



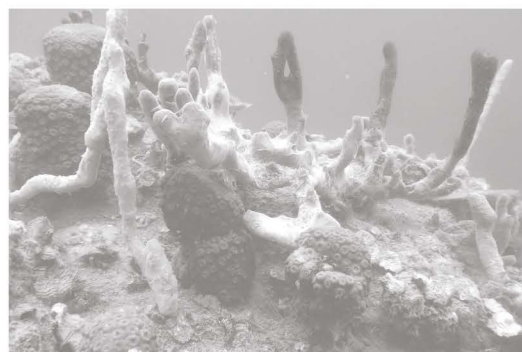
Ocean deoxygenation: Everyone's problem

Causes, impacts, consequences and solutions

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2.4 Evidence for ocean deoxygenation and its patterns: Eastern Boundary Upwelling Systems

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IUCN GLOBAL MARINE AND POLAR PROGRAMME



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Summary

- Eastern boundary upwelling systems (EBUS) are one of the ocean's most productive biomes, supporting one-fifth of the world's wild marine fish harvest. These ecosystems are defined by ocean currents that bring nutrient-rich but oxygen-poor water to coasts that line the eastern edges of the world's ocean basins. As naturally oxygen-poor systems, EBUS are especially vulnerable to global ocean deoxygenation.
- The dynamics of EBUS are intimately linked to global alterations in ocean chemistry and circulation from climate change. Upwelling currents connect the vast region of the subsurface open ocean that is experiencing declines in dissolved oxygen with the productive coastal waters of EBUS. The strength and location of upwelling currents depend on wind fields that are also affected by climate change. For some systems, this combination of changes will result in an intensification and expansion of coastal low oxygen zones.
- In comparison to the open ocean, long-term changes in dissolved oxygen availability in dynamic EBUS are much more challenging to resolve. Nonetheless, important trends have started to emerge. In a number of systems, dissolved oxygen (DO) has declined by approximately $10 \mu\text{mol kg}^{-1}$ per decade. This is of great concern because many EBUS locales already sit near if not pass the canonical threshold for hypoxia of $60 \mu\text{mol kg}^{-1}$. Observations of strengthening in winds that drive the upwelling delivery of low-oxygen and nutrient-rich waters in some systems portend heightened risks of ecosystem changes that outpace those expected from ocean deoxygenation alone.
- Because many EBUS are already exposed to low-oxygen conditions, the risk of crossing important biological thresholds that regulate the distribution and productivity of fishery-dependent stocks, and ecosystem functioning are heightened. Shallow water anoxia has already resulted in mass die-offs of fish and shellfish in some systems. Expansion of low oxygen zones have led to rapid, transient invasion of hypoxia-tolerant jumbo squid in others. Movement of fish away from low oxygen zones have also affected the accuracy of fishery-independent surveys even as the needs for tools for managing in the face of climate change grows.
- The intensification and expansion of low oxygen zones can have further ecosystem consequences as oxygen-dependent cycling of elements by microbes alter the supply of nutrients or in extreme cases, lead to increased production of toxic hydrogen sulphide gas (H_2S). Low oxygen EBUS are also regions of CO_2 enrichment as the loss of DO is coupled to the production of CO_2 . In combination with ocean uptake of human CO_2 emissions, CO_2 levels in some EBUS have already reached levels where the calcium carbonate shells of marine life are now being readily dissolved. Eastern boundary upwelling systems thus represent hotspots for both hypoxia and ocean acidification where development of mitigation and adaptation solutions are urgently warranted.

Ocean hypoxia effect	Potential consequences
Oxygen declines induces species range shifts, changes to vertical and across-shelf movement patterns, and losses in spawning habitats.	<ul style="list-style-type: none"> Altered ecological interaction rates among consumers and prey, and species that compete for resources. Altered ecological interactions as invasive hypoxia-tolerant species increase in abundance. Reduced fishery productivity as population replenishment declines for benthic spawning species and those that have strong habitat dependence for growth. Increased fishery conflicts as multiple targeted species are compressed into narrow oxygen refuges. Increased management uncertainty as fishery-independent surveys are compromised by reduced accessibility of fish to survey methodology.
Spatial and / or temporal expansion of areas currently affected by suboxia or anoxia as well as novel development of suboxic habitats in regions where they have previously been absent.	<ul style="list-style-type: none"> Increased loss of nitrogen nutrients as denitrification intensifies. Increased risk of water column hydrogen sulphide accumulation effects as sulphate reduction intensifies. Altered ratios of nutrient availability as the flux of iron and phosphorus from sediment increases.
Intensification of ocean acidification in conjunction with coastal hypoxia.	<ul style="list-style-type: none"> A wider array of taxa is affected and/or the effects of hypoxia are amplified if hypoxia and ocean acidification act as compounding or interactive stressors on organisms. More rapid shift to no-analogue state where multiple aspects of coastal ocean environment move away from natural ranges in exposure.

2.4.1 Introduction

Eastern boundary upwelling systems (EBUS) represent one of the ocean's most productive biomes. Even though EBUS comprise some 1% of the area of the ocean, the productivity of these coastal ecosystems supports one-fifth of the world's ocean wild fish harvests (Pauly & Christensen, 1995) and gives rise to important habitats for highly migratory seabirds, marine mammals and pelagic fishes (Block et al., 2011). Eastern boundary current upwelling systems can be readily identified from satellite maps of sea surface temperature (Figure 2.4.1A) as coastal regions that are colder than expected for their latitude. Ocean water gets colder with depth. The presence of cold water signals wind-driven ocean currents that transport or upwell cold, nutrient-rich waters from deeper layers of the ocean. Once upwelled nutrients reach the sunlight surface, they fuel dense blooms of phytoplankton (Figure 2.4.1B) that serve as the base of ocean food webs.

The dissolved oxygen (DO) content of ocean water also declines with depth. This reflects the loss of DO from

the respiration of microbes and other marine life as they consume the rain of organic matter that sinks down from the productive ocean surface. In fact, the high level of nutrients found in upwelled waters is a direct product of this process of organic matter remineralization. The surface ocean also acts as an insulating layer that keeps the waters below from replenishing lost DO with oxygen from the atmosphere (Figure 2.4.2). The atmosphere does have a central role to play, however, in structuring EBUS through the actions of coastal winds. Equatorward winds are common to coastlines that lie on the eastern edge of ocean basins. These winds serve as engines that move the surface ocean. These surface currents do not simply flow in the same direction as the equatorward winds, however. Known as the Coriolis effect, currents in motion on a rotating planet are deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The movement of surface waters away from the coastline lies at the heart of upwelling as seaward surface flows (known as the Ekman layer) must be counterbalanced by deep waters that rise and move toward the coast. As a consequence of coastal winds, eastern boundary current upwelling systems are one

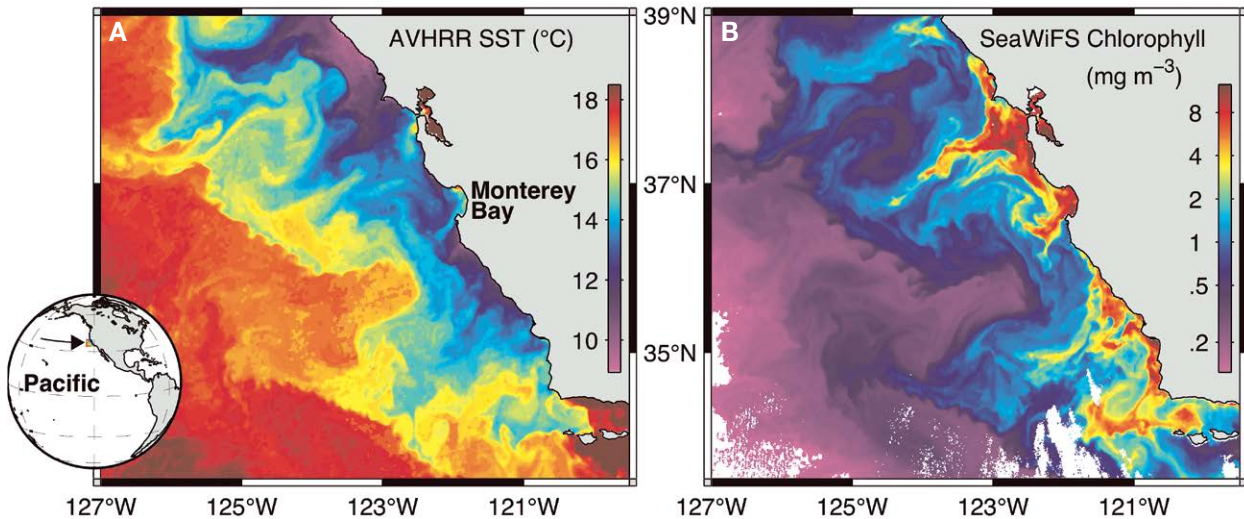


Figure 2.4.1 Satellite remote sensing imagery of Central California Current upwelling: A) Sea surface temperature (AVHRR), B) Surface chlorophyll (SeaWiFS) August 2000. © John P Ryan Monterey Bay Aquarium Research Institute.

of the few places in the ocean where nutrient-rich but oxygen-poor deep waters are transported to relatively shallow depths. Once upwelled, oxygen-poor waters can be further robbed of oxygen as the exceptional productivity of EBUS accelerates the rain of organic matter that fuels oxygen consumption above and in the sea floor (Adams et al., 2013).

As inherently oxygen-poor and in many instances, oxygen-deficient systems, it is not surprising that EBUS represent hot spots for ocean deoxygenation concern, particularly as relatively small DO declines can push systems past thresholds for hypoxia ($DO \leq 60 \mu\text{mol kg}^{-1}$) or suboxia ($DO \leq 5 \mu\text{mol kg}^{-1}$) (Deutsch et al., 2011). However, the sensitivity of EBUS to ocean deoxygenation is further compounded by strong climate-dependence in the very factors that give rise to low oxygen baseline conditions in EBUS. As systems that receive oxygen-poor oceanic waters, the global reductions in oxygen solubility from a warming ocean and slowing exchange of oxygen between the atmosphere and the deep ocean from a more strongly stratified sea can directly impact EBUS (Breitburg et al., 2018). In particular, because oceanic DO declines are not distributed evenly but can be most strongly focused at depth layers where upwelling currents are often drawn from (Ito et al., 2017), EBUS face disproportionate impacts from oceanic deoxygenation. The sensitivity of EBUS is also structured by the effects of climate change on coastal upwelling (Bakun et al., 2010). Winds are generated by spatial differences in atmospheric pressure, and coastal upwelling winds arise from pressure differences between a cool ocean and a warm continental land mass. As

land masses are projected to warm more quickly than the ocean, a long-standing hypothesis posits that atmospheric pressure differences are expected to intensify and/or shift spatially to alter upwelling winds (Bakun, 1990). For systems where upwelling winds intensify or lengthen seasonally, increased flux of low DO, high nutrient waters can directly and indirectly strengthen local deoxygenation (Bakun, 2017).

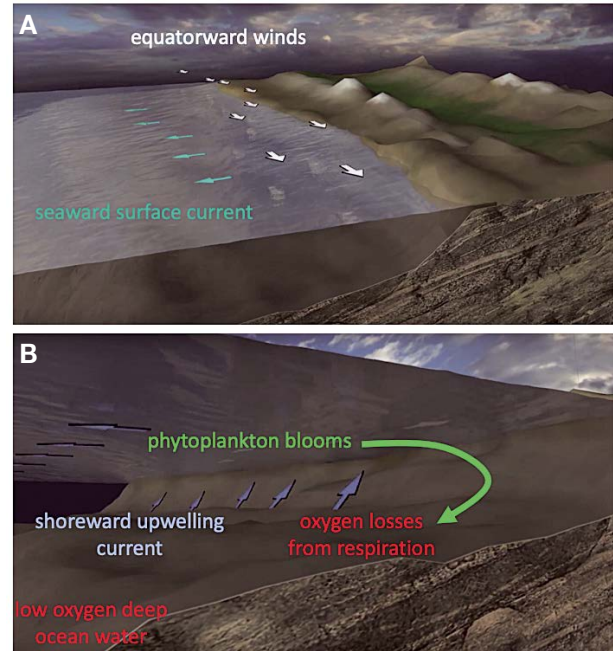


Figure 2.4.2 A) Equatorward coastal winds drives the flow of surface water away from the coast. B) Seaward surface flows are compensated by the upwelling of deep ocean water across the continental shelf. This deep water is deficient in dissolved oxygen and enriched in nutrients that promotes phytoplankton blooms. Low DO ocean water can experience further losses of DO as sinking phytoplankton cells are respired by microbes. © College of Earth, Ocean and Atmospheric Sciences - Oregon State University.

2.4.2 Geographic definition

As their name indicates, EBUS are located at the eastern margins of ocean basins where equatorward winds give rise to coastal upwelling currents (Figure 2.4.3). There are four major EBUS. In the Eastern Pacific Basin, the Humboldt or Peru-Chile Current System (HCS) lies off the coasts of Peru and Chile, while the California Current System (CCS) stretches from British Columbia, Canada to Baja, Mexico. In the Eastern Atlantic, the Benguela Current System (BCS) is situated along the coasts of Angola, Namibia and South Africa while the Canary Current System (CaCS) extends along the coasts of the Iberian Peninsula into North-west Africa.

Although the four major EBUS share broad commonalities in physical and ecological structures their locations give rise to notable differences in their present, as well as future, exposure to low-oxygen conditions (Chavez & Messie, 2009). One important factor is the location of EBUS relative to oceanic oxygen minimum zones (OMZ) (Figure 4.2.3) (Monteiro et al., 2011). In the eastern Pacific Ocean, broad expanses of the ocean interior are marked by vertical profiles where the DO-rich surface waters give way to a DO minimum layer that can reach anoxia ($\text{DO} = 0 \mu\text{mol kg}^{-1}$) (Thamdrup et al., 2012). Where the OMZ is particularly shallow such as along reaches of the HCS, coastward upwelling flows can directly draw suboxic

or even anoxic water to compensate for the seaward movement of the surface Ekman layer. In the Northern Pacific, OMZ is typically found too deep to serve as the source water for upwelled water (Connolly et al., 2010). Instead, upwelled water is drawn from the top of the OMZ where DO concentrations often range near hypoxic levels (Adams et al., 2013). In contrast, DO levels in the OMZ of the Atlantic Ocean do not reach the extremely low values found in Pacific Ocean OMZ. This does not mean that low DO and deoxygenation are less of a concern in the Atlantic systems. The development of anoxia in portions of the Benguela Current System highlight the importance of coastal biogeochemical processes in driving DO loss (Pitcher et al., 2014).

2.4.3 Trends and impacts

Future changes in oxygen dynamics in EBUS can be thought of as the product of global-scale decline in oceanic oxygen content, and system-scale alterations in upwelling-favourable winds that affect the delivery of oxygen-poor water to the coast and the supply of nutrients that fuel further local drawdown of DO. To understand the likely scope for future changes, we can consider the observed changes in oceanic and atmospheric forcings. The ability to resolve trends in DO is strongly dependent on the background variability of a given system (Long et al., 2016). Thus, it is not surprising that considerable attention has been paid

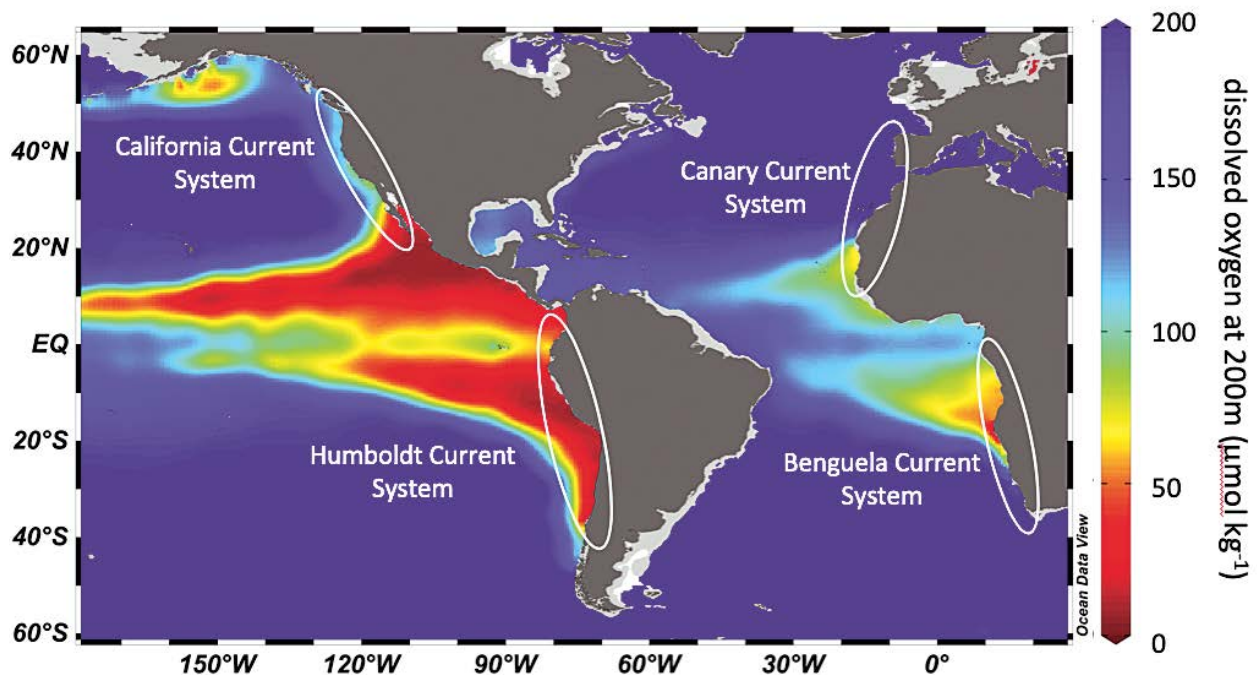


Figure 2.4.3 Location of the four major eastern boundary upwelling systems and dissolved oxygen values at 200m depth illustrating the location of each system relative to oceanic oxygen minimum zones. Data from World Ocean Atlas 2013.

to time trends of DO change in the global open ocean (Stramma et al., 2008). Analyses of the past five decades of available data point to a globally averaged DO loss rate of 0.4% per decade (Schmidtko et al., 2017). At first glance, this appears to be a minor rate of loss but because DO declines are not distributed evenly across the ocean, certain regions experience rates of loss that are considerably higher. In particular, areas near OMZ exhibit DO loss rates in excess of 4% per decade (Schmidtko et al., 2017). The volume of suboxic waters in OMZ increases non-linearly with DO decline, doubling in size with a 1% drop in mean DO (Ito & Deutsch, 2013). This expansion of oceanic suboxia has immediate consequences for the Humboldt Current System as it draws upwelled water directly from the Eastern Tropical Pacific OMZ. In absolute terms, DO losses in the upper 300m of the ocean (the range of depth from where upwelling water is typically drawn from) range upwards of $5 \mu\text{mol kg}^{-1}$ per decade (Stramma et al., 2012). This is notable as oceanic source waters that arrive in the CCS may hold only $90 \mu\text{mol kg}^{-1}$ of DO, already close to the canonical hypoxia threshold of $60 \mu\text{mol kg}^{-1}$ (Adams et al., 2013).

Model simulations of the impacts of climate change on ocean oxygen inventory point to the need for long, five decades plus records for deoxygenation trends to emerge from background natural variability (Long et al., 2016). This imposes important constraints on our ability to resolve deoxygenation trends in temporally-dynamic coastal EBUS where requirements for time series lengths can be even greater. Nevertheless, for the CCS, where long-term observations have been the most extensive, multi-decadal declines in DO (Pierce et al., 2012) and shoaling of hypoxia horizons (Bograd et al., 2008) have been reported. Declines in DO in recent decades (past 20-30 years) have been the strongest on record, with rates in the order of $10 \mu\text{mol kg}^{-1}$ per decade (Crawford & Pena, 2013). It is noteworthy that long-term measurements from multiple independent programmes have reported rates of DO decline that have been quite uniform across the CCS. Oxygen loss trends from Vancouver Island to Southern California range narrowly between 8 to $13 \mu\text{mol kg}^{-1}$ per decade (Crawford & Pena, 2013). We have fewer long-term records of DO from other EBUS. In the Benguela Current System, Moloney et al. (2013) report a decline of $9 \mu\text{mol kg}^{-1}$ per decade between 1957 and 2007 for St. Helena Bay, - a large productive bay that has been subject to episodic anoxia events (Pitcher & Probyn, 2017). For EBUS, changes can also manifest as increases in the

frequency or severity of oxygen-deficiency events. In the CCS, nearshore suboxia and anoxia events have occurred in the past decade that have no precedence in the observational record that extends back to the 1960s (Chan et al., 2008).

Insights from time-series analyses also highlight the climate sensitivity of oxygen dynamics in EBUS across inter-annual to inter-decadal time scales (Buil & Di Lorenzo, 2017). The impacts of ENSO variability on ecosystem dynamics in the HCS has been particularly well characterized where strong El Niño events results in a marked rise in oxygen levels that reorganizes the structures of benthic and pelagic communities (Bertrand et al., 2011; Escribano et al., 2004; Gutiérrez et al., 2008). In the southern CCS, decrease and subsequent increase in the volume of suboxic water between 1960 and 2005 were found to be well correlated with shifts in the phases of the Pacific Decadal Oscillation (PDO) index (Deutsch et al., 2011). During cool PDO periods, the thermocline and OMZ shoals increasing the supply of nutrients and resultant rain of organic materials that drive respiratory losses of DO. A larger fraction of this respiration also takes place within shallower OMZ reinforcing the tendency toward suboxia expansion. In the North Atlantic, declines in DO have been linked to weakening of trade winds associated shifts in the Atlantic Multidecadal Oscillation index (AMO) that are important to the ventilation of subsurface water masses (Montes et al., 2016). Anomalous events detected from long-term time series have also provided insights into the linkages between basin-scale changes in ocean conditions. In 2002, increased supply of nutrients from the Gulf of Alaska heightened coastal productivity, oxygen demand and the eventual emergence of nearshore hypoxia in the northern CCS (Grantham et al., 2004). These observations highlight the sensitivity of EBUS oxygen to climate through changes in winds, stratification and circulation that takes place over local to basin scales.

Because oxygen dynamics in EBUS are the product of complex interplay of physical and biogeochemical factors, future changes are unlikely to simply follow mean ocean trends in deoxygenation. Indeed, model projections suggest that DO in the tropical ocean may rise in the future if warming slackens trade winds that drive equatorial upwelling (Bianchi et al., 2018). In contrast, coastal upwelling as first proposed by Bakun (1990) is projected to increase as enhanced continental warming strengthens low-pressure systems on land

relative to high-pressure systems in the sea. The growth of modelling and observational studies stimulated by Bakun (1990) has provided general support for increased upwelling in recent decades (Sydeman et al., 2014; Varela et al., 2015; Wang et al., 2015). The deepening of this literature has also revealed important differences in how different EBUS and regions within EBUS are changing (Aravena et al., 2014). Support for upwelling increases are strongest for the HCS, CCS, and BCS (Sydeman et al., 2014). Within systems, evidence of upwelling increases is strongest for poleward regions of individual EBUS (Garcia-Reyes et al., 2015). Our understanding of the potential mechanisms behind upwelling intensification has similarly diversified. Recent work has emphasized the importance of poleward movement and expansion of oceanic high-pressure systems from climate change in structuring upwelling winds (Garcia-Reyes et al., 2015; Rykaczewski et al., 2015).

Eastern boundary upwelling systems are oceanic in their nature and discussions of future trends have most often focused on offshore and wind-forced changes. However, studies illustrating the effects of terrestrial nitrogen inputs on the nitrogen budget (Howard et al., 2014) and phytoplankton bloom formation (Beman et al., 2005) in coastal upwelling systems, as well as the effects of atmospheric iron deposition in enhancing primary production that contribute to DO losses in OMZ (Ito et al., 2016) suggests that projections of future oxygen trends in EBUS will need to consider the role of pollution from human activities across a variety of scales. As the ocean warms, increasing stratification will inhibit the resupply of oxygen into the ocean but this change can also reduce primary production that fuels oxygen loss, and suppress the upwelling of oxygen-poor waters to the coast. In addition, increased upwelling may lead to greater flushing of nearshore waters so that local development of hypoxia may be moderated in intensity and/or location. These are important factors that can serve to offset deoxygenation trends faced by EBUS. At the moment, their influence is poorly resolved and contributes to real uncertainties in the projections of future conditions in EBUS.

2.4.4 Ecosystem consequences

Oxygen availability is a major environmental gradient that organizes the structure and functioning of marine ecosystems. The effects of oxygen change also tends to be highly non-linear with declines when DO is

already low, soliciting disproportionately large biological responses (Pörtner, 2010). As a result, the expression of ocean deoxygenation in EBUS is expected to bring about important changes in ecosystem dynamics. The potential ecosystem consequences of future ocean deoxygenation can be illustrated by how marine ecological communities and biogeochemical cycles shift in response to currently observed gradients in DO.

Oceanic microbes play a dominant role in elemental cycles that govern the supply of nutrients, availability of trace elements, DO levels, and seawater pH. The metabolic processes of microbes can exhibit sharp thresholds in activity as DO declines from hypoxic to suboxic and anoxic conditions (Ulloa et al., 2012). One particularly important suite of processes is the cycling of nitrogen – a fundamental nutrient that limits ocean productivity. The natural decay of organic matter releases nitrogen compounds that are readily used by phytoplankton to fuel their growth. In DO-deficient waters, however, microbes that transform biologically-available nitrogen compounds into inert N_2 gas become active (Wright et al., 2012). This important loss of nutrients only occurs when DO drops below $5 \mu\text{mol kg}^{-1}$ (Babbin et al., 2014) or less (Bristow et al., 2017) and the expansion of anoxic water in the eastern tropical North Pacific OMZ has been linked to increasing loss of nitrogen between 1972 and 2012 (Horak et al., 2016). The onset of suboxia can also have global consequences for the climate system. Nitrous oxide (N_2O) is an important greenhouse gas with 298 times the heat trapping capacity of CO_2 and oxygen-deficient waters are a major oceanic source of N_2O to the atmosphere. This greenhouse gas is produced by microbes as a byproduct of nitrogen metabolism and its production is accentuated as DO levels decline. Because low DO and active nitrogen cycling converge in productive EBUS, resulting in exceptional hotspots for N_2O emissions (Arevalo-Martínez et al., 2015), it has been postulated that the ocean deoxygenation and upwelling intensification will strengthen this positive feedback in the climate system (Codispoti, 2010). For EBUS, local feedbacks will also be important to consider. The production of hydrogen sulphide by microbes is an active process in oxygen-poor environments particularly in sediments. In the HCS and the BCS, transient water column accumulation of hydrogen sulphide, a compound toxic to wide suites of marine fish and invertebrates, have been observed (Schunck et al., 2013) and enhancement of coastal upwelling by climate change has been postulated to increase

the likelihood of such events (Bakun, 2017). Concerns for potential cascading ecosystem consequences of expanding hypoxia also arise from the supply of iron and phosphorus from sediments (Capone & Hutchins, 2013). The fluxes of these elements are tightly linked to oxygen concentration and can be instrumental in regulating productivity directly or indirectly through controls on nitrogen fixation (Moore et al., 2013).

The sensitivity of microbial communities to DO declines is mirrored by the sensitivities exhibited by the broader suite of marine animals that also make up the water column and seafloor ecological communities of EBUS. Compilations of laboratory studies on marine life from around the world highlight the potential for lethal and sub-lethal effects of low DO to occur well above the canonical threshold of $60 \mu\text{mol kg}^{-1}$ (Vaquer-Sunyer & Duarte, 2008). This holds particular concerns for EBUS as DO levels readily range below $60 \mu\text{mol kg}^{-1}$ in the HCS, CCS, BCS (Helly & Levin, 2004). Considerable details are lost of course when the DO sensitivities of many species are aggregated into global mean values, and myriad of life history adaptations that are known from inhabitants of OMZ (Gibson & Atkinson, 2003) might suggest the potential for reduced biological vulnerability to future oxygen declines in systems where exposure to low oxygen conditions is already common. While some taxa that inhabit such systems can exhibit dampened sensitivity to oxygen declines, this sensitivity is by no means universal (Chu & Gale, 2017; Seibel et al., 2016). Across oxygen-deficient water column and seafloor habitats in EBUS, episodes of DO decline are often accompanied by community changes. For example, seasonal formation of hypoxia in Saanich Inlet in British Columbia results in rapid reorganization of fish and crustacean communities as more oxygen sensitive species are displaced from zones of hypoxia (Chu & Tunnicliffe, 2017). In the central CCS, ENSO-driven changes in DO are propagated into estuaries and can drive population level changes in fishes that rely on estuarine nursery habitats (Hughes et al., 2015). To the south, long-term observations that began in 1951 have revealed that abundance of deep-water fishes is reduced by 63% as the system transitions into periods of relatively low DO levels (Koslow et al., 2011). The authors speculated that the decline reflects the loss of refuge from visual predators as oxygen-loss makes deeper habitats become in accessible to prey fishes. This habitat compression effect has also been proposed as a contributor to the opposing fluctuations in the abundance of sardines and anchovies in the HCS

where the avoidance of oxygen-poor waters by sardines represent a loss of foraging habitat that can be utilized by more hypoxia-tolerant anchovies (Bertrand et al., 2011).

2.4.5 Societal consequences and implications of continuing ocean deoxygenation

Eastern Boundary Upwelling Systems link oxygen changes that are taking place in the global ocean with impacts to some of the world's most productive coastal ecosystems. While this represents one of the most direct connections between ocean deoxygenation and society, considerable uncertainties remain. In fact, one immediate societal consequence of ocean deoxygenation may be increasing uncertainty in our ability to rely on the productivity of coastal ecosystems or in the efficacy of management approaches that have been employed to date.

In the CCS, coast-wide fishery-independent surveys of groundfish stocks used to inform stock assessments have revealed a strong influence of DO on where fish are distributed, their condition and catch per unit effort – a key metric used to estimate abundance (Keller et al., 2010, 2015). Expansion and intensification of low oxygen zones can thus not only influence where fish will be caught but also how managers estimate population size and management targets. In the BCS, episodic hypoxia events that drive mass strandings of the rock lobsters (*Jasus lalandii*) results in destabilization of catch and mortality rates that challenges the management of a commercially important fishery that is already under pressure from over-exploitation (Branch & Clark, 2006; Cockcroft, 2001). Continuing ocean deoxygenation can also bring ecological surprises, whose occurrence and consequence may be difficult to forecast. Recent rapid poleward range expansion by the hypoxia-tolerant jumbo (Humboldt) squid (*Dosidicus gigas*) (Figure 2.4.4) in the California and Humboldt Current Systems have been putatively linked to expansion of low oxygen zones (Gilly et al., 2013; Stewart et al., 2014). While the expansion appears to be have abated at the moment, the sudden introduction of a large, highly active predator presented considerable challenges to projections of ecosystem changes in EBUS.

The uncertainty imposed by continued ocean deoxygenation will amplify and be amplified by challenges from future ocean warming and the progression of ocean acidification (Breitbart et al., 2015). The former

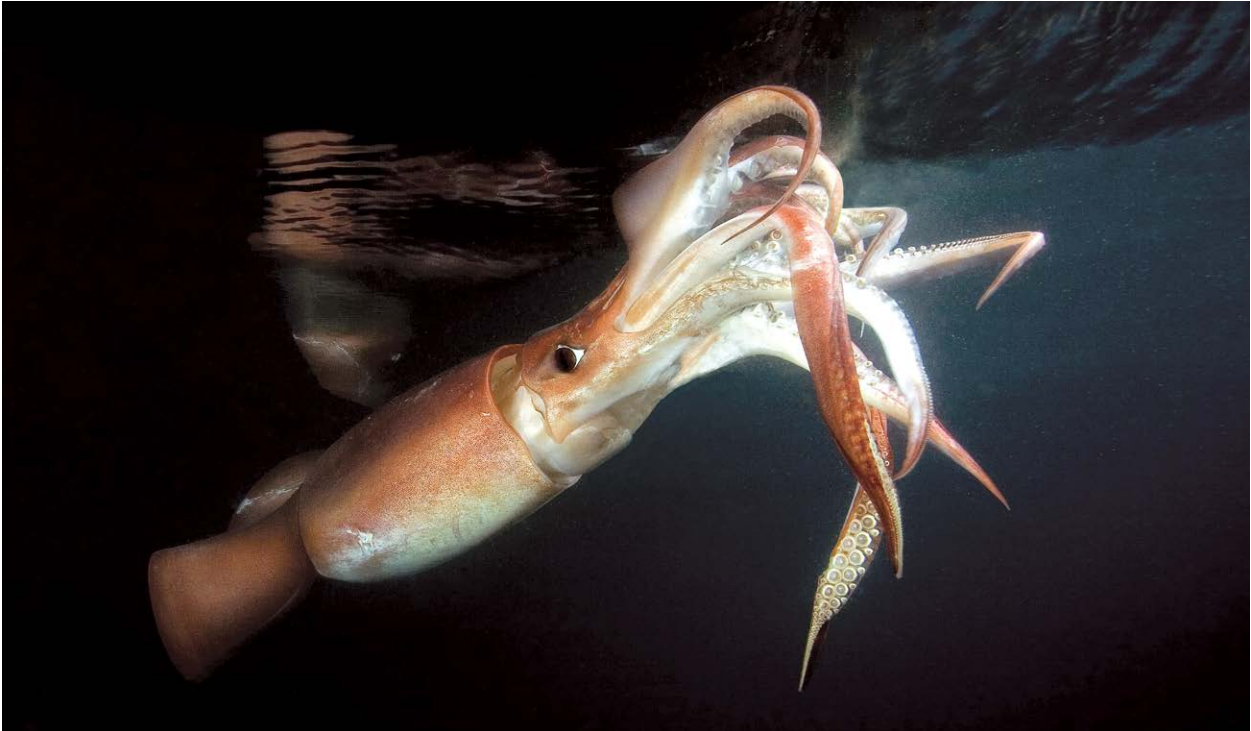


Figure 2.4.4 Jumbo (Humboldt) squid *Dosidicus gigas* in Sea of Cortez, Baja, California © Waterframe / Alamy stock photo.

reflects the effects of greenhouse gases on the earth's heat balance and the latter reflects the impacts of CO₂ emissions and its storage in the ocean on seawater pH and associated chemistry. Termed “hot, breathless, and sour,” these multiple stressors can act synergistically to impact marine life (Gruber, 2011). For example, ocean warming accentuates the impacts of hypoxia in part by increasing organismal demand for oxygen (Pörtner & Knust, 2007; Vaquer-Sunyer & Duarte, 2011). The effects of ocean acidification are similarly magnified when organisms are also challenged by the stress of hypoxia (Gobler & Baumann, 2016; Miller et al., 2016).

Although ocean acidification is a globally pervasive problem, the interaction between ocean acidification and ocean deoxygenation represent particularly acute coupled stressors for EBUS. Low DO levels common to EBUS ultimately reflect the cumulative breakdown of organic matter which consumes oxygen and releases CO₂. As a result, some of the highest levels of pCO₂ measured in surface ocean waters can be found in EBUS (Emeis et al., 2017; Feely et al., 2016; González-Dávila et al., 2017; Shen et al., 2017), reaching levels not expected for the mean global ocean until the middle of the 21st century. As a result, EBUS represent hotspots for both ocean deoxygenation and ocean acidification (Chan et al., 2017; Reum et al., 2015), where the addition of CO₂ from society's emissions to an already

elevated background level of CO₂ puts systems close to, and in many instances past, thresholds for biological impacts (Bednaršek et al., 2017).

2.4.6 Conclusions / Recommendations

The impacts of ocean deoxygenation will not be evenly felt across ecosystems. For EBUS where DO values already centre near thresholds for biological impacts, the confluence of future OMZ expansion and climate modulation of coastal upwelling and local productivity point to their risk as early impact systems to global deoxygenation. The true scope of ocean deoxygenation's impacts, however, will be dependent on the realized rate and scale of oxygen changes, and the capacity of natural and human systems to mitigate and adapt to a more hypoxia- or anoxia-prone ocean. Thus, even as global greenhouse gas emissions lie at the heart of ocean deoxygenation, recent assessments have highlighted the importance of local to regional-scale actions in fostering global-change readiness in EBUS (Chavez et al., 2017). Actions include sustaining and growing integrated ocean observing capacity to detect and track the progression of hypoxia and covarying ocean acidification and warming stressors. Integration in observing reflects the need for not only physical and biogeochemical measurements but also those for physiological, population, and ecological metrics. In turn,

the ability to identify and provide early warning of which species or parts of the system are most vulnerable can guide priorities in protection. There is also a premium need for research and partnerships that will grow new local mitigation and adaptation solutions that decision-makers can draw on to ostensibly buy time until the root causes of ocean deoxygenation are addressed (Klinger et al., 2017).

While the challenges are great, efforts are underway to enhance monitoring and research to support ocean deoxygenation actions. For example, because of the strong covariation between carbonate and oxygen chemistry and the utility of combined measurements in assessing data accuracy, the growth of ocean acidification observing networks regionally (e.g. OA-Africa, California Current Acidification Network, etc.) and globally (i.e. Global Ocean Acidification Observing Network) is also leading to enhanced observing capacity, intellectual exchanges and partnerships for ocean oxygen monitoring. Recognition of coastal vulnerability to ocean acidification and oxygen deoxygenation has also stimulated new research into the use of green infrastructure in the form of seagrass and kelp beds to locally mitigate oxygen and pH declines (Duarte et al., 2017). Equally important are efforts to identify and promote sources of biological and ecological resilience to ocean chemistry changes. Examples of local adaptation to ocean acidification highlight the potential for evolutionary rescue (Munday et al., 2013) in conferring some level of resilience to ocean deoxygenation. The use of available management tools such as marine protected areas to support climate change adaptation is another arena of active research (Roberts et al., 2017). In Baja California, marine reserves supported larger and more demographically-diverse populations of pink abalone (*Haliotis corrugate*) that were better able to withstand and recover from hypoxia events relative to unprotected areas (Micheli et al., 2012). Ocean deoxygenation has quickly emerged as a leading pathway for climate change impacts. Monitoring and research efforts that identify what's most at risk and grows our portfolio of mitigation and adaptation solutions will play a disproportionate role in preparing coastal EBUS communities and nations for the changes ahead.

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