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# Dynamics and distribution of natural and human-caused coastal hypoxia

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## Abstract

Water masses can become undersaturated with oxygen when natural processes alone or in combination with anthropogenic processes create enough carbon that is aerobically decomposed faster than the rate of oxygen re-aeration. The dominant natural processes usually involved are photosynthetic carbon production and microbial respiration. The re-aeration rate is indirectly related to its isolation from the surface layer. Hypoxic water masses ($<2\,\mathrm{mg\,L^{-1}}$, or approximately 30% saturation) can form, therefore, under "natural" conditions, and is more likely to occur in marine systems when the water residence time is extended, water exchange and ventilation is minimal, stratification occurs, and where carbon production and export to the bottom layer are relatively high. Hypoxia has occurred throughout geological time and naturally occurs in oxygen minimum zones, deep basins, eastern boundary upwelling systems and fjords. Hypoxia development and continuation in many areas of the world's coastal ocean is accelerated by human activities, especially where nutrient loading increased in the Anthropocene. This higher loading set in motion a cascading set of events related to eutrophication. Nutrient loading is likely to increase further as population growth and resource intensification rises, especially in developing countries dependent on crops using fertilizers, and it is likely that the occurrence and persistence of hypoxia will be even more widespread and have more impacts than presently observed. Climate change will further complicate the causative factors.

## 1 Introduction

Over the past five to ten years, changes in the ocean's dissolved oxygen content have become a focal point of oceanic research. The oxygen content in the open ocean appears to have decreased in most (but not all) areas (Gilbert et al., 2009). At the same time, low oxygen areas, also known as "dead zones," have spread in the coastal oceans during the last three decades. These changes in oxygen are an increasingly important topic due to large impacts on the ecosystems and biogeochemical cycles. Hypoxic (low dissolved oxygen) and anoxic (no oxygen) aquatic environments have occurred through geologic time. In fact, the biological and physical processes that formed large deposits of oil in geologic formations are the same as those occurring in oil rich areas where hypoxia and anoxia exist presently, such as the Santa Barbara Basin, California, USA, a silled basin interacting with an oxygen minimum zone, and the continental shelf of the northern Gulf of Mexico adjacent to the outflow of the Mississippi River (Rabalais et al., 2007). Low oxygen waters in the world oceans are normal, or naturally formed, in areas such as oxygen minimum zones (OMZs), deep basins, upwelling areas of eastern boundary currents and fjords (Helly and Levin, 2004). But in the 19th, 20th and 21st centuries, the activities of humans have resulted in many more areas of hypoxia than occurred historically and aggravated conditions in areas that were already low in oxygen (Fig. 1) (Díaz and Rosenberg, 1995, 2008; Vaquer-Sonyer and Duarte, 2008; Gooday et al., 2009).

In contrast to what occurs in the OMZs and upwelling zones, much of the hypoxia and anoxia in shallow coastal marine areas has developed within the last 50 yr and is closely associated with anthropogenic activities. Díaz and Rosenberg (1995) noted that no other environmental variable of such ecological importance to estuarine and coastal marine ecosystems has changed so drastically, in such a short period of time. They noted consistent trends of increasing severity in duration, intensity, or extent in areas where hypoxia has a long history, which were coincidental with an increase in human activities. In 1995 there were 195 literature-documented areas of human-caused coastal hypoxia. In their most recent compilation, Diaz and Rosenberg (2008) documented just over 400 such areas in the world's coastal ocean covering more than 245 000 $\mathrm{km}^2$ of sea bottom (Fig. 1). The worldwide distribution of coastal hypoxia is related to major population centers or is closely associated with developed watersheds that export large quantities of nutrients, specifically nitrogen and phosphorus. Up to 1970, there were scattered reports of hypoxia in North America and northern Europe. By the 1990s, coastal hypoxia was prevalent in North America, northern Europe, and Japan. By the 2000s, there were increased reports of hypoxia in South America, southern Europe, and Australia (Fig. 1). Considering the close association of human population and hypoxia, it is unlikely that Asia and the Indo-Pacific have no hypoxia.

Many systems that are currently hypoxic were not when first studied. For systems with historical data from the first half of the 20th century, declines in oxygen concentrations started in the 1950s and 1960s for the northern Adriatic Sea, between the 1940s and 1960s for the northwest continental shelf of the Black Sea (see case studies), and in the 1970s for the Kattegat (Baden et al., 1990a). Declining dissolved oxygen levels were noted in the Baltic Sea as early as the 1930s (Fonselius, 1969), but it was in the 1950s that hypoxia became widespread (Karlson et al., 2002). Other systems have experienced hypoxia since the beginning of oxygen data collection in the 1930s for the Chesapeake Bay and the 1970s for the northern Gulf of Mexico (see case studies).

Hypoxia now occurs in many areas where it did not occur before, including estuarine and coastal areas. For example in the Kattegat, between Sweden and Denmark, classic descriptions of well-developed, advanced successional stage benthic communities were made by Petersen (1911). In the mid 1980s Petersen's stations were revisited, and it was determined that the benthic communities at most of the stations have undergone substantial changes over the 70 yr period (Pearson et al., 1985). The general decrease of large species and increase of small-sized species was consistent with the hypothesis that intermittent low oxygen conditions affected many of the shallower areas of the Kattegat and kept the system in an early successional stage (Pearson et al., 1985). Moreover, in the deeper northeastern area where there is no evidence of low oxygen, both total biomass and mean individual size of the dominant species were greater than before, possibly suggesting that in these areas only enrichment had taken place (Pearson et al., 1985). By the mid 1980s severe seasonal hypoxia had developed in the southeastern Kattegat and negatively affected both the benthos and Norway lobster (*Nephrops norvegicus*) fisheries (Rosenberg and Loo, 1988; Baden et al., 1990a,b).

The negative effects of hypoxia include loss of suitable and required habitat for many bottom-dwelling fishes and benthic fauna, habitat compression for pelagic fishes, direct mortality, increased predation, decreased food resources, altered trophic energy transfer, altered bioenergetics (physiological, development, growth, and reproductive abnormalities) and altered migration. These result in reduced fisheries, including valuable finfishes and crustaceans (Rabalais and Turner, 2001; Eby and Crowder, 2002; Wu, 2002; Baird et al., 2004; Breitburg et al., 2009). Increasing nutrient loads that also change the nutrient ratios can affect the composition of the phytoplankton community and can shift trophic interactions. Hypoxia also alters or interrupts ecosystem functions and services such as nutrient cycling and bioturbation (Turner et al., 1998; Dortch et al., 2001; Childs et al., 2001; Turner, 2001; Rabalais, 2004; Weissberger et al., 2009; Middelburg and Levin, 2009).

This paper is one of several in this special issue culminating from the activities of the SCOR (Scientific Committee on Ocean Research) Working Group 128 on "Natural and Human-Induced Hypoxia and Consequences for Coastal Areas." The focus of the Working Group was coastal waters, but other marine systems with hypoxia, such as anoxic basins, oxygen minimum zones, or areas of upwelling are of interest for understanding comparative processes and dynamics. Hypoxic waters from these systems may also impinge on coastal areas with similar affects as human-induced hypoxia (Levin et al., 2009).

## 2 Definitions and terminology

Aquatic ecologists have borrowed the term hypoxia (low oxygen) from the medical community but the meaning and processes are the same. The medical condition is where the body or a part of the body is deprived of adequate oxygen. Similarly, a water body can be deprived of adequate oxygen. Strenuous physical exercise may reduce concentrations of oxygen in the blood as a normal course of activity. Similarly, dissolved oxygen concentrations will fall below 100% saturation through the process of respiration. Hypoxia, or anoxia (no oxygen), as considered in this paper, becomes detrimental to aerobic, aquatic organisms when behavioral and physiological responses result in altered behavior or negative impacts, such as reduced growth, loss of reproductive capacity, mortality, reduced biodiversity, and loss of secondary production, including fisheries. There is no single defined concentration at which marine, coastal or estuarine waters become hypoxic to the resident organisms, nor is there consistency in units of oxygen used to express hypoxia.

Most definitions of hypoxia consider that there are sublethal biological implications. Marine animals, however, vary in their responses to declines in dissolved oxygen. For example, cod growth is reduced below 70% dissolved oxygen saturation (Chabot and Dutil, 1999). Shrimp and fish avoid dissolved oxygen below 2 mg $L^{-1}$ (approximately 30% oxygen saturation) in the northern Gulf of Mexico hypoxic zone, while sharks and rays emigrate from the area at a dissolved oxygen concentration of 3 mg $L^{-1}$ (Rabalais et al., 2001a). For this reason, Rabalais et al. (1991) and others (Renaud et al., 1986) defined hypoxia in the northern Gulf of Mexico as <2 mg $L^{-1}$ based on the behavioral response of most demersal fish, crabs and shrimp that move away from these low dissolved oxygen waters.

It is clear that behavioral, physiological and reproductive responses among taxa differ by taxon, stage of life, and history of exposure to low oxygen of varying levels (Rabalais and Turner, 2001 and chapters therein on taxa from foraminiferans to fishes; Gray et al., 2002). Vaquer-Sunyer and Duarte (2008) more recently surveyed the literature on benthic organisms to evaluate experimentally-derived oxygen thresholds for lethal and sublethal responses to hypoxia. They found an order-of-magnitude variability in the lethal thresholds of oxygen concentrations among benthic marine organisms, which does not lend support for a single universal threshold value. They concluded that the conventional definition of 2 mg $L^{-1}$ to designate hypoxia waters is below the empirical sublethal and lethal dissolved oxygen thresholds for half of the species tested. Most aquatic ecologists and oceanographers would agree that there is no "conventional" definition of hypoxia and that the relevant thresholds are context-dependent.

The SCOR WG 128 agreed to consider a general value for hypoxia of 30% oxygen saturation, or about 2 $\mathrm{mg\,L^{-1}}$ (1.4 $\mathrm{ml\,L^{-1}}$ or 63 µM) (Fig. 2). This value is most relevant as a threshold to coastal waters and organisms that have evolved in normoxic waters without severe hypoxia as an evolutionary stress. However, throughout this article there will be multiple definitions and multiple units. Where possible, we will to provide suitable conversions and put the terminology in context within the habitats of concern.

The most commonly used units for dissolved oxygen are $\mathrm{mg\,L^{-1}}$, $\mathrm{ml\,L^{-1}}$, and µM, and different concentrations are considered to be the definition of hypoxia (Fig. 2). As mentioned above, a commonly used definition of hypoxia is dissolved oxygen $<2\,\mathrm{mg\,L^{-1}}$, the equivalent of 1.4 $\mathrm{ml\,L^{-1}}$ or 63 µM. Others define hypoxia as 2 $\mathrm{ml\,L^{-1}}$ (2.8 $\mathrm{mg\,L^{-1}}$ or 91.4 µM) (Díaz and Rosenberg, 1995). Physical oceanographers commonly use units of $\mathrm{ml\,L^{-1}}$ with a definition of hypoxia at dissolved oxygen $<2\,\mathrm{ml\,L^{-1}}$. Many aquatic ecologists use $\mathrm{mg\,L^{-1}}$, but not all, and at different concentrations to define hypoxia. Researchers who focus on oxygen minimum zones and anoxic basins prefer to work in µM or $\mathrm{ml\,L^{-1}}$ units, with multiple levels of low oxygen, hypoxia, sub-hypoxia, and anoxia at $<22$ µM or 0.5 $\mathrm{ml\,L^{-1}}$. An attempt to harmonize geologists and ecologists on definitions of oxygen deficiency was fraught with inconsistencies within the groups and definitely between the groups (Tyson and Pearson, 1991). In fact, Tyson and Pearson (1991) were the first to "blame" researchers from the northern Gulf of Mexico as the perpetrators of the use of the medical term "hypoxia" to connote low dissolved oxygen waters in marine ecosystems. The symposium coordinated by Tyson and Pearson (1991) on modern and ancient continental shelf anoxia was the first agreement to disagree about a conventional definition of hypoxia.

The "measure" of hypoxia varies by location around the globe. The extent of Chesapeake Bay, USA hypoxia, defined as $<2\,\mathrm{mg\,L^{-1}}$ dissolved oxygen, is most often expressed as volume (Hagy et al., 2004) instead of areal extent on the sea bed, which differs from most measures of hypoxia that employ the area of bottom-water hypoxia, e.g., northern Gulf of Mexico, western shelf of India, and Baltic Sea. There is, however, a close relationship between volume and areal extent of hypoxia in the Chesapeake Bay, USA and the northern Gulf of Mexico (Fig. 3), so that there is some basis for

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comparison (Fig. 4). As long as a geographic area is consistent in the use of the measure of hypoxia and/or provides multiple metrics, then efforts to mitigate hypoxia will have a baseline against which progress can be measured, e.g., a goal of 5000 $\mathrm{km}^2$ areal extent of hypoxia for the northern Gulf of Mexico on a five-year running average by 2015 (Hypoxia Task Force, 2009).

Area or volume estimates of dissolved oxygen concentrations less than a certain value (e.g., $<2\,\mathrm{mg\,L^{-1}}$, moderate hypoxia and $<1\,\mathrm{mg\,L^{-1}}$, severe hypoxia) do not adequately describe the total oxygen deficiency of the water column. Dissolved oxygen "condition" can be determined as a deficit (or a surplus) of oxygen below (or above) 100% saturation at a known temperature and salinity (Justić, 1991). Calculations of oxygen deficiency would be useful in models that link oxygen budgets with carbon budgets and nutrient loads. Changes in nutrient management that can be expressed in altered loads of nitrogen and phosphorus and subsequent changes in carbon and oxygen budgets provide a more comprehensive status of ecosystem condition than area or volume calculations.

For other considerations, such as the effect of hypoxia on benthic macroinfauna, the severity, or period of time that the dissolved oxygen level is below 1 or $2\,\mathrm{mg\,L^{-1}}$, is a more appropriate measure of hypoxia effects on organisms. For example, the period of exposure to hypoxia on the inner shelf of the northern Gulf of Mexico was an important factor in the response of the benthic macroinfaunal community (Fig. 5) (Rabalais et al., 2001b; Baustian and Rabalais, 2009). The New York State Department of Environmental Conservation, besides mapping area of dissolved oxygen levels, determines the weekly bottom-water oxygen levels, and from that determines the number of days of hypoxia ($<3.5\,\mathrm{mg\,L^{-1}}$) each year (National Research Council, 2000).

Adequate characterization of hypoxia calls for measurements on multiple temporal and spatial scales, which are not always possible because of limited resources. Determination of the extent of hypoxia, usually maximal in summer through early fall, depends on a quasi-synoptic survey over a consistent grid. More frequent mapping of the extent, however, would help ensure that the metric is representative of the system in any given year. A series of years with active tropical storm seasons or abnormal oceanographic currents may skew the area or volume data, such as when the water column is disrupted by a storm within a week or two of the mapping survey (Rabalais et al., 2007, 2009). Depending on the size of the anticipated area, the costs of multiple mapping ventures may be cost prohibitive. Representative areas, within the larger area affected, may take less time to document thereby reducing the cost, so that greater temporal coverage over the annual cycle may be possible. Finer temporal resolution can be recorded with deployed or profiling instrumentation; finer spatial resolution can be achieved with towed instrumentation or autonomous underwater vehicles. In most cases, however, the resources are not available for adequate assessments, especially with the appropriate physical, chemical and biological parameters. A monitoring or assessment program would need to balance the information gained with the funds available to optimize data acquisition

The hypoxia literature has generated terms, in some cases used inconsistently. Where did the term "dead zone" originate? In Louisiana coastal waters (northern Gulf of Mexico) when the oxygen falls below $2\,\mathrm{mg\,L^{-1}}$, a shrimper dragging a bottom trawl will not catch any shrimp (or demersal organisms) (Renaud, 1986; Rabalais et al., 2001a). Bottom waters with dissolved oxygen $<2\,\mathrm{mg\,L^{-1}}$ were first documented with measurements in these waters in August 1972 (Ward et al., 1979) and increasingly in years to come. The shrimpers called these "dead waters" along with another term they coined "red waters" when they encountered blooms of dinoflagellates, suspected to be *Noctiluca scintillans* or *Akashiwo sanguinea*. Ecologists examining low oxygen waters in brackish Lake Pontchartrain north of New Orleans, Louisiana would encounter dead blue crabs in traps and dead *Rangia* clams in the sediments. A reporter from the Baton Rouge, Louisiana paper "The Morning Advocate" asked if these dead areas could be described as a dead zone, and the term first appeared in print in 1981 or 1982. Perhaps there was co-evolution of this term elsewhere during the period of increasing eutrophication and hypoxia in coastal waters in the early 1970s. In their initial paper outlining the increase in areas of hypoxia, Díaz and Rosenberg (1995) did not use the term “dead zone,” but it appeared in Diaz (2001), and eventually in the title of their most recent review (Díaz and Rosenberg, 2008). In Chesapeake Bay, USA, where hypoxia has been present since at least the 1920s, Truitt used the term oxygen desert in 1938 to report that “studies are in progress, which aim to show how important these so-called oxygen deserts are in the economy of the Bay” (Hagy et al., 2004).

In reality the link between hypoxia and “dead zone” best applies to coastal waters where oxygen depletion occurs in otherwise normoxic (oxygenated) waters, with evident migration or mortality of charismatic megafauna (fishes and large invertebrates). Even in the absence of larger fauna, smaller meiofaunal taxa (foraminiferans, nematodes) and microbes persist, such that the regions are not truly “dead.” And, the upper water column is fully oxygenated and supports diverse metazoan communities, including productive fisheries. The term “dead zone” is not appropriate for shelf and slope regions with permanent oxygen depletion (oxygen minimum zones or oxygen minimum layers), where fully developed, diverse animal communities with megafauna can occur at concentrations $<2\,\mathrm{mg\,L^{-1}}$ (Levin, 2003; Gooday et al., 2009).

## 3 Dynamics

The combination of physical, chemical and biological processes that lead to hypoxia differ in magnitude and importance by water body, but there is one basic response. Hypoxia, or oxygen deficiency, occurs when the amount of dissolved oxygen in the water column is decreased by the process of respiration at a faster rate than resupply through air-sea exchange, photosynthetic production of oxygen, or advection. Hypoxia is usually associated with a density barrier, caused by temperature, salinity or both that prevents the diffusion of oxygen from a higher concentration layer to a lower concentration layer, but may occur without stratification in a smaller, highly eutrophic system, typical in the form of diel-cycling hypoxia (Flindt et al., 1997; Tyler et al., 2009), or in larger systems that have intense heterotrophic activity (Verity et al., 1999). Hypoxia may also be “introduced” by advection of low-oxygen source waters. In the eastern Pacific, changes in input of intermediate, low-oxygen water masses contribute to temporal observations of oxygen declines (Chan et al., 2008; Bograd et al., 2008). The Yaquina Bay system in Oregon, USA, experiences episodes of hypoxia when low dissolved oxygen water is advected into the system from the continental shelf (Brown et al., 2007). In Dokai and Ise Bays, Japan (Nakata et al., 1997; Ueda et al., 2000) and Chesapeake Bay, USA (Breitburg, 1992), local winds drive advection of seasonally hypoxic deep waters into shallow areas that reduce shellfish populations.

The carbon source that fuels the respiratory reduction of oxygen most often originates from settled phytoplankton production (autochthonous), but may have a natural terrestrial source (allochthonous) or be from discharge of sewage/industrial waste. Settled phytoplankton production in the form of senescent cells, zooplankton fecal pellets or marine aggregates will settle at a density gradient, as is the case with oxygen minimum zones and subsurface oxygen minima (e.g., off productive river deltas). In other instances the organic carbon sinks to the seabed in shallower waters, where the respiratory decomposition depletes the oxygen in the water column below a strong density gradient. Examples of subsurface oxygen minima, lower water column hypoxia, and both are illustrated for a series of stations near the Mississippi River plume in Fig. 6.

Human-induced coastal hypoxia is a symptom of eutrophication, the increase in the rate of carbon production and carbon accumulation in an aquatic ecosystem (Rabalais 2004, modified from Nixon, 1995). Eutrophication was initially a description for the natural aging process of freshwater systems, but has more recently been applied to estuarine and coastal systems. As noted above, the source of the increased organic carbon may come from within the system (autochthonous) or from outside the system (allochthonous). This distinction is relevant when management strategies are developed to reverse eutrophication, and its negative symptoms such as hypoxia, and to identify the sources and mechanisms of carbon accumulation. For example, a coastal system could become eutrophic from an increased delivery of organic carbon from terrestrial sources or from nutrient-enhanced primary production resulting from increased nutrient loads. Reducing organic loading from riverine sources would require different management strategies than that required to reduce nutrient loads.

The causes of eutrophication, and its negative outcome hypoxia, may include changes in physical characteristics of the system such as changes in hydrology, changes in biological interactions such as reduced grazing or an increase in the input of organic and inorganic nutrients. While the series of causes may include direct natural or anthropogenic carbon enrichment, eutrophication and hypoxia in the coastal ocean and in the 20th and 21st centuries is more often caused by excess nutrients that would otherwise limit the growth of phytoplankton. There is little doubt that there have been ecosystem-level changes in coastal systems as a result of eutrophication and associated hypoxia (Rabalais and Turner, 2001; Turner et al., 2008; Ekau et al., 2009; Kemp et al., 2009; Levin et al., 2009).

Coastal systems include estuaries and continental shelf waters that extend from the barrier island shoreface to the edge of the continental shelf at 100 m water depth. Estuaries vary in physiography, but those most conducive to the formation of hypoxia are characterized by longer water residence times that allow for accumulation of carbon and respiratory depletion of oxygen. Stratification is also a key factor for the development and maintenance of hypoxia in an estuary, for example Chesapeake Bay, USA (Kemp et al., 2005), Ardbear Salt Lake, Ireland (Henry et al., 2008) or Wilson Inlet, Australia (Brearley, 2005). Some estuaries, especially those with expanses of macroalgae or sea grasses, will have a diel cycle of high oxygen saturation during daylight photosynthesis but substantially reduced oxygen saturation during night or approaching dawn as respiratory consumption of oxygen exceeds the resupply from air-water exchange. Vegetated estuaries, however, may develop hypoxia as a result of nutrient-driven increases in phytoplankton biomass, epiphytic algae, or floating macroalgae (Schramm, 1999).

Fjords such Saanich Inlet, British Columbia, Canada are naturally susceptible to hypoxia and anoxia because water residence time is long, sills may prevent exchange with oceanic waters, and thermal stratification will form in warmer months preventing reoxygenation from surface waters (Tunnicliffe, 1981). Fjords, however, are also subject to excess nutrients and carbon leading to human-induced hypoxia in a conducive environment, for example Hood Canal in Puget Sound, USA (Parker-Stetter and Horne, 2008) and Himmerfjarden, Sweden (Savage et al., 2002).

Human-caused coastal hypoxia is most likely to occur within the 100-m isobath (Rabalais 2004), for example the northern Adriatic Sea (Justić et al., 1987) or the East China Sea (Li et al., 2002). Often, shallower waters on the open coast are not conducive to the carbon accumulation that supports hypoxia because turbidity caused by resuspension of sediments or delivery of sediments from coastal rivers limits the production of phytoplankton. However, at times during unusual calm weather events large areas of hypoxia can develop, as have been documented on the inner continental shelf of the New York Bight, USA (Boesch and Rabalais, 1991) and the west coast of Florida, USA (FWRI, 2005). Deeper, more open, coastal waters are not as conducive to eutrophication-caused hypoxia as intermediate-depth waters, because nutrients are depleted with distance offshore, less primary production occurs, and less organic matter is exported from surface waters. Along continental margins deeper than 100 m, hypoxia is most likely to occur as a response to coastal upwelling driven primary production or the advection of deeper water hypoxia closer to shore. Examples are shorelines along eastern boundary currents, such as the California Current System off Oregon and Washington states (Grantham et al., 2004; Chan et al., 2008).

Many semi-enclosed seas – entire basins or portions of systems – are conducive to the process of eutrophication because of a physiography that enhances water residence times and because they receive land-based runoff or atmospheric deposition that is less likely to dilute as quickly as on an open coast (Rabalais and Gilbert, 2009). Coastal systems adjacent to many large rivers have received increasingly nutrient-rich discharges. Swift currents that move materials away from the river discharge and that do not permit the development of stratification are not conducive to the accumulation of biomass or depletion of oxygen, for example in the Amazon and Orinoco plumes. The Bay of Bengal sustains a strong potential for impacts from riverine nutrient loads due to the very high nutrient yields in its catchment basins, e.g., via the Ganges/Brahmaputra, Godvari and Mahanadi rivers. The Bay of Bengal is the most open of all the systems receiving high nutrient inputs, with no physical barriers separating its coastal zone from the open ocean. In this case, the rate of exchange between coastal and open ocean water masses will dictate the system's response to the riverine loads. Off east India in the Bay of Bengal hypoxic conditions ($<1.42\,\mathrm{ml\,L^{-1}}$) are persistent on the outer shelf beyond 100 m and the oxygen minimum zone ($<0.5\,\mathrm{ml\,L^{-1}}$) is present from 150 m to about 500 or 600 m (Helly and Levin, 2004; Rao et al., 1994). Similar ocean circulation processes off the Changjiang (Yantze River) and high turbidity in the plume of the Huanghe (Yellow River), China were once thought to be reasons why hypoxia did not develop in those coastal systems. Incipient indications of the beginning of symptoms of cultural eutrophication are becoming evident at the terminus of both these systems as nutrient loads increased (Zhang, 1994; Liu et al., 2003). Presently, there is a well-documented area of hypoxia off the Changjiang in summer (Li et al., 2002). There is increasing likelihood that more and more coastal systems, where the physical conditions are appropriate, will become eutrophic, especially in developing countries.

The volume of freshwater discharge, exclusive of the nutrient load, can influence residence time, stratification, turbidity and nutrient dilution. High flow years of the Mississippi River result in intensified stratification on the continental shelf, higher chlorophyll biomass and more widespread bottom-water hypoxia (Rabalais et al., 1998). Within the Hudson River estuary higher discharge years result in lower residence time, increased turbidity, less stratification, lower primary production and less eutrophication (Howarth et al., 2000). The same Hudson River discharge onto the continental shelf of the New York bight, however, would be expected to have similar results as Mississippi River outflow with increased stratification, chlorophyll biomass and bottom-water hypoxia. Low oxygen events and shellfish mortalities have been reported from the New York bight historically and during an extreme episode in 1976 (Swanson and Sindermann, 1979; Whitledge, 1985).

## 4 Oxygen minimum zones

Similar to the coastal hypoxia debate, there is no consensus on what oxygen concentration defines oxygen minimum zones (OMZs). This term is generally used to refer to permanent midwater, low-oxygen features in the eastern Pacific, off western Africa and the northern Indian Ocean (Kamykowski and Zentara, 1990; Paulmier and Ruiz-Pino, 2009). Upper oxygen thresholds for OMZ regions range from dissolved oxygen of 9 µmol kg$^{-1}$ (0.2 ml L$^{-1}$) (Kamykowski and Zentara, 1990), to 20 µmol kg$^{-1}$ (0.5 ml L$^{-1}$) (Helly and Levin, 2004; Fuenzalida et al., 2008; Paulmier and Ruiz-Pino, 2009) to 45 µmol kg$^{-1}$ (1.0 ml L$^{-1}$) or 90 µmol g$^{-1}$ (2.0 ml L$^{-1}$) (Karstensen et al., 2008). The term "oxygen minimum layer" is sometimes used to refer to midwater layers exhibiting reduced oxygen relative to waters above and below. Such layers are ubiquitous in the global ocean due to isolation from sources of oxygenation, but often do not reach "hypoxic" oxygen levels, e.g., open Gulf of Mexico midwater layers (Conseil, 1936; Rabalais et al., 2002).

A series of terms describes the overall oxygen conditions. Anoxic refers to waters with virtually no oxygen. Often high levels of sulfide are released into the water column, as in the Black Sea where anoxic conditions extend to within 100–200 m of the surface (Richards, 1965; Konovalov and Murray, 2001). When oxygen is detectable but below about 4.5 µmol kg$^{-1}$ (0.1 ml L$^{-1}$) the water is suboxic, a threshold determined from nitrate remineralization to nitrite (Karstensen et al., 2008).

By any definition, oceanic OMZs are the largest low dissolved oxygen areas on earth (Paulmier and Ruiz-Pino, 2009). The principal factors that lead to the formation of OMZs are high surface productivity, old water mass age, and limited circulation. Increased upwelling, higher productivity, and greater oxygen demand along with isolation of intermediate water masses contribute to a thicker OMZ and lower oxygen concentrations (Helly and Levin, 2004). OMZs are widespread and stable oceanic features occurring at intermediate depths (typically 100–1200 m). They persist for long periods of time (at greater than decadal scales), but have variable upper and lower boundaries that are controlled by natural processes and cycles (Wyrtki, 1966; Kamykowski and Zentara, 1990; Helly and Levin, 2004). Minimum oxygen values typically occur from 300 to 500 m in the southeastern Pacific (Fig. 7) and Arabian Sea with lowest oxygen declining to <1 µmol kg$^{-1}$. Minimum oxygen concentrations are higher (~8–10 µmol kg$^{-1}$) and occur deeper in the northeastern Pacific (600–1000 m) (Fuenzalida et al., 2008; Paulmier and Ruiz-Pino, 2009). The eastern Atlantic has relatively high oxygen minimum values of about 17 µmol kg$^{-1}$ in the South Atlantic and >40 µmol kg$^{-1}$ in the North Atlantic (Karstensen et al., 2008). Globally, where OMZs contact the bottom, about 1 148 000 km$^2$ of continental margin sea floor is estimated to have bottom-water oxygen concentrations <0.5 ml L$^{-1}$. Using the stricter definition of <0.2 ml L$^{-1}$, about 764 000 km$^2$ of seafloor are affected (Helly and Levin, 2004). Oxygen minimum zones often occur beneath upwelling regions, and upwelling-induced productivity is a primary contributor to persistent hypoxia. However, inshore regions subject to seasonal or episodic upwelling can also develop severe hypoxia as deep-water nutrients are added to surface waters increasing production that eventually sinks and decomposes. Hypoxia associated with this type of coastal upwelling is not as long-lived and stable as that associated with OMZs. Upwelling can interact strongly with low-oxygen water masses (OMZs) to produce intense shelf hypoxia; this is observed off of Oregon, USA (Grantham et al., 2004) and Chile (Fuenzalida et al., 2008).

Demersal and pelagic fisheries benefit from the enhanced organic production associated with upwelling when oxygen concentrations remain high. For example, demersal fisheries (scallop, hake, and octopus) flourish under better oxygenated El Niño conditions. This results from the convergence of a combination of land and ocean processes that make these upwelling-influenced coastal zones incredibly productive (Pauly and Christensen, 1995). On the other hand, the resulting oxygen depletion, whether from mid-ocean OMZs or coastal upwelling or their interaction, affects mid-water plankton and pelagic organisms creating areas of low biodiversity, and hostile environmental condition for many commercially valued fisheries resources (Ekau et al., 2009). Monsoon-driven upwelling along the western coast of India migrates onshore and offshore depending on winds and currents. When pushed onshore, there are declining catches of fishes and prawns, and diminished demersal fisheries (Banse, 1959, 1968; Kurian et al., 2007). Intensification of the severity of low oxygen in coastal component of the upwelling zone, most likely due to human-sourced nutrients in recent years (Naqvi et al., 2006), would further impact the fisheries.

Where OMZs and upwelling zones impinge on the seafloor of continental slopes and shelves, and sea mounts, specialized low biodiversity communities have evolved to survive at dissolved oxygen concentrations as low as 0.1 mg $L^{-1}$ or less (Graham, 1990; O'Toole and Bartholomae, 1995; Childress and Seibel, 1998; Levin, 2003; Levin et al., 2009). These OMZ assemblages differ from shallower communities exposed to seasonal or episodic hypoxia in having much lower oxygen tolerance thresholds, morphological adaptations to maximize respiratory surface, specialist rather than opportunistic lifestyles, and potential to utilize chemosynthesis-based nutritional pathways (Levin, 2003).

While there are similarities and differences in the benthic response to both natural and human-induced hypoxia, a key difference between them is the time-scales of their development (Fig. 8). OMZ and upwelling-related hypoxia have been features of our global oceans for centuries (Helly and Levin, 2004). Historically, human-induced hypoxia appears to have been confined to highly modified watersheds or harbors and was not a global phenomenon (for the contrast in time, see Brongersma-Sanders, 1957; Díaz and Rosenberg, 2008). In the relatively short period from about 1950 to today, in response to increasing reliance on industrial fertilizers to feed an ever expanding population, humans have drastically altered global nitrogen cycling and coastal primary production to the point that now human-induced hypoxia has become a principal stressor of coastal systems on a global scale (Galloway et al., 2004, 2008). The relationship between increased fertilizer use and increased number of human-induced hypoxic systems lagged about 10 yr (Fig. 8), the time it took for excess organic matter from primary production to buildup and overwhelm an ecosystem's assimilative capacity to the detriment of higher trophic levels and the benefit of microbes (Díaz and Rosenberg, 2008;

Turner et al., 2008).

The upper limit of the OMZ can undergo large vertical excursions in the southeastern Pacific (Helly and Levin, 2004; Fuenzalida et al., 2008) and in the Arabian Sea (Brand and Griffiths, 2009) with major implications for fisheries and biogeochemical cycling. Forcing can be linked to seasonal winds (Brand and Griffiths, 2009), El Niño Southern Oscillation cycles (Fuenzalida et al., 2009), or longer term trends (Bograd et al., 2008). OMZ shoaling in the eastern tropical Pacific has compressed the depth distribution of tropical pelagic marlins, sailfish, and tunas into a narrow surface layer of oxic water about 50 to 100 m thick. The hypoxic water is a lower habitat boundary in the Pacific, but not in the western North Atlantic, where oxygen is not limiting. Eastern Pacific and eastern Atlantic sailfish are larger than those in western North Atlantic, where the OMZ is much deeper or absent. Larger sizes may reflect increased prey encounters within the compressed habitat, but habitat compression also makes them more vulnerable to over-exploitation. Predictably, the long-term landings of tropical pelagic tunas from areas of habitat compression have been far greater than in surrounding areas (Prince and Goodyear, 2006). For biogeochemical cycling, OMZs are the main areas of nitrogen loss, as $N_2$ and $N_2O$, to the atmosphere through denitrification and anammox, and could even indirectly mitigate the oceanic biological sequestration of $CO_2$ (Paulmier and Ruiz-Pino, 2009).

## 5 Upwelling areas

Upwelling systems associated with eastern boundary currents are among the most productive marine ecosystems in the global ocean with $973\,\mathrm{g\,C\,m^{-2}\,yr^{-1}}$ (Nixon, 1995) and, while occurring over less than 0.5% of the ocean area, are responsible for 7% of the global fishery yield (Pauly and Christensen, 1995). Much of the organic carbon produced by the upwelled nutrients is not grazed and sinks to a pycnocline where respiratory processes deplete oxygen. These oxygen deficient areas can form along continental shelf margins leading to hypoxia or anoxia on the continental shelf. Others form in offshore waters and currents bring the low oxygen waters onto the continental margin during appropriate current conditions. In either case, the continental shelf benthic communities are subject to the risk of hypoxia or anoxia. Water column shelf anoxia is well documented for the Humboldt (Escribano and Schneider, 2008) and Benguela current systems (Monteiro, 2009).

Upwelling systems are typical of eastern boundary currents, along the Pacific coast of Central and South America (Figs. 1 and 7) and along the eastern Atlantic Ocean off the southwestern coast of Africa. Upwelling systems cycle through phases of increased nutrient availability, high primary and secondary productivity and often hypoxia in the lower water column. The trophic status of upwelling systems would be considered "eutrophic" (an organic carbon supply of 300–500 $\mathrm{g\,C\,m^{-2}\,yr^{-1}}$, as defined by Nixon, 1995), but upwelling systems are not thought to be affected by anthropogenic eutrophication. Thus the hypoxia that develops would be considered the result of natural processes, not aggravated by human activity.

The upwelling system off the west Indian shelf, however, may be the first to be influenced by anthropogenic eutrophication. The west Indian shelf typically experiences moderate upwelling during June through November, but was particularly well developed in September through October in 1999 (Naqvi et al., 2000). The cold, saline upwelled water is usually capped by a 5- to 10-m thick warm, lower-salinity layer arising from high land runoff and local precipitation, contributing to the stratification and its intensification. The upwelled water is derived from the shelf break, which has a dissolved oxygen concentration of about 0.5 $\mathrm{ml\,L^{-1}}$ (22 µM) at 15° N latitude. Once impinged on the shelf, the dissolved oxygen content is quickly depleted to near anoxia by the decomposition of the high primary production (500 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$) stimulated by the upwelled nutrients. In the 1999 event, hypoxia (0.5 $\mathrm{ml\,L^{-1}}$, the most commonly used definition of an OMZ) covered an area of the shelf of about 180 000 $\mathrm{km^2}$, which is a large expression of upwelling on an eastern boundary. Naqvi et al. (2000) indicated that the severity of the 1999 event was possibly related to increasing anthropogenic nutrients to the western shelf of India.

## 6 Eastern boundary currents and changes in circulation

Changes in ocean circulation may also have a strong effect on the supply of oxygen to the bottom waters of affected shorelines. Five decades of available inner-shelf observations along the Oregon Pacific coast showed little evidence of severe inner shelf hypoxia (defined by researchers in this area as $<0.5\,\mathrm{ml\,L^{-1}}$, or $0.7\,\mathrm{mg\,L^{-1}}$, Chan et al., 2008). More recent studies documented hypoxia along the Oregon and Washington coasts beginning in 2000 related to basin-scale fluctuations in atmosphere-ocean processes that altered the oxygen content of upwelled water, the intensity of upwelling wind stress, and productivity-driven increases in coastal respiration (Grantham et al., 2004; Bakun et al., 2004). Strong persistent upwelling winds fueled intense biological production, leading to hypoxia in near-bottom waters. These changes were attributed to climate change in the jet stream caused by global warming and shifting of north winds in summer (Chan et al., 2008). Beginning in 2006 there was an intensification of severe inner shelf hypoxia associated with the California Coastal Current along the west coast of the U.S. and upwelling-favorable winds. At this point, hypoxia was evident at <50 m water depth on the inner shelf within 2 km of shore. The hypoxia expanded across broad sections of the central Oregon coast from 44.15° N to 45.00° N and extended from the shelf break to the inner shelf over a $3000\,\mathrm{km^2}$ area. Mass mortality of crabs and other megafauna was observed, along with loss of rockfish and growth of bacterial mats. The absence of prior records of anoxia or mass mortality of the fauna from the Oregon shelf suggests the fauna had no pre-adaption to hypoxic events and points to possible links with climate change. Global warming has altered ocean-atmospheric coupling with strong effects on upwelling wind stress as well as respiration that may influence oxygen saturation of source waters (Chan et al., 2008).

The shifts in circulation along the Oregon/Washington coast point to the strong coupling with atmospheric processes that shift wind-driven upwelling systems and intersect hypoxic waters with ecosystems and increase their vulnerability to modulations of upwelling from climate warming (Barth et al., 2007; McGregor et al., 2007). Also, inherent in these processes is a rise in discontinuous, unexpected, and aperiodic hypoxia in the California Coastal Current Large Marine Ecosystem and elsewhere as coastal shelf ecosystems subject to productive upwelling systems are also exposed to hypoxia (Chan et al., 2008).

In another example, the deep waters in the Lower St. Lawrence estuary are presently hypoxic, with low oxygen waters ($<2.0\,\mathrm{mg\,L^{-1}}$) covering a $1300\,\mathrm{km^2}$ area (Fig. 9) (Gilbert et al., 2005). Historical data indicate that dissolved oxygen concentrations in the 300 to 355 m depth range decreased by nearly 50% over the last 70 yr, from $4.0\,\mathrm{mg\,L^{-1}}$ in the 1930s to an average of $2.1\,\mathrm{mg\,L^{-1}}$ for 1984–2003 (Gilbert et al., 2005). Two-thirds of the $1.9\,\mathrm{mg\,L^{-1}}$ oxygen decline and a concomitant 1.7°C warming of the bottom waters are attributed to a decreasing proportion of oxygen-rich Labrador Current Water in the water mass entering the Gulf of St. Lawrence from the northwest Atlantic Ocean. The remaining one-third of the change could be due to increased oxygen consumption resulting from increased primary production and greater carbon flux to the deep water (see Coastal eutrophication and hypoxia below). The organic carbon content and the accumulation rates of dinoflagellate cysts and benthic foraminiferans have increased over the last four decades, and a shift in the stable carbon isotope signature of the organic carbon suggests enhanced accumulation of marine organic carbon (Thibodeau et al., 2006). As in the northern Gulf of Mexico, Long Island Sound, and Chesapeake Bay (reviewed by Rabalais et al., 2007b), there has been a shift in dominance towards benthic foraminifera that are tolerant to low dissolved oxygen concentrations and those that prefer high organic content of sediments.

## 7 Enclosed and semi-enclosed deep basins

There have been periods through the geologic record when hypoxia and anoxia were widespread in ocean basins (Tyson and Pearson, 1991; Wilson and Norris, 2001; Neretin, 2006). For example, an oceanic anoxic event that marked the Permian-Triassic boundary was coincident with mass extinction in both the marine and terrestrial environments, which may have been driven by the release of hydrogen sulfide from the deep waters of anoxic ocean basins (Kump et al., 2005). Today there are a number of isolated ocean basins that are anoxic due to restricted water circulation combined with high primary production in surface waters. Among them are the permanently anoxic Black Sea and Cariaco Basin and several smaller basins along the California, USA coast (Santa Monica, Santa Barbara, San Pedro, San Nicolas basins) that occasionally are reoxygenated (Berelson, 1991). Flow variation of the California Undercurrent ventilates the San Pedro and San Nicolas basins, enriching them with oxygen and allowing for recolonization of organisms until anoxic conditions return (Savrda et al., 1984).

In the case of the Black Sea, anoxic conditions occur beneath 100–200 m of oxygenated surface waters in the open sea area. The restricted connection with the Mediterranean Sea, strong stratification, and geomorphology make the Black Sea a sink for nutrients and other materials (Richards, 1965; McQuatters-Gollop et al., 2008). The persistent anoxic basin in the open Black Sea is the largest naturally occurring anoxic area on earth. It covers about 75% of the basin's area and is distinct from and not related to the eutrophication-related seasonal hypoxia that has occurred on the shallow northwestern continental shelf (Tolmazin, 1985; Zaitsev, 1992; Mee, 2001, 2005). The impact of eutrophication on the open Black Sea has been considerably less severe than that occurring in the shelf area. But, there is also evidence that the suboxic zone of the open Black Sea enlarged and expanded towards the surface by about 10 m since the 1970s (Oguz, 2005). These changes in the anoxic basin appear to be related to both climate change and nutrient loading factors (McQuatters-Gollop et al., 2008; Oguz, 2005).

The Cariaco Basin also supports a large persistent anoxic area below 250 m. Restricted circulation and high primary production within the basin (~500 g C $m^{-2}$ $yr^{-1}$) support this anoxia (Muller-Karger et al., 2001; 2004). This naturally occurring anoxic basin allows for sediments to be deposited without bioturbation, forming varves of alternating light and dark color, which correspond to the dry or rainy season (Haug et al., 2001).

Because of water column anoxia, basins like the Cariaco have a unique biogeochemistry that resembles that of anaerobic sediments (Madrid et al., 2001; Stoeck et al., 2003). Bacteria inhabit both the oxic and anoxic portions of the water column, with a maximum in the suboxic interface (Taylor et al., 2001). The suboxic layer oscillates between 200 and 300 m in the Cariaco Basin and 150 to 200 m in the Black Sea (Oguz, 2005).

## 8 Coastal eutrophication and hypoxia

### 8.1 Water column processes

Reactive nitrogen ($N_r$) has increased substantially over the last century through artificial fixation of nitrogen into fertilizers, the emission of nitrogen oxides from the consumption of fossil fuels and the transformation of reduced forms through volatilization processes. The creation of reactive nitrogen increased by a factor of 20 since 1860 to the 1995 level of anthropogenic production of $N_r$ of ~150–165 Tg N $yr^{-1}$ (Galloway and Cowling, 2002). The change was enormous, and it increased further from 156 Tg N $yr^{-1}$ in 1995 to 187 Tg N $yr^{-1}$ in 2005 (Galloway et al., 2008). Phosphorus additions to the landscape enter via phosphorus-containing fertilizers manufactured from mined phosphorus, animal manures and waste products from animals supplemented with phosphorus-enriched feed, and enter rivers and streams via wastewater effluents and soil erosion. Phosphorus is accumulating in the soil with important implications for increased runoff from the landscape to surface waters (Bennett et al., 2001). Increased flux of phosphorus eroded from the landscape or carried in wastewater effluents to the world's rivers has increased the global flux of phosphorus to the oceans almost 3-fold above historic levels of ~8 Tg P $yr^{-1}$ to end of the 20th century loadings of ~22 Tg P $yr^{-1}$ (Bennett et al., 2001). Accumulation in landscapes of developed countries is declining somewhat, but that of developing countries is increasing (Bennett et al., 2001). Compared to increased runoff of nitrogen and phosphorus, river concentration or loads or both of dissolved silicon have remained the same or decreased, so that the relative proportions of silicon to nitrogen and silicon to phosphorus in river effluents have decreased over time (Justić et al., 1995a,b; Turner et al., 2003). The result is increased primary production where nitrogen or phosphorus or both were limiting, and perhaps a limitation of diatom growth where the relative proportion of silicate to nitrogen has approached the Redfield ratio of 1:1 (Officer and Ryther, 1981; Turner et al., 1998). The overall result is increased primary production (e.g., Turner and Rabalais, 1994; Lohrenz et al., 1997; Cloern, 2001; Rabalais, 2004), flux of a high proportion of the organic carbon, and subsequent bottom-water hypoxia (where the physical structure necessary is present) (Fig. 10).

Marine sources provide more nutrients than from the land in some coastal ecosystems, such as the East China Sea (Chen and Wang, 1999), by various physical processes, such as coastal upwelling and slope water intrusion (Liu et al., 2009a). The estimated marine dissolved inorganic nitrogen and dissolved inorganic phosphorus fluxes to continental margins globally are larger than the total riverine loads by a factor of six to nine for nitrogen and by two to three for phosphorus (Walsh, 1991; Liu et al., 2009b), but the oceanic supply is probably more uniform over the long term. Like the riverine nutrient loads, the oceanic nutrient supply is subject to alteration due to changes in circulation from climate change (e.g., Vecchi and Soden, 2007). Depending on the coastal environment, the residence time of river-supplied fresh water, and rates of nutrient uptake and regeneration, the changes observed in the Anthropocene (Meybeck, 2003) from increased nutrient loads are most likely to lead to human-caused eutrophication and associated hypoxia. This does not preclude the potential for climate driven changes in oceanic nutrient sources from exacerbating an already precarious situation, as might be the case for the Oregon coast.

Oxygen is consumed in both the water column and the sediments. The relative contributions of water column and benthic respiration in the hypoxic part of the northern Gulf of Mexico have been estimated with oxygen concentration and stable isotope measurements (Quiñones-Rivera et al., 2007). Their model indicated that the severe bottom water oxygen depletion in mid-summer 2001 was due predominantly to benthic respiration (73%). High contributions of benthic respiration (52 to 77%) in association with bottom-water hypoxia have been observed in eutrophic Lake Erie during summer stratification (Ostrom et al., 2005). Subsequent work by Quiñones-Rivera et al. (2009) indicated that the relative proportion of water column and benthic respiration varied with year and season. Seasonal oxygen dynamics in surface waters were explained in a regression model by the concentration of algal biomass (as indicated by particulate organic carbon and C/N ratio), while physical factors, such as salinity, temperature, water column stability and station depth were only of minor importance. Seasonal oxygen dynamics were more pronounced in bottom waters than in surface waters. Despite considerable overlap in oxygen saturation values, the relative contributions of benthic respiration in bottom waters during summer season (75%) were noticeably higher and more variable compared to the winter season (42%) (Quiñones-Rivera et al., 2009). Similarly in Lake Erie, benthic respiration became more important as summer progressed (Wang et al., 2008), increasing from 33% in early summer to 53% in late summer. The physical mixing of the water column caused by Hurricane Claudette prior to mid-summer mapping of the 2003 Gulf of Mexico hypoxic zone diminished the relative importance of benthic respiration, so that the system resembled typical wintertime conditions (Quiñones-Rivera et al., 2009).

## 8.2 Sedimentary processes

The excess organic material in the form of senescent phytoplankton, fecal pellets and marine aggregates sinks to the lower water column and the seabed where the carbon is remineralized by aerobic and anaerobic processes or buried. As aerobic bacteria decompose the increased organic matter settling onto the seabed and in the lower water column, the dissolved oxygen concentration overlying the sediments becomes hypoxic and approaches anoxia. In this transition, numerous biological and geochemical shifts occur in the benthic community, many with negative feedback into the cycle of eutrophication and declining oxygen levels (see Middleburg and Levin, 2009).

With continued accumulation of organic carbon at the seabed, microbial biomass and microbial decomposition potential of substrates and community oxygen consumption increase, but not in simple linear relationships (Meyer-Reil and Köster, 2000). The redox potential discontinuity layer migrates upward to the sediment-water interface, sulfate respiration replaces oxygen respiration, hydrogen sulfide is generated from the sediments and oxygen penetrates less deeply into the sediments as the bioturbation potential of the macroinfauna decreases during their demise due to sulfide toxicity or lack of sufficient oxygen. The sediments become less cohesive, more susceptible to resuspension and contribute to turbidity of the overlying water, which in turn reduces the potential for growth of the photosynthetic microphytobenthic community and generation of oxygen into the lower water column. Some shifts in the benthic microbial community are visible at the sediment-water interface (Jørgensen, 1980; Harper et al., 1981; Rosenberg and Díaz, 1993; Rumohr et al., 1996). Typical black spots from iron sulfide precipitated from intense microbial degradation of organic matter, lacey white colonies and denser, yellowish colonies of sulfur-oxidizing bacteria (*Beggiatoa* and *Thiovulum*), and reddish to violet carpets of sulfur-purple bacteria can be observed as oxygen levels decline in the Wadden Sea (North Sea), in shallow water areas of the Baltic, on the Louisiana continental shelf, and many areas of the world's ocean where the oxygen minimum zone intersects the seabed (reviewed in Rabalais et al., 2001). When the bottom water is depleted of oxygen, or anoxic, hydrogen sulfide builds up in the bottom waters as anaerobic bacteria metabolism reduces sulfate to $H_2S$ (Jørgensen, 1980), the sediment becomes almost uniformly black, and there are no signs of aerobic life. Hydrogen sulfide is toxic to most metazoans and contributes to the overall benthic infaunal mortality.

The nitrification/denitrification cycle of estuarine and continental shelf sediments, which returns $N_2$ to the atmosphere is an ameliorating mechanism to excess $N_r$, but is disrupted by the limited availability of oxygen in sediments. Denitrification proceeds much of the year but is dependent on the nitrate supplied by nitrification, a process that is dependent on the presence of oxygen, which for extended periods and over broad areas of the seabed may be absent (Kemp et al., 1990; Childs et al., 2002). With the shift in redox potential in the sediments with decreasing oxygen concentration, there is an increase in the flux of inorganic nutrients, ammonium and particularly phosphate, into the overlying water (Fig. 11). These inorganic nutrients become available to fuel further phytoplankton production in the overlying water. The degree to which these nutrients diffuse upward through the water column and across strong pycnoclines is not known. High waves or strong currents will likely break down the physical structure of the water column, re-aerate the water column and resuspend sediments and adsorbed and dissolved nutrients into the water column with a subsequent stimulatory effect for primary production (Dagg, 1988). These events, however, are short-lived and may occur during seasons when restratification is unlikely to occur and subsequent accumulation of phytoplankton biomass may be limited, i.e., a short-term stimulation of phytoplankton response but not a long-term accumulation of carbon that would lead to further negative feedbacks on the system.

### 8.3 The Baltic Sea and Scandinavian waters

The Baltic Sea contains the largest anthropogenic hypoxic zone in the world (Díaz and Rosenberg, 2008). The combined hypoxic areas of the Baltic Sea (defined as 2 ml $L^{-1}$ or 2.8 mg $L^{-1}$) averaged 42 000 $km^2$ but expanded to about 60 000 $km^2$ over the past few years (Conley et al., 2009a). Hypoxic areas rim the Baltic Sea where increased nutrient loads have led to eutrophication and decreased dissolved oxygen concentrations. The Baltic Sea, a brackish ecosystem, is isolated from the North Sea by a series of shallow sills in the Danish Straits, and inflows of higher salinity oxygenated waters are limited but do improve the oxygen conditions (Gerlach, 1994). In the eastern and northern parts of the Baltic Sea the surface salinity is low, but increases towards the Danish Straits. A strong halocline separates the brackish surface water from the saltier bottom water.

Few major oceanic inflows have taken place since the mid-1970s (HELCOM, 2001), resulting in areas of hypoxia in the lower water column between 12 000 km$^2$ and 70 000 km$^2$, or 5 to 27% of the total bottom area (Fig. 12) (Conley et al., 2002a). Although hypoxia has occurred in the Baltic Sea during previous climatic warm periods (Zillén et al., 2008), nutrient-driven eutrophication is believed to be the primary cause of increases in hypoxia over the last 50 to 100 yr (Conley et al., 2009b). Nitrogen and phosphorus concentrations in the bottom waters of the Baltic Sea tripled from 1960 to 1990 (Fig. 13). The hypoxia and anoxia in the deep basins of the Baltic Sea enhance phosphorus release from the sediments and increase nitrogen loss (Conley et al., 2002a). Coastal areas within the Baltic that are hypoxic have increased over the last several decades (Díaz and Rosenberg, 2008).

The amount of dissolved inorganic phosphate released from sediments during hypoxia is approximately one order of magnitude greater than the anthropogenic total phosphorus loading (Conley et al., 2002a), and increases with decreasing dissolved oxygen concentrations. Removal of nitrogen through denitrification increases in the Baltic Sea with more hypoxia. The high phosphorus and low nitrogen concentrations are favorable for blooms of $N_2$-fixing cyanobacteria, thus increasing the available nutrients leading to more eutrophication (Vahtera, 2007) and more hypoxia.

Danish marine waters display all the classic symptoms associated with eutrophication, including periods of hypoxia and anoxia in bottom waters (Conley and Josefson, 2001; Conley et al., 2002a,b) and the mortality of benthic-dwelling organisms during anoxia (Fallesen et al., 2000; Conley et al., 2002b). Hypoxia is widespread in areas of the Skagerrak and Kattegat, the coastal waters of the North Sea, and many coastal embayments and fjords. Nutrient loading from Denmark ranks among the highest in Europe per unit area (Conley et al., 2002b) and reflects the density of the population and the intensity of the agriculture and animal production.

## 8.4 Mississippi River-influenced Gulf of Mexico

The Mississippi River watershed at $3.27 \times 10^6$ km$^2$, encompasses 41% of the lower 48 United States, and includes 13 states and numerous regulatory authorities, both state and federal. The Mississippi River system ranks among the world's top ten rivers in watershed size (third), freshwater discharge (seventh) and sediment delivery (seventh) (Milliman and Meade, 1983). Thus, the dimensions of the problem and the drainage system that affect it reflect a much greater magnitude than most nutrient-driven eutrophication problems elsewhere. The hypoxic zone on the continental shelf of the northern Gulf of Mexico is the second largest human-caused hypoxic zone in the coastal ocean (Rabalais et al., 2007a). The Mississippi River creates a strongly stratified coastal system west of the delta for much of the year and delivers two to three times more nutrients now than in the 1950s. The area of bottom covered by hypoxic water can reach 22 000 km$^2$ and averaged 13 500 km$^2$ between 1985 and 2009 (updated from Rabalais et al., 2007a). The hypoxia occurs from February through November and is nearly continuous from mid-May through mid-September (Fig. 14 depicts widespread hypoxia in July 2008).

In mid-summer, the size of the hypoxic zone is most closely related to the nitrate load of the Mississippi River in the two months prior to the mapping (Fig. 15; Turner et al., 2006), and the same size hypoxic area is now formed with a lower nitrate load than historically (Turner et al., 2008; Greene et al., 2009). Changes in the nitrate loads over time are due mostly to the change in nitrate concentration in the Mississippi River (80%) and the remainder is due to increased freshwater discharge (20%) (Donner et al., 2002; Justić et al., 2002). More precisely, the U.S. Geological Survey has identified specific activities for sources of nitrogen and phosphorus yields from the Mississippi River watershed and locations (Fig. 16) (Alexander et al., 2008). Most of the excess nutrients from the basin come from extensive corn and soybean row crops in tile-drained fields, which are the recommended sources for reductions to mitigate Gulf of Mexico hypoxia (Fig. 16) (Alexander et al., 2008; US Environmental Protection

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Evidence from paleoindicators (reviewed by Rabalais et al., 2007b) and models that relate the size and frequency of the hypoxic zone to nitrate-N load of the Mississippi River (Justić et al., 2002; Scavia et al., 2003; Turner et al., 2006) converge on the period of the early 1970s as to when hypoxia as a large scale phenomenon began. Community changes in hypoxia-intolerant benthic foraminiferans began early in the 1900s with landscape alterations (Rabalais et al., 1996, 2007b), but most shifts in eutrophication/hypoxia foramiferan indicators began in the mid-1950s consistent with increases in sediment total organic carbon, biogenic silica remains of diatoms, nitrate-N loads from the Mississippi River, and fertilizer use in the Mississippi River watershed (Sen Gupta et al., 1996).

## 8.5 East China Sea

The Changjiang (Yangtze River) and East China Sea provide another example of the complex problems that arise with major changes in the watershed and result in dramatic impacts on the development of eutrophication in surface waters and hypoxia in near-bottom waters of the adjacent shelf region. The historical water discharge of the Changjiang is $928.5\times10^9\,\mathrm{m}^3\,\mathrm{yr}^{-1}$ and accounts for 90–95% (ca. $1.0\times10^{12}\,\mathrm{m}^3\,\mathrm{yr}^{-1}$) of the freshwater input to the East China Sea. Other large rivers contributing fresh water are the Qiantangjiang ($40.4\times10^9\,\mathrm{m}^3\,\mathrm{yr}^{-1}$), Minjiang ($53.6\times10^9\,\mathrm{m}^3\,\mathrm{yr}^{-1}$), and Jiulongjiang ($13.9\times10^9\,\mathrm{m}^3\,\mathrm{yr}^{-1}$). Through human intervention over the watersheds of the Changjiang (e.g., hydraulic engineering and change in land use), there has been a step-wise reduction of sediment input to the East China Sea (Qu et al., 2005; Yang et al., 2006). However, an increase in human-sourced nitrogen and phosphorus influx from the Changjiang since the 1960s has been reported (Zhang, 1994; Liu et al., 2003). For instance, the Changjiang discharged approximately $30\,\mathrm{Gmol\,N\,yr}^{-1}$ to the East China Sea in the 1960s and more recently $72.2\,\mathrm{Gmol\,N\,yr}^{-1}$ at beginning of the 21st century, or a 260% increase (Table 1). Over the same period, the N/P ratio increased from 30–40 to ca. 100. This increase in nitrogen flux sustains the high productivity of the shelf, as is the case of most river effluents.

In the region of 20–75 m water depth off the Changjiang, near-bottom hypoxia has been observed with dissolved oxygen concentrations as low as 1–2 mg L$^{-1}$ (Tong and Zhang, 2007). The hypoxia on the inner shelf is usually found in summer and sometimes in early autumn (i.e., July–September). The hypoxic area increased from ca. 1800 km$^2$ in August, 1959 to >15 400 km$^2$ in August, 2006 (Chen et al., 2007; Zhu, 2007). The temporal distribution and configuration of hypoxia on the shelf is driven by the dynamic circulation in the East China Sea. The core area, where hypoxia is most concentrated, changed from year-to-year, but generally was observed in 40- to 50-m water depths from 29° N to 33° N. Sometimes, two hypoxic core areas were identified.

### 8.6 Northern Adriatic Sea

Historical data from the northern Adriatic Sea, going back to 1911, show increasing oxygen concentrations in the surface water and decreasing oxygen concentrations in the bottom water as well as decreasing water clarity associated with an increase in phytoplankton biomass. These trends are consistent with increasing nutrient loads in the Po River and development of hypoxia in the northern Adriatic Sea (Fig. 17) (Justić et al., 1987; Justić, 1988, 1991b). The loss of taxa of pelagic medusae, which depend on benthic stages as part of their life cycle, is also consistent with the expanding hypoxia (Benović et al., 1987). Paleoindicators in sediment cores from the vicinity of the Po River delta plume indicate a gradual increase in eutrophication at the end of the 19th century, which accelerated after 1930. Seasonal hypoxia began in 1960 and became more intense and prolonged in 1980 (Barmawidjaja et al., 1995).

### 8.7 Chesapeake Bay

Eutrophication and, especially, hypoxia and anoxia are well-documented and recognized as an environmental problem in the Chesapeake Bay, USA (Kemp et al., 2005). Chesapeake Bay is the largest estuary in the United States and one of the largest in the world with tidal waters extending over 11 400 $\text{km}^2$. The Chesapeake drainage basin covers $1.7\times10^5$ $\text{km}^2$ and six United States. The Susquehanna River, the largest tributary to the bay, drains 71 200 $\text{km}^2$, accounting for 41% of the Chesapeake Bay watershed and 80% of the area draining directly into the bay. The bay is relatively shallow (mean depth 6.5 m), and the area of its catchment is large in comparison to the estuarine volume. This, coupled with its modest tidal exchange, makes the bay susceptible to inputs of fresh water, sediments and dissolved materials from the land (Horton, 2003; Hagy et al., 2004).

The Chesapeake Bay ecosystem has undergone substantial human-induced changes since colonization by Europeans almost 400 yr ago, including increased sedimentation resulting from clearing of its previously forested watershed (Brush, 1994, 2009; Cooper and Brush, 1991). During the period of agricultural expansion extending into the early 19th century, more plant nutrients – forms of nitrogen and phosphorus that the native forests efficiently retained – also began to wash down into the Bay. Although there were numerous and cumulative, human-induced changes in Chesapeake Bay through the early 20th century, during the last half of the 20th century even more dramatic changes occurred from a growing population, runoff of agricultural fertilizers and animal wastes, and atmospheric deposition of nitrogen oxides from fossil fuel combustion and volatilization of ammonium from fertilizers and animal wastes. By the mid-1980s Chesapeake Bay was receiving about seven times more nitrogen and 16 times more phosphorus than when English colonists arrived (Boynton et al., 1995). The result of these increased nutrient loads and multiple stressors, such as overharvesting of fisheries resources, was high concentrations of phytoplankton biomass, reduced water clarity, loss of submerged aquatic vegetation, loss of shellfish and fisheries production, and an increase in stressful low-oxygen conditions during the summer (Fig. 18) (Hagy et al., 2004; Kemp et al., 2005). The extent of Chesapeake Bay hypoxia, defined as $<2\,\text{mg}\,\text{L}^{-1}$ dissolved oxygen is most often expressed as volume instead of areal extent (Figs. 18 and 19).

Hagy et al. (2004) examined a 52-yr record of dissolved oxygen in Chesapeake Bay

(1950–2001) and a record of nitrate-N loading from the Susquehanna River spanning a longer period (1903, 1945–2001) to describe the long-term pattern of hypoxia and anoxia in Chesapeake Bay and its relationship to nitrate-N loading. Year-to-year variability in river flow accounted for some of the observed changes in hypoxic volume, but the long-term increase was not due to increased river flow. From 1950–1985, the volume of hypoxic water in mid-summer averaged about 4.5 $\text{km}^3$. After 1985, the hypoxic volume increased substantially and averaged over 8 $\text{km}^3$ from 1986 to 2007 (Fig. 19). Nitrate-N concentrations in the Susquehanna River at Harrisburg, Pennsylvania, increased up to 3-fold from 1945 to a 1989 maximum and declined through 2001. On a decadal average basis, the combination of changes in river flow on the long-term and increase in nitrate-N resulted in a 2-fold increase in nitrate-N loading from the Susquehanna River during the 1960s to 1970s. Decadal average loads were subsequently stable through the 1990s. Hypoxia was positively correlated with nitrate-N, but more extensive hypoxia was observed in recent years than would be expected from the observed relationship. The results (Hagy et al., 2004) suggested that the bay may have become more susceptible to nitrate-N loading similar to the situation observed for the Mississippi River-influenced northern Gulf of Mexico hypoxic zone (Turner et al., 2008; Greene et al., 2009; Kemp et al., 2009).

Although heavy fishing pressure and other stressors in Chesapeake Bay make it difficult to isolate the effects of eutrophication-driven hypoxia on fish (Breitburg et al., 2009), an increased ratio of pelagic to demersal fish species documented for the Chesapeake Bay (Kemp et al., 2005) is indicative of increased eutrophication. The shift from demersal-dominated to pelagic-dominated fisheries has been observed for other coastal systems and is attributed, in part, to bottom-water hypoxia (Caddy, 1993). It is not clear how hypoxia influences the habitat requirements of particular fish and invertebrates, but some species, such as sturgeons, cannot reproduce in Chesapeake Bay due to a lack of habitat with adequate oxygen and temperature levels (Secor and Gunderson, 1998). Habitat for Chesapeake Bay striped bass, which prefers cooler, deep waters, is also severely curtailed during summer months due to a combination of bottom water hypoxia and temperature (Coutant and Benson, 1990). During summer, striped bass adults selected narrow lenses within the water column characterized by <25°C water and normoxia (Cheek et al., 1985). Using bioenergetic models Brandt and Kirsch (1993) mapped habitat suitability for striped bass in Chesapeake Bay and predicted substantial habitat restrictions in summer, leading to negative or static growth during summer months (Secor and Niklitschek, 2001).

Whereas impacts to fisheries have been difficult to determine, the impacts of hypoxia on benthic animals have been well documented (Holland et al., 1987; Dauer et al, 1992; Rabalais et al., 2001b). Loss of habitat and prey biomass as a result of hypoxia can have profound effects on ecosystem energetics and services (Díaz and Rosenberg, 2008). Estimates of biomass lost due to hypoxia are approximately 10 000 megatons of carbon per year, or 5% of the Chesapeake Bay's total secondary production. Under normoxic conditions, an estimated 60% of benthic energy would be passed up the food chain, thus the seasonal hypoxic conditions in the bay possibly result in a loss of 6000 megatons of carbon (Díaz and Rosenberg, 2008).

### 8.8 Other coastal and oceanic areas

There are many other well-documented expressions of eutrophication over a long period leading to coastal hypoxia and anoxia in estuarine and coastal ecosystems (Rosenberg, 1990; Cloern, 2001). The ability to document changes over time is complicated by lack of suitable data sets over sufficient time periods (Gilbert et al., 2009). Yet, Cloern (2001) provided several examples of decreasing low oxygen over the period of 1950–1990 in several areas of the northeastern Atlantic (Fig. 20). A similar long record of decreasing dissolved oxygen concentrations is available from the northern Gulf of Mexico (Fig. 20). There is also considerable evidence of decreasing levels in dissolved oxygen concentrations for many areas of the coastal ocean and the open ocean (Gilbert et al., 2009).

## 9 Reversal, or not, of human-caused hypoxia

Most often increases in nutrient loads to estuarine and coastal waters are the cause of eutrophication-associated hypoxia. In some cases the increase in carbon loading and accumulation result from allochthonous sources, such as organic waste from pulp mill factories (such as Idefjord, Sweden – Rosenberg, 1980 or Perdido Bay, Florida, USA – Flemer et al., 1999), fish aquaculture (such as Loch Ailort, Scotland – Gillibrand et al., 1996), or seafood processing plants (such as Saldanha Bay, South Africa – Christie and Moldan, 1977). It seems simple enough to remove the excess carbon inputs or to reduce nutrient loads to reverse the associated hypoxia. The process of eutrophication, however, has occurred over long time periods during which basic characteristics of the aquatic ecosystem change, such as composition of the phytoplankton community, sequential demise of components of the benthic community, shifts in sedimentary structure and composition, and biogeochemical processes.

Thus, it is not surprising that reductions in nutrients or allochthonous carbon do not immediately, if ever, return an aquatic ecosystem to its "prior" state and, seldom, a pristine state (for example pre-agricultural condition of the Mississippi River watershed). Part of this unsuccessful restoration is that eutrophication and subsequent development of hypoxia do not occur in a vacuum. Other components of the ecosystem have changed along with the development of hypoxia. Overfishing, habitat destruction, invasive species, aspects of climate change, and continued release of stored nutrients from the sediments of the coastal ecosystem (Conley et al., 2002a) and continued input of excess nutrients from land runoff (Bennett et al., 2001) prevent a similar reversed trajectory as the one leading to hypoxia. Often nutrient reductions focus on a single nutrient, e.g., nitrogen through improved agricultural practices or tertiary removal of nitrate-N or phosphorus through improved urban waste water treatment. This results in an imbalanced ratio of the nitrogen and phosphorus to each other, which can affect phytoplankton community composition.

In addition, reductions are often required in several nutrients, e.g., both phosphorus and nitrogen. While some reduction mechanisms, such as reduced fertilizer use, would apply to both phosphorus and nitrogen loadings, the mechanisms required to reduce nitrogen, primarily as dissolved nitrate-N in runoff or groundwater or dissolved orthophosphate in runoff differ from that of phosphorus control through prevention of soil erosion. The relative proportion of nitrogen to phosphorus and nitrogen and phosphorus to silicate are important in defining phytoplankton community composition, trophic structure, and determination of which phytoplankton species are likely to dominate and/or contribute to carbon flux to the lower water column and seabed.

There have been improvements in oxygen depletion conditions in many areas around the world, and other examples of diminished symptoms of eutrophication through reductions in nutrient loadings (Fig. 21).

## 9.1 Northwestern shelf of the Black Sea

A socio-ecological experiment that tested if increased nutrients leading to hypoxia and worsening hypoxia could be reversed by a reduction in nutrients that alleviates hypoxia occurred on the northwestern Black Sea shelf. The Black Sea shelf adjacent to the Danube River experienced hypoxia historically, but anoxic events became more frequent and widespread in the 1970s and 1980s (Tolmazin, 1985; Zaitsev, 1992; Mee, 2001; Mee et al., 2005), reaching over areas of the seafloor up to 40 000 km$^2$ in depths of 8 to 40 m (Fig. 22). The deep central part of the Black Sea is a permanently anoxic basin, but the northwestern margin that receives the freshwater inflows of the Danube River and the smaller Dneiper and Dneister rivers to the north experienced a surge of increased nitrogen and phosphorus loads in the 1960s to 1970s, similar to many other coastal areas downstream of developed nations (Galloway and Cowling, 2002; Rabalais, 2004a). Hypoxia developed along with the nutrient load and covered areas of the seafloor up to 40 000 km$^2$, in depths of 8 to 40 m (Tolmazin, 1985; Zaitsev, 1992; Mee, 2001; Mee et al., 2005) (Fig. 24). Beginning in the 1960s under the communist system the republics of the Soviet Union in the Danube River watershed, land was intensively farmed with large-scale production of livestock and heavy applications of nitrogen and phosphorus fertilizers, industrialization was on the rise, and human population – the source of waste water – was on the rise. The hypoxia was first recorded in 1973 and was present in increasing size through 1990 (Fig. 23).

The conditions on the northwestern shelf of the Black Sea improved from 1990 to 2000 when nutrient loads from the Danube River decreased substantially because of economic decline following the dissolution of the Soviet Union. Within 5 yr following the end of intensive farming, decreasing nutrient loads were followed by a decrease in the extent of bottom-water hypoxia (Mee, 2001) (Fig. 23) from 1991 on until it was essentially absent barring a relapse in the hot summer of 2001 (Mee, 2006). Not all parts of the ecosystem recovered so rapidly with mussel communities re-establishing by 2002, but planktonic communities not at 1960s composition. As the central and eastern economies recover, especially with membership in the European Union, cultural eutrophication may again intensify with the recurrence of hypoxia. Being a member of the European Community, however, will require conformance to water quality standards and regulations. In the Black Sea, surrounding governments, aided by the United Nations Global Environment Facility, have agreed to pursue an initiative to maintain nutrient levels of the mid-1990s; the process appears to be aiding recovery through pilot-scale projects to improve farming practices and waste water treatment (Mee, 2006).

### 9.2 Tampa Bay

The Tampa Bay estuary (Florida, USA) is often heralded as an example as a successful nutrient abatement program with restoration of seagrass beds and alleviation of hypoxia (National Research Council, 2000; Duarte et al., 2009; Waycott et al., 2009). In the 1960s Tampa Bay was experiencing reduced water clarity, loss of seagrass beds, reduction in landings of speckled sea trout and areas of hypoxia (Simon, 1974; Johansson and Lewis, 1992). The Tampa Bay National Estuary Program developed an environmental goal of restoring water clarity and therefore seagrass beds in the Tampa Bay estuary (TBEP, 2006). Phytoplankton production and biomass in Tampa Bay are strongly nitrogen limited, due in part to drainage from the phosphate-rich Bone Valley deposits located within the watershed (Johansson, 2005). As is often the case, control of one nutrient is often more feasible and cost effective than control of another or both.

Improvements in wastewater treatment facilities starting in 1980 resulted in a total nitrogen reduction from this source of 90%. Coupled with improved stormwater treatment and reductions in atmospheric deposition from local electric utilities, annual external nitrogen loads to Tampa Bay since 2000 are estimated to be less than one-half that estimated for the 1970s (Greening and Janicki, 2006). Average annual chlorophyll *a* concentrations and water quality targets have been met 85% of the time since 1995 (Sherwood, 2009). With increased water clarity, seagrass extent has expanded from 21 650 acres in 1982 to 29 650 acres in 2008 (SWFWMD, personal communication). Hypoxia still occurs in the deeper areas and dredged channels in summer months, but the spatial and temporal extent has been much reduced since 1980. All seems a good success story. With the increase in the population of the Tampa Bay area (the second largest metropolitan area in Florida – TBEP, 2006), the Tampa Bay estuary is now experiencing additional loading from changes in land uses, from natural or low-intensity agriculture to residential. This nonpoint nutrient source is more difficult to control than point sources from wastewater and nitrogen point source controls.

### 9.3 Chesapeake Bay

In the 1970s the scientific community in the Chesapeake Bay area began to understand and document the pervasive changes in the ecosystem that had taken place and to identify their causes. This led to growing awareness by the public and political leaders, which in turn resulted in the evolution of regional management structures and restoration objectives (Hennessey, 1994; Boesch et al., 2001). A simple agreement among three states, the national capital district and the federal government in 1983 "to assess and oversee the implementation of coordinated plans to improve and protect the water quality and living resources of the Chesapeake Bay estuarine systems," led to the Chesapeake Bay Program. A series of directives and agreements proposed reductions of nutrient and toxicant loads, habitat restoration, living resource management and landscape management. Reduction of hypoxia and anoxia was seen as the corner stone of the restoration effort.

A landmark agreement of the Chesapeake Bay Program was the 1987 commitment to reduce controllable inputs of nitrogen and phosphorus entering the bay by 40% by the year 2000. As that year approached it was clear that this goal would not be reached and, furthermore, a complex array of interrelated issues of environmental quality, living resources, and human activities needed to be addressed in a more comprehensive manner. The Chesapeake 2000 Agreement was reached, which included over one hundred goals and commitments that together comprise one of the most ambitious ecosystem management programs for a large coastal area.

Progress in restoring the Chesapeake Bay ecosystem has been mixed (Boesch, 2006). Although eutrophication is no longer growing, there is a public debate concerning the amount of nutrient load reductions that have been achieved, and few clear signs that the symptoms have been alleviated, except locally. The concentrations of a number of potentially toxic substances in sediments and organisms declined. Seagrasses have returned in some regions but cover only a small portion of the habitat occupied in the 1950s. Oyster populations have not recovered. Populations of several anadromous fishes have increased modestly. Populations of striped bass have increased. On the other hand, the productive blue crab fishery has shown some decline in recruitment and signs of over-fishing. Small improvements in dissolved oxygen conditions are expected for the mainstem of the Chesapeake Bay, and the outlook for rivers draining into the Bay ranges from slight deterioration to slight improvement (Bricker et al., 2007).

### 9.4 Baltic Sea

Since the late 1980s, the Convention on the Protection of the Marine Environment of the Baltic Sea (Helsinki Commission – HELCOM) has worked to implement a 50% reduction target for anthropogenic nutrient loads. In 2007, HELCOM reached an international agreement to reduce the effects of eutrophication, with targeted nutrient reductions for each country (HELCOM, 2007). By reducing specific point sources, such as industrial and municipal wastewater, there has been some local improvement in eutrophication and associated hypoxia. However, the reductions needed are far from reaching the targeted goals. The immense size and volume of the Baltic Sea hypoxia and the negative feedbacks to the nutrient cycles from the anoxic bottom layer will hamper remediation of hypoxia unless there are substantial, long-term efforts to reduce nutrient loads (Conley et al., 2009a).

Because of the immensity of the hypoxia problem in the Baltic Sea, the Swedish Environmental Protection Agency, other national agencies, and engineering concerns are examining ways to oxygenate the bottom layers or to increase the precipitation and sequestration of phosphorus, resulting in less release from the sediments (Conley et al., 2009a). Large-scale engineering schemes include aeration by bubbling oxygen into the bottom hypoxic layer and manipulation of currents so that there is increased flushing of the water into and out of the Baltic Sea and thus reduced residence time. The volume needed and the technology involved are seemingly insurmountable obstacles to bringing in higher concentration dissolved oxygen. And, while there is a general understanding of nutrient biogeochemical cycles in the Baltic Sea and the effects of those on hypoxia (Conley et al., 2009b), the effects of remediation efforts on phosphorus, nitrogen and silica biogeochemical cycles raise concerns and require additional knowledge (Conley et al., 2009a). Chemical precipitation has been suggested as a method to remove phosphorus from the water column and enhance permanent phosphorus burial, but little is known about the effects of brackish water usage compared to freshwater lakes and the effects of precipitation of aluminum and silica with regard to toxicity and trophic interactions, respectively. Biomanipulations should be approached cautiously because of unintended, negative results on ecosystem health and function.

Any solution to hypoxia in the Baltic Sea must include large reductions in the inputs of nutrients and consideration of balanced nutrient concentrations. Engineered, or otherwise manipulated, solutions may be beneficial on more local scales, but probably unfeasible in the Baltic Proper (Conley et al., 2009a).

### 9.5 Mississippi River and northern Gulf of Mexico

Social and political acceptance of the linkages between increased loads of Mississippi River nutrients and the expansion and worsening of hypoxia on the northern Gulf of Mexico continental shelf lags behind that of the Chesapeake Bay, Tampa Bay and the Baltic Sea, and there is to date no identified target nutrient reduction embodied in an action plan. The evidence was easily ignored or "denied" at first, and continues, because of the distance from the area of application, half a continent away, and a natural tendency for agri-business to protect their interests in this economically important region. Many studies and subsequent peer-reviewed publications continue to enhance the scientific understanding of changes in the watershed, landscape use, nutrient inputs, and linkages of increased nutrient loads and relative ratios on important biological processes of the Mississippi River-influenced shelf, and increasing hypoxia.

The U.S. Geological Survey has identified specific activities for sources of nitrogen and phosphorus yields from the Mississippi River watershed and locations (Fig. 16; Alexander et al., 2008). Most of the excess nutrients from the watershed come from extensive corn and soybean row crops in tile-drained fields in the middle of the U.S., an area commonly known as the "Corn Belt." These sources were recommended for specific targeting and reduction for an overall decrease in both nitrogen and phosphorus loads by 45% of current levels to mitigate Gulf of Mexico hypoxia (U.S. Environmental Protection Agency, Science Advisory Board, 2009). A 2001 Action Plan (Mississippi River/Gulf of Mexico Watershed Nutrient Task Force, 2001), based on a thorough review of the available literature and data, was delivered to Congress and President Clinton in early January 2001. This plan proposed an environmental goal of reaching an areal size of hypoxia of 5000 km$^2$, over a five-year running average by 2015, with implementation through voluntary actions and incentives. The Plan floundered between 2001 and 2009 with no input of necessary funds to implement nutrient reductions or to monitor any responses. A subsequent review of old and new data and information was completed by the U.S. Environmental Protection Agency, Science Advisory Board

(2008) that reinforced the science and the need for nutrient reductions of both nitrogen and phosphorus. The 2008 Gulf Hypoxia Action Plan (Mississippi River/Gulf of Mexico Watershed Nutrient Task Force, 2008), completed in parallel to the science assessment, is limping along without much support, as yet, from most federal agencies and the signatory states, with the exception of NOAA and more progressive states with regard to nutrient mitigation, such as Minnesota in the head waters.

The Mississippi River is seen as an "orphan" with regard to a united, coordinated effort to reduce nutrients (NRC, 2008). There is a lack of coordinated and consistent monitoring of nutrients and differing state approaches to assessing waters for the Clean Water Act Section 303(d) designations for meeting water quality standards. The lack of a suitable framework for making decisions, a less than adaptive 2008 Gulf Action Plan, a negative governmental response to the issue from 2001–2008, and lack of adequate funds for mitigation and monitoring stymie progress.

### 9.6 Approaches and implications of nutrient reduction

Nutrient control efforts are usually more successful in reducing point sources of nitrogen and phosphorus than with the multiple nonpoint sources of high solubility and growing atmospheric inputs of nitrogen. The success with point sources and the lack of success with nonpoint sources, may lead to critical imbalances of nitrogen to phosphorus, or silica to both, so that there is no decrease in phytoplankton biomass, but instead a shift to another species composition, potentially more noxious or toxic. On the other hand, a reduction in silica compared to nitrogen could result in silica-limited phytoplankton communities that may be less desirable to zooplankton grazers, affect the flux of organic matter to the seabed, or affect trophic interactions (Officer and Ryther, 1980; Turner, 2001; Turner et al., 1998). Closer attention should be paid to the importance of balanced nutrient composition as well as nutrient supply dynamics for the development of eutrophication versus efficient trophic transfer and fish production in nutrient-enriched systems (Snelgrove et al., 2009). In the process of nutrient management by reducing loads, consideration should also be given to all ecosystem services that are likely to be influenced by intervention (e.g., food, oxygen production through photosynthesis, primary production, nutrient cycling and habitat provisioning (Snelgrove et al., 2009).

The fear of reducing fishery yields adjacent to the Mississippi River delta is sometimes raised as a reason not to reduce nutrient loads into the river (Grimes, 2001). The waters adjacent to the Mississippi River delta have long been known as the Fertile Fisheries Crescent (Gunter, 1963), which emanates from the nutrient-enriched region in a bull's-eye pattern of sequentially reduced fisheries yields. The closure of the Aswan High Dam on the Nile River in 1965 resulted in a reduction in freshwater discharge and sediment and nutrient loads to the Nile delta with a subsequent loss of coastal diatom blooms and the fisheries the blooms supported (Nixon and Buckley, 2002). This reduction, however, has reversed to an increase in fisheries yields as agricultural fertilizer use and waste water from dense human populations in the delta region led to increased nutrient flux (Nixon, 2004; Oczkowskia et al., 2009). A comparison with the Nile River delta is not appropriate for the Mississippi River, because the increased nutrient load and associated bottom-water hypoxia in the northern Gulf of Mexico have turned the fertile bull's-eye pattern to altered distributions and abundances of demersal species, particularly the brown shrimp (Craig et al., 2005). It is feasible to maintain a high fisheries production with the same nutrient loads of the 1950s, given other factors holding constant, e.g., fishing pressure. Whether the Gulf of Mexico fisheries adjacent to the Mississippi River could return to the fertile fisheries crescent of the early 1950s depends on the form of the reversed trajectory of nutrient loads (quantity and ratios), nutrient concentrations, phytoplankton biomass, trophic interactions and food webs (see Kemp et al., 2009).

The growing decline of coastal water quality and expansion of symptoms of eutrophication and also the proven successes of reducing nutrients, however, are reasons enough to continue and to expand efforts to reduce nutrient over-enrichment. This is most important for systems that are currently eutrophic and have oxygen problems, where unusually warm and calm weather can tip the system from one with highly productive fisheries to one dominated by microbes. An example of weather exacerbated anoxia can be seen in the Mariager Fjord, Denmark. The 1997 summer in Denmark was unusually warm, sunny and calm. The high temperatures increased the rate of oxygen consumption, and the lack of wind mixing limited ventilation of deeper waters. As a result, the mussels at 8 to 10-m water depth died and contributed to the oxygen depletion. Then, the mussels at shallower depths died and the anoxic and hydrogen sulfide-rich zone gradually moved upward. After the massive phytoplankton bloom in the middle of August, the oxygen demand increased further due to the decomposition of phytoplankton. The long period of calm and unusual warm weather reducing the mixing in the surface was the triggering factor, but the geomorphology of the fjord and the high concentrations of nutrients mostly discharged from land were the basic drivers of the anoxia (Fallesen et al., 2000). To restore systems like this that are naturally prone to development of low oxygen, causal factors such as nutrient loadings need to be reduced beyond the level that will reduce hypoxia.

## 10 Global change and future expectations

### 10.1 Increases in eutrophication-driven hypoxia

The occurrence of hypoxia in coastal areas is increasing, and the trend is consistent with the increase in human activities that result in increased fluxes of nutrients to coastal waters. More and more coastal systems, especially in areas of increased industrialization and mechanized farming, where the physical conditions are appropriate and where nutrient loads are predicted to increase, will likely become eutrophic with accompanying hypoxia (Fig. 24).

The continued and accelerated export of nitrogen and phosphorus to the world's coastal ocean is the trajectory to be expected unless societal intervention in the form of controls or changes in culture are pursued. The largest increases are predicted for southern and eastern Asia, associated with predicted large increases in population, increased fertilizer use to grow food to meet the dietary demands of that population and increased industrialization (Seitzinger et al., 2002). Increased production of biofuels in many countries will further amplify nutrient delivery from the land to the sea (National Research Council, 2007; Runge and Senauer, 2007; Simpson et al., 2007). The implications for coastal eutrophication and subsequent ecosystem changes such as worsening conditions of oxygen depletion are significant.

Depending on the coastal environment, the residence time of river-supplied freshwater and rates of nutrient uptake and regeneration, the changes observed in the Anthropocene (Meybeck, 2003) from increased nutrient loads are most likely to lead to human-caused eutrophication and associated hypoxia. This does not preclude the potential for climate driven changes in oceanic nutrient sources from exacerbating an already precarious situation.

### 10.2 Climate change and coastal waters

The world's climate has changed and human activities will continue to contribute to the acceleration of greenhouse gases and temperature rise. The regional outcomes of the various GCC scenarios (IPCC, 2007) will be manifested in many different and synergistic effects on various components of ecosystems. These outcomes may be difficult to detect in the near-term, because they currently exist within the "background" variability inherent in natural systems, and also because there are effects from other human-caused changes. As in the sense of multiple stressors, it will be difficult to differentiate completely the global climate change aspect and the anthropogenic activities aspect of global change on eutrophication and associated hypoxia.

If the average global temperature rises 0.4°C over the next three decades and 1.8 to 4.0°C over the next century (IPCC, 2007), then there will be major changes in ecosystem structure and function. The major drivers of these changes are increased temperature, enhanced hydrological cycles and shifts in wind patterns that might alter coastal currents (Fig. 25) (Wolock and McCabe, 1999; Rabalais et al., 2009). Increasing temperatures alone have the potential to strengthen pycnoclines in estuarine and coastal waters, but lower surface salinity (e.g., from increased freshwater runoff) would be more of a factor in stratifying the water column. For example, at 20°C and 32 psu, a 1 psu decline in salinity changes seawater density by a factor of about 3 relative to a 1°C increase in temperature (Foofonoff and Millard, 1983). A stronger pycnocline (density differences dictated by a combination of temperature and salinity) may result in less diffusion of oxygen from the upper water column to the lower water column, leading to less dissolved oxygen in bottom waters.

Increased temperatures may also affect regional wind patterns, leading to changes in circulation and mixing. If wind patterns or intensity change, then coastal currents and their effects on coastal waters might change to either aggravate a low oxygen condition or to, in fact, alleviate it. An example of the former can be found off the Oregon and Washington coasts, where wind-driven shifts in the California current system in 2001 and subsequent years altered the long-term pattern of upwelling and resulted in extensive hypoxia along the inner shelf (Grantham et al., 2004; Chan et al., 2008).

If the frequency of tropical storms and hurricanes increases as a result of increased water temperatures in the lower- to mid-latitudes as predicted in GCC, the vulnerability of coastal habitats will be greatly increased. Warmer Atlantic Ocean sea surface temperatures could increase the frequency and severity of tropical storms (Trenberth, 2005; Hoyos, 2006; Kerr, 2006; Mann and Emanuel, 2006). Recent analyses (reviewed by Kerr, 2008), however, suggested only a modest increase, or even a decrease, in the frequency and intensity of future Atlantic cyclones. Knutson et al. (2007) predicted a modest 18% decrease in the frequency of Atlantic hurricanes by the end of the 21st century and a few percent increase in the intensity of storms. On the other hand, Emanuel et al. (2008) forecasted a few percent increase in storm frequency and a 7.5% increase in storm intensity.

The impacts of hurricanes and tropical storms on stratification and how much they temporarily alleviate hypoxia in the coastal waters of the northern Gulf of Mexico depend on several factors (Rabalais et al., 2007), which include barometric pressure, wind speed, storm trajectory, forward speed and circumference of the wind and wave field. If hurricanes occur during the hypoxia season, then there will only be transient dissipation of hypoxia. If they occur later in the summer, restratification may be slowed by an increasing frequency of fall-winter storms (mixing and thermal turnover).

The rates of biological processes, including both photosynthesis and respiration, are expected to increase with higher water temperatures up to a point. For instance, primary production may proceed at a faster rate but will eventually become limited by light (self-shading) or lack of nutrients. In addition, organisms will be brought closer to their thermal limits. These two factors, when combined, may result in a decrease in the physiological capacity of aerobic organisms (Pörtner and Knust, 2007).

Increased precipitation will result in more water, sediments and nutrients reaching the coastal zone where they are likely to enhance eutrophication through nutrient-enhanced production, increased stratification, or both (Cloern, 2001; Rabalais, 2004). The upper part of the Mississippi River watershed falls within the area of North America predicted to experience increased runoff (higher precipitation) and earlier peak discharge (increased temperature) (IPCC, 2007; Milliman et al., 2008). An increase in precipitation (10 to 30% per century) should result in increased erosion and loss of phosphorus and increased flux of dissolved inorganic nitrogen, particularly nitrate, through the soils and artificially drained agriculture areas. An increase in the water base flow, due to enhanced seasonal snow melt, is predicted to result in a 1 to 2 week earlier peak stream flow. The spring peak delivery of nutrients and freshwater to the northern Gulf of Mexico begins a seasonal cycle of increased primary production and flux of organic matter to bottom waters that subsequently lead to low oxygen concentrations (Rabalais et al., 2007). The combination of increased nutrient loads (from human activities) and increased freshwater discharge (from GCC) will aggravate the already high loads of nutrients from the Mississippi River to the northern Gulf of Mexico, strengthen stratification (all other factors remaining the same), and worsen the hypoxia situation (Justić et al., 2003a,b; Rabalais et al., 2009). Reduced precipitation, on the other hand, will lower the amount of nutrients and water reaching the coastal zone and, perhaps, lead to oligotrophication and reduced fisheries productivity (Nixon, 2004), or perhaps alleviate hypoxia. The increase or decrease in flow (whichever occurs), flux of nutrients and water temperature are likely to have important, but as yet not clearly identifiable, influences on hypoxia.

### 10.3 Oceanic waters

Oxygen-poor waters occupy large volumes of the intermediate-depth eastern tropical oceans. Oxygen-poor conditions have far-reaching impacts on ecosystems because important mobile macro-organisms avoid or cannot survive in hypoxic zones. Climate models predict declines in oceanic dissolved oxygen produced by global warming or increased atmospheric $CO_2$ or both. Gilbert et al. (2009) determined that the number of negative trend oxygen records for ocean waters >100 km from shore has significantly increased in the 1976–2000 period relative to 1951–1975.

As a consequence of global climate change, predictions are for OMZs to expand in area and extend into shallower depths, and for oxygen concentrations to decline. There could be a 50% increase in OMZs by 2100 (Riebesell et al., 2007; Arrigo, 2007). The core of tropical eastern Pacific OMZs is getting lower and had reached 40 $\mu$mol kg$^{-1}$ in 2008 (Stramma et al., 2008). With climate change, predictions are for global ocean oxygen outgassing to occur at a rate of 0.23±0.10 to 0.40 mol m$^{-2}$ yr$^{-1}$ by 2100 (Matear et al., 2000; Bopp et al., 2002; Frölicher et al., 2009). Stramma et al. (2008) constructed 50-year time series of dissolved-oxygen concentration for select tropical oceanic regions by augmenting a historical database with recent measurements. These time series revealed vertical expansion of the intermediate-depth low-oxygen zones in the eastern tropical Atlantic and the equatorial Pacific during the past 50 yr. The oxygen decrease in the 300- to 700-m layer was 0.09 to 0.34 $\mu$mol kg$^{-1}$ yr$^{-1}$. These declines, however, are barely significant given the cumulative oxygen content for the 300- to 700-m layer. Using the data in Stramma et al. (2008), it would take 38 yr to reduce the oxygen content in this layer by 5%. Organisms are not likely to respond to this small shift, but might if the trend continued for another 200 yr. Over longer time scales (ca. 100 000 yr), Shaffer et al. (2009) project severe, long-term ocean oxygen depletion, as well as a great expansion of ocean oxygen-minimum zones for scenarios with high emissions or high climate sensitivity. They predict that decreased oxygen solubility from surface-layer warming will account for most of the enhanced oxygen depletion in the upper 500 m of the open ocean.

Ocean scientists typically ignore the $CO_2$ side of the respiration equation, on the unspoken assumption that $pCO_2$ levels are low and are inversely proportional to the $O_2$ concentration via bacterial oxidation of marine organic matter. However, this may no longer be the case as Brewer and Peltzer (2009) predict that increasing atmospheric $CO_2$ concentrations rise and reset ocean chemical relations so that oceanic oxygen minimum zones will grow in size. Rising temperatures will further exacerbate the growth of low oxygen areas by decreasing the oxygen solubility of the water.

At global scales ocean oxygen is closely linked to climate (Pearce et al., 2008; Frölicher et al., 2009; Schaffer et al., 2009). While anthropogenic forcing of climate is well documented (IPCC), whether anthropogenic nutrients may further intensify declining oxygen at ocean basin scales and lead to expanded OMZ and anoxic condition is an issue of great concern. There is evidence for a higher proportion of declining oxygen trends near coastlines (<30 km distance) than in the open ocean, most probably due to runoff of nutrients in the coastal region (Gilbert et al., 2009). Anthropogenic nutrients have also been implicated in expanding the OMZ along the west coast of India (Naqvi et al., 2000). The Baltic Sea also has many physical characteristics that make it sensitive to development of hypoxia and anoxia. Over the last 50 to 100 yr the Baltic Sea has been tipped into semi-permanent anoxia by human action (Karlson et al., 2004). Similarly in the Black Sea, eutrophication is responsible for part of the expansion and shallowing of the suboxic zone (Oguz, 2005).

## 11 Conclusions

The ecosystems considered by SCOR WG 128 on “Natural and Human-Induced Hypoxia and Consequences for Coastal Areas” have different dominant time scales of variability, from estuaries (24 h to days), the northern Gulf of Mexico (weeks to months), to the Peru-Chile upwelling system (2 to 8 years with El Niño Southern Oscillation forcing) and even longer time scales (the Baltic and Black seas). The ecosystems in which hypoxia occurs range from nearshore estuaries, through the coastal ocean and into ocean waters, over depths of 1- to 2-m in estuaries to 600- to 700-m in the open ocean, and vary in physiography, physical processes, organic and nutrient loading and ecosystem structure. The richness of temporal and spatial scales helps us understand the basic dynamics of hypoxia, changes in relation to anthropogenic and climate forcing and the potential for recovery of ecosystems from hypoxia.

There is no doubt that the increase in the areas in the global ocean with hypoxia is real. Coastal water quality with regard to hypoxia is currently on the decline, and the future, based on the continued increase in the global occurrence of hypoxia and current and projected increased loads of nutrients, is trending to more hypoxia. The formation of hypoxic areas has been exacerbated by any combination of interactions that increase primary production and accumulation of organic carbon leading to increased respiratory demand for oxygen below a seasonal or permanent pycnocline. The likelihood of strengthened stratification alone, from increased surface water temperature as the global climate warms, is sufficient to worsen hypoxia where it currently exists and facilitate its formation in additional coastal waters. The interplay of increased nutrients and stratification may be offset temporarily by the potential for more frequent and/or severe tropical storm systems, but the tropical storm projections with increased global temperatures remain under debate. The overall forecast, however, is for hypoxia to worsen, with increased occurrence, frequency, intensity and duration. The consequences of global warming and climate change are effectively uncontrollable at least in the near term. On the other hand, the consequences of eutrophication-induced hypoxia can be reversed if long-term, broad-scale, and persistent efforts to reduce substantial nutrient loads are developed and implemented. The need for water and resource managers to reduce nutrient loads even, if at a minimum, to maintain the current status, is critical in view of globally expanding hypoxia.

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**Table 1.** Changes in water discharge and nitrogen flux from the Changjiang over three time periods.

| Period | Fresh Water Discharge ($\times 10^9$ m$^3$ yr$^{-1}$) | Nitrate Concentration (µM) | Nitrogen Flux ($\times 10^9$ Mol yr$^{-1}$) | References |
|---|---|---|---|---|
| 1960s | 898.9 | 30–40 | 27–36 | Gu et al., 1981a,b |
| 1980s | 889.7 | 63.4 | 56.4 | Zhang et al., 1999 |
| 2001–2005 | 887.7 | 81.3 | 72.2 | J. Zhang, Unpubl. data |

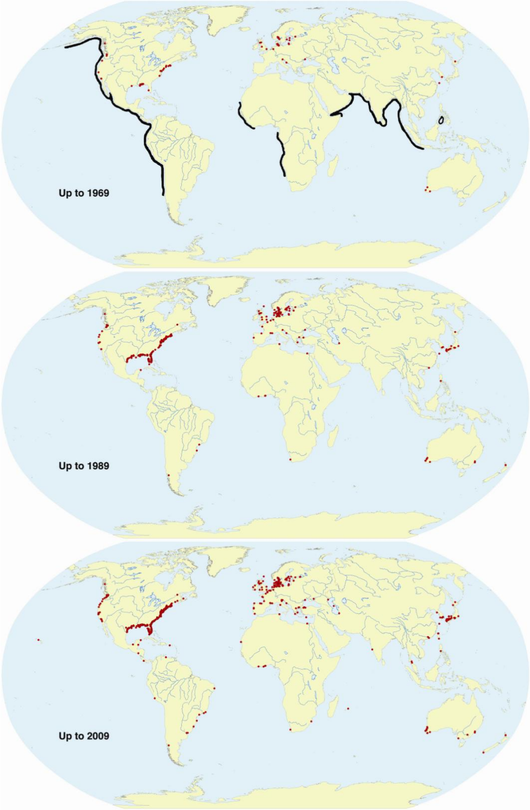

**Fig. 1.** Global pattern in the development of coastal hypoxia. Each red dot represents a documented case related to human activities. Number of hypoxic sites is cumulative through time. Black lines represent continental shelf areas threatened with hypoxia from expansion of OMZ and upwelling. Modified from Díaz and Rosenberg (2008) and Levin et al. (2009).

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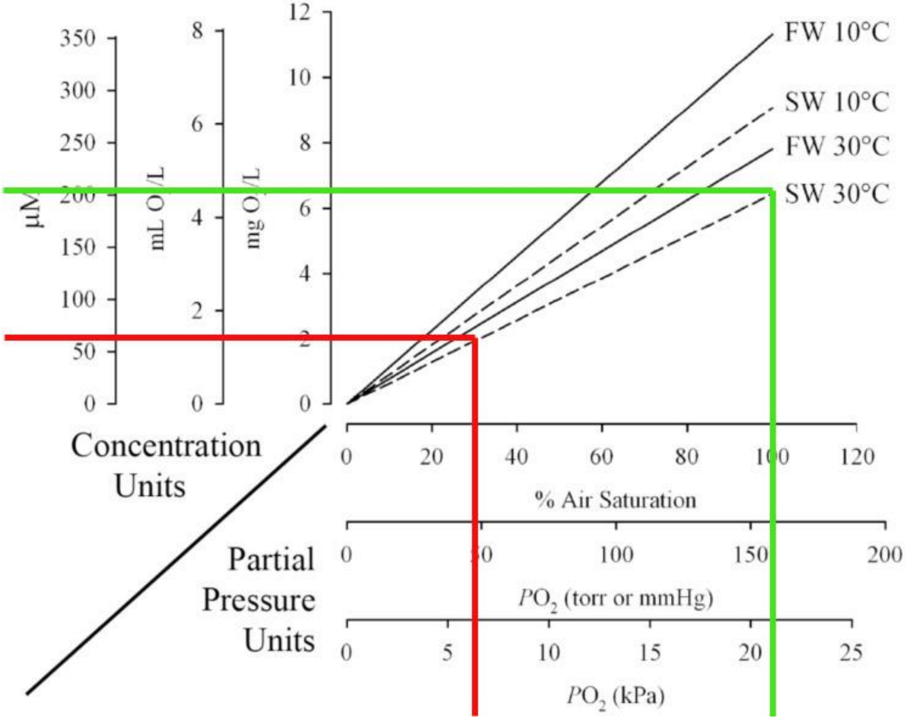

**Fig. 2.** Nomogram for dissolved oxygen in fresh water (FW) and sea water (SW) at 10°C and 30°C (modified from Díaz and Breitburg, 2009). Concentration units are on Y-axis, and partial pressure units are on X-axis. Red line is the SCOR definition of hypoxia and green line is the solubility of oxygen in sea water at 30°C.

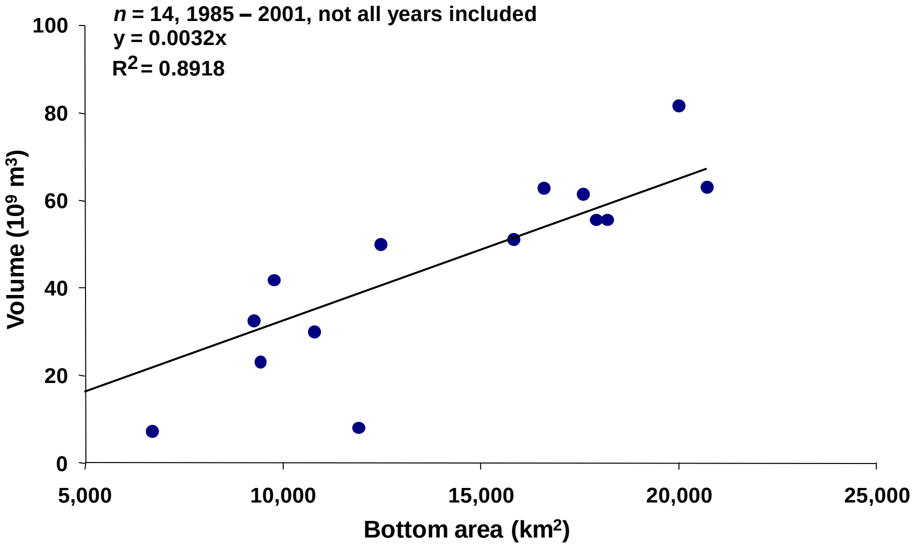

**Fig. 3.** The relationship between area covered by hypoxia (dissolved oxygen <2 $\mathrm{mg\,L^{-1}}$) and volume of hypoxia for the northern Gulf of Mexico (unpubl. data from N. N. Rabalais and R. E. Turner).

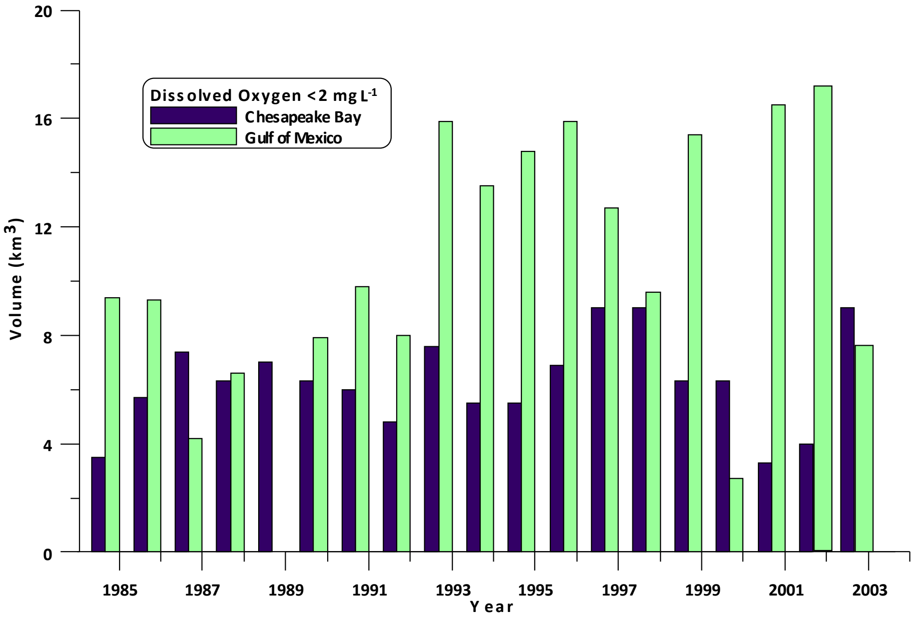

**Fig. 4.** Comparison of volume data for the Chesapeake Bay (http://www.eco-check.org/forecast/chesapeake/2009/i) and volume data for Gulf of Mexico, unpubl. data from N. N. Rabalais and R. E. Turner.

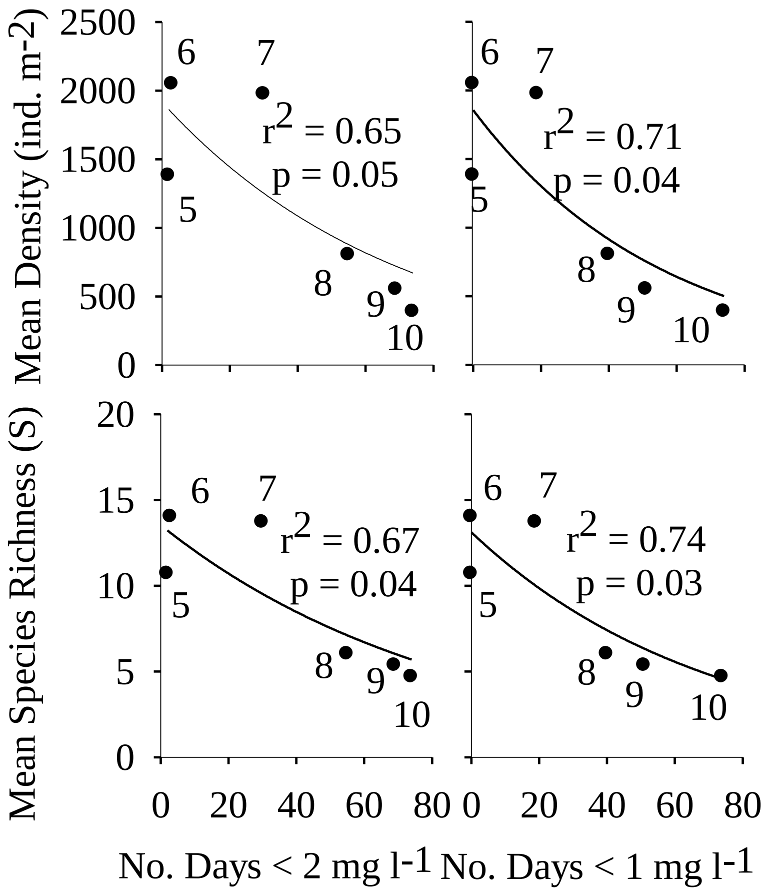

**Fig. 5.** Monthly mean density and species richness of macroinfauna from (5) May to (10) October 2004 after increasing exposure to oxygen levels $\leq 2\,\mathrm{mg\,O_2\,L^{-1}}$ and $\leq 1\,\mathrm{mg\,O_2\,L^{-1}}$ (modified from Baustian and Rabalais, 2009).

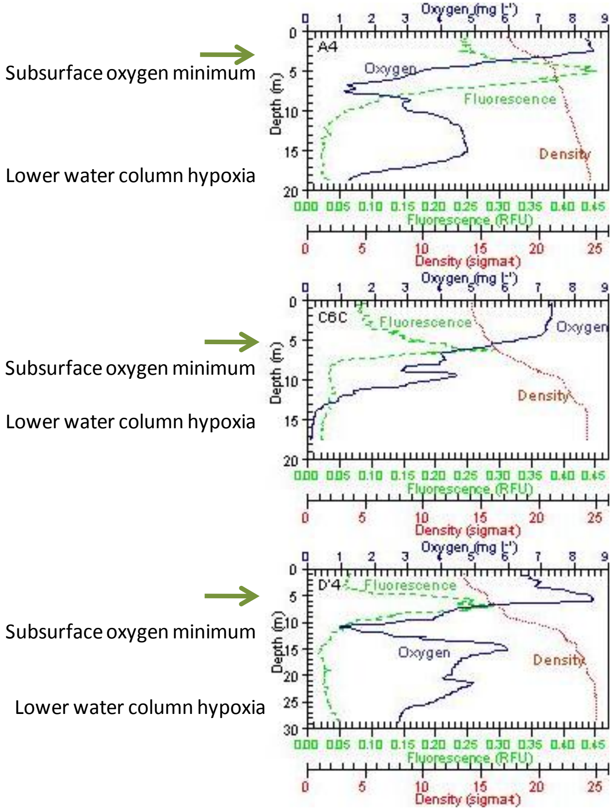

**Fig. 6.** Hypoxia below a chlorophyll biomass accumulation as shown by in vivo fluorescence, indicated by a green arrow, and lower water column hypoxia. Stations are in 20 to 30 m water depth within 50 km west of the Mississippi River delta (modified from Rabalais and Turner, 2006).

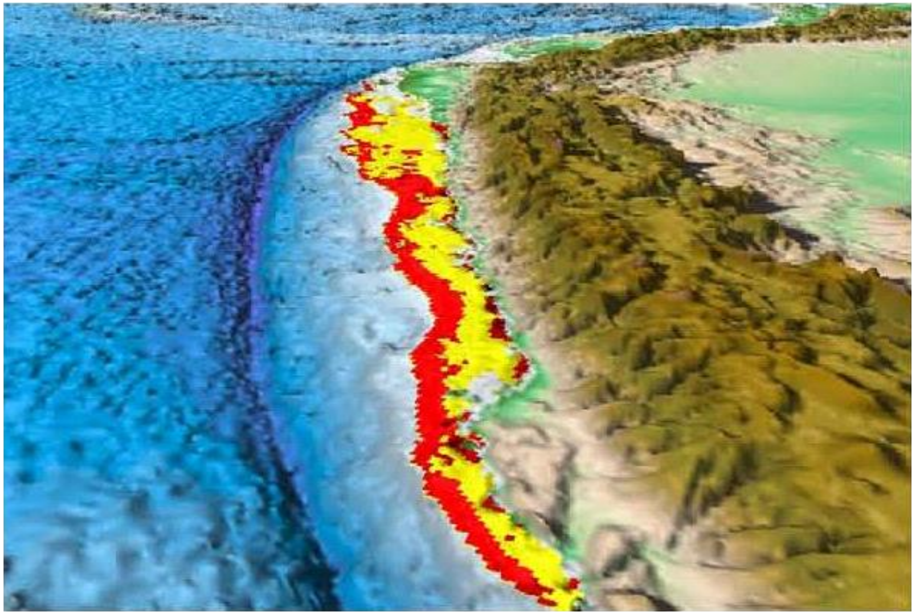

**Fig. 7.** The coast of Peru and northern Chile looking north. Central and North America are in the distant background. The red area is the estimated area of the OMZ during an extreme El Niño event, while the union of red and yellow indicates the extent during non-El Niño periods (from Helly and Levin, 2004).

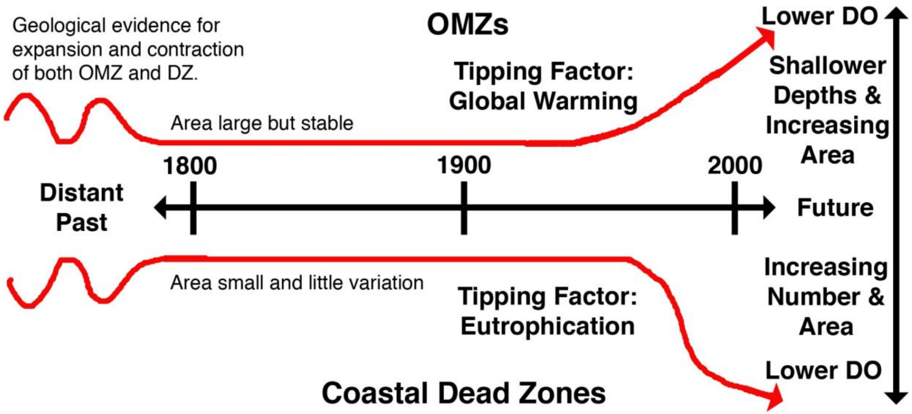

**Fig. 8.** Comparison of the history of oxygen minimum zones (OMZs) and coastal dead zones (DZ) from the past into the future. Both are expanding from anthropogenic activities, but from different processes. OMZs are expanding from global warming and DZs from eutrophication.

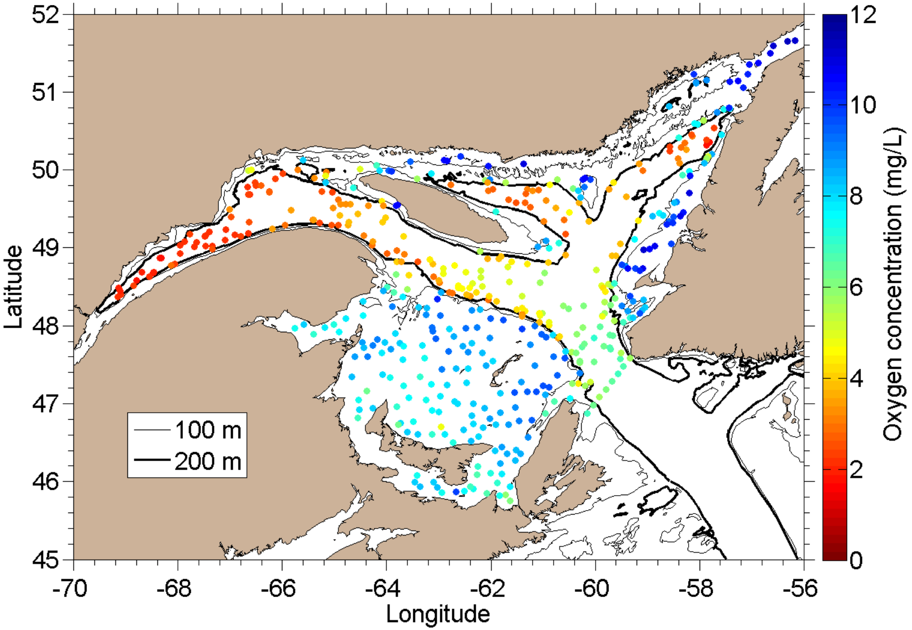

**Fig. 9.** Near-bottom oxygen concentration (mg $L^{-1}$) measured during the 2004 (southern and northern Gulf of St. Lawrence) and 2005 (only the northern part of the Gulf of St. Lawrence) fish stock assessment surveys. Modified from data published in Gilbert et al. (2007). Used with permission from Société Provancher.

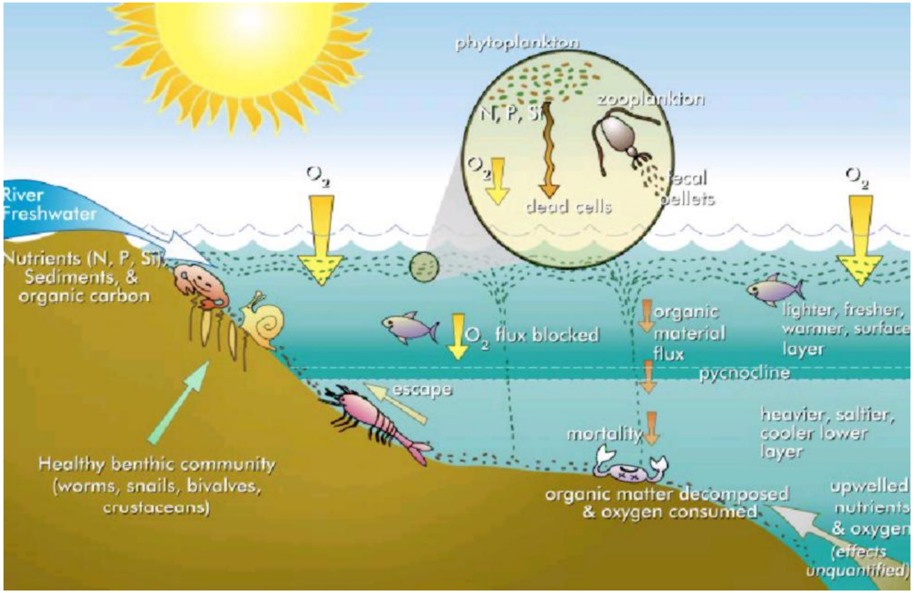

**Fig. 10.** Processes involved in the development and maintenance of hypoxia on the northern Gulf of Mexico continental shelf where there is sustained and high freshwater discharge from the Mississippi River, year-round haline and seasonal thermal stratification, nutrient-enhanced primary production, and accumulation and decomposition of organic matter (CENR, 2000).

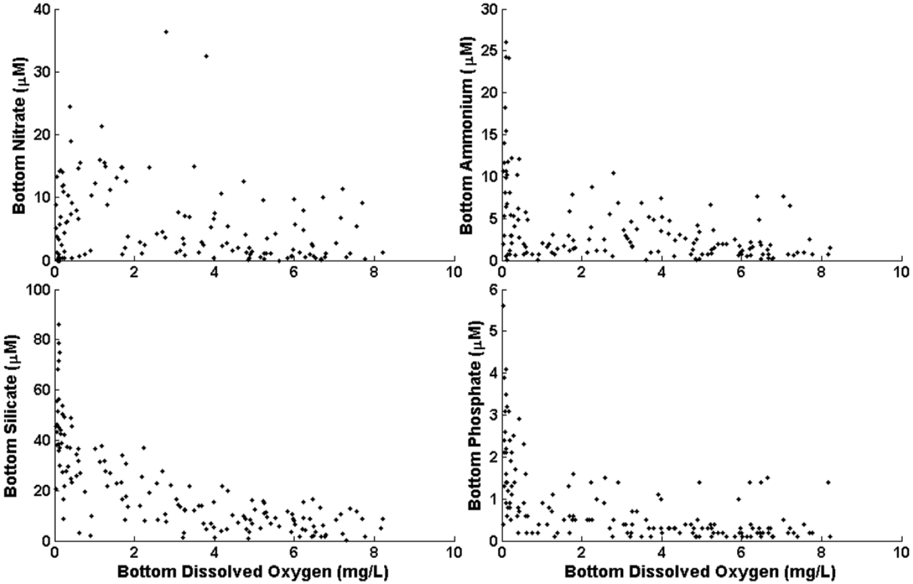

**Fig. 11.** Comparison of bottom water dissolved oxygen concentration and bottom-water dissolved inorganic nutrient concentrations for stations within the hypoxic area of the northern Gulf of Mexico (see Fig. 14), unpubl. data N. N. Rabalais and R. E. Turner.

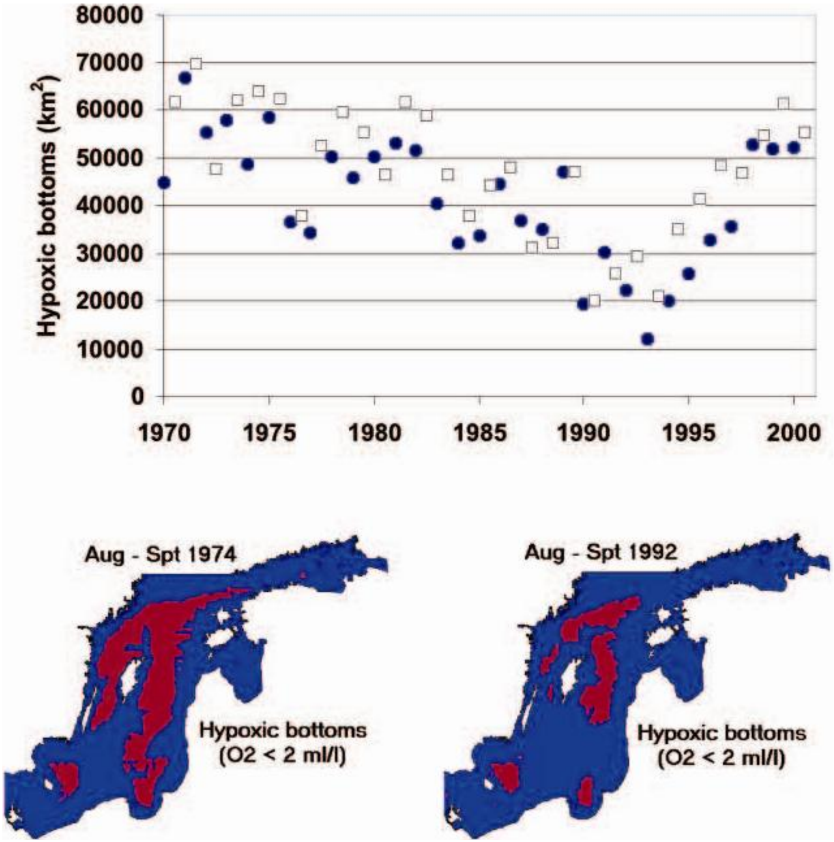

**Fig. 12.** Long-term variations of the bottom area covered with waters containing less than 2 ml L$^{-1}$ oxygen (upper panel). Oxygen fields were averaged over August–September (open squares) and January–March (closed circles) for each calendar year for the Baltic Proper, which includes the Gulf of Finland and the Gulf of Riga. Spatial distributions for August–September in 1974 and 1992 are shown in the bottom panel (from Conley et al., 2002a).

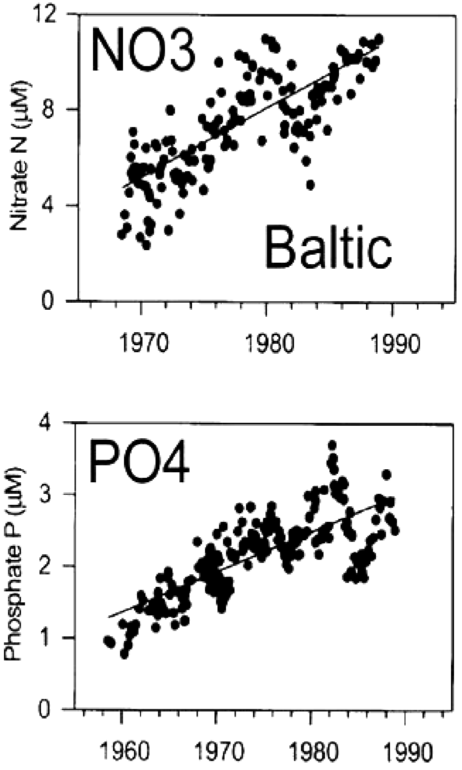

**Fig. 13.** Nitrate and phosphate concentrations in the central Baltic Sea (Nehring, 1992, redrawn by Cloern, 2001).

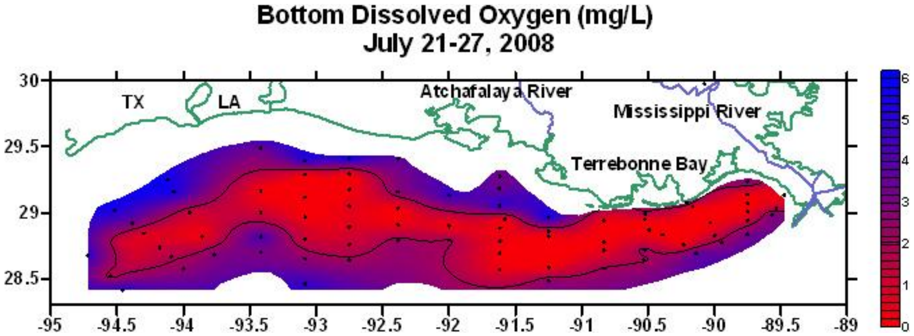

**Fig. 14.** Gulf of Mexico hypoxia in mid-summer 2008. Line depicts oxygen less than 2 mg $L^{-1}$. Dots indicate stations. Data source: N. N. Rabalais and R. E. Turner.

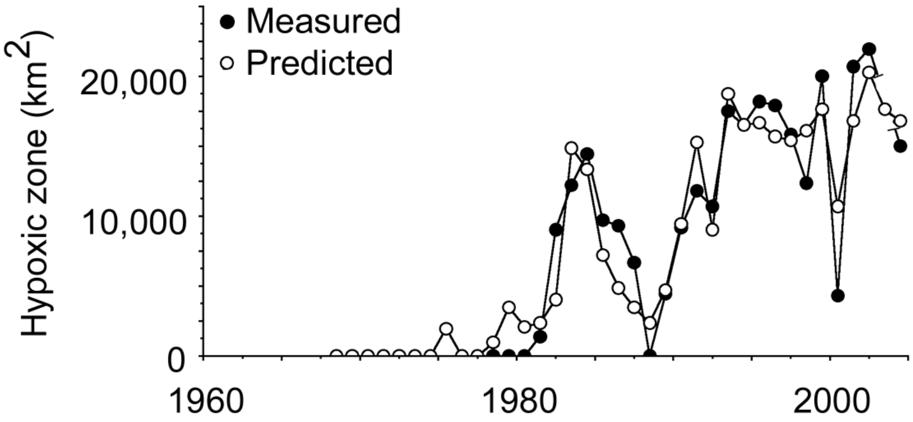

**Fig. 15.** The results of a final model predicting the size of the hypoxic zone from 1968 to 2004, and the estimates over the whole shelf. The equation is $y$ (km$^2$)$=-1\,337\,953.4+672.1589\times$Year$+0.0998\times$(May flux as $NO_3^{-2}$). The hindcasted values plotted as zero before 1978 are negative values in the model (from Turner et al., 2006).

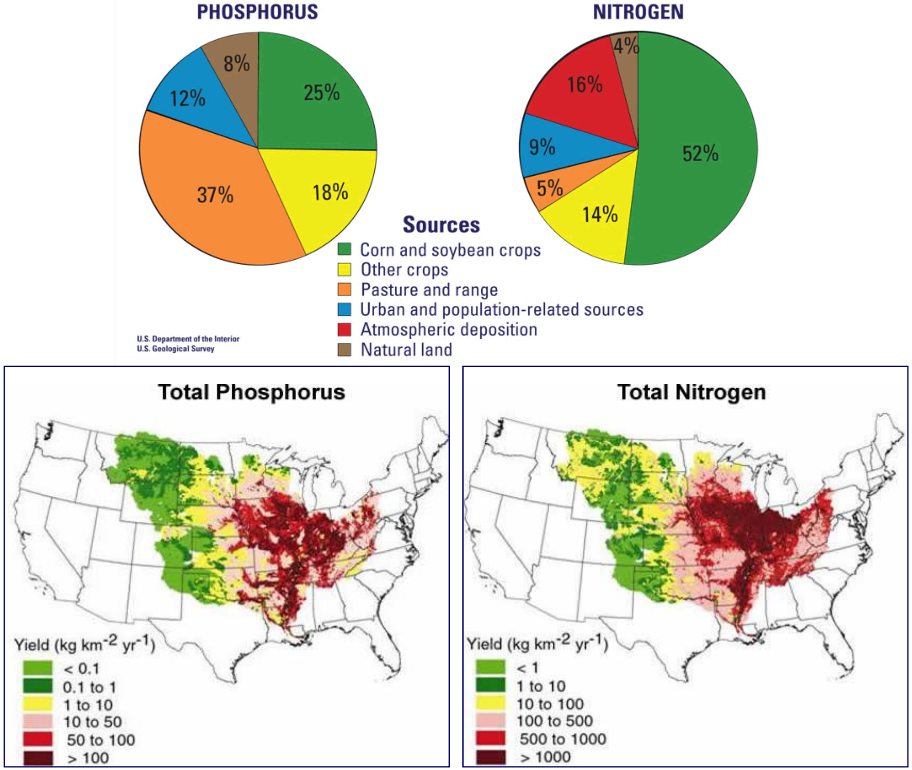

**Fig. 16.** Relative contribution of nitrogen and phosphorus sources based on land-use and activity (upper panel) and distribution of nitrogen and phosphorus yields throughout the Mississippi River basin (from http://water.usgs.gov/nawqa/sparrow/gulf_findings/) and Alexander et al. (2008).

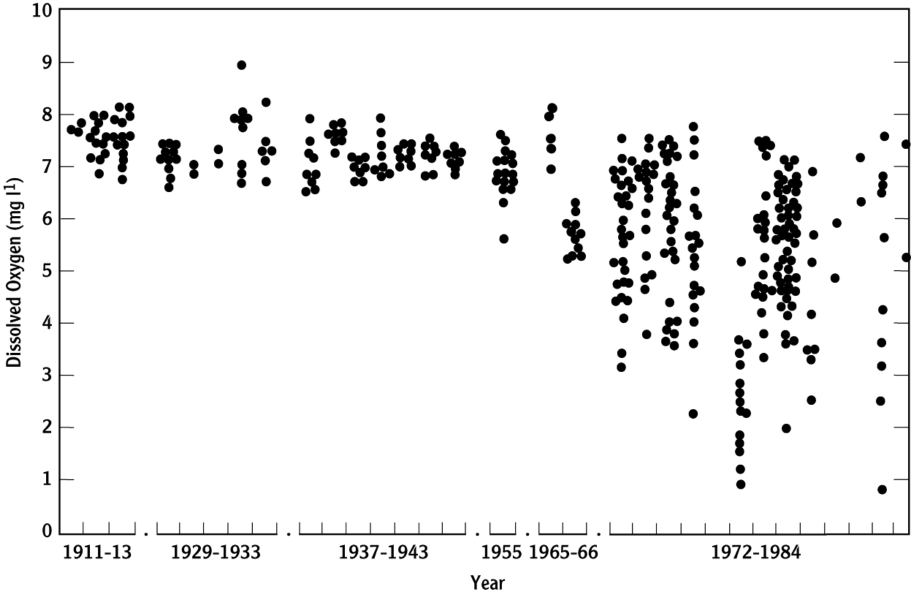

**Fig. 17.** Oxygen content 2 m above the bottom during August-September in the northern Adriatic Sea from 1911 to 1984 for the periods indicated. Surface water oxygen content increased in the 1972–1984 summer period, and Secchi disk depth during April–May decreased dramatically in 1966 and in the 1972–1984 period. Statistical trends for the bottom water oxygen (shown), surface water oxygen and Secchi disk depth were all significant at $P=0.05$. Redrawn from Justić (1991) with permission.

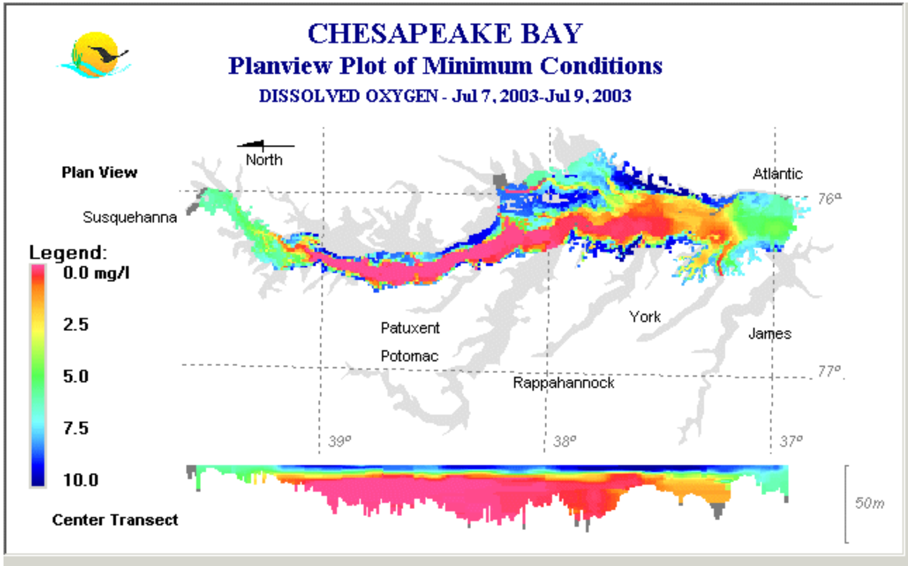

**Fig. 18.** Areal extent of low dissolved oxygen water in Chesapeake Bay and the concentrations along a central Bay transect, July 2003, a typical oxygen year (http://www.chesapeakebay.net).

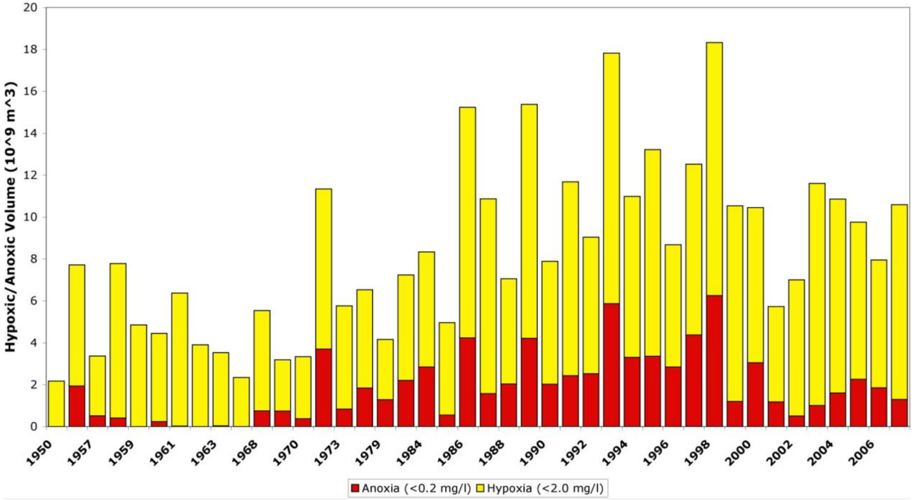

**Fig. 19.** Volume of hypoxic water in Chesapeake Bay (data from Hagy et al., 2004 and http://www.eco-check.org/forecast/chesapeake/2009/indicators/hypoxia/).

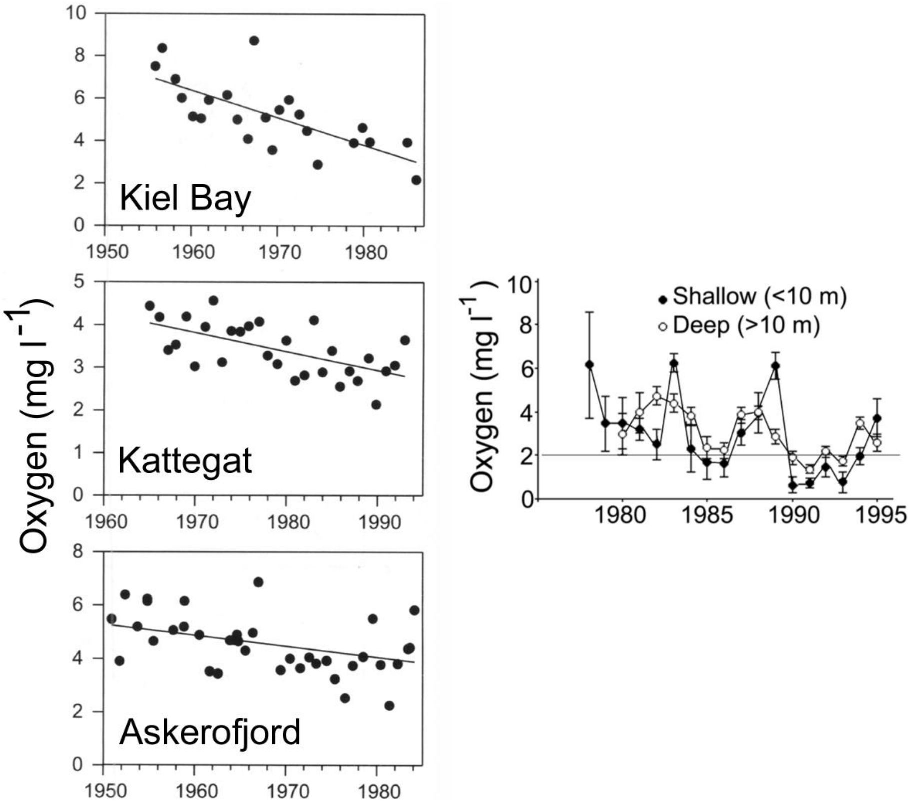

**Fig. 20.** Decline in oxygen concentrations within several coastal systems for similar time periods (left panel) (from Cloern, 2001), and bottom water oxygen concentration for shallow and deep water stations for June, July and August in the Louisiana bight of the northern Gulf of Mexico adjacent to the Mississippi River discharge (mean±1 S.E.) (right panel) (from Turner et al., 2005).

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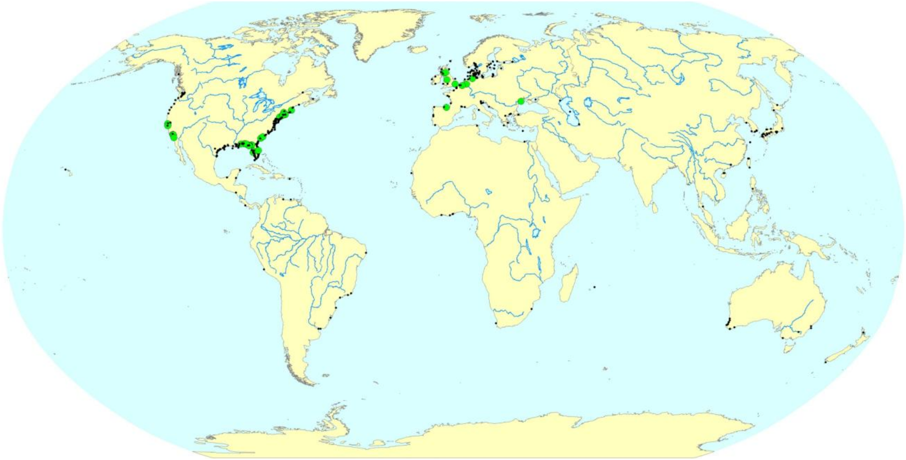

**Fig. 21.** Location of systems that have recovered from hypoxia (green circles), primarily through reduction of nutrient loads. All sites are in northern Europe and the United States, except the Black Sea. Black dots are systems that remain hypoxic.

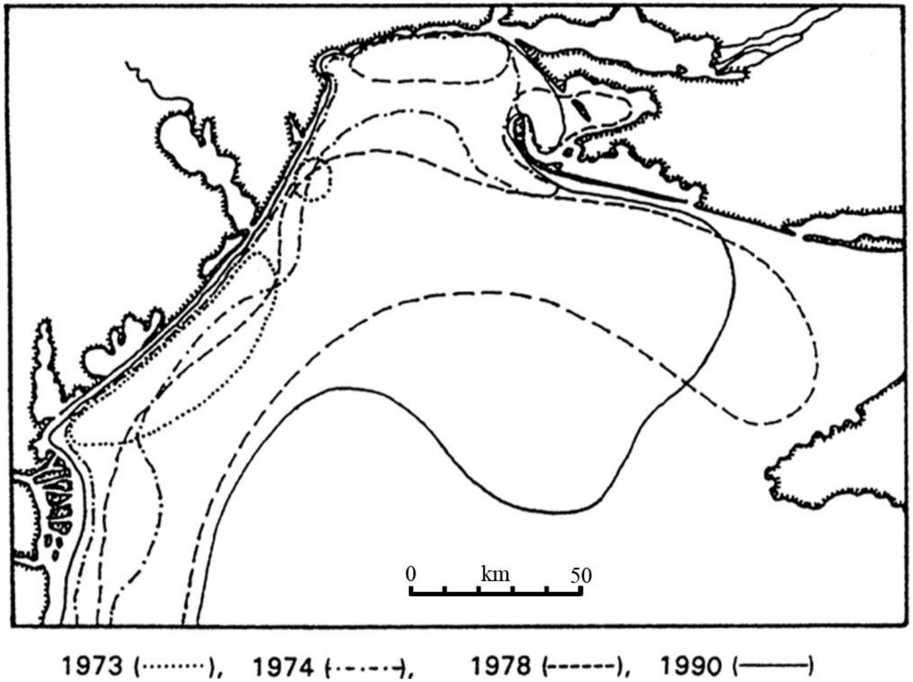

**Fig. 22.** Development of hypoxia on the northwestern shelf of the Black Sea from the 1970s to the 1990 (modified from Zaitsev, 1992).

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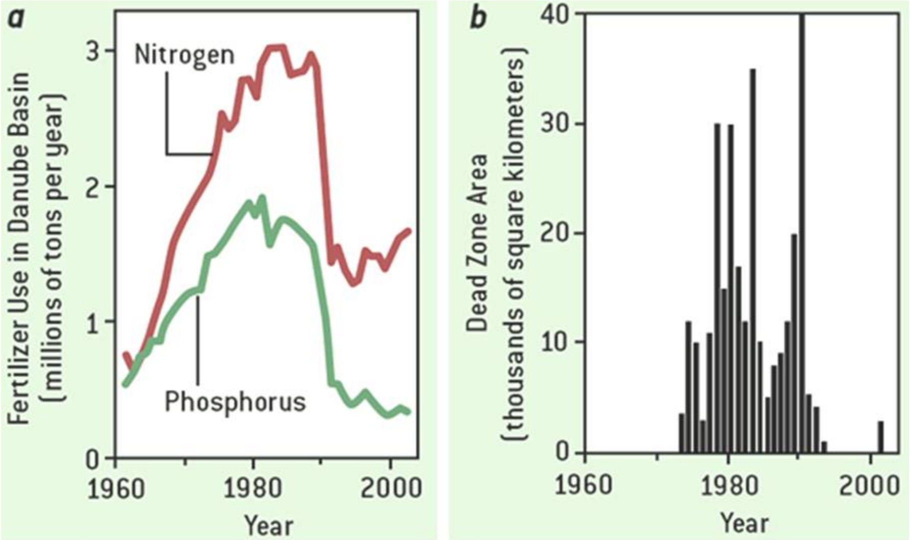

**Fig. 23.** Change in the extent of bottom-water hypoxia on the northwestern shelf of the Black Sea in relationship to groupings of years with differing use of nitrogen fertilizer in the Danube River watershed (data redrawn from Mee, 2001 in Mee, 2006). Nitrogen and phosphorus fertilizer trends for use in the watersheds draining to the northwestern shelf of the Black Sea. The nitrogen and phosphorus loads corresponded to fertilizer use (Mee, 2001).

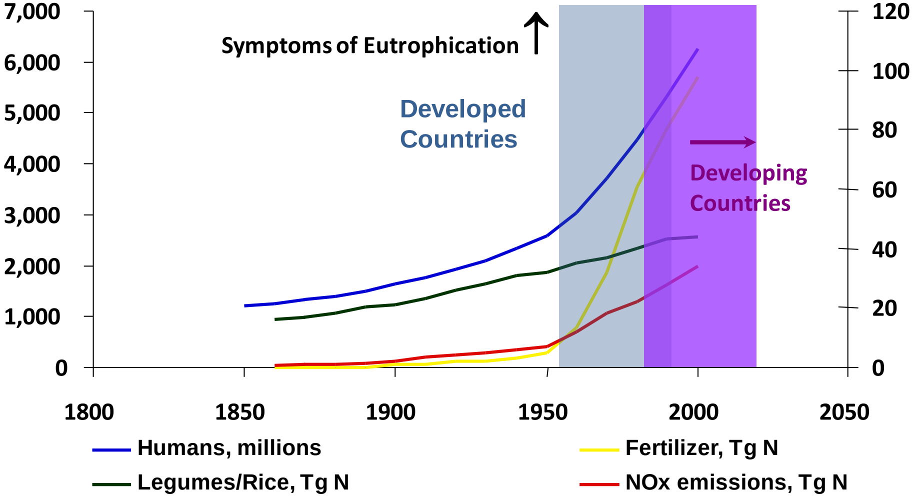

**Fig. 24.** Period in which the symptoms of eutrophication and hypoxia/anoxia began in developed countries and how the symptoms are shifted to more recent years for developing countries (modified by N. N. Rabalais from Galloway and Cowling, 2002; Boesch, 2002).

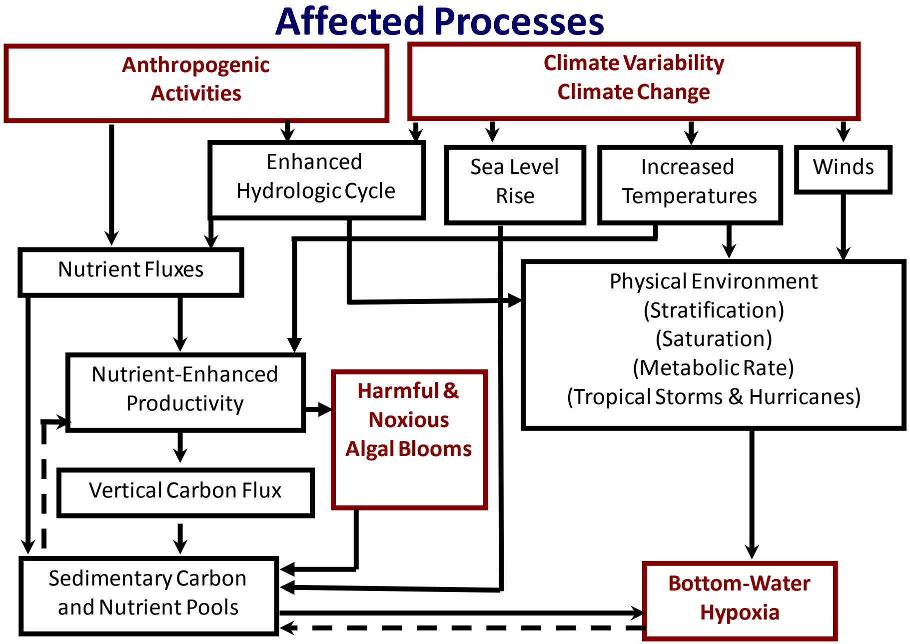

**Fig. 25.** Potential physical and hydrological changes resulting from climate change and their interaction with current and future human activities. The dashed lines represent negative feedback to the system (from Rabalais et al., 2009).