Tmol O}_2 \text{ yr}^{-1}$  between the 1970s and 1990s (Helm et al., 2011) is probably related to the strength of the Pacific Decadal Oscillation during this time period.

Circulation in the thermocline (~ upper few hundred metres of the ocean) is predominantly wind-driven and therefore particularly sensitive to changes in wind patterns, which in turn are highly associated with major climate modes such as Pacific Decadal Oscillation (PDO) and El Niño-Southern Oscillation (ENSO) in the Pacific Ocean and North Atlantic Oscillation (NAO) and Atlantic Meridional Mode (AMM) in the Atlantic. Hence these climate modes have a strong potential influence on the changes in the upper ocean oxygen content. For example for the overall time period of positive and negative PDO (1960 to 2008) in the tropical eastern Pacific ( $5^{\circ}\text{S}$  to  $5^{\circ}\text{N}$ ,  $105^{\circ}\text{W}$  to  $115^{\circ}\text{W}$ ; area D in Stramma et al. (2008); see Figure 2.1.1) the oxygen decrease is  $0.23 \mu\text{mol O}_2 \text{ kg}^{-1} \text{ yr}^{-1}$  in the 200 to 600 m depth layer but for the time period of mainly positive PDO (1979 to 2008) a much stronger oxygen decrease occurs ( $0.64 \mu\text{mol O}_2 \text{ kg}^{-1} \text{ yr}^{-1}$ , Figure 2.1.2).

Global warming leads to upper ocean warming, and less dissolved oxygen can therefore be stored in the surface layer due to a reduced oxygen solubility in warmer water. A tight relationship between oxygen inventories and ocean heat content was observed from measurements for the upper 1000 m (Ito et al., 2017). Temperature

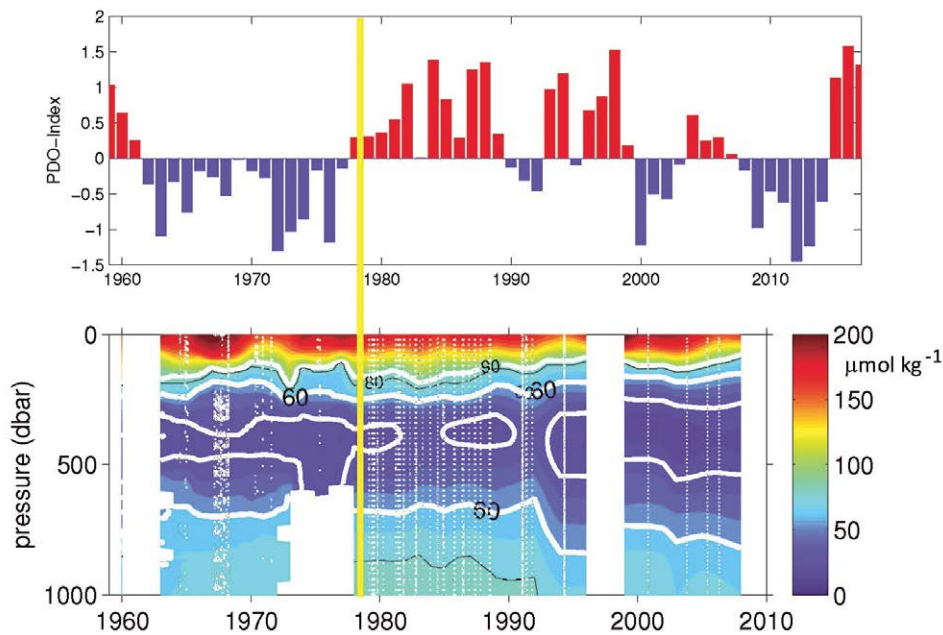


Figure 2.1.2 Annual mean Pacific Decadal Oscillation Index (<http://ds.data.jma.go.jp/tcc/tcc/products/elnino/decadal/annpdo.txt>) and dissolved oxygen concentration ( $\mu\text{mol kg}^{-1}$  versus time in the eastern equatorial Pacific Ocean ( $5^{\circ}\text{S}$  to  $5^{\circ}\text{N}$ ,  $105^{\circ}$  to  $115^{\circ}\text{W}$  (area D in Stramma et al., 2008)). The yellow line marks the shift from the negative to the positive PDO phase.

profiles in the Japan Sea show a warming by as much as  $0.1 - 0.5^{\circ}\text{C}$  in the upper 1000 m and by  $0.01^{\circ}\text{C}$  below 2000 m for a 30 year period and an associated oxygen decrease in the deep layer (Watanabe et al., 2003). Furthermore, warmer waters are less dense and enhance ocean stratification and consequently less oxygen-rich waters reach the deep ocean. A reduced oxygen supply in midwater depths, in the absence of any major change in the midwater microbial community, leads to a decreased oxygen concentration at a given depth, and ultimately to vertical expansion of the OMZs and the low-oxygen zone above them. In effect, this vertical expansion reduces the volume of the well-oxygenated epipelagic surface zone, thus reducing habitat extent of many fishes with high oxygen demand (Figure 2.1.3).

Another important driver of decreased oxygen content in coastal oceans is eutrophication-induced hypoxia that results from addition of excess nutrients into the ocean. The number of such coastal zone sites has increased by an order of magnitude since the 1960s (Diaz & Rosenberg, 2008) and oxygen decline rates are more severe in a 30 km band near the coast than in the open ocean (Gilbert et al., 2010). Near eastern shelf-breaks, poleward undercurrents such as the California and Peru-Chile Undercurrent in the Pacific Ocean and the Mauritania and Angola Currents in the Atlantic carry oxygen-poor water from expanding tropical OMZs

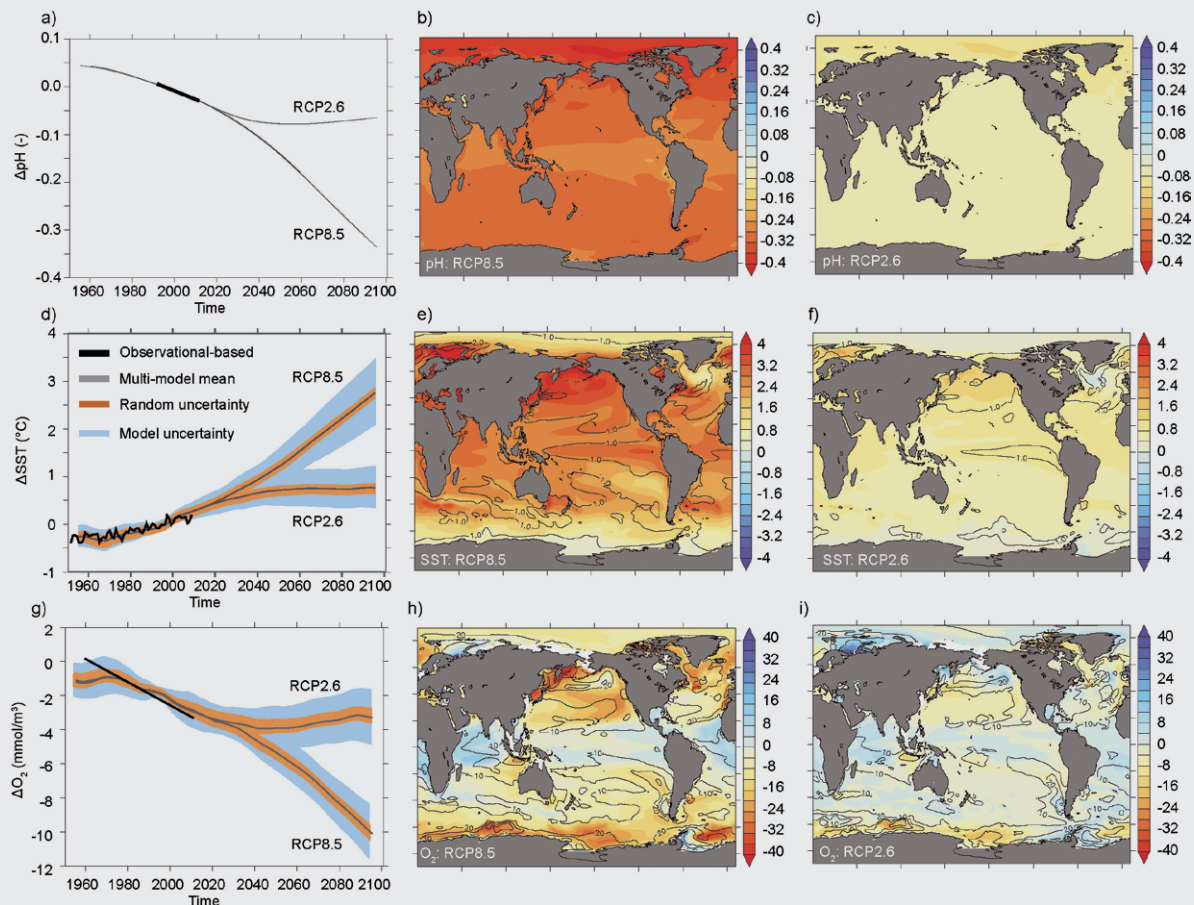
toward the subtropics, where it can interact, often through wind-driven upwelling processes, with both natural and eutrophication-induced hypoxic zones on the inner shelves. This occurs for example off Peru, Chile, Namibia and the western Indian margin (Figure 2.1.3; Stramma et al., 2010). In the eastern Pacific these oxygen-poor upwelled waters are corrosive (low pH) and undersaturated with respect to aragonite, exacerbating the stress imposed on the exposed ecosystem (Feely et al., 2008). It is important to point out that the unusually high productivity of such shelf regions depends on wind-driven upwelling. Upwelling of deoxygenated and corrosive water will impact the largest fisheries on the planet, e.g. that for Peruvian anchoveta.

Results of numerical biogeochemical Earth system models reveal that the magnitude of the observed change is consistent with  $\text{CO}_2$ -induced climate change. However, the correlation between pattern of simulated and observed oxygen change is negative for the subtropical and tropical upper ocean, indicating that the models do not correctly reproduce the current processes responsible for observed regional oxygen changes (Stramma et al., 2012a). Particularly in the tropical thermocline, a region which features open-ocean oxygen minimum zones, observations indicate a general oxygen decline, whereas most of the state-of-the-art models simulate increasing oxygen levels in these OMZs. Wind-driven changes in the low-latitude

### Box 2.1.2 Future projections based on projected greenhouse gas emissions

Model projections into the future depend on the projected greenhouse gas emissions. Representative Concentration Pathways (RCPs) are four greenhouse gas concentration trajectories which describe possible climate futures that are considered possible depending on the scale of greenhouse gas emissions in the years to come. They are RCP2.6, RCP4.5, RCP6 and RCP8.5 reflecting a range of possible greenhouse gas-caused radiative forcing values in the year 2100 relative to pre-industrial values of +2.6, +4.5, +6.0 and +8.5  $\text{W m}^{-2}$  (difference between insolation absorbed by the Earth and energy radiated back to space). RCP2.6 is compatible with the Paris Agreement and assumes that global annual greenhouse gas emissions peak between 2010-2020, with emissions declining substantially thereafter. In RCP8.5 emissions continue to rise throughout the 21st century, often called 'business as usual'. The key greenhouse gases emitted by human activities are 76% carbon dioxide, 16% methane, 6% nitrous oxide and 2% fluorinated gases (IPCC, 2014).

To assess internal variability and model scenario uncertainties, a combination of several Earth System Models (Frölicher et al., 2016) was used, and the results project a decrease of surface pH, increase of sea surface temperature (SST), and oxygen loss in most ocean basins in the 100 to 600 m layer by 2100 for the RCP2.6 and the RCP8.5 scenarios (Box Figure 2.1.2). Hence even the most optimistic scenario (RCP2.6) would lead to lower ocean surface pH, as well as ocean warming and deoxygenation in the future (Box Figure 2.1.2 c, f, i). By the end of the 21st century pH is projected to decrease (in pH units) by  $0.29 \pm 0.005$  (RCP8.5) and  $0.07 \pm 0.005$  (RCP2.6), SST is projected to increase by  $2.39 \pm 0.50^\circ\text{C}$  (RCP8.5) and  $0.75 \pm 0.30^\circ\text{C}$  (RCP2.6), and oxygen concentrations averaged over 100-600 m depth are projected to decrease by  $6.98 \pm 1.02 \text{ mmol m}^{-3}$  (RCP8.5) and  $1.29 \pm 1.06 \text{ mmol m}^{-3}$  (RCP2.6). (Those oxygen units are equivalent to  $\mu\text{mol L}^{-1}$  and  $\sim\mu\text{mol kg}^{-1}$ ).



Box Figure 2.1.2 Time series and spatial patterns of simulated change in surface pH, sea surface temperature ( $^\circ\text{C}$ ), and oxygen concentration ( $\text{mmol m}^{-3}$ ) averaged over 100 to 600 m depth. Simulated spatial patterns of multi-model mean changes in year 2095 for the (middle column) RCP8.5 scenario and right column RCP2.6 scenario (Reprinted with permission from Frölicher et al., 2016 (© Wiley and Sons)).



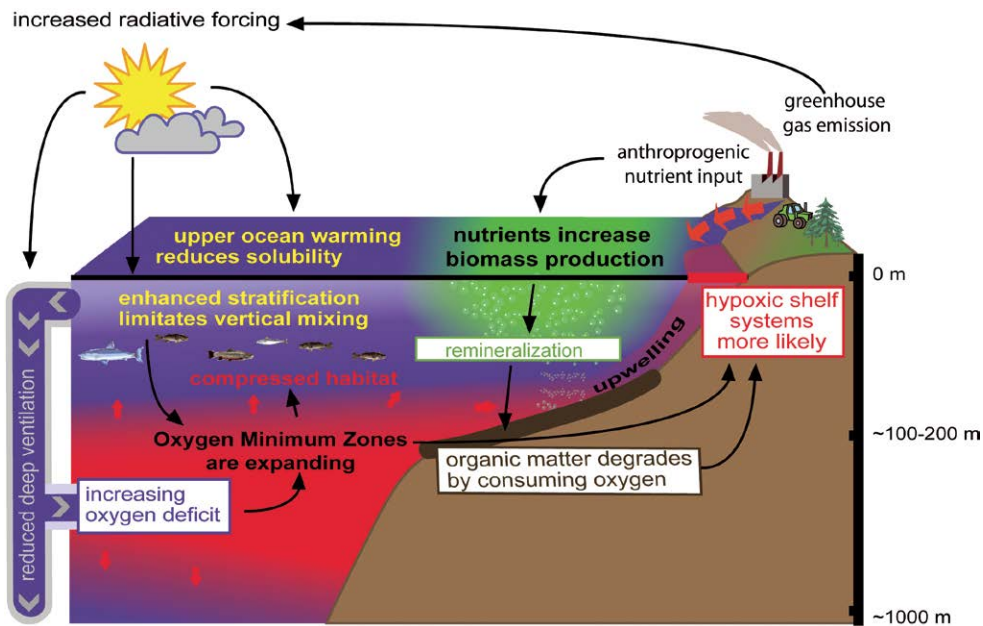


Figure 2.1.3 Schematic of interactions of open ocean oxygen minimum zone (OMZ, red) with hypoxic shelf system and dead zones on continental shelves of eastern ocean boundaries (modified after Stramma et al., 2010).

oceanic ventilation are identified as a possible factor contributing to the apparent discrepancies during the past decades. Other factors relevant to model-data discrepancies are the roles of equatorial jets, lateral and diapycnal mixing processes, changes in the wind-driven circulation, atmospheric nutrient supply and some poorly constrained biogeochemical processes (Oschlies et al., 2017). The process associated with the largest changes in simulated oxygen trend was variability in the wind forcing, particularly during the period prior to the oxygen content evaluation. Thus, a better observational constraint on the wind fields before the 1960s will help to yield more robust simulations of oxygen changes during the 50 years between 1960 and 2010 (Oschlies et al., 2017). Continued development and refinement of this more nuanced modelling approach thus holds great promise to more accurately predict future changes



Figure 2.1.4 Humboldt (jumbo) squid (*Dosidiscus gigas*) at night - Gulf of California' © robertharding/Alamy stock photo.

in ocean oxygen on a regional scale and resolve some observation-model discrepancies.

#### 2.1.4 Ecosystem consequences

Oxygen influences biological processes at the most fundamental level. Changes in temperature, oxygen content, and to a lesser degree some other ocean biogeochemical properties directly affect the ecophysiology of marine water-breathing organisms with regard to distribution, phenology and productivity (Cheung et al., 2012). Expanding OMZs benefit microbes and life forms adapted to hypoxia while restricting the ranges of most other species. Eutrophication as a result of coastal pollution exacerbates this problem, through organic matter which increases metabolic consumption in deeper coastal areas (Gattuso et al., 2015). Moreover, higher temperatures lead to community-level shifts to smaller Eukarya, Bacteria and Archaea under conditions of diminished oxygen (see e.g. Gattuso et al., 2015) and also increase sensitivity of various fish and invertebrates to hypoxia, thereby limiting the depth distribution of those species not adapted to hypoxic conditions. In the tropical Atlantic, for example, the expansion of the OMZ reduced the available habitat for tropical pelagic fishes (Stramma et al., 2012b). In addition, the maximum size of exploited fishes decreases with rising temperature and decreasing oxygen level, ultimately reducing potential fish yields (Baudron et al., 2014) which is in agreement with current model predictions (e.g. Cheung et al., 2012).



Figure 2.1.5 Owlfish (*Bathylagus miller*) Monterey Bay © 2001 MBARI.

Conversely, hypoxia-adapted species may benefit, as illustrated by the range-expansion of Humboldt squid (Figure 2.1.4), a species which is well adapted to hypoxia (Stewart et al., 2014). Increases in jellyfish blooms have been linked to eutrophication and hypoxia and lowered pH; such blooms may indirectly promote hypoxia by preying on zooplankton, leaving more unconsumed phytoplankton to sink and degrade via microbial consumption (Purcell et al., 2001). Some species will potentially benefit from a habitat expansion, these include, for example, larger animals that permanently inhabit the OMZ and include species with extremely low metabolic rates, e.g. the vampire squid and the mysid shrimp which have enhanced oxygen extraction and transport abilities and the bathylagid owlfishes (Figure 2.1.5), the pelagic worm *Poebobius meseres* (Christiansen et al., 2018) and filter-feeding tunicates (e.g. Pacific transparent sea squirt, or common sea grape) (Gilly et al., 2013). The impact of habitat compression was also quantified by the change in proportion of zooplankton biomass in the upper water column (Wishner et al., 2013). The vertical repositioning of biomass layers and the increased depth range of low oxygen water that diel vertical migrators and sinking particles must transit in an expanded OMZ, could have widespread effects on species distribution, the biological pump, and benthic-pelagic coupling (Wishner et al., 2013).

Future projections of potential stressors of ocean ecosystems, such as deoxygenation, warming, acidification, and changes in ocean productivity, remain vague due to incomplete understanding of fundamental mechanisms behind these processes, internal climate variability, and divergent carbon-emission scenarios (Frölicher et al., 2016). Nevertheless, it is clear that future oxygen changes will be strongly dependent on and

mostly negatively impacted by future global greenhouse gas emissions (see Box 2.1.2).

### 2.1.5 Societal consequences

Deoxygenation especially in shelf and coastal waters has and will lead to more societal consequences. Productivity will be decreased both for capture fisheries and aquaculture. Fish and shellfish species that support regional human needs for protein are scarce and vulnerable to hypoxic waters, a problem particularly severe for artisanal fisheries without the capability of relocating in response to hypoxic or anoxic events that cause mortality and changes in distributions of fisheries species (Levin & Breitburg, 2015). Human communities, especially in developing nations, that depend heavily on coastal fishery resources for food, economic security, and traditional culture are at particular risk from shifts in ocean primary production and species habitat ranges (Gattuso et al., 2015).

Ecosystem services comprise the entire range of benefits that nature provides to human communities and can be classified in four major categories: supporting services, provisioning services, regulating services and cultural services (see Cooley, 2012). Specific ecosystem services could be negatively affected by pollution, warming, ocean acidification, and deoxygenation. Services that are very likely to be negatively impacted by all stressors include: the supporting services of biodiversity, biological regulation, nutrient cycling, and fertility of marine organisms; the provisioning services of food, building materials, ornamental resources (e.g. coral, pearls, shell material, aquarium fish); and all the cultural ecosystem services of spiritual benefits, tourism and recreation, aesthetic benefits, and opportunities for education and research (Cooley, 2012).

Established, emerging, and new ocean industries are directly influenced by oceanic deoxygenation. Negative influences are listed with regard to fisheries, aquaculture, pharmaceuticals and chemical industries, tourism, and assimilation of nutrients; whilst positive influences are seen only with regard to technology R & D (Allison & Bassett, 2015).

Societal responses to climate change range from individual adaptive decisions, to attempts by representatives of sovereign states to reach a global consensus on reducing greenhouse gas emissions (Allison & Bassett, 2015). O'Neill and Nicholson-Cole



Figure 2.1.6 A menhaden (*Brevoortia tyrannus*) fish kill due to severe hypoxia – near anoxia in Greenwich Bay (Narragansett Bay, Rhode island) © Chris Deacutis.

(2009) have shown that non-threatening images that engage everyday emotions, such as a picture of a community mobilizing to protect their local environment, are more likely to inspire wider citizen action to address climate change impacts than are fear-inducing representations of a climate catastrophe. For example, pollutants mostly from wastewater treatment facilities resulted in poor water quality and deoxygenation in Narragansett Bay over the past decades. The pollution led to algal blooms and a lack of dissolved oxygen in the bay. Following a large fish kill in 2003 (Figure 2.1.6) wastewater treatment plants reduced the amount of nitrogen discharge into the water and in some parts of the Bay the oxygen returned to normal levels by 2014 leading to better water quality and a healthier marine habitat that resulted in increased home prices in this region (Jepsen, 2017).

### 2.1.6 Implication of continuing ocean deoxygenation

As deoxygenation is closely connected to global warming and CO<sub>2</sub> increase, a continuation and possibly acceleration of the current rate of ocean deoxygenation is expected. An area in the eastern tropical Atlantic,

where repeated measurements have been made in recent years indicates that the oxygen concentration continuously decreased in the open ocean in the 50 to 300 m layer by 0.49  $\mu\text{mol kg}^{-1}\text{yr}^{-1}$  for the period 1960 to 2016 (Figure 2.1.7). Excess nutrients from land can stimulate further biogeochemical activity and push open-ocean ecosystems into more severe hypoxic conditions. At the same time, future warmer estuarine and ocean waters resulted in an increasing number of eutrophic coastal sites approaching hypoxic tipping points, thus worsening the severity and spatial extent of oxygen depletion in systems with historically occurring hypoxia (Levin & Breitburg, 2015). Harmful algal blooms might be augmented from nutrients released in bottom water due to hypoxia (Davidson et al., 2014) with severe consequences for pelagic fishes and invertebrates that provide important fisheries for humans and prey for marine mammals.

Shoaling of the upper boundaries of the OMZs accompanies horizontal OMZ expansion, and decreased oxygen at shallower depths can affect all marine organisms through multiple direct and indirect mechanisms. Effects include altered microbial processes that produce and consume key nutrients and gases, changes in predator-prey dynamics, and shifts in the abundance and accessibility of commercially fished species (Gilly et al., 2013). Microbial communities will be greatly affected by the expansion and shoaling of OMZs. Three key changes in microbial communities and microbe-mediated biogeochemical cycles are likely to be consequences of OMZ expansion, intensification, and shoaling: alteration of the N cycle, the emergence of sulphur (S) cycling, and the creation of new low-oxygen habitats (Gilly et al., 2013). The removal of nitrogen can be a limiting nutrient in upwelling areas, phytoplankton growth and surface production would be reduced. In OMZs sometimes sulphate seems to be reduced to sulphide, which is toxic to most organisms. Deeper low-oxygen habitats will expand as OMZ shoals with effects on carbon, nitrogen and sulphur cycling, but the ultimate biogeochemical outcome is difficult to predict (Gilly et al., 2013).

### 2.1.7 Conclusions / Recommendations

Deoxygenation is an ongoing process and is expected to continue as increasing global temperatures reduces the capacity of the ocean to hold oxygen, decreases the degree of mixing in the upper water column and reduces the ocean overturning circulation. The current

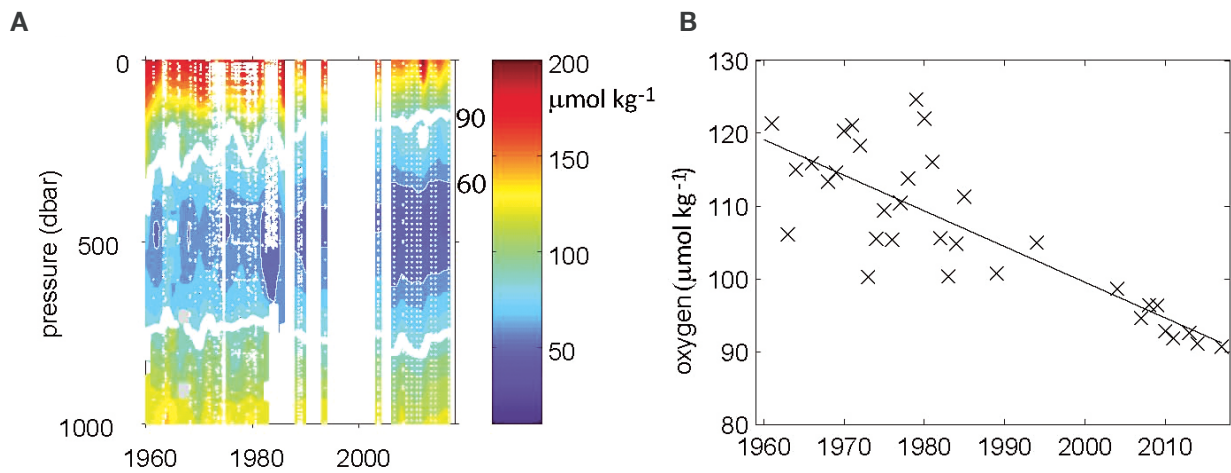


Figure 2.1.7 A) Dissolved oxygen concentration ( $\mu\text{mol kg}^{-1}$ ) versus time in the eastern tropical Atlantic Ocean ( $10^{\circ}$  to  $14^{\circ}\text{N}$ ,  $20^{\circ}$  to  $30^{\circ}\text{W}$  (area A in Stramma et al., 2008)) extended with cruises until September 2016; and B) mean oxygen content of the layer 50 to 300 m with a decreasing linear trend of  $0.49 \mu\text{mol kg}^{-1} \text{ yr}^{-1}$  for the period 1960 to 2016.

understanding of both the causes and consequences of reduced oxygen in the ocean and our ability to represent them in models especially in the low latitudes are inadequate. The reasons for the discrepancies between simulations and observations currently remain unclear (Shepherd et al., 2017).

Accelerated oxygen loss in both the open ocean and coastal waters is generating complex biological responses. The phenomena in the open ocean and in shelf regions are highly interconnected, and investigations in both areas need to be linked more closely in the future to address ocean deoxygenation.

Spatial patterns of oxygen change differ considerably between observed and modelled estimates. Within models the primary uncertainty of ocean deoxygenation in the subsurface tropical ocean is due to a compensation between decreasing oxygen saturation due to warming and decreasing apparent oxygen utilization (difference between oxygen saturation and measured oxygen concentration) due to increased ventilation of the corresponding water masses (Bopp et al., 2017). Model improvements are needed for better simulations to understand oxygen changes and consequently improve the quality of projections further into a changing future (Oschlies et al., 2017).

Detecting time-varying trends in ocean properties and processes requires consistent, high-quality measurements. Ocean time-series as well as spatial monitoring provide vital information needed for assessing oxygen-related ecosystem change. Some

of the most important contributions to emerge from the ocean time-series programmes are reconstructions of biogeochemical rate processes based on annual mass balance of properties such as dissolved oxygen, dissolved inorganic carbon, nitrate and nitrogen and carbon isotopes. Future studies that seek mechanistic understanding of ecological interactions underlying the biogeochemical dynamics should be encouraged (Church et al., 2013).

As  $\text{CO}_2$  increases in the atmosphere and the ocean, this is a driver for ocean warming and acidification, and closely related with deoxygenation. Investigations of the biological effects of a single, isolated driver are problematic, in that they can produce misleading inferences about different organism's responses in a multivariate natural environment. Essentially, interactive (additive, synergistic, or antagonistic) effects are generally not predictable from single-driver studies. In some cases, hypoxic conditions can mediate negative effects of ocean acidification, however, ocean acidification and hypoxia increase heat sensitivity, and oxygen loss combined with warming is projected to lead to the contraction of metabolically viable habitats of marine animals on a global scale (see Gattuso et al., 2015). Growing evidence suggests that interactions with other environmental factors, such as irradiance, nutrient availability, geographical location and species community composition, can strongly modulate the biological effects of warming, ocean acidification and hypoxia (see Gattuso et al., 2015).

Understanding the impact of climate change, as well as society's diverse perspectives and multiscale responses to the changing ocean, requires a correspondingly diverse body of scholarship in the physical, biological, and social sciences as well as humanities. This can ensure that the research that informs climate policy leads to a plurality of values and viewpoints and may enable the concerns of maritime societies and economic sectors to be heard in key adaptation and mitigation discussions (Allison & Bassett, 2015).

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## 2.1.8 References

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