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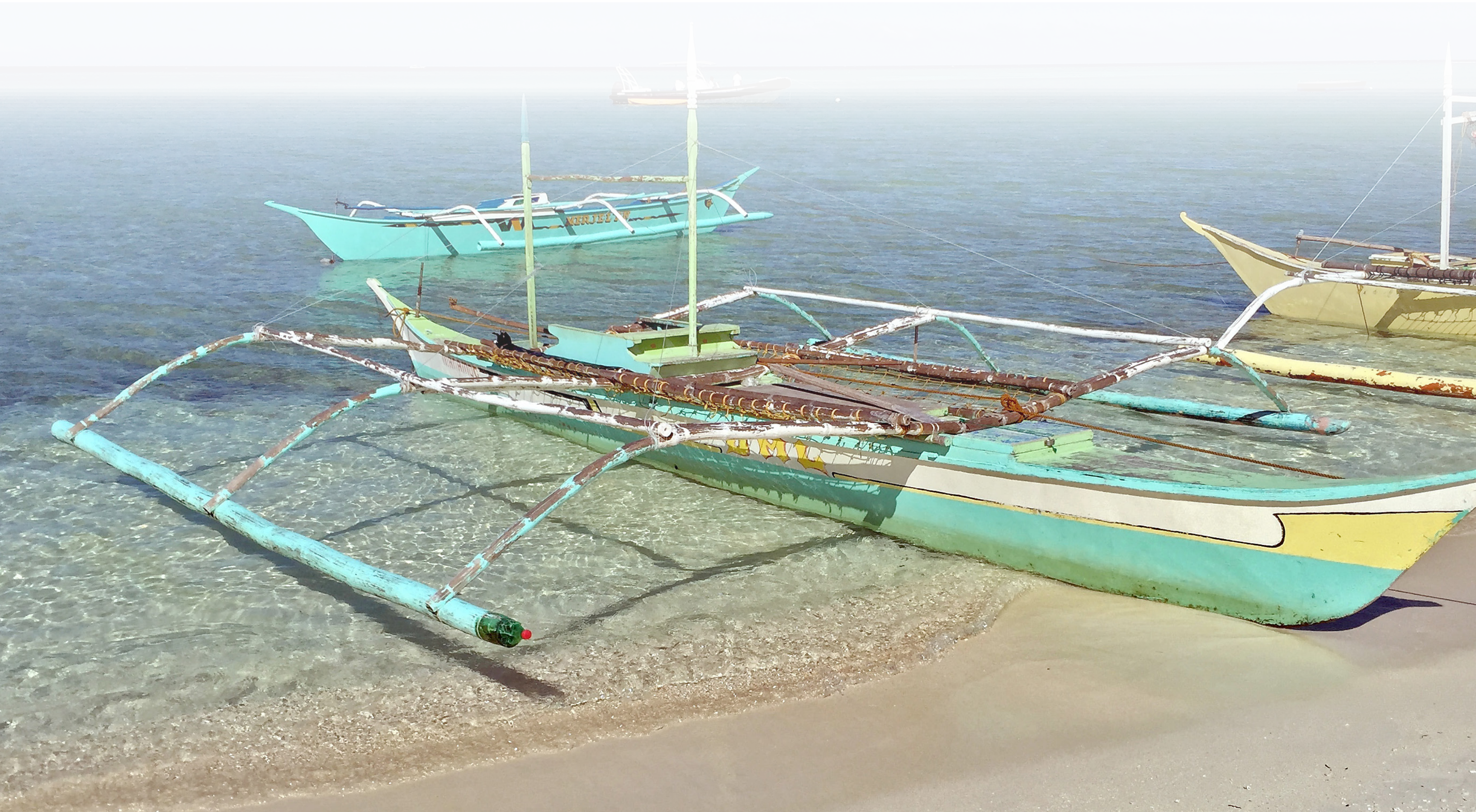

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Fifth Assessment Report of the
Intergovernmental Panel on Climate Change

Climate Change 2014: Impacts, Adaptation, and Vulnerability

5. Coastal Systems and Low-Lying Areas
6. Ocean Systems (+ supplementary material)
30. The Ocean (+ supplementary material)

Cross-Chapter Boxes (CR,HS,MB,OA,PP,UP)

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3. Observations: Ocean
   <http://www.climatechange2013.org/images/report/WG1AR5_Chapter03_FINAL.pdf>
6. Carbon and Other Biogeochemical Cycles
   <http://www.climatechange2013.org/images/report/WG1AR5_Chapter06_FINAL.pdf>
   <http://www.climatechange2013.org/images/report/WG1AR5_Ch06SM_FINAL.pdf>
13. Sea Level Change
   <http://www.climatechange2013.org/images/report/WG1AR5_Chapter13_FINAL.pdf>
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Coastal Systems and Low-Lying Areas

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Frequently Asked Questions

Executive Summary

Coastal systems are particularly sensitive to three key drivers related to climate change: sea level, ocean temperature, and ocean acidity (*very high confidence*). {5.3.2, 5.3.3.4, 5.3.3.5} Despite the lack of attribution of observed coastal changes, there is a long-term commitment to experience the impacts of sea level rise because of a delay in its response to temperature (*high confidence*). {5.5.8} In contrast, coral bleaching and species ranges can be attributed to ocean temperature change and ocean acidity. {5.4.2.2, 5.4.2.4} For many other coastal changes, the impacts of climate change are difficult to tease apart from human-related drivers (e.g., land use change, coastal development, pollution) (*robust evidence*, *high agreement*).

Coastal systems and low-lying areas will increasingly experience adverse impacts such as submergence, coastal flooding, and coastal erosion due to relative sea level rise (RSLR; *very high confidence*). In the absence of adaptation, beaches, sand dunes, and cliffs currently eroding will continue to do so under increasing sea level (*high confidence*). {5.4.2.1, 5.4.2.2} Large spatial variations in the projected sea level rise together with local factors means RSLR at the local scale can vary considerably from projected global mean sea level rise (GMSLR) (*very high confidence*). {5.3.2} Changes in storms and associated storm surges may further contribute to changes in sea level extremes but the small number of regional storm surge studies, and uncertainty in changes in tropical and mid-latitude cyclones at the regional scale, means that there is *low confidence* in projections of storm surge change {5.3.3.2} Both RSLR and impacts are also influenced by a variety of local processes unrelated to climate (e.g., subsidence, glacial isostatic adjustment, sediment transport, coastal development) (*very high confidence*).

Acidification and warming of coastal waters will continue with significant negative consequences for coastal ecosystems (*high confidence*). The increase in acidity will be higher in areas where eutrophication or coastal upwellings are an issue. It will have negative impacts for many calcifying organisms (*high confidence*). {5.4.2.2} Warming and acidification will lead to coral bleaching, mortality, and decreased constructional ability (*high confidence*), making coral reefs the most vulnerable marine ecosystem with little scope for adaptation. {5.4.2.4, Box CC-OA} Temperate seagrass and kelp ecosystems will decline with the increased frequency of heat waves and sea temperature extremes as well as through the impact of invasive subtropical species (*high confidence*). {5.4.2.3}

The population and assets exposed to coastal risks as well as human pressures on coastal ecosystems will increase significantly in the coming decades due to population growth, economic development, and urbanization (*high confidence*). The exposure of people and assets to coastal risks has been rapidly growing and this trend is expected to continue. {5.3.4.1, 5.4.3.1} Humans have been the primary drivers of changes in coastal aquifers, lagoons, estuaries, deltas, and wetlands (*very high confidence*) and are expected to further exacerbate human pressures on coastal ecosystems resulting from excess nutrient input, changes in runoff, and reduced sediment delivery (*high confidence*). {5.3.4.2, 5.3.4.3, 5.3.4.4}

For the 21st century, the benefits of protecting against increased coastal flooding and land loss due to submergence and erosion at the global scale are larger than the social and economic costs of inaction (*limited evidence*, *high agreement*). Without adaptation, hundreds of millions of people will be affected by coastal flooding and will be displaced due to land loss by year 2100; the majority of those affected are from East, Southeast, and South Asia (*high confidence*). {5.3.4.1, 5.4.3.1} At the same time, protecting against flooding and erosion is considered economically rational for most developed coastlines in many countries under all socioeconomic and sea level rise scenarios analyzed, including for the 21st century GMSLR of above 1 m (*limited evidence*, *high agreement*). {5.5.5}

The relative costs of adaptation vary strongly between and within regions and countries for the 21st century (*high confidence*). Some low-lying developing countries (e.g., Bangladesh, Vietnam) and small islands are expected to face very high impacts and associated annual damage and adaptation costs of several percentage points of gross domestic product (GDP). {5.5.5} Developing countries and small islands within the tropics dependent on coastal tourism will be impacted directly not only by future sea level rise and associated extremes but also by coral bleaching and ocean acidification and associated reductions in tourist arrivals (*high confidence*). {5.4.3.4}

The analysis and implementation of coastal adaptation toward climate-resilient and sustainable coasts has progressed more significantly in developed countries than in developing countries (*high confidence*). Given ample adaptation options, more proactive responses can be made and based on technological, policy related, financial, and institutional support. Observed successful adaptation includes major projects (e.g., Thames Estuary, Venice Lagoon, Delta Works) and specific practices in both developed countries (e.g., Netherlands, Australia) and developing countries (e.g., Bangladesh). {5.5.4.2} More countries and communities carry out coastal adaptation measures including those based on integrated coastal zone management, local communities, ecosystems, and disaster reduction, and these measures are mainstreamed into relevant strategies and management plans (*high confidence*). {5.5.4, 5.5.5}

5.1. Introduction

This chapter presents an updated picture of the impacts, vulnerability, and adaptation of coastal systems and low-lying areas to climate change, with sea level rise perceived as the most important risk for human systems. Unlike the coastal chapter in the previous assessment (Fourth Assessment Report, AR4), materials pertinent to the oceans are not covered here but in two new ocean chapters (Chapters 6 and 30). As in AR4, polar coasts are in another chapter (Chapter 28); small islands are also considered separately (Chapter 29) so an in-depth discussion is not provided herein.

The topics covered in this chapter follow the outline for sectoral chapters approved by the IPCC. An Executive Summary summarizes the key messages with a line of sight to the supporting sections in the chapter.

This chapter consists of six sections, with this first section dealing with progress in knowledge from AR4 to AR5 (Fifth Assessment Report), scope of chapter, and new developments. Section 5.2 defines the coastal systems and climate and non-climate drivers. The coastal systems include both natural systems and human systems, and this division is generally followed throughout the chapter. The climate and non-climate drivers are assessed in Section 5.3, followed by the impacts, vulnerabilities, and risks in Section 5.4. Section 5.5 deals with adaptation and managing risks. Information gaps, data gaps, and research needs are assessed in Section 5.6. There is one box on a specific example and reference to three cross-chapter boxes.

In AR4, the coastal chapter assessed the impact of climate change and a global sea level rise up to 0.59 m in the 2090s. The coastal systems were considered to be affected mainly by higher sea levels, increasing temperatures, changes in precipitation, larger storm surges, and increased ocean acidity. Human activities had continued to increase their pressure on the coasts with rapid urbanization in coastal areas and growth of megacities with consequences on coastal resources. Regionally, South, Southeast, and East Asia; Africa; and small islands were identified as most vulnerable. The AR4 chapter offered a range of adaptation measures, many under the Integrated Coastal Zone Management (ICZM) framework that could be carried out in both developed and developing countries, but recognized that the latter would face more challenges. Various issues on increasing the adaptive capacity or increasing the resilience of coastal communities were discussed. The unavoidability of sea level rise in the long term, even with stringent mitigation, was noted, with adaptation becoming an urgent issue.

A number of key issues related to the coasts have arisen since AR4. There is now better understanding of the natural systems, their ecosystem functions, their services and benefits to humanity, and how they can be affected by climate change. Their linkages landward to the watersheds and seaward to the seas and oceans need to be considered for a more integrated assessment of climate change impacts. The global mean sea level rise (GMSLR) is projected to be 0.28 to 0.98 m by 2100 (Table 5-2), although with regional variations and local factors the local sea level rise can be higher than that projected for the GMSLR. This has serious implications for coastal cities, deltas, and low-lying states. While higher rates of coastal erosion are generally expected under rising sea levels, the complex inter-relationships between the geomorphological and ecological attributes of the coastal system (Gilman et al., 2006; Haslett, 2009) and the relevant climate and oceanic processes need to be better established at regional and local scales. Such complex inter-relationships can be influenced by different methods and responses of coastal management.

Also of concern is ocean acidification. Together with warming, it causes coral reefs to lose their structural integrity, negatively implicating reef communities and shore protection (Sheppard et al., 2005; Manzello et al., 2008; see Boxes CC-OA, CC-CR). Acidification has potential impacts of reduced calcification in shellfish and impacts on commercial aquaculture (Barton et al., 2012). Since AR4, a significant number of new findings regarding the impacts of climate change on human settlements and key coastal systems such as rocky coasts, beaches, estuaries, deltas, salt marshes, mangroves, coral reefs, and submerged vegetation have become available and are reviewed in this chapter. However, uncertainties regarding projections of potential impacts on coastal systems remain generally high.

This chapter also provides advances in both vulnerability assessments and the identification of potential adaptation actions, costs, benefits, and trade-offs. A large number of new studies estimate the costs of inaction versus potential adaptation. Coastal adaptation has become more widely used, with a wider range of approaches and frameworks such as integrated coastal management, ecosystem-based adaptation, community-based adaptation, and disaster risk reduction and management.

Climate change will interact differently with the variety of human activities and other drivers of change along coastlines of developed and developing countries. For example, on the coastlines of developed countries, changes in weather and climate extremes and sea level rise may impact the demand for housing, recreational facilities, and construction of renewable energy infrastructure on the coast (Hadley, 2009), including critical infrastructures such as transportation, ports, and naval bases. Along the coasts of developing countries, weather and climate extremes affect a wide range of economic activities supporting coastal communities and pose an additional risk to many of the fastest growing low-lying urban areas, such as in Bangladesh and China (McGranahan et al., 2007; Smith, 2011).

5.2. Coastal Systems

Coastal systems and low-lying areas, also referred to as coasts in this assessment, include all areas near mean sea level. Generally, there is no single definition for the coast and the coastal zone/area, where the latter emphasizes the area or extent of the coastal ecosystems. In relation to exposure to potential sea level rise, the low-elevation coastal zone (LECZ) has been used in recent years with reference to specific area and population up to 10 m elevation (Vafeidis et al., 2011).

Coastal systems are conceptualized to consist of both natural and human systems (Figure 5-1). The natural systems include distinct coastal features and ecosystems such as rocky coasts, beaches, barriers and sand dunes, estuaries and lagoons, deltas, river mouths, wetlands, and coral reefs. These elements help define the seaward and landward boundaries of the coast. In spite of providing a wide variety of regulating, provisioning, supporting, and cultural services (MEA, 2005), they have

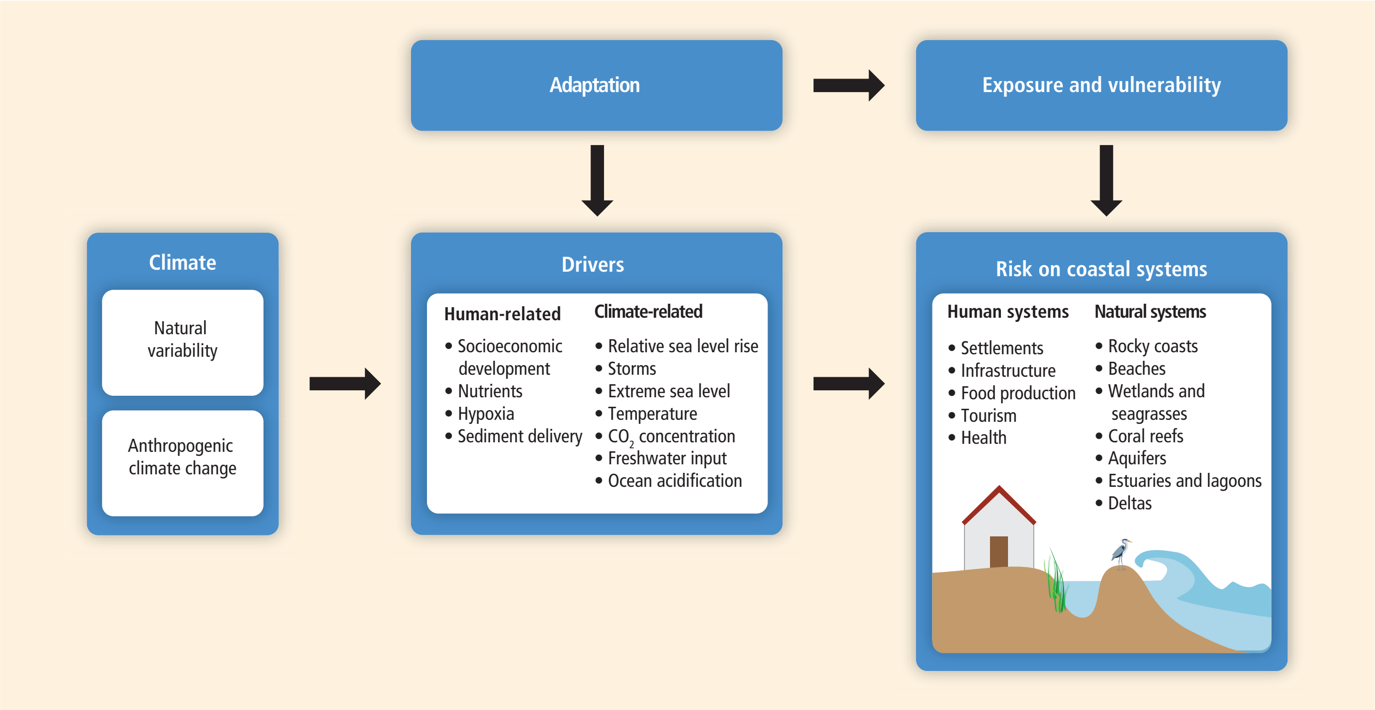

Figure 5-1 | Climate, just as anthropogenic or natural variability, affects both climate and human related drivers. Risk on coastal systems is the outcome of integrating drivers' associated hazards, exposure, and vulnerability. Adaptation options can be implemented either to modify the hazards or exposure and vulnerability, or both.

been altered and heavily influenced by human activities, with climate change constituting only one among many pressures these systems are facing. The human systems include the built environment (e.g., settlements, water, drainage, as well as transportation infrastructure and networks), human activities (e.g., tourism, aquaculture, fisheries), as well as formal and informal institutions that organize human activities (e.g., policies, laws, customs, norms, and culture). The human and natural systems form a tightly coupled socio-ecological system (Berkes and Folke, 1998; Hopkins et al., 2012).

5.3. Drivers

5.3.1. Introduction

In AR4, changes in climate drivers (i.e., any climate-induced factor that directly or indirectly causes a change), including sea level rise, were projected for different Special Report on Emissions Scenarios (SRES) emissions scenarios (IPCC, 2000). Consequently, to date, most of the impacts and vulnerability assessments of climate change in coastal areas are based on SRES A2, A1B, B2, and A1F1 scenarios. Since AR4 a new scenario process has been initiated to replace the SRES scenarios with Representative Concentration Pathways (RCPs) and Shared Socioeconomic Pathways (SSPs) (Moss et al., 2010). The RCPs are scenarios specifying concentrations, rather than emissions, thereby avoiding differences in concentrations of long-lived greenhouse gas (GHG) and aerosol concentrations for the same emissions scenarios that can arise from the use of different models (van Vuuren et al., 2011). For a comparison between RCP and SRES scenarios, see WGI AR5 Box 1.2. In addition, Extended Concentration Pathways (ECPs) have been introduced for the 2100–2300 period (Meinhausen et al., 2011), providing the opportunity to assess the long-term commitment to sea level rise, which is *virtually certain* to continue beyond 2500 unless global temperature declines (WGI AR5 Chapter 1; Section 13.5.2).

The SSPs provide representative qualitative story lines (narratives) of world development together with quantitative pathways of key socioeconomic variables such as gross domestic product (GDP) and population. A preliminary list of five SSPs has been proposed (Arnell et al., 2011; O'Neill et al., 2012), and work to further refine them is ongoing (Kriegler et al. 2012; Van Vuuren et al., 2012). SSPs do not include assumptions on mitigation policy and are thus independent from RCPs in the sense that the same SSP may lead to different concentration levels and consequently rises in sea level depending on the level of mitigation reached (Arnell et al., 2011; O'Neill et al., 2012). Table 5-1 summarizes the main climate-related drivers for the coastal systems.

5.3.2. Relative Sea Level Rise

Assessments of coastal impacts, vulnerability, and adaptation need to consider relative sea level rise (RSLR), which includes climate-induced GMSLR (Section 5.3.2.1) and regional variations (Section 5.3.2.2) as well as local non-climate-related sea level changes (Section 5.3.2.3). RSLR poses a significant threat to coastal systems and low-lying areas around the globe, leading to inundation and erosion of coastlines and contamination of freshwater reserves and food crops (Nicholls, 2010). Sea level rise due to thermal expansion as the oceans warm, together with meltwater from glaciers, icecaps, and ice sheets of Greenland and Antarctica, are the major factors that contribute to RSLR globally. However, regional variations in the rate of rise occur because of ocean circulation patterns and interannual and decadal variability (e.g., Zhang

Table 5-1 | Main climate-related drivers for coastal systems, their trends due to climate change, and their main physical and ecosystem effects.

Climate-related driver	Physical/chemical effects	Trends	Projections	Progress since AR4
Sea level	Submergence, flood damage, erosion; saltwater intrusion; rising water tables/impeded drainage; wetland loss (and change).	Global mean sea level *very likely* increase (Section 5.3.2.2; WGI AR5 Sections 3.7.2, 3.7.3).	Global mean sea level *very likely* increase (see Table 5.1; WGI AR5 Section 13.5.1). Regional variability (Section 5.3.2.2; WGI AR5 Chapter 13).	Improved confidence in contributions to observed sea level. More information on regional and local sea level rise.
Storms: tropical cyclones (TCs), extratropical cyclones (ETCs)	Storm surges and storm waves, coastal flooding, erosion; saltwater intrusion; rising water tables/impeded drainage; wetland loss (and change). Coastal infrastructure damage and flood defense failure.	TCs (Box 5-1, WGI AR5 Section 2.6.3): *low confidence* in trends in frequency and intensity due to limitations in observations and regional variability. ETCs (Section 5.3.3.1; WGI AR5 Section 2.6.4): *likely* poleward movement of circulation features but *low confidence* in intensity changes.	TCs (Box 5-1): *likely* decrease to no change in frequency; *likely* increase in the most intense TCs. ETCs (Section 5.3.3.1): *high confidence* that reduction of ETCs will be small globally. *Low confidence* in changes in intensity.	Lowering of confidence of observed trends in TCs and ETCs since AR4. More basin-specific information on storm track changes.
Winds	Wind waves, storm surges, coastal currents, land coastal infrastructure damage.	*Low confidence* in trends in mean and extreme wind speeds (Section 5.3.3.2, SREX, WGI AR5 Section 3.4.5).	*Low confidence* in projected mean wind speeds. *Likely* increase in TC extreme wind speeds (Section 5.3.3.2, SREX).	Winds not specifically addressed in AR4.
Waves	Coastal erosion, overtopping and coastal flooding.	*Likely* positive trends in Hs in high latitudes (Section 5.3.3.2; WGI AR5 Section 3.4.5).	*Low confidence* for projections overall but *medium confidence* for Southern Ocean increases in Hs (Section 5.3.3.2).	Large increase in number of wave projection studies since AR4.
Extreme sea levels	Coastal flooding erosion, saltwater intrusion.	*High confidence* of increase due to global mean sea level rise (Section 5.3.3.3; WGI AR5 Chapter 13).	*High confidence* of increase due to global mean sea level rise, *low confidence* of changes due to storm changes (Section 5.3.3.3; WGI AR5 Section 13.5).	Local subsidence is an important contribution to regional sea level rise in many locations.
Sea surface temperature (SST)	Changes to stratification and circulation; reduced incidence of sea ice at higher latitudes; increased coral bleaching and mortality, poleward species migration; increased algal blooms.	*High confidence* that coastal SST increase is higher than global SST increase (Section 5.3.3.4).	*High confidence* that coastal SSTs will increase with projected temperature increase (Section 5.3.3.4).	Emerging information on coastal changes in SSTs.
Freshwater input	Altered flood risk in coastal lowlands; altered water quality/salinity; altered fluvial sediment supply; altered circulation and nutrient supply.	*Medium confidence* (*limited evidence*) in a net declining trend in annual volume of freshwater input (Section 5.3.3.6).	*Medium confidence* for general increase in high latitudes and wet tropics and decrease in other tropical regions (Section 5.3.3.6).	Emerging information on freshwater input.
Ocean acidity	Increased CO_2 fertilization; decreased seawater pH and carbonate ion concentration (or "ocean acidification").	*High confidence* of overall increase, with high local and regional variability (Section 5.3.3.5).	*High confidence* of increase at unprecedented rates but with local and regional variability (Box CC-OA).	Coastal ocean acidification not specifically addressed in AR4. Considerable progress made in chemical projections and biological impacts.

SREX = IPCC 2012 Special Report on Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation.

and Church, 2012; Ganachaud et al., 2013) and glacial isostatic rebound and tectonic movement. Subsidence of coastal land from sediment compaction due to building loads, harbor dredging, changes in sediment supply that cause erosion/accretion, and subsurface resource extraction (e.g., groundwater, gas and petroleum; Syvitski et al., 2009) may also contribute to RSLR locally and therefore requires consideration in coastal impact studies. Sea level impacts are most pronounced during episodes of extreme sea levels and these are discussed in Section 5.3.3.

5.3.2.1. Global Mean Sea Level

It is *very likely* that global mean sea level rose at a mean rate of 1.7 [1.5 to 1.9] mm yr^{-1} between 1900 and 2010 and at a rate 3.2 [2.8 to 3.6] mm yr^{-1} from 1993 to 2010 (WGI AR5 Section 13.2.2). Ocean thermal expansion and melting of glaciers have been the largest contributors, accounting for more than 80% of the GMSLR over the latter period (WGI AR5 Section 13.3.1). Future rates of GMSLR during the 21st century are projected to exceed the observed rate for the period 1971–2010 of 2.0 [1.7 to 2.3] mm yr^{-1} for all RCP scenarios (WGI AR5 Table 13.1). Table 5-2 summarizes the *likely* ranges of 21st century GMSLR as established by the Working Group I contribution to this Assessment Report.

From a coastal risk management perspective (Nicholls et al., 2013) assessments of impacts, vulnerabilities, and adaptation have been using GMSLR scenarios above the ranges put forward by WGI reports of AR4 (Meehl et al., 2007; Table 10.7) and AR5 (WGI AR5 Table 13.5). The ranges estimated by WGI of AR4 and AR5 include only those components of GMSLR that can be quantified using process-based models (i.e., models derived from the laws of physics; WGI AR5 Glossary). The ranges given in AR4 thus explicitly excluded contributions to GMSLR resulting from changes in ice flows from the ice sheets of Greenland and Antarctica because at that time process-based models were not able to assess this with sufficient confidence (Meehl et al., 2007; WGI AR5 Section 4.4.5). Since then, understanding has increased and the *likely* range of GMSLR given in AR5 now includes ice sheet flow contributions. *Likely*, however, means that there is still a 0 to 33% probability of GMSLR beyond this range, and coastal risk management needs to consider this. WGI does not assign probabilities to GMSLR beyond the *likely* range, because this cannot be done with the available process-based models. WGI, however, assigns *medium confidence* that 21st century GMSLR does not exceed the likely range by several tenths of a meter (WGI AR5 Section 13.5.1). When using other approaches such as semi-empirical models, evidence from past climates and physical constraints on ice-sheet dynamics GMSLR upper bounds of up to 2.4 m by 2100 have been estimated, but there is *low agreement* on these higher estimates and no consensus on a 21st century upper bound (WGI AR5 Section 13.5.3). Coastal risk management is thus left to choose an upper bound of GMSLR to consider based on which level of risk is judged to be acceptable in the specific case. The Dutch Delta Programme, for example, considered a 21st century GMSLR of 1.3 m as the upper bound.

It is *virtually certain* that sea level rise will continue beyond the 21st century, although projections beyond 2100 are based on fewer and simpler models that include lower resolution coupled climate models for thermal expansion and ice sheet models coupled to climate models to project ice sheet contributions. The basis for the projections are the Extended Concentration Pathways (ECPs), and projections are provided for low, medium, and high scenarios that relate to atmospheric GHG concentrations <500, 500 to 700, and >700 ppm respectively (WGI AR5 Section 13.5.2). Projections of GMSLR up to 2500 are also summarized in Table 5-2.

5.3.2.2. Regional Sea Level

Sea level rise will not be uniform in space and time. Natural modes of climate variability influence sea levels in different regions of the globe and this will affect the rate of rise on interannual and interdecadal time periods. For example, in the equatorial Pacific, sea levels can vary from the global mean by up to 40 cm due to El Niño-Southern Oscillation (ENSO; e.g., Walsh et al., 2012) and this can strongly influence trends on decadal scales. Regional variations in the rate of sea level rise on the coast can arise from climate and ocean dynamic processes such as changes in winds and air pressure, air-sea heat and freshwater fluxes, and ocean currents and their steric properties (Timmermann et al., 2010; WGI AR5 FAQ 13.1). Although the vast majority of coastlines are experiencing sea level rise, coastlines near current and former glaciers and ice sheets are experiencing relative sea level fall (Milne et al., 2009; WGI AR5 FAQ 13.1). This is because the gravitational attraction of the ice sheet decreases as it melts and exerts less pull on the oceans and also because the land tends to rise as the ice melts, the shape of the sea floor changes under the reduced load of the ice sheets, and the change in mass distribution alters the Earth's rotation (WGI AR5 FAQ 13.1; Gomez et al., 2010). In terms of absolute sea level change, approximately 70% of the global coastlines are projected to experience sea level change that is within 20% of the global mean sea level change (WGI AR5 Section 13.6.5).

5.3.2.3. Local Sea Level

Besides the effect of long-term vertical land movement on regional sea level, RSLR can occur locally due to subsidence or uplifts of coastal plains as well as due to other natural causes. Natural subsidence can occur because of sediment compaction and loading, as in the Mississippi River, and other deltas (Törnqvist et al., 2008; Dokka, 2011; Marriner et al., 2012). Tectonic movements, both sustained and abrupt, have brought about relative sea level changes. The Great East Japan Earthquake in 2011 caused subsidence of up to 1.2 m of the Pacific coast of northeast Japan (Geospatial Information Authority of Japan, 2011). The Sumatra-Andaman earthquake in 2004 and subsequent earthquakes in 2005 produced vertical deformation ranging from uplift of 3 m to subsidence of 1 m (Briggs et al., 2006). These movements are especially important in coastal zones located near active plate margins.

Anthropogenic causes of RSLR include sediment consolidation from building loads, reduced sediment delivery to the coast, and extraction of subsurface resources such as gas, petroleum, and groundwater. Subsidence rates may also be sensitive to the rates of oil and gas removal (e.g., Kolker et al., 2011). Syvitski et al. (2009) estimate that the majority of the world's largest deltas are currently subsiding at rates that are considerably larger than the current rates of sea level rise because of coastal sediment starvation due to substantial dam building over the 20th century or sediment compaction through natural or anthropogenic activities. Many large cities on deltas and coastal plains have subsided during the last 100 years: ~4.4 m in eastern Tokyo, ~3 m in the Po delta, ~2.6 m in Shanghai, and ~1.6 m in Bangkok (Syvitski et al., 2009; Teatini et al., 2011). Loads from massive buildings and other large structures can also increase sediment compaction and subsidence (Mazzotti et al., 2009). RSLR can exceed GMSLR by an order of magnitude, reaching more than 10 cm yr^{-1}, and it is estimated that the delta surface

Table 5-2 | Projections of global mean sea level rise in meters relative to 1986–2005 are based on ocean thermal expansion calculated from climate models, the contributions from glaciers, Greenland and Antarctica from surface mass balance calculations using climate model temperature projections, the range of the contribution from Greenland and Antarctica due to dynamical processes, and the terrestrial contribution to sea levels, estimated from available studies. For sea levels up to and including 2100, the central values and the 5–95% range are given whereas for projections from 2200 onwards, the range represents the model spread due to the small number of model projections available and the high scenario includes projections based on RCP6.0 and RCP8.5. Source: WGI AR5 Summary for Policymakers and Sections 12.4.1, 13.5.1, and 13.5.4.

Emission scenario	Representative Concentration Pathway (RCP)	2100 CO_2 concentration (ppm)	Mean sea level rise (m)		Emission scenario	Mean sea level rise (m)		
			2046–2065	2100		2200	2300	2500
Low	2.6	421	0.24 [0.17–0.32]	0.44 [0.28–0.61]	Low	0.35–0.72	0.41–0.85	0.50–1.02
Medium low	4.5	538	0.26 [0.19–0.33]	0.53 [0.36–0.71]	Medium	0.26–1.09	0.27–1.51	0.18–2.32
Medium high	6.0	670	0.25 [0.18–0.32]	0.55 [0.38–0.73]	High	0.58–2.03	0.92–3.59	1.51–6.63
High	8.5	936	0.29 [0.22–0.38]	0.74 [0.52–0.98]				

area vulnerable to flooding could increase by 50% for 33 deltas around the world under the sea level rise as projected for 2100 by the IPCC AR4 (Syvitski et al., 2009).

Clearly large regional variations in the projected sea level rise, together with local factors such as subsidence, indicates that RSLR can be much larger than projected GMSLR and therefore is an important consideration in impact assessments (*very high confidence*).

5.3.3. Climate-Related Drivers

Increasing GHGs in the atmosphere produce changes in the climate system on a range of time scales that impact the coastal physical environment. On shorter time scales, physical coastal impacts such as inundation, erosion, and coastal flooding arise from severe storm-induced surges, wave overtopping, and rainfall runoff. On longer time scales, wind and wave climate change can cause changes in sediment transport at the coast and associated changes in erosion or accretion. Natural modes of climate variability, which can affect severe storm behavior and wind and wave climate, may also undergo anthropogenic changes in the future. Ocean and atmospheric temperature change can affect species distribution with impacts on coastal biodiversity. Carbon dioxide (CO_2) uptake in the ocean increases ocean acidity and reduces the saturation state of carbonate minerals, essential for shell and skeletal formation in many coastal species. Changes in freshwater input can alter coastal ocean salinity concentrations. Past and future changes to these physical drivers are discussed in this section (see also Table 5-1).

5.3.3.1. Severe Storms

Severe storms such as tropical and extratropical cyclones (ETCs) can generate storm surges over coastal seas. The severity of these depends on the storm track, regional bathymetry, nearshore hydrodynamics, and the contribution from waves. Globally there is *low confidence* regarding changes in tropical cyclone activity over the 20th century owing to changes in observational capabilities, although it is *virtually certain* that there has been an increase in the frequency and intensity of the strongest tropical cyclones in the North Atlantic since the 1970s (WGI AR5 Section 2.6). In the future, it is *likely* that the frequency of tropical cyclones globally will either decrease or remain unchanged, but there will be a *likely* increase in global mean tropical cyclone precipitation rates and maximum wind speed (WGI AR5 Section 14.6).

ETCs occur throughout the mid-latitudes of both hemispheres, and their development is linked to large-scale circulation patterns. Assessment of changes in these circulation features reveals a widening of the tropical belt, poleward shift of storm tracks and jet streams, and contraction of the polar vortex; this leads to the assessment that it is *likely* that, in a zonal mean sense, circulation features have moved poleward (WGI AR5 Sections 2.7.5 to 2.7.8) but there is *low confidence* regarding regional changes in intensity of ETCs (e.g., Seneviratne et al., 2012). With regard to future changes, a small poleward shift is *likely* in the Southern Hemisphere but changes in the Northern Hemisphere are basin specific and of *lower confidence* (WGI AR5 Section 14.6.3). Globally, it is *unlikely* that the number of ETCs will fall by more than a few percent due to anthropogenic climate change (*high confidence*; WGI AR5 Section 14.6.3).

5.3.3.2. Extreme Sea Levels

Extreme sea levels are those that arise from combinations of factors including astronomical tides, storm surges, wind waves and swell, and interannual variability in sea levels. Storm surges are caused by the falling atmospheric pressures and surface wind stress associated with storms such as tropical and ETCs and therefore may change if storms are affected by climate change. To date, however, observed trends in extreme sea levels are mainly consistent with mean sea level (MSL) trends (e.g., Marcos et al., 2009; Haigh et al., 2010; Menendez and Woodworth, 2010; Losada et al., 2013) indicating that MSL trends rather than changes in weather patterns are responsible.

Assuming that sea level extremes follow a simple extreme value distribution (i.e., a Gumbel distribution), and accounting for the uncertainty in projections of future sea level rise, Hunter (2012) has developed a technique for estimating a sea level allowance, that is, the minimum height that structures would need to be raised in a future period so that the number of exceedances of that height remains the same as under present climate conditions (Figure 5-2). Such an allowance can be factored into adaptive responses to rising sea levels. It should be noted, however, that extreme sea level distributions might not follow a simple Gumbel distribution (e.g., Tebaldi et al., 2012) owing to different factors influencing extreme levels that may not be measured by tide gauges (e.g., Hoeke et al., 2013).

Regarding future changes to storm surges, hydrodynamic models forced by climate models have been used in several extratropical regional studies such as the northeast Atlantic (e.g., Debenard and Roed, 2008; Wang et al., 2008; Sterl et al., 2009) and southern Australia (Colberg and McInnes, 2012). These studies show strong regional variability and sensitivity to the choice of Global Climate Model (GCM) or Regional Climate Model (RCM). The effect of future tropical cyclone changes on storm surges has also been investigated in a number of regions using a range of different methods. These include methods to stochastically generate and/or perturb cyclones within background environmental conditions that represent historical (e.g., Harper et al., 2009) and GCM-represented future conditions (e.g., Mousavi et al., 2011; Lin et al., 2012). Regional studies include Australia's tropical east coast (Harper et al., 2009), Louisiana (Smith et al., 2010), Gulf of Mexico (Mousavi et al., 2011), India (Unnikrishnan et al., 2011), and New York (Lin et al., 2012), and the details of the methods and findings vary considerably between the studies. While some studies indicate for some regions increase to extreme sea levels due to changes in storms, others indicate the opposite. In general, the small number of regional storm surge studies together with the different atmospheric forcing factors and modeling approaches means that there is *low confidence* in projections of storm surges due to changes in storm characteristics. However, observed upward trends in MSL together with projected increases for 2100 and beyond indicate that coastal systems and low-lying areas will increasingly experience extreme sea levels and their adverse impacts (*high confidence*) (see also WGI AR5 Section 13.7).

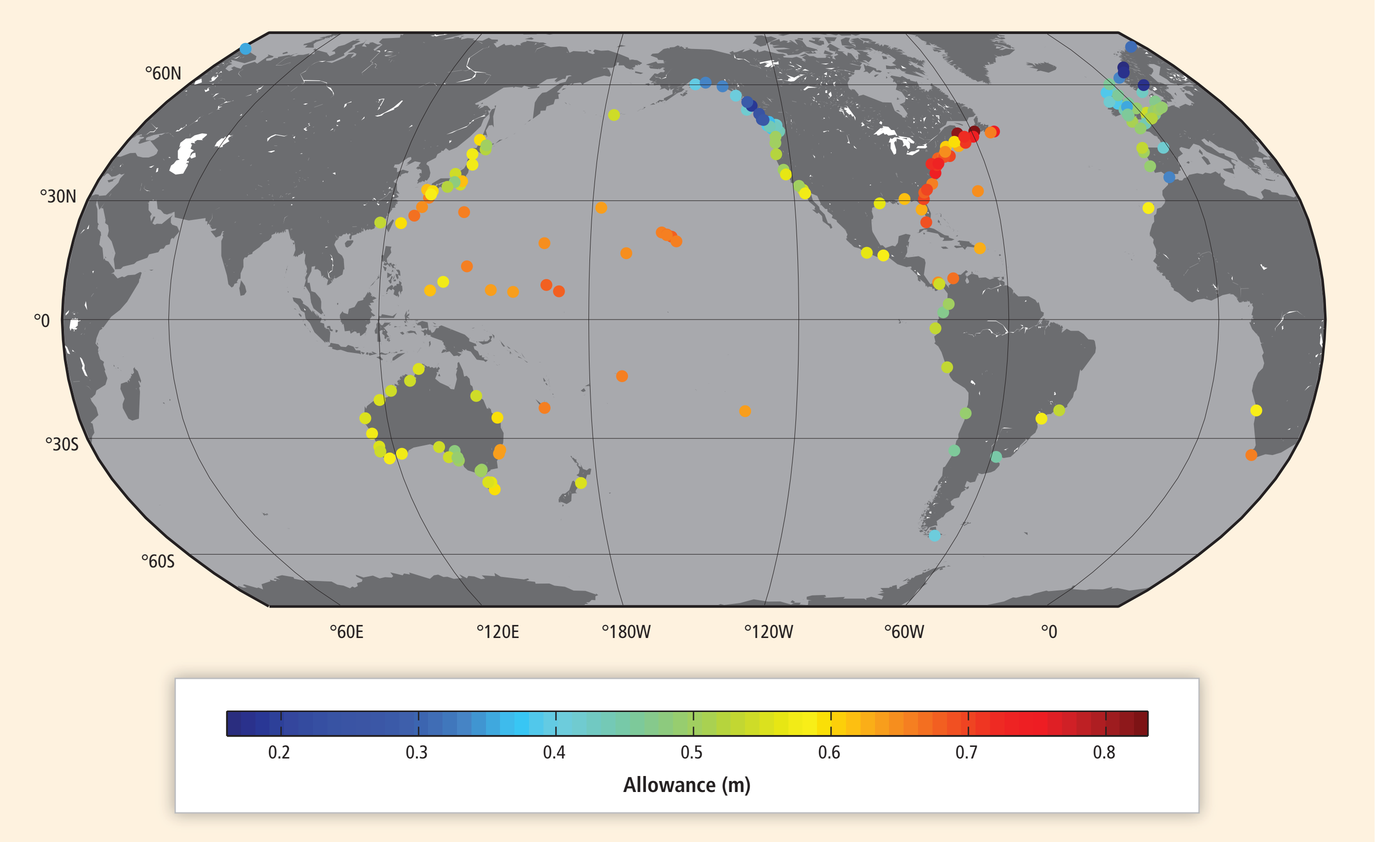

Figure 5-2 | The estimated increase in height (m) that flood protection structures would need to be raised in the 2081–2100 period to preserve the same frequency of exceedances that was experienced for the 1986–2005 period, shown for 182 tide gauge locations and assuming regionally varying relative sea level rise projections under an Representative Concentration Pathway 4.5 (RCP4.5) scenario (adapted from Hunter et al., 2013).

5.3.3.3. Winds and Waves

Changes in wind climate affect large-scale wave climate. Winds also influence longshore current regimes and hence upwelling systems (Narayan et al., 2010; Miranda et al., 2012; see also Sections 6.3.3, 6.3.5). Energy dissipation via wave breaking contributes to longshore and cross-shore currents, elevated coastal sea levels through wave set-up, and run-up and beach erosion. Changes to wind and wave climate therefore can affect sediment dynamics and shoreline processes (e.g., Aargaard et al., 2004; Reguero et al., 2013), and extreme winds and waves are a threat to coastal populations. The coastal impacts of wave climate change are also a function of wave direction and period as well as the coastline itself, which can influence shoaling and refraction. Long period swell, which dominates the wave energy field, poses a significant danger to coastal and offshore structures and shipping (e.g., Semedo et al., 2011) and can causes significant flooding of coastlines with steep shelf margins (Hoeke et al., 2013).

There is *low confidence* in trends calculated from measurements of mean and extreme winds and their causes due to the limited length of records and uncertainties associated with different wind measurement techniques (Seneviratne et al., 2012). However, there is increasing evidence for a strengthening wind stress field in the Southern Ocean since the early 1980s from atmospheric reanalyses, satellite observations, and island station data (WGI AR5 Section 3.4.5). Positive trends in wave height have been detected in the Northeast Atlantic over the 1958–2002 period based on reanalyses and ship observations and in the Southern Ocean between 1985 and 2008 based on satellite data (*medium confidence*) (WGI AR5 Section 3.4.6; see Table 5-2).

Projected changes in mean and extreme winds and waves were assigned *low confidence* (Seneviratne et al., 2012) owing to limited studies. Although there has been an increase in studies addressing future wave climate change (Hemer et al., 2013), generally *low confidence* remains in projected wave climate change (except for *medium confidence* over the Southern Ocean), and this is due to uncertainties in future winds, particularly those associated with storms (see WGI AR5 Section 13.7).

5.3.3.4. Sea Surface Temperature

Sea surface temperature (SST) has significantly warmed during the past 30 years along more than 70% of the world's coastlines, with highly heterogeneous rates of change both spatially and seasonally (Lima and Wethey, 2012). The average rate is 0.18 ± 0.16°C per decade and the average change in seasonal timing was -3.3 ± 4.4 days per decade. These values are larger than in the global ocean where the average of change is 0.11 [0.09 to 0.13]°C per decade in the upper 75 m of the ocean during the 1971–2010 period (WGI AR5 Section 3.2.2) and the seasonal shift is -2.3 days per decade (Lima and Wethey, 2012). Extreme events have also been reported. For example, the record high ocean temperatures along the western Australian coast during the austral summer of 2010/2011, with nearshore temperatures peaking at about 5°C above average, were unprecedented (Pearce and Feng, 2013). In summary, positive trends in coastal SSTs are seen on the majority of coastlines, and the rate of rise along coastlines is higher on average than the oceans (*high confidence*). Based on projected temperature increases there is *high confidence* that positive coastal SST trends will continue.

5.3.3.5. Ocean Acidification

Anthropogenic ocean acidification refers to the changes in the carbonate chemistry primarily due to the uptake of atmospheric CO_2 (Box CC-OA). Seawater pH exhibits a much larger spatial and temporal variability in coastal waters compared to open ocean owing to the variable contribution of processes other than CO_2 uptake (Duarte et al., 2013a) such as upwelling intensity (Feely et al., 2008; Box CC-UP), deposition of atmospheric nitrogen and sulfur (Doney et al., 2007), carbonate chemistry of riverine waters (Salisbury et al., 2008; Aufdenkampe et al., 2011), as well as inputs of nutrients and organic matter (Borges, 2011; Cai et al., 2011). For example, pH (NBS scale) ranges from 6 to 9 in 24 estuaries (Borges and Abril, 2011) and short-term (hours to weeks) changes of up to 0.5 pH units are not unusual in coastal ecosystems (Hofmann et al., 2011).

Few high-quality ocean acidification time series exceed 5 years in the coastal ocean (Wootton et al., 2008; Provoost et al., 2010; Waldbusser et al., 2010). Some exhibit considerable differences compared to open ocean stations, illustrating that anthropogenic ocean acidification can be lessened or enhanced by processes such as primary production, respiration, and calcification (Borges and Gypens, 2010; Kleypas et al., 2011).

Under the IS92a CO_2 emission scenario, the global pH (total scale) of coastal waters has been projected to decrease from about 8.16 in the year 1850 to 7.83 in 2100 (Lerman et al., 2011) but with considerable spatial variability. For example, using the same CO_2 emission scenario, Cai et al. (2011) projected an overall decline of pH in the Northern Gulf of Mexico of 0.74 over the same period, a value that is much greater than that of the open ocean (Box CC-OA).

To summarize, seawater pH exhibits considerable temporal and spatial variability in coastal areas compared to open ocean owing to additional natural and human influences (*very high confidence*). Coastal acidification is projected to continue but with large and uncertain regional and local variations (*high confidence*).

5.3.3.6. Freshwater Input

Changes in river runoff arise from changes in climate drivers such as precipitation, complex interactions between changing levels of CO_2, plant physiology, and, consequently, evapotranspiration (e.g., Gedney et al., 2006; Betts et al., 2007) as well as human drivers such as land use change, water withdrawal, dam building, and other engineered modifications to waterways (see more detailed discussion in Chapter 3). An assessment of runoff trends in 925 of the world's largest ocean-reaching rivers, which account for about 73% of global total runoff, indicates that from 1948–2004 statistically significant trends were present in only one-third of the top 200 rivers and, of these, two-thirds exhibited downward trends and one-third upward trends (Dai et al., 2009). While precipitation changes dominate freshwater flows, decreasing trends in river discharges may be further enhanced as a result of human pressures (Dai et al., 2009; Section 3.2.3).

Average annual runoff is generally projected to increase at high latitudes and in the wet tropics and to decrease in most dry tropical regions (Section 3.4.5). Shifts to earlier peak flows are also projected in areas affected by snowmelt (Adam et al, 2009). However, there are some regions where there is considerable uncertainty in the magnitude and direction of change, specifically South Asia and large parts of South America. Both the patterns of change and the uncertainty are largely driven by projected changes in precipitation.

To summarize, there is *medium confidence* (*limited evidence*, *high agreement*) in a net declining trend in freshwater input globally, although large regional variability exists. Trends are dominated by precipitation changes although human pressures on water supply may enhance downward trends (*medium confidence*). Uncertainty in future changes in runoff is linked to precipitation uncertainty. Runoff is generally projected to increase in high latitudes with earlier peak flows and in the wet tropics and decrease in other tropical regions, however, with large uncertainty (*medium confidence*).

5.3.4. Human-Related Drivers

Coastal systems are subject to a wide range of human-related or anthropogenic drivers (e.g., Crain et al., 2009) that interact with climate-related drivers and confound efforts to attribute impacts to climate change. Some of the major terrestrially based human drivers that directly or indirectly cause changes are briefly reviewed. Related drivers in the marine environment are discussed in Sections 6.4 and 30.6.

5.3.4.1. Socioeconomic Development

Socioeconomic development (SED) drives coastal impacts in several ways. SED influences the number of people and the value of assets exposed to coastal hazards. Since AR4, a number of studies have estimated the influence of future sea level rise and associated hazards on coastal population and assets. Although these estimates are subject to uncertainties associated with global elevation and population data sets (Lichter et al., 2011; Mondal and Tatem, 2012), all the studies indicate high and growing exposure of low-lying coastal areas. The Low Elevation Coastal Zone (LECZ) constitutes 2% of the world's land area but contains 10% of the world's population (600 million) and 13% of the world's urban population (360 million), based on year 2000 estimates (McGranahan et al., 2007). About 65% of the world's cities with populations of greater than 5 million are located in the LECZ (McGranahan et al., 2007). The global population exposed to the 1-in-100-year extreme sea level (i.e., the sea level that has a 1% chance of being exceeded every year) has increased by 95% from 1970 to 2010, with about 270 million people and US$13 trillion worth of assets being exposed to the 1-in-100-year extreme sea level in 2010 (Jongman et al., 2012). In 2002, about US$1.9 trillion worth of assets below the 1-in-100-year extreme sea level were concentrated in the following 10 port cities: Miami (USA), New York-Newark (USA), New Orleans (USA), Osaka-Kobe (Japan), Tokyo (Japan), Amsterdam (Netherlands), Rotterdam (Netherlands), Nagoya (Japan), Virginia Beach (USA), and Guangzhou (China) (Hanson et al., 2011). Compared to other regions, Asia exhibits the greatest exposure in terms of population and assets (Jongman et al., 2012).

For many locations, population and assets exposure is growing faster than the national average trends owing to coastward migration, coastal industrialization, and urbanization (e.g., McGranahan et al., 2007; Seto, 2011; Smith, 2011; see also Chapter 8; *high confidence*). Coastal net migration has largely taken place in flood- and cyclone-prone areas, which poses a challenge for adaptation (de Sherbinin et al., 2011). These processes and associated land use changes are driven by a combination of many social, economic, and institutional factors including taxes, subsidies, insurance schemes, aesthetic and recreational attractiveness of the coast, and increased mobility (Bagstad et al., 2007; Palmer et al., 2011). In China, the country with the largest exposed population, urbanization and land reclamation are the major drivers of coastal land use change (Zhu et al., 2012). Although coastal migration is expected to continue in the coming decades, it is difficult to capture this process in global scenarios, as the drivers of migration and urbanization are complex and variable (Black et al., 2011).

SED also influences the capacity to adapt. Poor people living in urban informal settlements, of which there are about 1 billion worldwide, are particularly vulnerable to weather and climate impacts (de Sherbinin et al., 2011; Handmer et al., 2012). The top five nations classified by population in coastal low-lying areas are developing and newly industrialized countries: Bangladesh, China, Vietnam, India, and Indonesia (McGranahan et al., 2007; Bollman et al., 2010; Jongman et al., 2012). SED and associated land reclamation are also major drivers of the destruction of coastal wetlands, which also makes human settlements more vulnerable because wetlands act as natural buffers reducing wave and storm impacts on the coast (e.g., Crain et al., 2009; Shepard et al., 2011; Arkema et al., 2013; Duarte et al., 2013b). Finally, socioeconomic development is expected to exacerbate further a number of human pressures on coastal systems related to nutrient loads, hypoxia, and sediment delivery, which is discussed in the following subsections.

5.3.4.2. Nutrients

Increased river nutrient (nitrogen, phosphorus) loads to coasts in many regions are observed, and simulated by regional and global models (Alexander et al., 2008; Seitzinger et al., 2010). Anthropogenic global loads of dissolved inorganic nutrients (DIN, DIP) are two to three times larger than those of natural sources (Seitzinger et al., 2010), causing coastal ecosystem degradation (Sections 5.3.4.3, 5.4.2.6). Large variations exist in magnitude and relative sources of nutrient loads. Anthropogenic sources are related primarily to fertilizer use in agriculture and fossil fuel emissions (NO_x) (Galloway et al., 2004; Bouwman et al., 2009).
Future trends depend on measures available to optimize nutrient use in crop production and minimize loss to rivers from agriculture (crop, livestock), sewage, and NO_x emissions. In scenarios with little emphasis on nutrient management, global nutrient discharge increases (DIN 29%, DIP 64%) between 2000 and 2050 (Seitzinger et al., 2010). With ambitious nutrient management, global DIN loads decrease slightly and DIP increases (35%). Climate change is projected to change water runoff (Chapter 3) that influences river nutrient loads. Studies of climate change effects related to increased watershed nutrient sources are needed. In summary, nutrient loads have increased in many world regions (*high confidence*); future increases will depend largely on nutrient management practices (*medium confidence*).

5.3.4.3. Hypoxia

The presence of excessive nutrients in coastal waters, which causes eutrophication and the subsequent decomposition of organic matter, is the primary cause of decreased oxygen concentration (hypoxia). Globally, upwelling of low oxygen waters (e.g., Grantham et al., 2004) and ocean warming, which decreases the solubility of oxygen in seawater (Shaffer et al., 2009), are secondary drivers but can be locally important. The oxygen decline rate is greater in coastal waters than in the open ocean (Gilbert et al., 2010). Hypoxia poses a serious threat to marine life, which is exacerbated when combined with elevated temperature (Vaquer-Sunyer and Duarte, 2011; see also Section 6.3.3). The number of so-called "dead zones" has approximately doubled each decade since 1960 (Diaz and Rosenberg, 2008). Fishery catches from these areas are generally lower than predicted from nutrient loading alone (Breitburg et al., 2009). Although non-climate anthropogenic factors are responsible for virtually all hypoxia in estuaries and inner continental shelves, climate drivers such as ocean warming, altered hydrological cycles, and coastal current shifts and changes in upwellings may interact with eutrophication in the next decades (Rabalais et al., 2010; Meire et al., 2013; *high confidence*).

5.3.4.4. Sediment Delivery

Human activities in drainage basins and coastal plains have impacted the coastal zone by changing the delivery of sediment to the coast. Sediment trapping behind dams, water diversion for irrigation, and sand and gravel mining in river channels all contribute to decrease sediment delivery, whereas soil erosion due to land use changes helps increase it (Syvitski, 2008; Walling, 2006). It is estimated that the global discharge of riverine sediment was 16 to 19 Gt yr^{-1} in the 1950s before widespread dam construction (e.g., Syvitski et al., 2005; Milliman and Farnsworth, 2011) and it has decreased to 12 to 13 Gt yr^{-1} (Syvitski and Kettner, 2011). Out of 145 major rivers with mostly more than 25 years of record, only seven showed evidence of an increase in sediment flux while 68 showed significant downward trends (Walling and Fang, 2003). The number of dams has increased continuously and their distribution has expanded globally. As of early 2011, the world has an estimated 16.7 million reservoirs larger than 0.01 ha (Lehner et al., 2011). Globally, 34 rivers with drainage basins of 19 million km^2 in total show a 75% reduction in sediment discharge over the past 50 years (Milliman and Farnsworth, 2011). Reservoir trapping of sediments is estimated globally as 3.6 Gt yr^{-1} to more than 5 Gt yr^{-1} (Syvitski et al., 2005; Milliman and Farnsworth, 2011; Walling, 2012). Human pressure is the main driver of the observed declining trend in sediment delivery to the coast (*high agreement*).

5.4. Impacts, Vulnerabilities, and Risks

5.4.1. Introduction

This subsection briefly introduces the diverse approaches and methods applied in the literature on coastal impact, vulnerability, and risk. The following subsections then assess this literature related to coastal natural systems (Section 5.4.2) and coastal human systems (Section 5.4.3). Much of this literature focuses on RSLR and extreme sea level events as the main drivers. The main biophysical impacts of this driver are increasing flood damage, dry-land loss due to submergence and erosion, wetland loss and change, saltwater intrusion into surface and ground water, and rising water tables and impeded drainage (Table 5-3).

Impacts and risks are assessed using a wide variety of approaches from the local to global scale. Sea level rise exposure approaches are applied at all scales to assess values exposed to sea level rise (e.g., people, assets, ecosystems, or geomorphological units). Submergence exposure approaches assess exposure to permanent inundation under a given sea level rise (e.g., Dasgupta et al., 2009; Boateng, 2012) whereas flood exposure approaches assess exposure to temporary inundation during a coastal flood event by combining the extreme water level of the flood event with a given level of sea level rise (e.g., Dasgupta et al., 2011; Kebede and Nicholls, 2012).

Indicator-based approaches are also used at all scales to aggregate data on the current state of the coastal systems into vulnerability indices (Gornitz, 1991; Hinkel, 2011), based on either biophysical exposure or hazard variables (e.g., Bosom and Jimenez, 2011; Yin et al., 2012), socioeconomic variables representing a social group's capacity to adapt (e.g., Cinner et al., 2012), or both kinds of variables (e.g., Bjarnadottir et al., 2011; Li and Li, 2011; Yoo et al., 2011).

At local scales (<100 km coastal length), process-based models are applied to assess flooding, erosion, and wetland impacts. Approaches include assessments of flood damage of single extreme water level events using numerical inundation models (e.g., Lewis et al., 2011; Xia et al., 2011). Erosion impacts are assessed using either numerical morphodynamic models (e.g., Jiménez et al., 2009; Ranasinghe et al., 2012) or simple geometric profile relationships such as the Bruun Rule (Bruun, 1962). For ecosystem impacts ecological landscape simulation models are used to predict habitat change due to sea level rise and other factors (e.g., Costanza et al., 1990).

At regional to global scales, numerical process-based models are not available for assessing the impacts of RSLR and extreme sea level events due to data and computational limits. Global scale assessments of coastal impacts have been conducted with the models Climate Framework for Uncertainty, Negotiation and Distribution (FUND) and Dynamic and Interactive Coastal Vulnerability Assessment (DIVA). FUND is an integrated assessment model with a coastal impact component that includes country-level cost functions for dry-land loss, wetland loss, forced migration, and dike construction (Tol, 2002). DIVA is a dedicated coastal impact model employing subnational coastal data (Vafeidis et al., 2008) and considering additional impacts such as coastal flooding and erosion as well as adaptation in terms of protection via dikes and nourishment (Hinkel and Klein, 2009). DIVA assesses coastal flood risk based on hydrologically connected elevation and extreme water level distributions

Frequently Asked Questions

FAQ 5.1 | How does climate change affect coastal marine ecosystems?

The major climate-related drivers on marine coastal ecosystems are sea level rise, ocean warming, and ocean acidification.

Rising sea level impacts marine ecosystems by drowning some plants and animals as well as by inducing changes of parameters such as available light, salinity, and temperature. The impact of sea level is related mostly to the capacity of animals (e.g., corals) and plants (e.g., mangroves) to keep up with the vertical rise of the sea. Mangroves and coastal wetlands can be sensitive to these shifts and could leak some of their stored compounds, adding to the atmospheric supply of these greenhouse gases.

Warmer temperatures have direct impacts on species adjusted to specific and sometimes narrow temperature ranges. They raise the metabolism of species exposed to the higher temperatures and can be fatal to those already living at the upper end of their temperature range. Warmer temperatures cause coral bleaching, which weakens those animals and makes them vulnerable to mortality. The geographical distribution of many species of marine plants and animals shifts towards the poles in response to warmer temperatures.

When atmospheric carbon dioxide is absorbed into the ocean, it reacts to produce carbonic acid, which increases the acidity of seawater and diminishes the amount of a key building block (carbonate) used by marine 'calcifiers' such as shellfish and corals to make their shells and skeletons and may ultimately weaken or dissolve them. Ocean acidification has a number of other impacts, many of which are still poorly understood.

Table 5-3 | Main impacts of relative sea level rise. Source: Adapted from Nicholls et al. (2010).

Biophysical impacts of relative sea level rise	Other climate-related drivers	Other human drivers
Dryland loss due to erosion	Sediment supply, wave and storm climate	Activities altering sediment supply (e.g., sand mining)
Dryland loss due to submergence	Wave and storm climate, morphological change, sediment supply	Sediment supply, flood management, morphological change, land claim
Wetland loss and change	Sediment supply, CO_2 fertilization	Sediment supply, migration space, direct destruction
Increased flood damage through extreme sea level events (storm surges, tropical cyclones, etc.)	Wave and storm climate, morphological change, sediment supply	Sediment supply, flood management, morphological change, land claim
Saltwater intrusion into surface waters (backwater effect)	Runoff	Catchment management and land use (e.g., sand mining and dretching)
Saltwater intrusion into groundwaters leading to rising water tables and impeded drainage	Precipitation	Land use, aquifer use

(Hinkel et al., 2013) and erosion based on a combination of the Bruun Rule and a simplified version of the Aggregated Scale Morphological Interaction between a Tidal inlet and the Adjacent coast (ASMITA) model for tidal basins (Nicholls et al., 2011). The results of these models are discussed in Sections 5.4.3.1 and 5.5.5.

For impacts on natural systems, the key climate-related drivers considered are temperature, ocean acidification, and sea level. A variety of approaches are applied including field observations of ecosystem features (e.g., biodiversity, reproduction) and functioning (e.g., calcification, primary production), remote sensing (e.g., extent of coral bleaching, surface area of vegetated habitats), and perturbation experiments in the laboratory and in the field.

5.4.2. Natural Systems

Coastal ecosystems are experiencing large cumulative impacts related to human activities (Halpern et al., 2008) arising from both land- and ocean-based anthropogenic drivers. Anthropogenic drivers associated with global climate change are distributed widely and are an important component of cumulative impacts experienced by coastal ecosystems. There is no wetland, mangrove, estuary, rocky shore, or coral reef that is not exhibiting some degree of impact. Overexploitation and habitat destruction are often the primary causes of historical changes in coastal systems leading to declines in diversity, structure, and functioning (Lotze et al., 2006). Further, extreme climate events generate changes to both the mean and the variance of climatic variables over ecological time scales.

5.4.2.1. Beaches, Barriers, and Sand Dunes

Beaches, barriers, and sand dunes are about half as common as rocky coasts (Bird, 2000; Davis and FitzGerald, 2004) and often exhibit distinct and seasonal changes. Owing to their aesthetic qualities, they are highly valued for recreation and residences.

5.4.2.1.1. Observed impacts

Globally, beaches and dunes have in general undergone net erosion over the past century or longer (e.g., for an overview, see Bird, 2000). A number of studies have investigated shoreline change by comparing historical maps and imagery, available since about the mid-19th century with more recent maps and imagery to quantify combined climate and non-climate changes. For example, along the U.S. Mid-Atlantic and New England coasts the long-term rate of erosion, based on 21,184 transects equally spaced along more than 1000 km of coast, is 0.5 ± 0.09 m yr^{-1}, with 65% of transects showing net erosion (Hapke et al., 2011). A similar study by Webb and Kench (2010) in the central Pacific utilized historical aerial photographs and satellite images to show physical changes in 27 islets located in four atolls over a 19- to 61-year period. The analysis highlighted the dynamic nature of sea level rise response in the recent past, with physical changes in shoreline progradation and displacement influencing whether the island area increased (46%), remained stable (46%), or decreased (14%).

Attributing shoreline changes to climate change is still difficult owing to the multiple natural and anthropogenic drivers contributing to coastal erosion. For example, rotation of pocket beaches (i.e., where one end of the beach accretes while the other erodes and then the pattern reverses) in southeast Australia is closely related to interannual changes in swell direction (Harley et al., 2010). Additional processes, unrelated to climate change, that contribute to coastal change include dams capturing fluvial sand (e.g., in Morocco; Chaibi and Sedrati, 2009). Statistically linking sea level rise to observed magnitudes of beach erosion has had some success, although the coastal sea level change signal is often small when compared to other processes (e.g., Leatherman et al., 2000a,b; Sallenger et al., 2000; Zhang et al., 2004). A Bayesian network incorporating a variety of factors affecting coastal change, including RSLR, has been successful in hindcasting shoreline change, and can be used to evaluate the probability of future shoreline change (Gutierrez et al., 2011).

While some coastal systems may be able to undergo landward retreat under rising sea levels, others will experience coastal squeeze, which occurs when an eroding shoreline approaches hard, immobile structures such as seawalls or resistant natural cliffs. In these instances the beaches will narrow owing to the resulting sediment deficit and produce adverse impacts such as habitat destruction, impacting the survivability of a variety of organisms (Jackson and McIlvenny, 2011). With such a manifestation of coastal squeeze, sand dunes will ultimately be removed as the beach erodes and narrows. Extreme storms can erode and completely remove dunes, degrading land elevations and exposing them to inundation and further change if recovery does not occur before the next storm (Plant et al., 2010). Even in the absence of hard obstructions, barrier island erosion and narrowing can occur, as a result of rising sea level and recurrent storms, as in the Chandeleur Islands and Isles Dernieres, Louisiana, USA (Penland et al., 2005).

5.4.2.1.2. Projected impacts

With projected GMSLR (see Section 5.3.3), inundation and erosion may become detectable and progressively important. In the first instance,

Frequently Asked Questions

FAQ 5.2 | How is climate change influencing coastal erosion?

Coastal erosion is influenced by many factors: sea level, currents, winds, and waves (especially during storms, which add energy to these effects). Erosion of river deltas is also influenced by precipitation patterns inland which change patterns of freshwater input, runoff, and sediment delivery from upstream. All of these components of coastal erosion are impacted by climate change.

Based on the simplest model, a rise in mean sea level usually causes the shoreline to recede inland due to coastal erosion. Increasing wave heights can cause coastal sand bars to move away from the shore and out to sea. High storm surges (sea levels raised by storm winds and atmospheric pressure) also tend to move coastal sand offshore. Higher waves and surges increase the probability that coastal sand barriers and dunes will be over-washed or breached. More energetic and/or frequent storms exacerbate all these effects.

Changes in wave direction caused by shifting climate may produce movement of sand and sediment to different places on the shore, changing subsequent patterns of erosion.

the impacts will be apparent through sea level rise which, combined with storm surge, will make extreme water levels higher and more frequent and therefore enable greater attack on beaches and dunes (Tebaldi et al., 2012).

The Bruun rule (a simple rule based on the assumption that to maintain an equilibrium cross-shore profile under rising sea levels, the coastline will move landwards a distance of approximately 100 times the vertical sea level rise; Bruun, 1962) has been used by many researchers to calculate erosion by sea level rise. However, there is disagreement about whether the Bruun rule is appropriate (Cooper and Pilkey, 2004; Woodroffe and Murray-Wallace, 2012), and how to calculate the amount of retreat remains controversial (Gutierrez et al., 2011; Ranasinghe et al., 2012). An increase in storm intensity and ocean swell may accelerate erosion of beaches, barriers, and dunes, although in some places beach response to sea level rise could be more complex than just a simple retreat (Irish et al., 2010).

Coastal squeeze is expected to accelerate with a rising sea level. In many locations, finding sufficient sand to rebuild beaches and dunes artificially will become increasingly difficult and expensive as present supplies near project sites are depleted (*high confidence*). New generation models are emerging to estimate the costs of saving oceanfront homes through beach nourishment relative to the structures cost (McNamara et al., 2011). In the absence of adaptation measures, beaches and sand dunes currently affected by erosion will continue to be affected under increasing sea levels (*high confidence*).

5.4.2.2. Rocky Coasts

Rocky coasts with shore platforms form about three-fourths of the world's coasts (Davis and FitzGerald, 2004; Jackson and McIlvenny, 2011) and are characterized by very strong environmental gradients, especially in the intertidal zone where both marine and atmospheric climate regime changes can pose challenges.

5.4.2.2.1. Observed impacts

Cliffs and platforms are erosional features and any change that increases the efficiency of processes acting on them, such as RSLR, storminess, wave energy, and weathering regimes, increases erosion (Naylor et al., 2010). Their responses vary, owing to different lithology (e.g., hard rock vs. non-lithified soft rock) and profiles (e.g., plunging cliffs or cliffs with shore platforms). Cliffs and platforms have reduced resilience to climate change impacts; once platforms are lowered or cliffs have retreated, it is difficult to rebuild them (Naylor et al., 2010). On the decadal scale, for example, the retreat of soft rock cliffs in East Anglia, UK, has been linked to the North Atlantic Oscillation (NAO) phases with high energetics (Brooks and Spencer, 2013).

Changes in the abundance and distribution of rocky shore animals and algae have long been recognized (Hawkins et al., 2008), and perturbation experiments provide information about environmental limits, acclimation, and adaptation, particularly to changes in temperature (Somero, 2012). The challenge is to attribute the changes to climate-related drivers, human-related drivers, and to natural fluctuations.

The range limits of many intertidal species have shifted by up to 50 km per decade over the past 30 years in the North Pacific and North Atlantic, much faster than most recorded shifts of terrestrial species (Helmuth et al., 2006; Box CC-MB). However, the distribution of some species has not changed in recent decades, which may be due to weak local warming (Rivadeneira and Fernández, 2005) or overriding effects of variables such as timing of low tide; hydrographic features; lack of suitable substrate; poor larval dispersal; and effects of food supply, predation, and competition (Helmuth et al., 2002, 2006; Poloczanska et al., 2011).

The dramatic decline of biodiversity in mussel beds of the Californian coast has been attributed to large-scale processes associated with climate-related drivers (59% mean loss in species richness, comparing 2002 to historical data (1960s to 1970s); Smith et al., 2006) (*high* *confidence*). Warming reduced predator-free space on rocky shores, leading to a decrease of the vertical extent of mussel beds by 51% in 52 years in the Salish Sea, and to the disappearance of reproductive populations of mussels (Harley, 2011). Unusually high air or water temperature led to mass mortalities, for example, of mussels on the California coast (Harley, 2008) and gorgonians in the northwestern Mediterranean (Garrabou et al., 2009).

Rocky shores are one of the few ecosystems for which field evidence of the effects of ocean acidification is available. Observational and modeling analysis have shown that the community structure of a site of the northeast Pacific shifted from a mussel to an algal-barnacle dominated community between 2000 and 2008 (Wootton et al., 2008), in relation with rapidly declining pH (Wootton and Pfister, 2012).

5.4.2.2.2. Projected impacts

Modeled relationships suggest that soft-rock recession rates depends on the relative change in sea level rise while cliff retreat depends both on total elevation change of sea level and on the rate of sea level rise (Ashton et al., 2011). In a modeling study, Trenhaile (2010) found sea level rise to trigger faster rates of cliff recession, especially in coasts that are already retreating fast. In addition, based on modeling cliff dynamics with contemporary and historic data of soft cliff retreat along Suffolk Coast, UK, rapid retreat is associated with accelerating sea level rise (Brooks and Spencer, 2013). However, coasts currently retreating slowly would experience the largest proportional increase in retreat rates. Increases in storminess have smaller effects on rocky shores (Dawson et al., 2009; Trenhaile, 2011).

Few projections of the effect of climate change on rocky shores have considered the effects of direct and indirect species interactions (Poloczanska et al., 2008; Harley, 2011) and the effects of multiple drivers (Helmuth et al., 2006). The abundance and distribution of rocky shore species will continue to change in a warming world (*high confidence*). For example, the long-term consequences of ocean warming on mussel beds of the northeast Pacific are both positive (increased growth) and negative (increased susceptibility to stress and of exposure to predation) (Smith et al., 2006; Menge et al., 2008; *medium confidence*). Extrapolations of ecosystem change based on temperature-focused studies alone are likely to be conservative, as hypoxia (Grantham et al., 2004) or ocean acidification (Feely et al., 2008) are also known to occur in this region.

Observations performed near natural CO_2 vents in the Mediterranean Sea show that diversity, biomass, and trophic complexity of rocky shore communities will decrease at future pH levels (Barry et al., 2011; Kroeker et al., 2011; *high confidence*). An abundant food supply appears to enable mussels of the Baltic Sea to tolerate low pH (Thomsen et al., 2010, 2013) at the cost of increased energy expenditure. Model projections that include the interactive effects of ocean warming and acidification suggest that a population of barnacle of the English Channel will become extinct 10 years earlier than it would with warming alone (Findlay et al., 2010; *medium confidence*). Ocean acidification may also exacerbate mass mortality events in the Mediterranean Sea (Rodolfo-Metalpa et al., 2011; *limited evidence*, *medium agreement*).

In summary, rocky shores are among the better-understood coastal ecosystems in terms of potential impacts of climate variability and change. The most prominent effects are range shifts of species in response to ocean warming (*high confidence*) and changes in species distribution and abundance (*high confidence*) mostly in relation to ocean warming and acidification.

5.4.2.3. Wetlands and Seagrass Beds

Vegetated coastal habitats and coastal wetlands (mangrove forests, salt marshes, seagrass meadows, and macroalgal beds) extend from the intertidal to the subtidal areas in coastal areas, where they form key ecosystems.

5.4.2.3.1. Observed impacts

Vegetated coastal habitats are declining globally (Duarte et al., 2005), rendering shorelines more vulnerable to erosion due to increased sea level rise and increased wave action (e.g., Alongi, 2008) and leading to the loss of carbon stored in sediments. Together, the loss of coastal wetlands and seagrass meadows results in the release of 0.04 to 0.28 PgC annually from organic deposits (Pendleton et al., 2012). Recognition of the important consequences of the losses of these habitats for coastal protection and carbon burial (Duarte et al., 2013a) has led to large-scale reforestation efforts in some nations (e.g., Thailand, India, Vietnam).

The response of saltmarshes to sea level rise involves landward migration of salt-marsh vegetation zones, submergence at lower elevations, and drowning of interior marshes. Ocean warming is leading to range shifts in vegetated coastal habitats. The poleward limit of mangrove forests is generally set by the 20°C mean winter isotherm (Duke et al., 1998). Accordingly, migration of the isotherm with climate change (Burrows et al., 2011) should lead to a poleward expansion of mangrove forests, as observed in the Gulf of Mexico (Perry and Mendelssohn, 2009; Comeaux et al., 2011; Raabe et al., 2012) and New Zealand (Stokes et al., 2010), leading to increased sediment accretion (*medium confidence*).

Seagrass meadows are already under stress due to climate change (*high confidence*), particularly where maximum temperatures already approach their physiological limit. Heat waves lead to widespread seagrass mortality, as documented for *Zostera* species in the Atlantic (Reusch et al., 2005) and *Posidonia* meadows in the Mediterranean Sea (Marbà and Duarte, 2010) and Australia (Rasheed and Unsworth, 2011; *high confidence*). Warming also favors flowering of *P. oceanica* (Diaz-Almela et al., 2007), but the increased recruitment rate is insufficient to compensate for the losses resulting from elevated temperatures (Diaz-Almela et al., 2009).

Kelp forests have been reported to decline in temperate areas in both hemispheres (Fernández, 2011; Johnson et al., 2011; Wernberg et al., 2011a,b), a loss involving climate change (*high confidence*). Decline in kelp populations attributed to ocean warming has been reported in southern Australia (Johnson et al., 2011; Wernberg et al., 2011a,b) and the North Coast of Spain (Fernández, 2011). The spread of subtropical invasive macroalgal species may be facilitated by climate change, adding to the stresses experienced by temperate seagrass meadows due to ocean warming (*medium evidence*, *high agreement*).

5.4.2.3.2. Projected impacts

Ocean acidification (Section 5.3.3.5; Box CC-OA) is expected to enhance the production of seagrass, macroalgae, salt-marsh plants, and mangrove trees through the fertilization effect of CO_2 (Hemminga and Duarte, 2000; Wu et al., 2008; McKee et al., 2012; *high confidence*). Increased CO_2 concentrations may have already increased seagrass photosynthetic rates by 20% (Hemminga and Duarte, 2000; Hendriks et al., 2010; *limited evidence, high agreement*).

Coupling of downscaled model projections using the SRES A1B scenario in the western Mediterranean with relationships between mortality rates and maximum seawater temperature led Jordá et al. (2012) to conclude that seagrass meadows may become functionally extinct by 2050–2060 (*high confidence*). Poleward range shifts in vegetated coastal habitats are expected to continue with climate change (*high confidence*).

Although elevated CO_2 and ocean acidification are expected to increase productivity of vegetated coastal habitats in the future, there is *limited evidence* that elevated CO_2 will increase seagrass survival or resistance to warming (Alexandre et al., 2012; Jordá et al., 2012).

Coastal wetlands and seagrass meadows experience coastal squeeze in urbanized coastlines, with no opportunity to migrate inland with rising sea levels. However, increased CO_2 and warming can stimulate marsh elevation gain, counterbalancing moderate increases in sea level rise rates (Langley et al., 2009; Kirwan and Mudd, 2012). Climate change is expected to increase carbon burial rates on salt marshes during the first half of the 21st century, provided sufficient sediment supply, with carbon-climate feedbacks diminishing over time (Kirwan and Mudd, 2012; *medium confidence*).

In summary, climate change will contribute to the continued decline in the extent of seagrasses and kelps in the temperate zone (*medium confidence*) and the range of seagrasses, mangroves, and kelp in the Northern Hemisphere will expand poleward (*high confidence*). The limited positive impact of warming and increased CO_2 on vegetated ecosystems will be insufficient to compensate the decline of their extent resulting from other human drivers such as land use change (*very high confidence*).

5.4.2.4. Coral Reefs

Coral reefs are shallow-water ecosystems made of calcium carbonate secreted by reef-building corals and algae. They are among the most diverse ecosystems and provide key services to humans (Box CC-CR).

5.4.2.4.1. Observed impacts

Mass coral bleaching coincided with positive temperature anomalies over the past 30 years, sometimes followed by mass mortality (Kleypas et al., 2008; *very high confidence*). More than 80% of corals bleached during the 2005 event in the Caribbean and more than 40% died (Eakin et al., 2010). Bleaching events and their recovery are variable in time and space: 7% of the reef locations exhibited at least one bleaching between 1985 and 1994 compared to 38% in the 1995–2004 period, most of which occurred during the 1997–98 El Niño event (Figure 5-3). Recovery from the 1998 global bleaching event was generally variable in the Indian Ocean, absent in the western Atlantic, and no clear trends elsewhere (Baker et al., 2008). Warming has caused a poleward range expansion of some corals (Greenstein and Pandolfi, 2008; Yamano et al., 2011; *high confidence*).

Persistence of coral reefs depends on the balance between the production and erosion of calcium carbonate and on coral settlement, both of which are affected by ocean acidification (Section 5.3.3.5; Box CC-OA). Experimental data show that ocean acidification generally decreases calcification (Andersson et al., 2011; Kroeker et al., 2013) and promotes dissolution of calcium carbonate and bioerosion (Tribollet et al., 2009; Wisshak et al., 2012), leading to poorly cemented reefs (Manzello et al., 2008); it also negatively affects early life history stages, which could reduce the number of larval settlers (Albright, 2011).

Coral cover and calcification have decreased in recent decades (e.g., Gardner et al., 2003; De'ath et al., 2009, 2012; Manzello, 2010; Box CC-CR; *very high confidence*) but attribution to climate-related and human-related drivers is difficult. Globally, the primary climate-related driver appears to be ocean warming rather than ocean acidification, cyclonic activity, and changes in freshwater input (Cooper et al., 2012; De'ath et al., 2012; *medium confidence*). Sea level rise also controls reef growth but, within the uncertainties of past sea level rise and coral reef growth, most coral reefs seem to have kept pace with the recent sea level rise (Buddemeier and Smith, 1988; Brown et al., 2011).

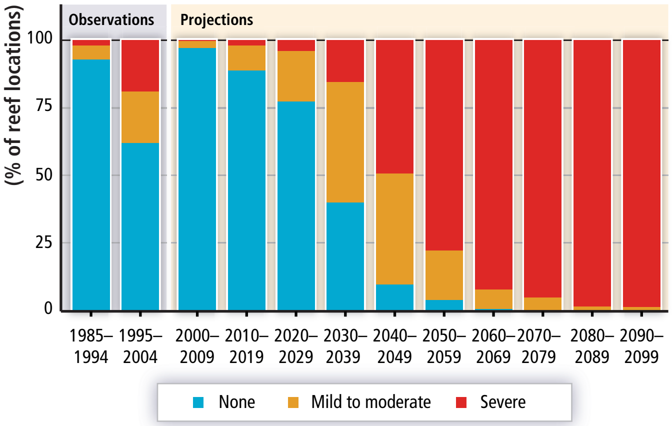

Figure 5-3 | Percent of reef locations (1° × 1° grid cells which have at least one reef) that experience no bleaching, at least one mild bleaching event, or at least one severe bleaching event for each decade. Observed bleaching events are summarized from the ReefBase data set (Kleypas et al., 2008). In the observations, some of the "no bleaching" cells may have experienced bleaching but it was either not observed or not reported. Modeled bleaching events are averages of data from four ensemble runs of the Community Climate System Model version 3 using the Special Report on Emissions Scenarios (SRES) A1B scenario and the standard degree heating month formula (Teneva et al., 2011). The labels of values ≤1% are not shown.

5.4.2.4.2. Projected impacts

Coral bleaching and mortality will increase in frequency and magnitude over the next decades (*very high confidence*). Under the A1B CO_2 emission scenario, 99% of the reef locations will experience at least one severe bleaching event between 2090 and 2099 (Figure 5-3), with *limited evidence* and *low agreement* that coral acclimation and/or adaptation will limit this trend (Logan et al., 2014). The onset of annual bleaching event under RCP8.5 is delayed by more than 2 decades in about 23% of reef locations compared to RCP6.0 (van Hooidonk et al., 2013).

Ocean warming and acidification have synergistic effects in several reef-builders (Reynaud et al., 2003; Anthony et al., 2008). They will increase coral mortality, reduce calcification and the strength of calcified organisms, and enhance skeletal dissolution (Manzello et al., 2008; *high confidence*). Reefs will transition from a condition of net accretion to one of net erosion (Andersson and Gledhill, 2013; *high confidence*) and will be more susceptible to breakage. The onset of global dissolution is at an atmospheric CO_2 of 560 ppm (Silverman et al., 2009; *medium confidence*) and dissolution will be widespread in 2100 (RCP8.5 emission scenario, Dove et al., 2013; *medium confidence*). The observed poleward range extension will be limited by ocean acidification (Yara et al., 2012; Couce et al., 2013) and may be followed by equatorial range retractions (Kiessling et al., 2012).

The maximum rate of vertical accretion has been variable regionally during the last deglaciation (about 20 mm yr^{-1}; Dullo, 2005; Montaggioni, 2005) and has not enabled all coral reefs to keep up with sea level rise. Some reefs kept up, even when the eustatic sea level rise exceeded 40 mm yr^{-1} (Camoin et al., 2012). A number of coral reefs could therefore keep up with the maximum rate of sea level rise of 15.1 mm yr^{-1} projected for the end of the century (WGI AR5 Table 13.5; *medium confidence*) but a lower net accretion than during the Holocene (Perry et al., 2013) and increased turbidity (Storlazzi et al., 2011) will weaken this capability (*very high confidence*).

In summary, ocean warming is the primary cause of mass coral bleaching and mortality (*very high confidence*), which, together with ocean acidification, deteriorates the balance between coral reef construction and erosion (*high confidence*). The magnitude of these effects depends on future rates of warming and acidification (*very high confidence*), with a limited moderating role owing to biological acclimation and adaptation (*medium confidence*).

5.4.2.5. Coastal Aquifers

Coastal aquifers are of strategic importance for the water supply of highly populated coastal areas, especially in small islands (Section 29.3).

5.4.2.5.1. Observed impacts

Temperature and evaporation rise, precipitation changes, and extended droughts affecting aquifer recharge can contribute to saltwater intrusion (Section 3.2.4). Rising sea levels and overwash from waves or storm surge are also relevant, especially in low-lying areas and islands (Terry and Falkland, 2010; White and Falkland, 2010; see also Section 29.3).

Aquifers on the coasts of the USA have experienced increased levels of salinity largely due to excessive water extraction (Barlow and Reichard, 2010). Natural drivers combined with over-extraction, pollution, mining, and erosion compound groundwater supply problems in small islands in the Pacific, Indian, and Atlantic Oceans (White et al., 2007; White and Falkland, 2010). This increased usage of groundwater resources globally has, over the last century, led to a reduction in groundwater quality, including increased salinization (*very high confidence*).

Attribution of saline intrusion to incremental sea level rise is still not sufficiently supported (Rozell and Wong, 2010; White and Falkland 2010). In small islands, observed saltwater intrusion due to flooding and overwash under storm events cannot be attributed to climate change (Section 29.3.2; *limited evidence*, *high agreement*).

5.4.2.5.2. Projected impacts

Available information on projected impacts on coastal aquifers is limited (Section 3.4.6). Rozell and Wong (2010) assessed the impact of rising sea levels on fresh water resources on Shelter Island (USA) for two different combinations of precipitation change and sea level rise. Projected impacts were highly dependent on local conditions. Ferguson and Gleeson (2012) concluded that the direct impact of groundwater extraction in the USA has been and will be much more significant than the impact of a 0.59 m sea level rise by the end of the 21st century under a wide range of hydrogeological conditions and population densities.

Saltwater intrusion is generally a very slow process; as a consequence, reaching equilibrium may take several centuries limiting the reversibility of the process in the near term (Webb and Howard, 2011).

Human-induced pressure will continue to be the main driver for aquifer salinization during the next century (*high confidence*). Changing precipitation, increased storminess, and sea level rise will exacerbate these problems (*limited evidence*, *high agreement*).

5.4.2.6. Estuaries and Lagoons

Coastal lagoons are shallow water bodies separated from the ocean by a barrier and connected at least intermittently to the ocean, while estuaries, where fresh and saltwater mix, are the primary conduit for nutrients, particulates, and organisms from land to the sea.

5.4.2.6.1. Observed impacts

Sediment accumulation in estuaries is high, heterogeneous, and habitat-specific and directly affected by human drivers, such as dredging and canalization, and indirectly via habitat loss, changes in sea level, storminess, and freshwater and sediment supply by rivers (Syvitski et al., 2005; Swanson and Wilson, 2008). Coastal lagoons are also susceptible to alterations of sediment input and erosional processes driven by changes in sea level, precipitation, and storminess (Pickey and Young, 2009). Droughts, floods, and sea level rise impact estuarine circulation, tidal characteristics, suspended matter, and hence turbidity with consequences for biological communities, particularly in microtidal systems. Climate change and habitat modification (e.g., dams and obstructions) impact fish species such as salmon and eels that pass through estuaries (Lassalle and Rochard, 2009).

Enhanced nutrient delivery (Section 5.3.4.3) has resulted in major changes in biogeochemical processes, community structure, metabolic balance, and CO_2 exchange (Howarth et al., 2011; Canuel et al., 2012; Statham, 2012), including enhanced primary production which has affected coastal fishery yield (Nixon, 1982; Savage et al., 2012). Eutrophication has modified the food-web structure (*high confidence*) and led to more intense and long lasting hypoxia (Section 5.3.4.4), more frequent occurrence of harmful algal blooms (Breitburg et al., 2009; Howarth et al., 2011; *medium confidence*), and to enhanced emission of nitrous oxide (de Bie et al., 2002; Kroeze et al., 2010; *high confidence*).

In summary, there is *very high confidence* that humans have impacted lagoons and estuaries.

5.4.2.6.2. Projected impacts

The increase of atmospheric CO_2 levels will reduce the efflux of CO_2 from estuaries (Borges, 2005; Chen and Borges, 2009; *high confidence*). Its impact on the pH of estuarine and lagoon waters will generally be limited because other drivers are usually more important (Section 5.3.3.4 and Box CC-OA; *high confidence*). For example, freshwater flow in the Scheldt estuary was the main factor controlling pH, directly via a decreased supply of dissolved inorganic carbon and total alkalinity, and, indirectly, via decreased input ammonia loadings and lower rates of nitrification (Hofmann et al., 2009).

Changes in sea level and hydrology could affect lagoons and estuaries in multiple ways. Sea level rise will impact sediment redistribution, the partitioning of habitats within estuaries, salinity, tidal range, and submergence periods (Anthony et al., 2009; *high confidence*). Lagoons may shrink because landward migration is restricted due to human occupation or extend due to the drowning of marshes (Anthony et al., 2009; Pilkey and Young, 2009; Stutz and Pilkey, 2011). Salinity, primary production, biodiversity, fisheries, and aquaculture may be impacted by changes in water discharge, withdrawals and precipitation-evaporation balance (Webster and Harris, 2004; Smith et al., 2005; Anthony et al., 2009; Canu et al., 2010). Altered riverine discharge and warming may lead to enhanced thermal and/or salinity stratification of estuaries and lagoons. This has consequences for biogeochemical processes, organism distribution patterns, and frequency and duration of hypoxia (Diaz and Rosenberg, 2008; Rabalais et al., 2009; Hong and Shen, 2012; *medium confidence*). Stronger winds and droughts may reduce the extent, duration, and frequency of estuarine stratification, counteracting the decrease in oxygen concentration (Rabalais et al., 2009; *medium confidence*).

Changes in storm events may also alter the sediment deposition-erosion balance of lagoons and estuaries (Pilkey and Young, 2009), the structure and functioning of biological communities via the transport of communities and/or of their resources, and the underwater light climate (Wetz and Paerl, 2008; Canuel et al., 2012; *medium confidence*). Changes in precipitation extremes and freshwater supply may induce fluctuations in salinity with the associated adverse impacts on biodiversity, benthic macrofauna, and ecosystem functions (Jeppesen et al., 2007; Fujii and Raffaelli, 2008; Levinton et al., 2011; Pollack et al., 2011). Warming may directly affect most biological processes and the trophic status of coastal ecosystems, and higher carbon dioxide emission (Canuel et al., 2012; *limited evidence*, *medium agreement*). Warming may lengthen the duration of phytoplankton production season (Cloern and Jassby, 2008; *medium confidence*).

Any change in the primary production of lagoons might impact fisheries, as primary production and fisheries yield are correlated (Nixon, 1982; *limited evidence*, *medium agreement*). For example, seawater warming and changes in seasonal patterns of precipitations projected in the Venice lagoon, using the SRES A2 emission scenario for the period 2071–2100, may lead to a reduction in plankton production, with a decline of habitat suitability for clam growth and aquaculture (Canu et al., 2010).

Finally, projected changes in climate-related drivers such as warming, storms, sea level, and runoff will interact with non-climate human drivers (e.g., eutrophication, damming) and will have consequences for ecosystem functioning and services of lagoons and estuaries (*high confidence*).

In summary, the primary drivers of change in lagoons and estuaries are human-related rather than climate-related drivers (*very high confidence*). Future changes in climate-related drivers such as warming, acidification, waves, storms, sea level, and runoff will have consequences on the functions and services of ecosystems in lagoons and estuaries (*high confidence*) but the impacts cannot be assessed at the global scale as the key drivers operate at a local to regional scale.

5.4.2.7. Deltas

Characterized by the interplay between rivers, lands, and oceans and influenced by a combination of river, tidal, and wave processes, deltas are coastal complexes that combine natural systems in diverse habitats (e.g., tidal flats, salt marshes, mangroves, beaches, estuaries, low-lying wetlands) and human systems (e.g., houses, agriculture, aquaculture, industry, and transport). They are low-lying coastal landforms formed by riverine sediments in the areas around river mouths, mostly during the last 6000–8000 years of relatively stable sea level and have a population density more than 10 times the world average (Ericson et al., 2006; Foufoula-Georgiou et al., 2011). As low-lying plains, deltas are highly sensitive to changes in sea level. They are subject to climatic impacts from rivers upstream (e.g., freshwater input) and oceans downstream (e.g., sea level changes, waves) as well as within the deltas themselves. At the same time, they are affected by human activities such as land use changes, dam construction, irrigation, mining, extraction of subsurface resources, and urbanization (Nicholls et al., 2007).

5.4.2.7.1. Observed impacts

The combined impact of sediment reduction, RSLR, and land use changes in delta and river management on channels and banks has led to the widespread degradation of deltas (*very high confidence*). The changes of sediment delivery from rivers due to dams, irrigation, and embankments/dikes create an imbalance in sediment budget in the coastal zones. Degradation of beaches, mangroves, tidal flats, and subaqueous delta fronts along deltaic coasts has been reported in many deltas (e.g., Nile and Ebro; Sanchez-Arcilla et al., 1998; Po, Simeoni and Corbau, 2009; Krishna-Godavari, Nageswara Rao et al., 2010; Changjiang, Yang et al., 2011; Huanghe, Chu et al., 1996; *very high confidence*). Deltaic coasts naturally evolve by seaward migration of the shoreline, forming a delta plain. However, decreasing sediment discharge during the last 50 years has decreased the growth of deltaic land, even reversing it in some locations (e.g., Nile, Godavari, Huanghe). Artificial reinforcement of natural levees also has reduced the inter-distributory basin sedimentation in most deltas, resulting in wetland loss.

The major impacts of sea level rise are changes in coastal wetlands, increased coastal flooding, increased coastal erosion, and saltwater intrusion into estuaries and deltas (Mcleod et al., 2010), which are exacerbated by increased human-induced drivers (*very high confidence*). Ground subsidence amplifies these hazards in farms and cities on deltaic plains through RSLR (Day and Giosan, 2008; Mazzotti et al., 2009). RSLR due to subsidence has induced wetland loss and shoreline retreat (e.g., the Mississippi delta; Morton et al., 2005; Chao Phraya delta, Saito et al., 2007; *high confidence*). Episodic events superimpose their effects on these underlying impacts and accelerate land loss (*high confidence*) (e.g., Hurricanes Katrina and Rita in 2005; Barras et al., 2008). To forestall submergence and frequent flooding, many delta cities now depend on a substantial infrastructure for flood defense and water management (Nicholls et al., 2010).

Deltas are impacted by river floods and oceanic storm surges (*very high confidence*). Tropical cyclones are noteworthy for their damages to deltas, for example, the Mississippi delta by Hurricane Katrina in 2005 (Barras et al., 2008), the Irrawaddy delta by Cyclone Nargis in 2008, and the Ganges-Brahmaputra delta by Cyclone Gorky in 1991 and Cyclone Sidr in 2007 (Murray et al., 2012; see also Box CC-TC). A detailed study of 33 deltas around the world found that 85% of them had experienced severe flooding in the past decade, causing the temporary submergence of 260,000 km^2 (Syvitski et al., 2009).

5.4.2.7.2. Projected impacts

The projected natural impacts on deltas under changing global climate are caused mainly by extreme precipitation-induced floods and sea level rise. These will result in increased coastal flooding, decreased wetland areas, increased coastal erosion, and increased salinization of cultivated land and groundwater (McLeod et al., 2010; Day et al., 2011; Box CC-TC; *high confidence*). The surface area of flooding in 33 deltas around the world is estimated to increase by 50% under sea level rise estimations as projected for 2100 by the IPCC AR4 (Syvitski et al., 2009). Non-climatic drivers (e.g., reduction in sediment delivery, subsidence, and land use changes) rather than climatic drivers have affected deltas for the last 50 years (Syvitski, 2008; *very high confidence*). Densely populated deltas are particularly vulnerable owing to further population growth together with the above-described impacts. The impacts beyond 2100 show a more complex and enhanced flood risk on deltas (e.g., Katsman et al., 2011). In summary, increased human drivers have been primary causes in changes of deltas (e.g., land use, subsidence, coastal erosion) for at least the last 50 years (*very high confidence*). There is *high agreement* that future sea level rise will exacerbate the problems of increased anthropogenic degradation in deltas.

5.4.3. Human Systems

5.4.3.1. Human Settlements

Important direct effects of climate change on coastal settlements include dry-land loss due to erosion and submergence, damage of extreme events (such as wind storms, storm surges, floods, heat extremes, and droughts) on built environments, effects on health (food- and water-borne disease), effects on energy use, effects on water availability and resources, and loss of cultural heritage (Hunt and Watkiss, 2010). Since AR4, a large number of regional, national, and subnational scale studies on coastal impacts have been conducted. These are covered in the respective regional chapters. At the global scale, studies have focused either on exposure to sea level rise or extreme water levels or on the physical impacts of flooding, submergence, and erosion.

5.4.3.1.1. Projected exposure

Coastal flood risks are strongly influenced by the growing exposure of population and assets. The population exposed to the 1-in-100-year coastal flood is projected to increase from about 270 million in 2010 to 350 million in 2050 due to socioeconomic development only (UN medium fertility projections) (Jongman et al., 2012). Population growth, economic growth, and urbanization will be the most important drivers of increased exposure in densely populated areas (Hanson et al., 2011; Seto, 2011; see also Chapter 14; *high confidence*). For 136 port cities above 1 million inhabitants, the number of people exposed to a 1-in-100-year extreme sea level is expected to increase from 39 million in 2005 to 59 million by 2070 through 0.5 m GMSLR alone and to 148 million if socioeconomic development (UN medium population projections) is considered (Hanson et al., 2011). Human-induced subsidence alone is expected to increase the global economic exposure of 136 major port cities by around 14% from 2005 to 2070 although this driver only applies to 36 of the cities (Hanson et al., 2011). As a result of socioeconomic development Asia is expected to continue to have the largest exposed population and sub-Saharan Africa the largest increases in exposure (Dasgupta et al., 2009; Vafeidis et al., 2011; Jongman et al., 2012).

5.4.3.1.2. Projected impacts and risks

Exposure estimates, however, give an incomplete picture of coastal risks to human settlements because they do not consider existing or future adaptation measures that protect the exposed population and assets against coastal hazards (Hallegatte et al., 2013; Hinkel et al., 2013). Although the global potential impacts of coastal flood damage and land loss on human settlements in the 21st century are substantial, these impacts can be reduced considerably through coastal protection (*limited evidence*, *high agreement*). Nicholls et al. (2011) estimate that without

Figure 5-4 | The 20 coastal cities where average annual losses (AALs) increase most (in relative terms in 2050 compared with 2005) in the case of optimistic sea level rise, if adaptation maintains only current defense standards or flood probability (PD) (Hallegatte et al., 2013).

protection 72 to 187 million people would be displaced due to land loss due to submergence and erosion by 2100 assuming GMSLRs of 0.5 to 2.0 m by 2100. Upgrading coastal defenses and nourishing beaches would reduce these impacts roughly by three orders of magnitude. Hinkel et al. (2013) estimate the number of people flooded annually in 2100 to reach 117 to 262 million per year in 2100 without upgrading protection and two orders of magnitude smaller with dike (levee) upgrades, given GMSLRs of 0.6 to 1.3 m by 2100. The major driver of increasing risks to human settlements in the next decades is socioeconomic development. When upgrading flood defenses to maintain a constant probability of flooding, average annual losses (AALs) in the 136 largest coastal cities are expected to increase ninefold from 2005 to 2050 due to socioeconomic development, only another 12% due to subsidence, and 2 to 8% due to GMSLRs of 0.2 to 0.4 m (Hallegate et al., 2013; Figure 5-4).

Despite the delayed response of sea level rise to global warming levels (WGI AR5 Section 13.5.4) mitigation may limit 21st century impacts of increased coastal flood damage, dry-land loss, and wetland loss substantially (*limited evidence*, *medium agreement*) albeit numbers are difficult to compare owing to differences in scenarios, baselines, and adaptation assumptions. Tol (2007) finds that stabilizing CO_2 concentration at 550 ppm reduces global impacts on wetlands and dry lands by about 10% in 2100 compared to a scenario of unmitigated emissions. Hinkel et al. (2013) report that stabilizing emissions at 450 ppm CO_2-eq reduces the average number of people flooded in 2100 by about 30% compared to a baseline where emissions increase to about 25 Gt C-eq in 2100. Arnell et al. (2013) find that an emissions pathway peaking in 2016 and declining at 5% per year thereafter reduces flood risk by 58 to 66% compared to an unmitigated A1B scenario. All three studies only consider the effects of mitigation during the 21st century and assume low or no contribution of ice sheets to GMSLR. Mitigation is expected to be more effective when considering impacts beyond 2100 and higher contributions of ice sheets (Section 5.5.8).

Global studies confirm AR4 findings that there are substantial regional differences in coastal vulnerability and expected impacts (*high confidence*). Most countries in South, Southeast, and East Asia are particularly vulnerable to sea level rise due to rapid economic growth and coastward migration of people into urban coastal areas together with high rates of anthropogenic subsidence in deltas where many of the densely populated areas are located (Nicholls and Cazenave, 2010). At the same time, economic growth in these countries increases the monetary capacity to adapt (Nicholls et al., 2010). In contrast, although many African countries experience a similar trend in rapid urban coastal growth, the level of economic development is generally lower and consequently the monetary capacity to adapt is smaller (Kebede and Nicholls, 2012; Hinkel et al., 2013).

In summary, while there is *high agreement* on some general findings, only a small fraction of the underlying uncertainty has been explored, which means evidence is limited. Gaps remain with respect to impacts of possible large contributions of the ice sheets of Greenland and Antarctica to GMSLR (WGI AR5 Sections 13.4.3, 13.4.4), regional patterns of climate-induced sea level rise, subsidence, and socioeconomic change and migration. Many studies rely on few or only a single socioeconomic scenario. Few studies consider adaptation and those that do generally ignore the wider range of adaptation measures beyond hard protection options. Integrated studies considering the interactions between a wide range of RSLR impacts (Table 5-3) as well as trade-offs between diverse adaptation options are missing.

5.4.3.2. Industry, Infrastructure, Transport, and Network Industries

Coastal industries, their supporting infrastructure including transport (ports, roads, rail, airports), power and water supply, storm water, and sewerage are highly sensitive to a range of extreme weather and climate events including temporary and permanent flooding arising from extreme precipitation, high winds, storm surges, and sea level rise (Horton et al. 2010; Handmer et al. 2012; Hanson and Nicholls, 2012; Aerts et al. 2013; *high confidence*). Most industrial facilities, infrastructure, and networks are designed for service lives extending over several decades. In fact, many bridges, ports, and road and railway lines remain in their original design location for centuries even if the infrastructure on them has been rehabilitated or replaced several times. Certain facilities, such as new nuclear power plants, are designed to last even well beyond the 22nd century (Wilby et al., 2011).

As the need to locate most of these industries and networks in coastal areas will remain and probably increase due to coastal development (Section 5.4.3.1), considering climate variability and climate change drivers in life cycle assessment of industry, infrastructure, and transport and network industries is of utmost importance (*high confidence*).

5.4.3.2.1. Observed impacts

Climate impacts on coastal industries and infrastructures vary considerably depending on geographical location, associated weather and climate, and specific composition of industries within particular coastal regions (*high confidence*).

Over the last 10 years an extensive number of climate-related extreme events (Coumou and Rahmstorf, 2012) illustrate the potential for impacts on coastal industry, infrastructure, transport, and network industry. Severe storms with associated winds, waves, rain, lightning, and storm surges have been particularly disruptive to transport and power and water supplies (Jacob et al., 2007; USCCSP, 2008; Horton et al., 2010; *high confidence*). In such network configurations, flooding of even the smallest component of an intermodal system can result in a much larger system disruption. Even though a transportation terminal may not be affected, the access roads to it could be, thus forcing the terminal to cease or reduce operation. Disruption to port activities in one location can disrupt supply chains, which can have far reaching consequences (Becker et al. 2012, 2013). Existing experience has also shown that impacts of hurricanes and flooding on underground infrastructure can have long-term effects (Chisolm and Matthews, 2012).

Hurricanes like Katrina (2005) caused US$100 million of damage to Mississippi's ports and Sandy (2012) led to a week-long shut-down of the Port of New York, generating economic damages reaching US$50 billion (Becker et al., 2012). These have shown the critical need to better prepare coastal human settlements and associated network infrastructures and industries for future extreme weather impacts and climate change (Aerts et al., 2013; *high confidence*).

5.4.3.2.2. Projected impacts

Although there is *robust evidence* of the impacts and consequences of extreme events on coastal infrastructure and industrial facilities, there are limited assessments on projected impacts of long-term changes (*high agreement*). Besides, while there is an important amount of non-journal literature on projected impacts of sea level rise and increasing flooding levels on certain coastal infrastructures (USCCSP, 2008; USACE, 2011; McEvoy and Mullet, 2013), limited peer review information is available.

Vulnerability to flooding of railroads, tunnels, ports, roads, and industrial facilities at low-lying areas will be exacerbated by rising sea levels or more frequent or intense storms, causing more frequent and more serious disruption of services and damages under extreme sea levels unless adaptation is enforced (Esteban et al., 2010, 2012; Wilby et al., 2011; Aerts et al., 2013; *high confidence*).

Furthermore, sea level rise will reduce extreme flood return periods and therefore increase the need for adaptation of infrastructure such as airports, tunnels, coastal protections, and ship terminals to extreme sea level impacts (Jacob et al., 2007; Becker et al., 2013).

It is estimated that a hypothetical 1 m RSLR projected for the Gulf Coast region between Alabama and Houston over the next 50 to 100 years would permanently flood a third of the region's roads as well as putting more than 70% of the region's ports at risk (USCCSP, 2008).

The estimated costs of climate change to Alaska's public infrastructure could add US$3.6 to 6.1 billion (+10 to 20% above normal wear and tear) from now to 2030 and US$5.6 to 7.6 billion (+10 to 12%) from now to 2080 (Larsen et al., 2008). Higher costs of climate change for coastal infrastructure are expected due to its proximity to the marine environment. Other projected impacts are beneficial for the transportation system. For example, decline of Arctic sea-ice coverage could extend seasonal accessibility to high-latitude shipping routes such as the northwest shipping route that connects the Atlantic to the North Pacific.

Hanson et al. (2011) presents a first estimate of the exposure of the world's large port cities to coastal flooding due to sea level rise and storm surge in the 2070s. The analysis suggests that the total value of assets exposed in 2005 across all cities considered is estimated to be US$3000 billion, corresponding to around 5% of global GDP in 2005. By the 2070s, and assuming a homogeneous global sea level rise of 0.5 m, increased extreme water levels up to 10%, and a fixed subsidence rate in susceptible cities with respect to today's values, asset exposure is estimated to increase to approximately 9% of projected global GDP in this period.

Coastal infrastructural instability may result from natural hazards triggered by groundwater-level (GWL) variations resulting from rising sea level. For earthquake-prone coasts, this could be exacerbated by earthquake liquefaction if GWL increases with sea level rise (Yasuhara et al., 2007). Increasing sea levels, surges, and waves can also lead to a stability loss of coastal structures (Headland et al., 2011).

Other impacts may arise in coastal industries in high latitudes affected by permafrost thaw causing ground instability and erosion, thereby affecting transport safety and the industries that rely on such travel in these regions (e.g., Pearce et al., 2010).

5.4.3.3. Fisheries, Aquaculture, and Agriculture

Fisheries and aquaculture and the associated post-harvest activities globally create millions of jobs (Daw et al., 2009; Sumaila et al., 2011) and contribute significantly to the dietary animal protein of millions of people and to the world merchandise trade (FAO, 2010, 2012; see also Section 6.4.1.1). In addition to small-scale fisheries and aquaculture, which are important for the food security and economy of coastal communities (Bell et al., 2009), coastal zones also support significant agricultural activities, for example, rice production in the low-lying deltaic regions of Asia (Wassmann et al., 2009).

5.4.3.3.1. Observed impacts

Climate variability and change impact both fishers' livelihoods (Badjeck et al., 2010) and fish production (Barange and Perry, 2009) (Section 6.5.3). In the North Sea, ocean warming over the 1977–2002 period led to relatively increased distribution ranges of some fish species (Hiddink and Hofstede, 2008), and demersal fish assemblage deepened in response to climate change (Dulvy et al., 2008). In southeastern Australia, Last et al. (2011) found an increasing abundance of 45 fish species of warm temperate origin, which they linked to the observed strengthening of the East Australian Current (EAC) bringing warm waters further south (Ridgeway, 2007). A study (Sherman et al., 2009) of the impact of sea surface temperature changes on the fisheries yields of 63 large marine ecosystems over a 25-year period shows a positive relationship for the northeast Atlantic large marine ecosystems, due to zooplankton biomass increases (Section 6.5.3). Distributional effects are very important for migratory pelagic fisheries, such as tuna (see Table 29-2). Impacts of climate change on aquaculture (*Mytilus edulis* and *Salmo salar*) in the UK and Ireland have been difficult to discern from natural environmental variability (Callaway et al., 2012).

Seawater inundation has become a major problem for traditional agriculture in Bangladesh (Rahman et al., 2009), and in low-lying island nations (e.g., Lata and Nunn, 2012). The combination of rice yield reduction induced by climate change and inundation of lands by seawater causes an important reduction in production (Chen et al., 2012).

5.4.3.3.2. Projected impacts

Fisheries may be impacted either negatively or positively (Hare et al., 2010; Meynecke and Lee, 2011; Cinner et al., 2012) depending on the latitude, location, and climatic factors. Climate change can impact the pattern of marine biodiversity through changes in species' distributions, and may lead to large-scale redistribution of global catch potential depending on regions (Cheung et al., 2009, 2010). Narita et al. (2012) estimated that the global economic costs of production loss of molluscs due to ocean acidification (Section 5.3.3.5) by the year 2100 based on IPCC IS92a business-as-usual scenario could be higher than US$100 billion. As a result of increased sea temperatures, the reduction in coral cover in the Caribbean basin and its associated fisheries production is expected to lead to a net revenue loss by 2015 (Trotman et al., 2009). Economic losses in landed catch value and the costs of adapting fisheries resulting from a 2°C global temperature increase by 2050 have been estimated at US$10 to 31 billion globally (Sumaila et al., 2011). For aquaculture, negative impacts of rising ocean temperatures will be felt in the temperate regions whereas positive impacts will be felt in the tropical and subtropical regions (De Silva and Soto, 2009). Changes to the atmosphere-ocean in the Pacific Island countries are likely to affect coral reef fisheries by a decrease of 20% by 2050 and coastal aquaculture may be less efficient (Bell et al., 2013).

In summary, changes have occurred to the distribution of fish species (*medium confidence*) with evidence of poleward expansion of temperate species (*limited evidence*, *high agreement*). Tropical and subtropical aquaculture has not been adversely affected by rising ocean temperatures to date (*limited evidence*, *high agreement*). Coastal agriculture has experienced negative impacts (*medium confidence*) due mainly to increased frequency of submersion of agricultural land by saltwater inundation (*limited evidence*, *high agreement*).

5.4.3.4. Coastal Tourism and Recreation

Coastal tourism is the largest component of the global tourism industry. Over 60% of Europeans opt for beach holidays and beach tourism provides more than 80% of U.S. tourism receipts (UNEP, 2009). More than 100 countries benefit from the recreational value provided by their coral reefs, which contributed US$11.5 billion to global tourism (Burke et al., 2011).

5.4.3.4.1. Observed impacts

Observed significant impacts on coastal tourism have occurred from direct impacts of extreme events on tourist infrastructure (e.g., beach resorts, roads), indirect impacts of extreme events (e.g., coastal erosion, coral bleaching), and short-term adverse tourist perception after the occurrence of extreme events (e.g., flooding, tropical storms, storm surges) (Phillips and Jones, 2006; Scott et al., 2008; IPCC, 2012, Section 4.3.5.3). Recent observed climate change impacts on the Great Barrier Reef include coral bleaching in the summers of 1997–1998, 2001–2002, and 2005–2006 and extreme events including floods and cyclones (Tropical cyclones Larry in 2006, Hamish in 2009, and Yasi in 2011). The stakeholders show a high level of concern for climate change, and various resilience initiatives have been proposed and developed by the Great Barrier Reef Marine Park Authority (Biggs, 2011; GBRMPA, 2012).

5.4.3.4.1. Projected impacts

To provide some idea of climate change impacts on coastal destinations, many studies have been carried out on projecting tourism demand, for example, in Europe (Perch-Nielson et al., 2010), in the Baltic region (Haller et al., 2011), in the Mediterranean (Moreno and Amelung, 2009a), and in 51 countries worldwide (Perch-Nielson, 2010). The studies provide varying details, although it is difficult to draw overarching conclusions on tourism demand for coastal destinations. With increased temperature in mid-latitude countries and coupled with increased storms in tropical areas, tourist flows could decrease from mid-latitude countries to tropical coastal regions with large developing countries and small island nations most affected (Perch-Nielson, 2010). The Mediterranean would likewise be affected in summer (Moreno and Amelung, 2009a). In contrast, less is known about the relationship between the impacts of climate change and specific tourist behavior, activities, or flows to coastal destinations (Moreno and Amelung, 2009b; see Section10.6.2). Usually tourists do not consider climate variability or climate change in their holidays (Hares et al., 2009) although there are a few studies that show the contrary (Cambers, 2009; Alvarez-Diaz et al., 2010).

As for future impacts on coastal tourism, there is *high confidence* in the impacts of extreme events and sea level rise aggravating coastal erosion. A scenario of 1-m sea level rise by 2100 would be a potential risk to Caribbean tourism (Scott et al., 2012). The presence of coastal tourism infrastructure will continue to exacerbate beach reduction and coastal ecosystems squeeze under rising sea levels, as exemplified in Martinique (Schleupner, 2008). Carbonate reef structures would degrade under a scenario of at least 2°C by 2050–2100 with serious consequences for tourism destinations in Australia, the Caribbean, and other small islands (Hoegh-Gulberg et al., 2007; see Box CC-CR).

The costs of future climate change impacts on coastal tourism are enormous. For example, in the Caribbean community countries, rebuilding costs of tourist resorts are estimated US$10 to 23.3 billion in 2050. A hypothetical 1-m sea level rise would result in the loss or damage of 21 airports, inundation of land surrounding 35 ports, and at least 149 multi-million dollar tourism resorts damaged or lost from erosion to the coastal beach areas (Simpson et al., 2010).

In summary, while coastal tourism can be related to climate change impacts, it is more difficult to relate tourism demand directly to climate change. Coastal tourism continues to be highly vulnerable to weather, climate extremes, and rising sea levels with the additional sensitivity to ocean temperature and acidity for the sectors that rely on reef tourism (*high confidence*). Developing countries and small island states within the tropics relying on coastal tourism are most vulnerable to present and future weather and climate extremes, future sea level rise, and the added impacts of coral bleaching and ocean acidification (*high confidence*).

5.4.3.5. Health

The relationship between health of coastal populations and climate change include direct linkages (e.g., floods, droughts, storm surges, and extreme temperatures) and indirect linkages (e.g., changes in the transmission of vector-, food-, and water-borne infectious diseases and increased salinization of coastal land that affects food production and freshwater supply and ecosystem health). Coastal and particularly informal settlements concentrate injury risk and death from storm surges and rainfall flooding (Handmer et al., 2012). This section deals with human health in the context of the coastal zone, while Chapter 11 addresses general health issues and Section 6.4.2.3 deals with health issues associated with ocean changes. Understanding the relationship between climate and health is often confounded by socioeconomic factors that influence coastal settlement patterns and the capacity of authorities to respond to health-related issues (Baulcomb, 2011).

5.4.3.5.1. Observed impacts

Mortality risk in coastal areas is related to exposure and vulnerability of coastal populations to climate hazards (e.g., Myung and Jang, 2011). A regional analysis of changes in exposure, vulnerability, and risk indicates that although exposure to flood and cyclone hazards has increased since 1980, the risk of mortality has generally fallen. The reductions reflect a strengthening of the countries' capacity to respond to disasters (Box 5-1). However, mortality is still rising in the countries with the weakest risk governance capacities (UNISDR, 2011).

Coastal regions face a range of climate-sensitive diseases. Increased saline intrusion is linked to increased hypertension disease (Vineis et al., 2011), with greater occurrence in pregnant women living in coastal regions compared to further inland (Khan et al., 2008). Increasing temperature, humidity, and rainfall can increase vector-borne diseases such as malaria, dengue, leishmaniasis, and chikungunya (Pialoux et al., 2007; Stratten et al., 2008; Kolivras, 2010; van Kleef et al., 2010) and diarrhea, infectious gastrointestinal disease, rotovirus, and salmonella (e.g., Hashizume et al., 2007; Zhang et al., 2007, 2010; Chou et al., 2010; Onozuka et al., 2010). The parasitic disease schistosomiasis, endemic in many tropical and small island coastal regions (Section 29.3.3.2), is also sensitive to temperature increase (Mangal et al., 2008). Vibrio outbreaks (e.g., cholera) are sensitive to rainfall and SST (e.g., Koelle et al., 2005), and recent increased vibrio outbreaks in the Baltic have been linked to heat waves and low salinity (Baker-Austin et al., 2013). Harmful algal blooms (HABs) outbreaks (e.g., ciguatera) have been linked to SST variability (e.g., Erdner et al., 2008; Jaykus et al., 2008). However, in general there is *limited evidence* and *low confidence* in how global climate change will impact HABs (Section 6.4.2.3), suggesting the need for increased monitoring (Hallegraeff, 2010). Nontoxic blooms of high biomass can reduce biodiversity through oxygen depletion and shading (Erdner et al., 2008), with consequences for ecosystem and human nutrition and health.

5.4.3.5.2. Projected impacts

Under future climate conditions, expansion of brackish and saline water bodies in coastal areas under projected sea level rise may increase the incidence of vector-borne diseases (Ramasamy and Surendran, 2011), diarrhea, and hypertension (Vineis et al., 2011). Human responses to climate change may also influence outcomes on health; however, limited empirical climate-health data increases uncertainties on such projections (Kolstad and Johansson, 2011).

Evidence continues to emerge of the relation between climate and diseases that affect human health in the coastal zone including air and water temperature, rainfall, humidity, and coastal salinity. However, the relations are often complex and vary between diseases and even regionally for the same disease. The interplay between climate and human systems with regard to health impacts is poorly understood and this continues to confound reliable projections of health impacts (*robust evidence*, *high agreement*).

5.4.4. Summary: Detection and Attribution

There is *high confidence* in the attribution to climate change of observed coastal impacts that are sensitive to ocean temperature change, such as coral bleaching and movements in species ranges. However, for many other coastal changes, the impacts of climate change are difficult to tease apart from human-related drivers (e.g., land use change, coastal development, pollution). Figure 5-5 shows changes of major phenomena observed in coastal systems and low-lying areas. Horizontal and vertical axes indicate the degree of confidence in detection of trends for phenomena, which are elements sensitive to climate change, and the degree of confidence in attribution of phenomena to climate change, respectively. Mainly phenomena with *high* to *very high confidence* in trend detection are illustrated in this figure.

The increase of coral bleaching and the shift in distribution and range limits of some species are attributed to climate change with *high confidence.* Mass coral bleaching coincided with positive temperature anomalies over the past 30 years. A poleward expansion of mangrove forests and some corals, and shifts of range limits of many intertidal species, are also attributed. Vegetated coastal habitats are declining globally. Coral cover and calcification have decreased in recent decades. Elevated temperatures along with ocean acidification reduce the calcification rate of corals. Although the attribution of decreased calcification to either climate- or human-related drivers is difficult, we have *medium confidence* that the primary climate-related driver is ocean warming globally. Seagrass meadows are already under stress due to climate change, particularly where maximum temperatures already approach their physiological limit. However, the decline of the distribution of mangroves and salt marshes is mainly linked with human activities, for example, deforestation and reclamation. Therefore the degree of their attribution to climate change is *very low*.

Globally beaches and shorelines have, in general, undergone net erosion over the past century or longer. There is *high confidence* in detection of increased beach erosion globally. However, attributing shoreline changes to climate change is still difficult owing to the multiple natural and human-related drivers contributing to coastal erosion (e.g., subsidence, decreased sediment delivery, land use change). There is *high confidence* that human pressures, for example, increased usage of surface water and groundwater resources for agriculture and coastal settlements, and river channel deepening, have led to increased saltwater intrusion and *low confidence* in attribution of saltwater intrusion to climate change.

The population living in coastal lowlands is increasing and more than 270 million people in 2010 are already exposed to flooding by the 1-in-100-year coastal flood (Mimura, 2013). Population growth and land subsidence in coastal lowlands are the major causes; therefore, there is *very low* attribution to climate change.

5.5. Adaptation and Managing Risks

5.5.1. Introduction

Coastal adaptation and risk management refer to a wide range of human activities related to the social and institutional processes of framing the

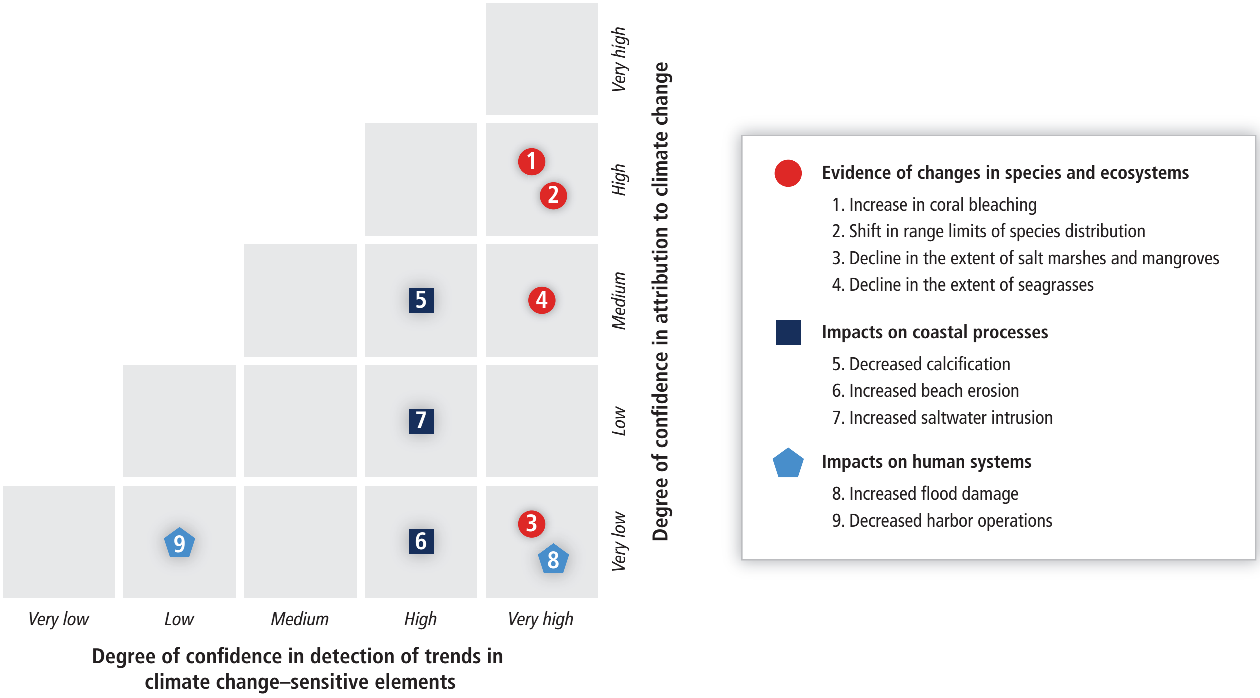

Figure 5-5 | Summary of detection and attribution in coastal areas.

Frequently Asked Questions

FAQ 5.3 | How can coastal communities plan for and adapt to the impacts of climate change, in particular sea level rise?

Planning by coastal communities that considers the impacts of climate change reduces the risk of harm from those impacts. In particular, proactive planning reduces the need for reactive response to the damage caused by extreme events. Handling things after the fact can be more expensive and less effective.

An increasing focus of coastal use planning is on precautionary measures, that is, measures taken even if the cause and effect of climate change is not established scientifically. These measures can include things like enhancing coastal vegetation and protecting coral reefs. For many regions, an important focus of coastal use planning is to use the coast as a natural system to buffer coastal communities from inundation, working with nature rather than against it, as in the Netherlands.

While the details and implementation of such planning take place at local and regional levels, coastal land management is normally supported by legislation at the national level. For many developing countries, planning at the grass roots level does not exist or is not yet feasible.

The approaches available to help coastal communities adapt to the impacts of climate change fall into three general categories:

1. Protection of people, property, and infrastructure is a typical first response. This includes "hard" measures such as building seawalls and other barriers, along with various measures to protect critical infrastructure. "Soft" protection measures are increasingly favored. These include enhancing coastal vegetation and other coastal management programs to reduce erosion and enhance the coast as a barrier to storm surges.
2. Accommodation is a more adaptive approach involving changes to human activities and infrastructure. These include retrofitting buildings to make them more resistant to the consequences of sea level rise, raising low-lying bridges, or increasing physical shelter capacity to handle needs caused by severe weather. Soft accommodation measures include adjustments to land use planning and insurance programs.
3. Managed retreat involves moving away from the coast and may be the only viable option when nothing else is possible.

Some combination of these three approaches may be appropriate, depending on the physical realities and societal values of a particular coastal community. The choices need to be reviewed and adjusted as circumstances change over time.

adaptation problem, identifying and appraising adaptation options, implementing options, and monitoring and evaluating outcomes (Chapters 2, 14, 15, 16, and 17). The governance of this process is challenging due to the complex, nonlinear dynamics of the coastal socio-ecological systems (Rosenzweig et al., 2011) as well as the presence of multiple management goals, competing preferences of stakeholders, and social conflicts involved (Hopkins et al., 2012). In many instances, coastal adaptation may thus be characterized to be a "wicked problem" (Rittel and Webber, 1973), in the sense that there is often no clear agreement about what exactly the adaptation problem is and there is uncertainty and ambiguity as to how improvements might be made (Moser et al., 2012).

Since AR4, the set of adaptation measures considered has been expanded specifically toward ecosystem-based measures (Section 5.5.2); novel approaches for appraising coastal adaptation decisions have been applied (Section 5.5.3.1) and the analysis of adaptation governance and the institutional context in which decisions are taken has progressed (Section 5.5.3.2). Progress has also been made in better integrating adaptation practices within existing policy frameworks (Section 5.5.4.1) as well as in implementing adaptation and identifying good practices (Section 5.5.4.2). A number of studies have also explored the global costs and benefits of coastal adaptation (Section 5.5.5); opportunities, constraints, and limits of coastal adaptation (Section 5.5.6); linkages between coastal adaptation and mitigation (Section 5.5.7); and the long-term commitment to coastal adaptation (Section 5.5.8).

5.5.2. Adaptation Measures

A detailed discussion on general adaptation needs and measures can be found in Chapter 14. As a first approximation, adaptation measures were classified into institutional and social measures (Section 14.3.2.1), technological and engineered measures (Section 14.3.2.2), and ecosystem-based adaptation measures (Section 14.3.2.3). In terms of coastal adaptation, most of the existing measures can be included within this classification.

The IPCC classification of coastal adaptation strategies consisting of retreat, accommodation, and protection (Nicholls et al., 2007) is now widely used and applied in both developed and developing countries (Boateng, 2010; Linham and Nicholls, 2012). This trilogy of strategies has expanded into broad approaches of retreat, defend, and attack (Peel, 2010). Protection aims at advancing or holding existing defense lines by means of different options such as land claim; beach and dune nourishment; the construction of artificial dunes and hard structures such as seawalls, sea dikes, and storm surge barriers; or removing invasive and restoring native species. Accommodation is achieved by increasing flexibility, flood proofing, flood-resistant agriculture, flood hazard mapping, the implementation of flood warning systems, or replacing armored with living shorelines. Retreat options include allowing wetlands to migrate inland, shoreline setbacks, and managed realignment by, for example, breaching coastal defenses allowing the creation of an intertidal habitat. The appropriate measure may depend on several factors requiring a careful decision-making and governance process (Section 5.5.3).

Since AR4, coastal adaptation options have been revised and summarized in several guidebooks (EPA, 2009; USAID, 2009; UNEP, 2010) including best practice examples. Especially relevant has been the growth of Community Based Adaptation (CBA) measures (*robust evidence*, *high agreement*). Table 5-4 compiles different examples of CBA measures in countries such as Bangladesh, India, and the Philippines.

Ecosystem-based adaptation is increasingly attracting attention (Munroe et al., 2011). Adaptation measures based on the protection and restoration of relevant coastal natural systems such as mangroves (Schmitt et al., 2013), oyster reefs (Beck et al., 2011), and salt marshes (Barbier et al., 2011) are seen as no- or low-regret options irrespective of future climate (Cheong et al., 2013; *medium evidence*, *high agreement*). Further work is still needed in order to make reliable quantitative estimates and predictions of the capability of some of these ecosystems to reduce wave, storm surge, and sea level rise impacts and in order to provide reliable cost-benefit analysis of how they compare to other measures based on traditional engineering approaches.

5.5.3. Adaptation Decision Making and Governance

Since AR4, progress has been made in understanding coastal adaptation decisions and governance. For a general treatment of adaptation decision making and governance, see Chapters 2, 15, and 17.

5.5.3.1. Decision Analysis

One specific quality of many coastal adaptation decisions is that these involve options with long (i.e., 30 and more years) investment time scales (e.g., land use planning, flood defenses, construction of housing, and transportation infrastructure; Section 5.5.2). For such decisions, standard methods that rely on probability distribution of outcomes, such as cost-benefit analysis under uncertainty, cannot be applied because of the difficulties, both in theory and practice, to associate probabilities to future levels of GHG emissions, which determine the level of impacts and outcomes (Lempert and Schlesinger, 2001; Hallegate, 2009; see also Section 17.3.6.2).

Alternative approaches that represent uncertainty not through a single probability distribution but through a range of scenarios have thus been applied to long-term coastal adaptation. Robust decision making (RDM), for example, refers to approaches where options that work well over a wide range of these scenarios are preferred (Lempert and Schlesinger, 2000; Lempert and Collins, 2007). RDM in this sense has been applied to, e.g., the Port of Los Angeles infrastructure (Lempert et al., 2012).

Another set of approaches uses the criterion of flexibility to decide between alternative strategies. Flexible and reversible options are favored over non-flexible and non-reversible ones and decisions are delayed to keep future options open (Hallegate, 2009). The adaptation pathways approach, for example, implements the criterion of flexibility by characterizing alternative strategies in terms of two attributes: (1) adaptation tipping points (ATPs), which are points beyond which strategies are no longer effective (Kwadijk et al., 2010); and (2) what alternative strategies are available once a tipping point has been reached (Haasnoot et al., 2013). Importantly, the exact time when an ATP is reached does not matter; it is rather the flexibility of having alternative strategies available that is driving the decision. Prominent applications of this approach include the Thames Estuary 2100 Plan (Penning-Roswell et al., 2012; Box 5-1), the Dutch Delta Programme (Kabat et al., 2009), and the New York City Panel on Climate Change (Rosenzweig et al., 2011).

5.5.3.2. Institution and Governance Analysis

Decisions are made within a context. Institution and governance analysis comprise a variety of approaches that aim at describing this context as well as at explaining the emergence and performance of institutions and governance structures (GS). Institution analysis is particularly relevant to coastal adaptation, because deciding between options and implementing them is an ongoing process involving complex inter-linkages between public and private decisions at multiple levels of decision making and in the context of other issues, existing policies, conflicting interests, and diverse GS (e.g, Few et al., 2007; Urwin and Jordan, 2008; Hinkel et al., 2010; see also Sections 2.2.2, 2.2.3). The non-consideration of this context may hinder or mislead adaptation decisions and implementations as reported by the emerging literature on barriers to adaptation (Section 5.5.5). Institution analysis strives to understand how this context shapes decisions, and insights gained may be employed to craft effective institutions and policies for adaptation.

For coastal adaptation, the effectiveness of existing GS is often hindered owing to a lack of horizontal (i.e., within the same level of decision making) and vertical (i.e., between different levels of decision making) integration of organizations and policies (*high confidence*). Storbjörk and Hedren (2011), for example, report on a weak vertical administrative interplay in coastal GS in Sweden. In the UK, the effectiveness of local GS of Coastal Partnership is found to be limited because these are poorly integrated with higher level policies (Stojanovic and Barker, 2008). In the UK, national level coastal recommendations are difficult to translate into local level actions (Few et al., 2007) and, in the USA, coastal policies often have ambiguous or contradictory goals (Bagstad et al., 2007). In a number of African cases, coastal policies are found not to take into account longer term climate change (Bunce et al., 2010).

Box 5-1 | London's Thames Estuary 2100 Plan: Adaptive Management for the Long Term

The Environment Agency in Britain has recently developed the Thames Estuary 2100 plan (TE2100) to manage future flood threat to London (Environment Agency, 2012). The motivation was a fear that due to accelerated climate change-induced sea level rise the time could already be too short for replacing the Thames Barrier (completed in 1982) and other measures that protect London, because such major engineering schemes take 25 to 30 years to plan and implement. An adaptive plan that manages risk in an iterative way was adopted based on the adaptation pathway approach (Penning-Rowsell et al., 2012; see also Section 5.5.3.1; Figure 5-6). This plan includes maintaining the existing system in the first 25 years, then enhancing the existing defenses in a carefully planned way over the next 25 to 60 years, including selectively raising defenses and possibly over-rotating the Barrier to raise protection standards. Finally, in the longer term (beyond 2070) there will be the need to plan for more substantial measures if sea level rise accelerates. This might include a new barrier, with even higher protection standards, probably nearer to the sea, or even a coastal barrage. In the meantime the adaptive approach requires careful monitoring of the drivers of risk in the Estuary to ensure that flood management authorities are not taken by surprise and forced into emergency measures.

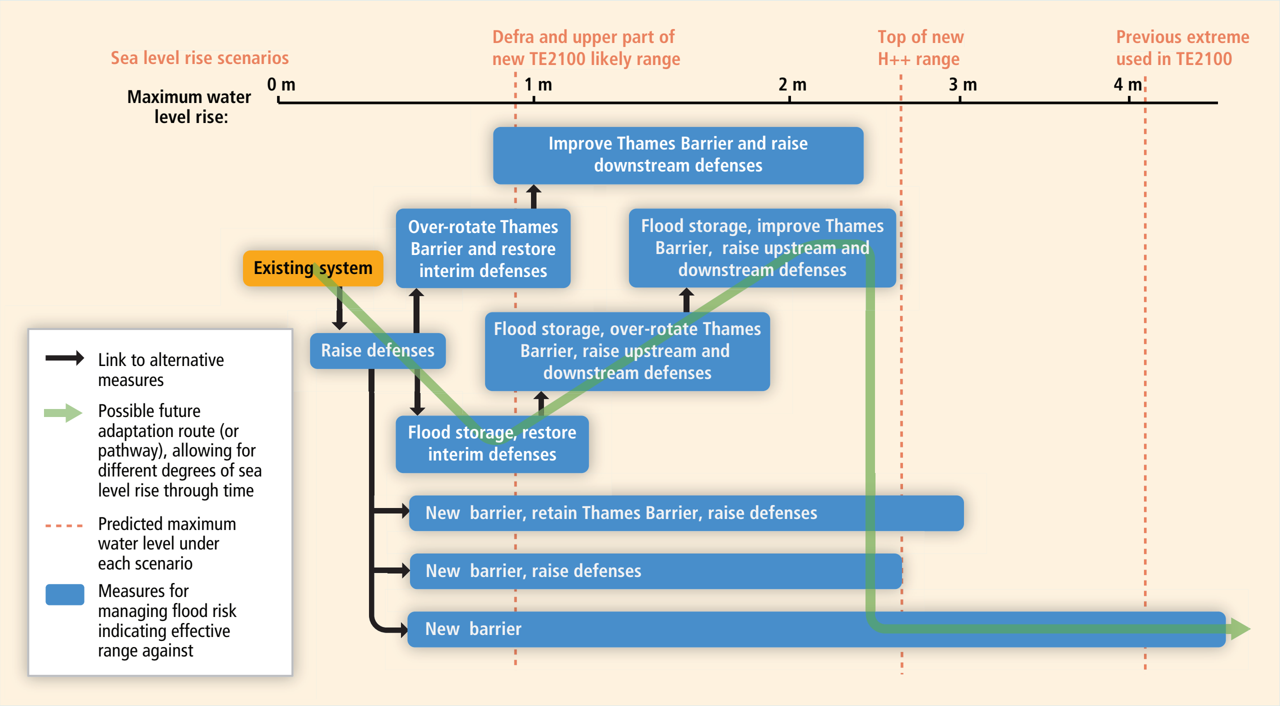

Figure 5-6 | Adaptation measures and pathways considered in the TE2100 project. The boxes show the measures and the range of sea level rise over which the measures are effective. The black arrows link to alternative measures that may be applied once a measure is no longer effective. The red lines show various 21st century sea level rise scenarios used in the analysis including a conservative estimate of about 0.9 m by the UK Department for Environment Food and Rural Affairs ('Defra and upper part of new TE2100 likely range'), a high-level scenario ('Top of new H++ range'), and an extreme scenario of over 4 meters ('previous extreme used in TE2100'). The fat green line shows a possible future adaptation route (or pathway), allowing for different degrees of sea level rise through time (adapted from Lowe et al., 2009).

Governance issues are particularly challenging when considering planned retreat (*medium evidence*). While managed realignment is on the political agenda in Germany and the UK, the political costs of doing so are high, as both the existing GS as well as public opinion are geared toward protection (e.g., Tunstall and Tapsell, 2007), so that short election cycles do not provide incentives for politicians to undertake actions that may produce benefits in the long term (Few et al., 2007; Rupp-Armstrong and Nicholls, 2007). Along the Queensland coast in Australia the option of planned retreat is disappearing because of rapid coastal development and liability laws favoring development. To prevent this, risks and responsibilities would need to be redistributed from the governments to the beneficiaries of this development (Abel et al., 2011).

While institutional factors are decisive in enabling coastal adaptation (*high confidence*), the role of institutions in coastal adaptation is generally under-researched. The majority of studies are descriptive. Institutional analysis striving to understand which GS emerge and are effective depending on both biophysical and social system characteristics as found in the fields of socio-ecological systems (Dietz et al., 2003; Folke et al., 2005; Ostrom 2007, 2009) and institutional economics (Hagedorn et al., 2002; Bougherara et al., 2009) are practically nonexistent.

5.5.4. Implementation and Practice

Since AR4, more experience has been gained in coastal adaptation implementation and practice. Generally, adaptation is not carried out stand-alone but in the context of already existing policy and practice frameworks. Section 5.5.4.1 assesses frameworks that are particularly relevant for coastal adaptation, and Section 5.5.4.2 assesses the experience as well as principles and compiled best practice guidelines.

5.5.4.1. Frameworks

The issues for coastal adaptation are not radically different from issues encountered within ICZM, which offers an enabling environment for adaptation practice (Celliers et al., 2013). ICZM is a long-term, institutionalized and iterative process that promotes the integration of coastal activities, relevant policymakers, practitioners, and scientists across coastal sectors, space and organizations with a view to use coastal resources in a sustainable way (Christie et al., 2005; Kay and Alder, 2005; Sales, 2009; WGII AR5 Glossary). Considering climate change in this framework does not mean radical changes to ICZM, because ICZM already emphasizes the integration of coastal issues across sectors and policy domains as well as the long-term perspective (e.g., Hofstede, 2008; Falaleeva et al., 2011). The major difference of coastal adaptation from ICZM is coping with greater uncertainty, longer time frames in planning (beyond 30 years), and long-term commitments inherent to climate change (Tobey et al. 2010). So far, however, there is *limited evidence* and *low agreement* on the effectiveness of ICZM alone or combined with climate change adaptation. Even though ICZM has been applied throughout the world for over 40 years, many obstacles to its successful implementation still remain (*high confidence*). Generally, there is a lack of empirical research evaluating ICZM (Stojanovic et al., 2004; Stojanovic and Ballinger, 2009). A recent review of ICZM in Europe concluded that the complexity of coastal regulations, demographic deficits, lack of sustainable finance and a failure to involve communities, business, and industry hinder its implementation (Shipman and Stojanovic, 2007). Developing countries in particular struggle to meet the goals of ICZM owing to a lack of qualified human resources, a lack of human, legal, and institutional capacities (Isager, 2008; González-Riancho et al., 2009), difficulties in integrating policy across multiple coastal agencies (Martinez et al., 2011; Ibrahim and Shaw, 2012), power (abuse) of the majority political party or political leaders (Isager, 2008; Tabet and Fanning, 2012), the lack of long-term financial commitment of donors (González-Riancho et al., 2009; Ibrahim and Shaw, 2012), and a lack of knowledge regarding the coastal system (González-Riancho et al., 2009).

Another prominent framework used for coastal adaptation practice is adaptive management (AM), which has been developed as a response to the deep uncertainty characterizing ecosystem management, where it is often impossible to predict outcomes of management interventions. AM thus aims to test management hypothesis by implementing them, monitoring their outcomes and learning from these to refine the management hypothesis to be applied (Holling, 1978; Walters, 1986). There are numerous applications of AM to coastal management (e.g., Walters, 1997; Marchand et al., 2011; Mulder et al., 2011), but there is *limited evidence* of its long-term effectiveness. Limitations of AM are also notable, such as the potential high cost of experimentation and a range of institutional barriers hindering the delivery of flexible management approaches (e.g., McLain and Lee, 1996).

Community-based adaptation (CBA) refers to the generation and implementation of locally driven adaptation strategies that address both climate change impacts and development deficits for the climate-vulnerable poor and that aim to strengthen the adaptive capacity of local people to climate and non-climate risk factors (Nicholls et al., 2007; Reid et al., 2009; Ayers and Dodman, 2010; Ayers and Huq, 2013; see also Sections 14.2.1, 15.4.3.1, 24.4.6.5). CBA is a bottom-up approach to adaptation involving all relevant stakeholders, especially local communities (Ayers and Huq, 2009; UNDP, 2010; Riadh et al., 2012) (Table 5-4). As such, CBA approaches have been developed through active participatory processes with local stakeholders (Ayers and Forsyth, 2009), and operated on a learning-by-doing, bottom-up, empowerment paradigm (Kates, 2000; Huq and Reid, 2007).

CBA experiences emphasize that it is important to understand a community's unique perception of its adaptive capacities in order to identify useful solutions (Parvin et al., 2008; Badjeck et al., 2010; Paul and Routray, 2010) and that scientific and technical information on anticipated coastal climate impacts needs to be translated into a suitable language and format that allows people to be able to participate in adaptation planning (Saroar and Routray, 2010). Furthermore, effective CBA needs to consider measures that cut across sectors and technological, social, and institutional processes, as technology by itself is only one component of successful adaptation (Pelling, 2011; Rawlani and Sovacool, 2011; Sovacool et al., 2011).

Efforts are also being made to integrate climate change adaptation into Disaster Risk Reduction (DRR) frameworks (Mercer, 2010; Polack, 2010; Romieu et al., 2010; Gero et al., 2011) and adaptation practice is likely to move forward as climate change adaptation (CCA) converges with disaster risk reduction (UNISDR, 2009; Setiadi et al., 2010; Tran and Nitivattananon, 2011; Hay, 2012). In Japan, for example, coastal climate change adaptation has been mainstreamed into the framework of Coastal Disaster Management in the aftermath of the 2011 Tohoku Earthquake Tsunami. The priority of upgrading coastal defenses in the face of sea level rise is thereby judged from the potential damage on the assets in predicted inundation areas on the one hand as well as from the age and earthquake resistance of the coastal structures on the other hand (Central Disaster Management Council, 2011; Committee on Adaptation Strategy for Global Warming in the Coastal Zone, 2011). Other important policy and practice frameworks in place in the coastal zone include poverty reduction and development (Mitchell et al., 2010).

5.5.4.2. Principles, Guidance, and Experiences

Much of the observed adaptation practice deals with the coastal hazards of erosion and flooding (Hanak and Moreno, 2012). In many

Table 5-4 | Community-based adaptation measures.

Impact	Type of option	Measures	Brief description	References
Increased salinity	New and diversified livelihoods	Saline-tolerant crop cultivation	Farmer production of saline-tolerant multi-vegetable varieties and non-rice crops	Ahmed (2010); Rabbani et al. (2013)
	New and diversified livelihoods	Keora nursery	Mangrove fruit production to develop local female entrepreneurship	Ahmed (2010)
	New and diversified livelihoods	Crab fattening	Collection, rearing, and feeding of crabs for 15 days to increase local market value	Pouliotte et al. (2009)
	Structural	Homestead protection	Houses constructed on raised foundations to mitigate salinity ingress	Ayers and Forsyth (2009)
Flooding/ inundation	Socio-technical	Disaster management committees	Multi-community stakeholder committees established to discuss disaster preparedness and response on a monthly basis	Ahammad (2011)
	Socio-technical	Early flood warning systems	Established systems converted into a language and format understood by local communities; warning dissemination through community radio services	Ahmed (2005); Saroar and Routray (2010)
	New and diversified livelihoods	Aquaculture: cage and integrated approaches	Small-scale fish culture in cages on submerged agriculture land; aquaculture integrated with other livelihood practices	Pomeroy et al. (2006); Pouliotte et al. (2009); Khan et al. (2012)
	New and diversified livelihoods	Embankment cropping	Growing different vegetable varieties around heightened shrimp enclosures/coastal polders for productive use of fallow land	Ahmed (2010)
	New and diversified livelihoods	Hydroponics	Cultivating vegetables and other crops on floating gardens	Ayers and Forsyth (2009); Ahmed (2010); Dev (2013)
Cyclones/ storm surges	Structural/hard	Homestead reinforcement	Low-cost retrofitting to strengthen existing household structures, especially roofs; strict implementation of building codes	Sales (2009); Ahmed (2010)
	Structural/soft	Homestead ecosystem protection	Plantation of specific fruit trees around homestead area	Haq et al. (2012)
	Structural/hard	Underground bunker construction	Underground bunker established, providing protected storage space for valuable community assets	Raihan et al. (2010)
Sea level rise	Institutional	Risk insurance mechanisms	Farmers educated on comprehensive risk insurance, focusing on sea level rise and coastal agriculture	Khan et al. (2012)
Multi-coastal impacts	Institutional	Integrating climate change into education	Formal and informal teacher training and curriculum development on climate change, vulnerability, and risk management	Ahmed (2010)
	Institutional	Integrated coastal zone management (ICZM) plan	ICZM plan development at local institutional level, including land and sea use zoning for ecosystem conservation	Sales (2009)
	Structural/soft	Restoration, regeneration and management of coastal habitats	Community-led reforestation and afforestation of mangrove plantations, including integration of aquaculture and farming to increase household income levels	Rawlani and Sovacool, (2011); Sovacool et al. (2012)
	Institutional	Community participation in local government decision-making	Active female participation in local government planning and budgeting processes to facilitate delivery of priority coastal adaptation needs	Faulkner and Ali (2012)
	Institutional/ socio-technical	Improved research and knowledge management	Establishment of research centers; community-based monitoring of changes in coastal areas	Sales (2009); Rawlani and Sovacool (2011)

parts of the world, small island indigenous communities address climate change consequences based on their own traditional knowledge (Percival, 2008; Langton et al., 2012; Nakashima et al., 2012). Long-term adaptation to sea level rise has been confined to a few major projects such as the Venice Lagoon project, the Thames Estuary 2100 project (Box 5-1), and the Delta Programme, Netherlands (Norman, 2009).

Through the Delta Programme, the Dutch government has set out far-reaching recommendations on how to keep the country flood-proof over the 21st century taking into account a sea level rise as high as 0.65 to 1.3 m by 2100. These recommendations constitute a paradigm shift from "fighting" the forces of nature with engineered structures to "working with nature" and providing "room for river" instead (Kabat et al., 2009). The recommendations include soft and environmentally friendly solutions such as preserving land from development to accommodate increased river inundation, maintaining coastal protection by beach nourishment, improving the standards of flood protection, and putting in place the necessary political-administrative, legal, and financial resources (Stive et al., 2011).

From adaptation experiences, good practices (practices that have shown consistently better results and could be used as benchmark) have been derived. For some European cases, for example, McInnes (2006) has collected good practices for coastlines facing coastal erosion, flooding, and landslide events. In the California adaptation study that includes coasts, the lessons learnt include using best available science, decision on goals and early actions, locating relevant partners, identification and elimination of regulatory barriers, and encouragement of introduction of new state mandates and guidelines (Bedsworth and Hanak, 2010). Boateng (2010) presented 15 case studies from 12 countries of best practice in coastal adaptation to help coastal managers and policymakers. Bangladesh provides good examples on awareness raising, disaster warning and control, and protective building measures (Martinez et al., 2011). In general, documentation on good adaptation practices for coasts is improving.

In addition, numerous principles have been set forward. In a broad-scale assessment of climate change threats to Australia's coastal ecosystems, seven principles in adaptation were suggested: clearly defined goals by location, thorough understanding of connectivity within and between ecosystems, consideration of non-climatic drivers, involvement of all relevant stakeholders, easily available and shared data, re-thinking of existing policy and planning constraints, and adaptation at local/regional scales (Hadwen et al., 2011). Based on Oxfam's adaptation programs in South Asia that include coastal communities, additional principles presented include a focus on the poor, vulnerable, and marginalized; community or local ownership; flexible and responsive implementation; and preparation for future and capacity building at multiple levels (Sterrett, 2011). An assessment of worldwide case studies indicates the importance of knowledge transfer of good practice methods for scaling up adaptation strategies in and between regions and beyond the national scale (Martinez et al., 2011).

Further principles reported include: Information on efficient adaptation options alone (as assessed through DA approaches) may not fully serve the needs of managers and must to be supplemented by financial and technical assistance as well as boundary organizations that serve as an interface between science and practice (Tribbia and Moser, 2008). The adaptation and decision-making processes should be participatory and inclusive, integrating all relevant stakeholders in a way that is culturally appropriate (Milligan et al., 2009; Nunn, 2009). The adaptation processes should foster mutual learning, experimentation, and deliberation among stakeholders and researchers (Fazey et al., 2010; Kenter et al., 2011). For example, neither scientific climate knowledge alone nor indigenous knowledge alone is considered sufficient for coastal adaptation (Sales, 2009; Dodman and Mitlin, 2011; Bormann et al., 2012). Finally, since coastal systems are complex, diverse, and dynamic, their governance requires experimentation and learning by doing (Jentoft, 2007).

In summary, a wealth of adaptation activities can now be observed in the coastal zone that depend on technology, policy, financial, and institutional support, and are supported by documentation on good practices (*very high confidence*). ICZM, with its emphasis on integration, is likely to remain a major framework for coastal adaptation. While there is *high agreement* on adaptation principles, there is to date little systematic review of and hence *limited evidence* on why a given principle or approach is effective in a given context (and not in another), which emphasizes the need for research to better understand this context (Section 5.5.3.2). Some of the literature on adaptation practice needs to be treated with caution, because normative principles that have been established *ex ante* are not systematically distinguished from *ex post* evaluations of the experiences carried out. Despite the wealth of coastal adaptation activities, it must, however, be emphasized that meeting the multiple goals of coastal adaptation, improving governance, accounting for the most vulnerable populations and sectors and fully integrating consideration of natural ecosystems is still largely aspirational. Meanwhile, development continues in high-risk coastal areas, coastal ecosystems continue to degrade in many regions, coastal freshwater resources are being overexploited in many highly populated areas, and vulnerability to coastal disasters grows (e.g., Shipman and Stojanovic, 2007; McFadden, 2008; Jentoft, 2009; Mercer, 2010).

5.5.5. Global Adaptation Costs and Benefits

This section reports on studies that provide internally consistent estimates of the direct costs of sea level rise impacts and adaptation at global scales. These studies have used the models FUND and DIVA, which are described in Section 5.4.1. Studies that use computable general equilibrium models and growth models to estimate the indirect and dynamic costs of climate change, including sea level rise, are reviewed in Chapter 10.

Generally, cost estimates are difficult to compare across studies owing to differences in scenarios used, impacts and adaptation options considered, methodologies applied, and baseline conditions assumed. Global adaptation costs have only been assessed for protection via dikes and nourishment. Nicholls et al. (2011) estimate annual adaptation cost in terms of dike construction, dike maintenance, and nourishment to be US$25 to 270 billion per year in 2100 under a 0.5 to 2.0 m GMSLR for 2005–2100. Anthoff et al. (2010) estimate the net present value of dike construction costs for 2005–2100 to be US$80 to 120 billion for 0.5 m GMSLR and US$900 to 1100 billion for a 2 m GMSLR, respectively.

The available global studies show that it is economically rational to protect large parts of the world's coastline during the 21st century against sea level rise impacts of increased coastal flood damage and land loss (Nicholls and Tol, 2006; Anthoff et al., 2010; Hinkel et al., 2013; *limited evidence*, *high agreement*). For dry land and wetlands loss, the FUND model shows that cost-benefit analysis would justify protecting 80% of the exposed coast in all but 15 countries under a GMSLR of 20 to 40 cm per century (Nicholls and Tol, 2006). Using the same method, Nicholls et al. (2008) show that under extreme GMSLR of up to 4 m in 2100, this fraction would drop to 30% to 50%. For coastal flooding, an application of DIVA shows that, for 21st century GMSLR scenarios of 60 to 126 cm, the global costs of protection through dikes (levees) are much lower than the costs of damages avoided through adaptation (Hinkel et al., 2013).

At the same time, costs and benefits of sea level rise impacts and adaptation vary strongly between regions and countries with some developing countries and small islands reaching limits of adaption or not being able to bear the costs of impacts and adaptation (*limited evidence*, *high agreement*) (Section 29.6.2.1). The cost of 1 m of GMSLR in 2100 (considering land loss due to submergence and protection costs) is projected to be above 1% of national GDP for Micronesia, Palau, the Bahamas, and Mozambique (Anthoff et al., 2010). For coastal flooding, annual damage and protection costs are projected to amount to several percentages of the national GDP for small island states such as Kiribati, the Solomon Islands, Vanuatu, and Tuvalu under GMSLR projections of 0.6 to 1.3 m by 2100 (Hinkel et al., 2013). Further substantial costs arise, particularly for developing countries owing to their current adaptation deficit (i.e., coastal defenses are not adapted to the current climate variability), which is not well understood and requires further analysis (Parry et al., 2009). For example, the adaption deficit of Africa with regard to coastal flooding is estimated at US$300 billion (Hinkel et al., 2011) and that of Bangladesh with respect to cyclones at US$25 billion (World Bank, 2011).

Several methodological gaps remain. As there are so few studies on the costs and benefits of sea level rise at a global level, uncertainties are largely unknown and the need for further research is great. The socioeconomic drivers, sea level rise scenarios, and impacts considered as well as damages and losses valued are incomplete. For example, costs of salinity intrusion, land loss due to increased coastal erosion, cost of forced migration due to permanent inundation, the backwater effect, and the impact of sea level rise in combination with other drivers on ecosystems have not been assessed at global scales (Section 5.5.5). Generally for sea level rise impacts, it is difficult to establish a "no adaptation" baseline and the choice of the baseline changes damage costs (Yohe et al., 2011).

Another gap is related to the fact that global studies have focused on protection via hard structures while many more potentially cheaper or socially preferable measures are available including "soft" protection, retreat, and accommodation measures (Section 5.1). Future work needs to consider trade-offs between all available measures. Hard protection measures, for example, may incur additional costs on adjacent unprotected coasts (Brown et al., 2013) or destroy coastal wetlands through coastal squeeze (Section 5.4.2.3). While the costs of "soft" protection measures such as ecosystem-based adaptation (EBA) are largely unknown (Linham and Nicholls, 2010; Engineers Australia, 2012), these may provide additional benefits in the form of a variety of ecosystem services (Alongi, 2008; IUCN, 2008; Anthony et al., 2009; Vignola et al., 2009; Pérez et al., 2010; Espinosa-Romero et al., 2011; McGinnis and McGinnis, 2011; Zeitlin et al., 2012). Finally, it must be noted that protection also further attracts people and development to the floodplain, which in turn increases the risk of potential catastrophic consequence in the case of defense failure. This is particularly true for many coastal cities such as London, Tokyo, Shanghai, Hamburg, and Rotterdam that already rely heavily on coastal defenses (Nicholls et al., 2007).

5.5.6. Adaptation Opportunities, Constraints, and Limits

There is a growing recognition of the potential co-benefits and new opportunities that can be achieved by mainstreaming adaptation with existing local to national goals and priorities (Section 14.3.4). DRR and adaptation share the common goals of reducing vulnerability against impacts of extreme events while creating strategies that limit risk from hazards (IPCC, 2012). This is especially true in coastal areas where extreme flooding events due to severe storm surges are one of the main sources of hazard. Besides, integrating adaptation with national and local planning can also contribute to building resilience in coastal areas.

EBA is considered to be an emerging adaptation opportunity (Munroe et al. 2011) (Section 16.6, Box CC-EA). In coastal areas, the conservation or restoration of habitats (e.g., mangroves, wetlands, and deltas) can provide effective measures against storm surge, saline intrusion, and coastal erosion by using their physical characteristics, biodiversity, and the ecosystem services they provide as a means for adaptation (Borsje et al., 2011; Jones et al., 2012; Cheong et al., 2013; Duarte et al., 2013b; see also Section 5.5.7).

Since AR4, a variety of studies have been published providing a better understanding of the nature of the constraints and limits to adaptation, both generally (Sections 16.3, 16.4) and more specifically in the coastal sector (e.g., Ledoux et al., 2005; Moser et al., 2008; Tribbia and Moser, 2008; Bedsworth and Hanak, 2010; Frazier et al., 2010; Saroar and Routray, 2010; Mozumber et al., 2011; Storbjörk and Hedrén, 2011; Lata and Nunn, 2012).

Constraints specific to coastal adaptation are polarized views in the community regarding the risk of sea level rise and concerns regarding the fairness of retreat schemes in Australia (Ryan et al., 2011); lack of awareness of sea level rise risks and spiritual beliefs in Fiji (Lata and Nunn, 2012); insufficient budget for the development of adaptation policies and other currently pressing issues in the USA (Tribbia and Moser, 2008; Mozumber et al., 2011); distinct preferences for retreat options depending on several social and exposure conditions in Bangladesh (Saroar and Routray, 2010); and the need to provide compensatory habitats under the Habitats Regulations and lack of local public support in the UK (Ledoux et al., 2005). Other relevant constraints include the lack of locally relevant information, resource tenure, and political will, especially critical in developing countries (*robust evidence*, *high agreement*). Besides, a gap exists between the useful climate information provided by scientists and the one demanded by decision makers.

Different constraints typically do not act in isolation, but in interacting bundles (*robust evidence*, *high agreement*). Therefore it is difficult to predict which constraints matter most in any specific context but instead multiple constraints need to be addressed if adaptation is to move successfully through the different stages of the management process (Moser and Ekstrom, 2010; Londsdale et al., 2010; Storbjörk, 2010; *medium evidence*, *high agreement*). Besides, some factors can act as enablers and add to the adaptation capacity, while acting as constraints for others (Burch, 2010; Storbjörk, 2010; *medium evidence*, *high agreement*).

Finally, a common concern emerging from the literature reviews (Biesbroek et al., 2010; Ekstrom et al., 2011) is that some critical constraints arise from the interactions across policy domains, existing laws and regulations, and long-term impacts of past decisions and policies (*low evidence*, *high agreement*).

A limit is reached when adaptation efforts are unable to provide an acceptable level of security from risks to existing objectives and values and prevent the loss of key attributes, components, or services of an ecosystem (Box 16-1; Sections 16.2, 16.5) and may arise as a result of most of the constraints described above.

Regarding coastal areas, it is widely recognized that biophysical limitations arise, for example, in small island developing states where adaptation through retreat to increasing impact of sea level rise in conjunction with storm surges and flooding is not an option due to limited high land availability, creating a temporary and eventually permanent human displacement from low-lying areas (Pelling and Uitto, 2001; *medium evidence*, *high agreement*). Nicholls et al. (2011) show that only a limited number of adaptation options are available for specific coastal areas if sea level exceeds a certain threshold (1 m) at the end of the century.

Regarding natural (unassisted) adaptation, several researchers have examined biophysical limits, for example, of coastal marshes (Craft et al., 2009; Langley et al., 2009; Mudd et al., 2009; Kirwan et al., 2010), and found that under certain nonlinear feedbacks among inundation, plant growth, organic matter accretion, and sediment deposition coastal wetlands can adapt to conservative rates of sea level rise (SRES A1B) if suspended sediment surpasses a certain threshold. In contrast, even coastal marshes with high sediment supplies will submerge near the end of the 21st century under scenarios of more rapid sea level rise (e.g., those that include ice sheet melting).

Increased ocean acidification is expected to limit adaptation of coral reefs to climate change (Boxes CC-OA and CC-CR).

5.5.7. Synergies and Trade-offs between Mitigation and Adaptation

Klein et al. (2007, p. 749) defined trade-offs between mitigation and adaptation as the "balancing of adaptation and mitigation when it is not possible to carry out both activities fully at the same time (e.g., due to financial or other constraints)." Successful adaptive coastal management of climate risks will involve assessing and minimizing potential trade-offs with other non-climate policy goals (e.g., economic development, enhancement of coastal tourism) and interactions between adaptation and mitigation (e.g., Brown et al., 2002; Tol, 2007; Barbier et al., 2008; Bunce et al., 2010).

Adaptation will be the predominant approach to reducing climate risks to coastal communities, populations, resources, and activities over the 21st century as large increases in sea level rise cannot be ruled out (WGI AR5 Section 13.5.2) and because of the time lag between emissions reductions, temperature changes, and impacts on global sea levels (Nicholls et al., 2007, 2011; see also Section 5.5.7). Still, positive synergies and complementarities between mitigation and adaptation in the coastal sector exist.

Since AR4, a series of studies have pointed out that marine vegetated habitats (seagrasses, salt marshes, macroalgae, or mangroves) contribute to almost 50% of the total organic carbon burial in ocean sediments leading to the so-called Blue Carbon (coastal carbon stocks) strategies (Nellemann et al., 2009; McLeod et al., 2011; Duarte et al., 2013b). These strategies aim at exploring and implementing the necessary mechanisms allowing Blue Carbon to become part of emission and mitigation protocols along with other carbon-binding ecosystems such as rainforests (Nellemann et al., 2009). Besides, marine vegetated habitats provide additional functions including the buffering of impacts against storm surges and waves, soil preservation, raising the seafloor, and shelter for fish nursery or habitat protection (Alongi, 2002; Kennedy and Björk, 2009; Duarte et al., 2013b). Consequently, restoration or ecosystem engineering of marine vegetated areas can be considered as a good example of positive synergies between adaptation and mitigation in coastal areas (Borsje et al., 2011; Jones et al., 2012; Duarte et al., 2013b) and should be further explored to be considered as a valid alternative in the portfolio of measures for climate change mitigation and adaptation. Only recently results have been presented on the role of a 1700 ha seagrass restoration in carbon storage in sediments of shallow coastal ecosystems in Virginia (USA). Restored seagrass meadows are expected to accumulate carbon at a rate comparable to ranges measured in natural seagrass meadows within 12 years of seeding, providing an estimated social cost of US\$4.10 ha^{-1} yr^{-1} (Greiner et al., 2013).

Many coastal zone-based activities and various coastal management strategies involve emissions of GHGs. Reduction or cessation of some of them may have positive implications for both mitigation and adaptation. Limiting offshore oil production may imply a net reduction in GHG emissions depending on what form of energy replaces it, but also a reduced risk of oil spills, a reduction of stresses on the marine/coastal ecosystems, and variable socioeconomic impacts on human communities and public health (O'Rourke and Connolly, 2003). This may result in reduced vulnerability or increased resilience and consequently could prove positive for adaptation. However, this measure would increase the vulnerability of countries whose economies are highly dependent on oil extraction.

Some coastal adaptation options may have potentially negative implications on mitigation. Relocation of infrastructure and development out of the coastal floodplains (retreat) will imply increase in one-time GHG emissions due to rebuilding of structures and possible increase in low-density urban development and ongoing transportation-related emissions (Biesbroek et al., 2010). The building or upgrading of coastal protection structures or ports will also imply an increased energy use and GHG emissions related to construction (e.g., cement production) (Boden et al., 2011).

Similarly, actions beneficial for mitigation may result in potential negative impacts for adaptation. A more compact coastal urban design, increasing development in floodplains (Giridharan et al., 2007), or the development of marine renewable energy (Boehlert and Gill, 2010) may introduce additional drivers on coastal systems reducing coastal resilience and adaptive capacity.

5.5.8. Long-Term Commitment to Sea Level Rise and Adaptation

In AR4, both WGI and WGII highlighted the long-term commitment to sea level rise (Meehl et al., 2007; Nicholls et al., 2007), which means that sea levels will continue to rise for centuries due to global warming until reaching equilibrium conditions even if climate forcing is stabilized, because there is a delay in the response of sea level rise to global warming (WGI AR5 Section13.4.1). WGI AR5 has now assessed GMSLR until 2500 and this shows that even with aggressive mitigation measures (RCP2.6), sea level continues to rise after 2100 (Table 5-1; see also WGI AR5 Sections 13.5.1, 13.5.4). With more moderate (RCP4.5.) and little (RCP8.5) mitigation, larger ongoing increases in sea level are expected, lasting for several centuries. Note that the ranges given after 2100 are only model spread and not likely ranges. Looking beyond 2500, Levermann et al. (2013) project that GMSL will rise on average by about 2.3 m per degree Celsius of global warming within the next 2000 years. Under present levels of global warming, this means that we have already committed to a long-term sea level rise of 1.3 m above current levels (Strauss, 2013). For other climate-related drivers, responses to global warming levels are more immediate. For ocean acidification, for example, pH rise would cease several decades after strict CO_2 emission reductions begin (Bernie et al. 2010; see also Section 19.7.1).

This long-term commitment to sea level rise means that there is also a long-term commitment to sea level rise impacts and adaptation. Few studies have considered this and, from a methodological point of view, it is difficult to look at socioeconomic conditions and human responses on such large temporal scales. A limited number of studies have estimated the effects of mitigation on coastal impacts on human settlements and adaptation for the 21st century (Section 5.4.3.1). These studies show that despite the delayed response of sea level rise to global warming, mitigation can reduce impacts significantly already during the 21st century. These studies also show that for most urban areas, coastal protection is cost-efficient in reducing impacts during the 21st century (Section 5.5.5). Past and current adaptation practice also confirms this: cities such as Tokyo and Shanghai have protected themselves against local sea level rise of several meters during the 20th century and the Dutch and UK governments have decided that they can protect urban Netherlands and London against 21st century sea level rise above 1 m (Section 5.5.4). Not protecting cities such as Amsterdam, Rotterdam, and London during the 21st century is not an option. On the other hand, there are coastal areas such as small islands where protecting against several meters of sea level rise in the long term is not a viable option. Failing to mitigate, thus increasingly commits us to a world where densely populated areas lock into a trajectory of increasingly costly hard defenses and rising residual risks on the one hand and less densely populated areas being abandoned on the other hand. Mitigation thus plays, in the long term, a very important role in avoiding climate change impacts in coastal areas by reducing the rate of sea level rise and providing more time for long-term strategic adaptation measures to be adopted. However, even if anthropogenic CO_2 emissions were reduced to zero, sea levels would continue to rise for centuries, making adaptation in coastal areas inevitable.

5.6. Information Gaps, Data Gaps, and Research Needs

This chapter has updated knowledge on the impacts of climate change on the coastal systems not in isolation but also from the perspective of overexploitation and degradation that have been responsible for most of the historical changes. There is a better understanding of the varying impacts of weather and climate extremes and long-term sea level rise on human systems.

That sea levels will rise is a confident projection of climate science but uncertainties around the magnitude of future sea level rise remain large. The rates and magnitude of sea level rise are summarized in Table 5-1 but, under present levels of global warming, we are already committed to 1.3 m future sea level rise above current levels (Section 5.5.8). However, many sea level rise assessments are not provided at spatial or temporal scales most relevant for decision makers who require information on baseline conditions and projections of change (Kettle, 2012) of RSLR (i.e., including local subsidence) for vulnerability assessment and adaptation planning.

Generally, quantitative predictions of future coastal change remain difficult despite the application of improvements in technology—for example, aerial photographs, satellite imagery, Light Detection And Ranging (LiDAR; Sesil et al., 2009; Revell et al., 2011; Pe'eri and Long, 2012)—to investigate and characterize large-scale shoreline changes. There is incomplete understanding of coastal changes over the decade and century time scales (Woodroffe and Murray-Wallace, 2012). Shoreline response is more complex than simple submergence because of factors such as sediment supply, mobilization and storage, offshore geology, engineering structures, and wave forcing (Ashton et al., 2011). The projection of the future impacts of climate change on natural systems is often hampered by the lack of sufficiently detailed data at the required levels of space and time. Although observations have been made on impacts on beaches, rocky coasts, wetlands, coastal aquifers, delta areas, or river mouths by multi-drivers of climate and human-induced origin, there is still an incomplete understanding of the relative role played by each of these drivers and, especially of their combined effect. Uncertainties are even higher when it comes to the evaluation of projected impacts.

For coastal ecosystems, more work needs to be done to develop predictive models based on findings from multi-stressor experiments, both in the field and in the laboratory. Reliable predictions require information on multifactorial experiments performed on communities (preferably in the field), and on time scales of months to years in order to take into consideration the processes of biological acclimation and adaptation.

Although sea level is projected to rise in the future, there are significant gaps in vulnerability assessment of other specific coastal impacts. For example, the modeling of diseases that could affect coastal areas is based mainly on the mean values of climate. Also, despite tourism being one of the most important industries in the coastal areas, not enough is known about tourists' reactions to projected climatic change (Moreno and Amelung, 2009b) or required adaptation measures for port facilities (UNCTAD, 2009).

A wide range of coastal management frameworks and measures is available and used in coastal adaptation to climate change, and the scope for their integration has increased by combining scenarios of climate change and socioeconomic conditions and risk assessment (Kirshen et al., 2012). While various adaptation measures are available, at the local level, there remains insufficient information on assessment of adaptation options, particularly in developing countries.

Data and knowledge gaps exist or their reliability is insufficient. Despite the availability of potentially useful climate information, a gap exists between what is useful information for scientists and for decision makers. For example, at the project level, engineers may have difficulties to "plug in" climate projections presented by scientists. The proposed actions to improve usability include varying levels of interaction, customization, value-adding, retailing, and wholesaling (Lemos et al., 2012) so that data and methods can be more openly accessible to fellow scientists, users, and the public (Kleiner, 2011).

Coastal systems are affected by human and climate drivers and there are also complex interactions between the two. In general, certain components of coastal systems are sensitive and attributable to climate drivers while others are not clearly discernible. For example, data are available on the range shift in coastal plant and animal species and the role of higher temperatures on coral bleaching (see Box CC-CR). However, in many cases in the human systems, the detectable changes can be largely attributed to human drivers (Section 5.3.4). Reducing our knowledge gaps on the understanding of the processes inducing changes would help to respond to them more efficiently.

The economics of coastal adaptation are under-researched. More comprehensive assessments of valuation of coastal ecosystem services, adaptation costs, and benefits that simultaneously consider both the gradual impact of land loss due to sea level rise and the stochastic impacts of extreme water levels (storm surges, cyclones) are needed, as well as other impacts such as saltwater intrusion, wetlands loss and change, and backwater effects. Assessments should also consider a more comprehensive range of adaptation options and strategies, including "soft" protection, accommodation, and retreat options as well as the trade-offs between these.

Governance of coastal adaptation and the role of institutions in the transition toward sustainable coasts are under-researched. While institutional factors are recognized to be decisive in constraining and enabling coastal adaptation, most work remains descriptive. There is a great need for dedicated social science research aimed at understanding institutional change and which institutional arrangements are effective in which socioeconomic and biophysical contexts (Kay, 2012; see also Sections 5.5.3, 5.5.4).

Developing a coastal adaptation knowledge network between scientists, policymakers, stakeholders, and the general public could be considered a priority area for large coastal areas or regional areas affected by climate change and sea level rise. This is well developed in the USA, European Union, the Mediterranean, and Australia but less so in the developing countries, except in certain regions, for example, Caribbean islands and the Pacific Islands.

Future research needs for coastal adaptation are identified by several developments in climate science. Based on the Li et al. (2011) survey of the foci of climate research in the 21st century, the implications for coasts would be on biodiversity and flooding. Future technological advances may be significant—for example, new forms of energy and food production and information and communication technology (ICT) for risk monitoring (Delta Commission, 2008; Campbell et al., 2009; Zevenbergen et al., 2013)—and these would be useful for flood risks and food production in deltas and coastal systems (aquaculture).

With recent adverse climatic and environmental events on coasts, adaptation demands different decision regimes (Kiker et al., 2010) but adaptation, mitigation, and avoidance measures still require integrating research that includes natural and social sciences (CCSP, 2009). Although many gaps still remain, there is nevertheless a greater foundation of climate change research on coasts across a wide range of fields (Grieneisen and Zhang, 2011) upon which scientists, policymakers, and the public may find improved solutions for coastal adaptation.

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6

Ocean Systems

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Frequently Asked Questions

Executive Summary

Ocean ecosystems have responded and will continue to respond to climate changes of different rates, magnitudes, and durations (*virtually certain*). Human societies depend on marine ecosystem services, which are sensitive to climate change (*high confidence*), in particular the provisioning of food (fisheries and aquaculture) and other natural resources; nutrient recycling; regulation of global climate including production of oxygen (O_2) and removal of atmospheric carbon dioxide (CO_2); protection from extreme weather and climate events; and aesthetic, cultural, and supporting services. {6.3, 6.4, 6.5}

Climate change alters physical, chemical, and biological properties of the ocean (*very high confidence*). Oceanic drivers include salinity, circulation, temperature, carbon dioxide (CO_2), oxygen (O_2), nutrients, and light. These drivers shape the physiological performance of individual cells and organisms and ultimately determine ecosystem composition, spatial structure, and functioning. {6.1.1, 6.3}

The fossil record and present field and laboratory observations confirm links between key environmental drivers and responses of ocean ecosystems to climate change (*high confidence*). For millions of years in Earth history, natural climate change at rates slower than today's anthropogenic change has led to significant ecosystem shifts (*high confidence*), including species emergences and extinctions (*high confidence*). Contemporary multi-decadal natural climate variations associated with regional transient warming periods by 1°C have led to fundamental restructuring of ecosystems and large socioeconomic implications (*high confidence*). {6.1.2, 6.3.1, 6.4}

Vulnerability of most organisms to warming is set by their physiology, which defines their limited temperature ranges and hence their thermal sensitivity (*high confidence*). Temperature defines the geographic distribution of many species and their responses to climate change. Shifting temperature means and extremes alter habitat (e.g., sea ice and coastal), and cause changes in abundance through local extinctions and latitudinal expansions or shifts (*very high confidence*). Vulnerability is greatest in polar animals owing to their narrow temperature ranges (*medium confidence*) and in tropical species living close to upper thermal limits (*medium confidence*). Although genetic adaptation occurs (*medium confidence*), the capacity of present-day fauna and flora to compensate for or keep up with the rate of ongoing thermal change is limited (*low confidence*). {6.3.1, 6.3.5, 6.5.2}

The warming-induced shifts in the abundance, geographic distribution, migration patterns, and timing of seasonal activities of species (*very high confidence*) have been and will be paralleled by a reduction in their maximum body size (*medium confidence*). This has resulted and will further result in changing interactions between species, including competition and predator-prey dynamics (*high confidence*). Numerous observations over the last decades in all ocean basins show global-scale changes including large-scale distribution shifts of species (*very high confidence*) and altered ecosystem composition (*high confidence*) on multi-decadal time scales, tracking climate trends. The distribution and abundance of many fishes and invertebrates have shifted poleward and/or to deeper, cooler waters (*high confidence*). Poleward displacements of phyto- and zooplankton have occurred by hundreds of kilometers per decade (*high confidence*). Some warm-water corals and their reefs have responded with species replacement, bleaching, and a decreased coral cover causing habitat loss (*high confidence*). While marine reptiles such as turtles encounter direct effects of warming, impacts to seabirds and marine mammals are mostly indirect through effects of warming on their prey (*high confidence*). {6.3.1, 6.3.7, 6.5, Boxes CC-CR, CC-MB}

In response to further warming by 1°C or more by the mid-21st century and beyond, ocean-wide changes in ecosystem properties are projected to continue (*high confidence*). Large irreversible shifts in the spatial distribution of species and seasonal timing of their activities (feeding, growth, development, behaviors, and productivity) will have implications for species composition, and ecosystem goods and services. {6.3.1, 6.4, 6.5, 6.6}

By the mid-21st century, the spatial shifts of marine species will cause species richness to increase at mid- and high latitudes (*high confidence*) and to decrease at tropical latitudes (*medium confidence*), resulting in global redistribution of catch potential for fishes and invertebrates, with implications for food security (*medium confidence*). Animal displacements are projected to lead to high-latitude invasions and high local extinction rates in the tropics and semi-enclosed seas. This will cause a 30 to 70% increase in the fisheries yield of some high-latitude regions by 2055 (relative to 2005), a redistribution at mid-latitudes, but a drop of 40–60% in the tropics and the Antarctic, based on 2°C warming above preindustrial values (*medium confidence* in the direction of trends in fisheries yields, *low confidence* in the magnitude of change). If a decrease in global net primary production (NPP) or a shift toward smaller primary producers occurs, the overall fisheries catch potential may also decrease. {6.3.1-4, 6.4.1, 6.5.1-4}

Open ocean NPP is projected to fall globally depending on RCP scenario (*medium confidence*). The estimated decrease will occur by up to 9% by 2100 under the RCP8.5 business-as-usual climate scenario (relative to 1990, *low confidence*). The oceans currently provide about half of global NPP. Environmental controls on NPP include temperature, CO_2, nutrient supply, and light (through cloud cover, mixed layer depth), all of which will be altered (WGI AR5 Section 6.3). Present observations indicate increasing NPP at high (Arctic) latitudes (*medium confidence*), projected to continue beyond 2100 (*medium confidence*). This increase is offset by a decrease at temperate and tropical latitudes (*medium confidence*). Poor representation of shelf and coastal regions hamper projections in global NPP models for near-shore waters, reducing confidence in global projections. {6.3.4, 6.5.1, Box CC-PP}

Large-scale processes and climatic feedbacks sustained by microbes (bacteria, archaea, unicellular algae, and protozoans) play key roles in marine ecosystems (e.g., carbon and nitrogen (N_2) fixation or nutrient recycling) and will be altered by climate change (*medium confidence*). Identifying which microbial species, groups, and processes are being affected and how these will be altered is difficult, as these organisms and their responses to environmental change are extremely diverse and often modulated by biological interactions or changes in circulation and nutrient supply (*limited evidence*, *low agreement*). Warming will cause species-specific responses, such as enhancing metabolic rates and exceeding thermal tolerances, which will affect abundance, distribution, and community structure. Warmer, CO_2- and nutrient-enriched coastal oceans may stimulate harmful algal blooms (*medium confidence*), and the redistribution of certain microbes causing diseases such as cholera (*medium confidence*). {6.3, 6.4.2}

Rising atmospheric CO_2 over the last century and into the future not only causes ocean warming but also changes carbonate chemistry in a process termed ocean acidification (WGI AR5 Sections 3.8.2, 6.4.4). Impacts of ocean acidification range from changes in organismal physiology and behavior to population dynamics (*medium* to *high confidence*) and will affect marine ecosystems for centuries if emissions continue (*high confidence*). Laboratory and field experiments as well as field observations show a wide range of sensitivities and responses within and across organism phyla (*high confidence*). Most plants and microalgae respond positively to elevated CO_2 levels by increasing photosynthesis and growth (*high confidence*). Within other organism groups, vulnerability decreases with increasing capacity to compensate for elevated internal CO_2 concentration and falling pH (*low* to *medium confidence*). Among vulnerable groups sustaining fisheries, highly calcified corals, mollusks, and echinoderms are more sensitive than crustaceans (*high confidence*) and fishes (*low confidence*). Trans-generational or evolutionary adaptation has been shown in some species, reducing impacts of projected scenarios (*low* to *medium confidence*). Limits to adaptive capacity exist but remain largely unexplored. {6.3.2, Box CC-OA}

Few field observations conducted in the last decade demonstrate biotic responses attributable to anthropogenic ocean acidification, as in many places these responses are not yet outside their natural variability and may be influenced by confounding local or regional factors. Shell thinning in planktonic foraminifera and in Southern Ocean pteropoda has been attributed fully or in part to acidification trends (*medium* to *high confidence*). Coastward shifts in upwelling CO_2-rich waters of the Northeast Pacific cause larval oyster fatalities in aquacultures (*high confidence*) or shifts from mussels to fleshy algae and barnacles (*medium confidence*), providing an early perspective on future effects of ocean acidification. This supports insight from volcanic CO_2 seeps as natural analogs that macrophytes (seaweeds and seagrasses) will outcompete calcifying organisms. During the next decades ecosystems, including cold- and warm-water coral communities, are at increasing risk of being negatively affected by ocean acidification, especially as ocean acidification will be combined with rising temperature extremes (*medium* to *high confidence*, respectively). {6.1.2, 6.3.2, 6.3.5}

The expansion of hypoxic regions termed Oxygen Minimum Zones (OMZs) and anoxic "dead zones," observed over the last 50 years and projected into the future under climate change, especially if combined with nutrient enrichment (eutrophication), will constrain the habitat of O_2-dependent organisms and benefit anaerobic microbes (*medium confidence*). Hypoxia tolerance varies among species and is influenced by temperature, elevated CO_2, food consumption, and O_2 demand (*high confidence*). Warming-induced stratification limits the exchange of gases between water layers. Enhanced oxygen consumption by heterotrophic organisms depletes the oxygen further, causing a community shift toward lower species richness and hypoxia-tolerant specialists. Under extreme hypoxia ecosystems are dominated by microbes. These OMZs are also characterized by microbial removal of fixed nitrogen (denitrification), which can significantly reduce the low-latitude nutrient inventories with implications for regional productivity. {6.3.3, 6.3.5}

The climate-change-induced intensification of ocean upwelling in some eastern boundary systems, as observed in the last decades, may lead to regional cooling rather than warming of surface waters and cause enhanced productivity (*medium confidence*), but also enhanced hypoxia, acidification, and associated biomass reduction in fish and invertebrate stocks. Owing to contradictory observations there is currently uncertainty about the future trends of major upwelling systems and how their drivers (enhanced productivity, acidification, and hypoxia) will shape ecosystem characteristics (*low confidence*). {6.1.1, 6.3.2, 6.3.3, 6.3.5-6, Box CC-UP}

Environmental drivers acting simultaneously on ocean biota* often lead to interactive effects and complex responses (*high confidence*). Interactions of temperature, ocean acidification, and hypoxia narrow thermal ranges and enhance sensitivity to temperature extremes in organisms such as corals, coralline algae, mollusks, crustaceans, and fishes (*high confidence*). In primary producers, light and individual nutrients can also interact with temperature and acidification. Combined warming and ocean acidification reduce calcification in warm-water corals (*high confidence*). Ocean acidification will alter availability of trace metals (*low confidence*). (*The term biota encompasses the organisms of a region, habitat, or geological period.) {6.3.2.2, 6.3.5, 6.5.2}

The combination and often amplification of global and regional climate change and local anthropogenic drivers result in enhanced vulnerability of natural and human systems (*high confidence*). Major regional and local drivers include fishing, pollution, and eutrophication. {6.3.5, 6.4, 6.5}

The progressive redistribution of species and the reduction in marine biodiversity in sensitive regions and habitats puts the sustained provision of fisheries productivity and other ecosystem services at risk, which will increase due to warming by 1°C or more by 2100 compared to the present (*high confidence*). Human societies respond with limited adaptive capacity. Socioeconomic vulnerability is highest in developing tropical countries involving a risk of reduced supplies, income, and employment from marine fisheries (*high confidence*). This emphasizes disparities in food security between developed and underdeveloped nations. {6.4.1, 6.5}

With continuing climate change, local adaptation measures (such as conservation) or a reduction in human activities (such as fishing) may not sufficiently offset global-scale effects on marine ecosystems (*high confidence*). Effects of climate change will thus complicate management regimes such as of marine protected areas once species undergo distributional shifts. This increases the vulnerabilities of marine ecosystems and fisheries. {6.4.2.1}

Geoengineering approaches involving manipulation of the ocean to ameliorate climate change (such as nutrient fertilization, binding of CO_2 by enhanced alkalinity, or direct CO_2 injection into the deep ocean) have very large environmental and associated socioeconomic consequences (*high confidence*). Some actually require purposeful alteration of ocean ecosystems for implementation. Alternative methods focusing on solar radiation management (SRM) leave ocean acidification largely unabated as they cannot mitigate CO_2 emissions. {6.4.2}

6.1. Introduction: Point of Departure, Observations, and Projections

The oceans cover about 71% of Earth's surface to an average depth of 3700 m. Their importance for life on Earth, including humans, is vast (FAQ 6.1). Marine habitats display natural variability on various spatial and temporal scales but a dearth of long-term observational data from the vast open oceans limits our understanding of the causes and ecological consequences of this variability. The available information indicates that climate controls ocean temperatures, chemistry, circulation, upper ocean stratification, nutrient supply, and sunlight exposure. These drivers affect marine ecosystems through direct effects on organisms, amplified by their changing interactions with other species. Food webs are modified by changes in phytoplankton growth and the availability of live organisms or their decomposing bodies, that is, debris or dissolved organic matter, as food to (chemo-)heterotrophs (organisms gaining energy by feeding on organic matter). Organismal responses lead to changes in biogeochemical processes, such as the carbon cycle, and in biological diversity and the services the oceans provide.

Some impacts of climate change on marine ecosystems and their services were addressed in the IPCC Fourth Assessment Report (AR4): WGII Chapters 4 to 6 (ecosystems, food, coastal areas), and regional chapters, for example, 15 (polar regions) and 16 (small islands). The ecosystem assessment in WGII AR4 Chapter 4 focused on terrestrial, coastal, and marine systems, their properties, goods, and services. It emphasized the difficulty in assessing future ecosystem responses as a result of ecosystem complexity, different vulnerabilities of species, and ecosystem-specific, critical thresholds associated with nonlinear responses to environmental change. Focusing on terrestrial ecosystems, WGII AR4 Chapter 4 concluded that more than 2°C to 3°C warming above preindustrial levels causes high extinction risks to 20 to 30% of present-day species (*medium confidence*), paralleled by substantial changes in ecosystem structure and functioning (*high confidence*). The authors projected that a wide range of planktonic and benthic calcifiers will be impacted by ocean warming (*very high confidence*) and acidification (*medium confidence*), particularly in the Southern Ocean. They characterized sea ice and coral reef biomes as highly vulnerable. Key uncertainties identified in AR4 were the incomplete knowledge of ocean acidification (addressed in present Section 6.3.2), synergistic effects and their mechanisms (Section 6.3.5), biotic feedbacks to the climate system (Section 6.4), and the impacts of interactions between climate change, human uses, and ecosystem management (Section 6.4.2).

Much more than in previous IPCC reports (Figure 1-2), impacts on the oceans are a focus in AR5. This allows for a more comprehensive discussion of phenomena and impacts, as well as the associated uncertainties and the levels of confidence in observed and projected changes. The present chapter focuses on the general principles and processes characterizing climate change impacts on ocean systems and on the uses of these systems by human societies. For projections of responses to climate change, the chapter also assesses our understanding of underlying functional mechanisms causing change across all levels of biological organization, from molecules to organisms to ecosystems. As the ocean is a heterogeneous environment, the comparison of major ocean regions is required to understand variability and differences in key processes and carbon inventories (Box CC-PP, Figure 1). We discuss the changes and variability in the ocean's principal physical and chemical properties and assess knowledge drawn from paleo- and historical to present observations. We develop a conceptual framework for analyzing

Frequently Asked Questions

FAQ 6.1 | Why are climate impacts on oceans and their ecosystems so important?

Oceans create half the oxygen (O_2) we use to breathe and burn fossil fuels. Oceans provide about 17% of the animal protein consumed by the world's human population, or almost 20% of that protein consumed by 3 billion people. Oceans are home to species and ecosystems valued in tourism and for recreation. The rich biodiversity of the oceans offers resources for innovative drugs or biomechanics. Ocean ecosystems such as coral reefs and mangroves protect the coastlines from tsunamis and storms. About 90% of the goods the world uses are shipped across the oceans. All these activities are affected by climate change.

Oceans play a major role in global climate dynamics. Oceans absorb 93% of the heat accumulating in the atmosphere, and the resulting warming of oceans affects most ecosystems. About a quarter of all the carbon dioxide (CO_2) emitted from the burning of fossil fuels is absorbed by oceans. Plankton convert some of that CO_2 into organic matter, part of which is exported into the deeper ocean. The remaining CO_2 causes progressive acidification from chemical reactions between CO_2 and seawater, acidification being exacerbated by nutrient supply and with the spreading loss of O_2 content. These changes all pose risks for marine life and may affect the oceans' ability to perform the wide range of functions that are vitally important for environmental and human health.

The effects of climate change occur in an environment that also experiences natural variability in many of these variables. Other human activities also influence ocean conditions, such as overfishing, pollution, and nutrient runoff via rivers that causes eutrophication, a process that produces large areas of water with low oxygen levels (sometimes called "dead zones"). The wide range of factors that affect ocean conditions and the complex ways these factors interact make it difficult to isolate the role any one factor plays in the context of climate change, or to identify with precision the combined effects of these multiple drivers.

effects on organisms and ecosystems and assess present knowledge derived from experiments, field studies, and numerical model projections mostly using Representative Concentration Pathways (RCPs) of climate change scenarios to provide trajectories of climate change drivers (Moss et al., 2010). Finally, we assess the implications of such changes for ecosystem services, and identify plausible socioeconomic consequences.

Assessing climate change impacts on coastal systems is the topic of Chapter 5. An integrative treatment of regional climate changes and impacts in seven key ocean regions is the focus of regional Chapter 30. Marine issues are also included in regional Chapters 22 to 29, with a focus on polar oceans (Chapter 28) and small islands (Chapter 29). Topics important to several chapters, such as ocean acidification, upwelling systems, primary productivity, changes in biogeography, and coral reefs, are discussed in joint assessments presented in the respective cross-chapter boxes.

6.1.1. Changes in Physical and Chemical Variables

Trends in ocean conditions over the last 60 years reflect significant human impacts beyond natural variability on temperature, salinity, dissolved inorganic carbon and oxygen content, pH, and other properties of the upper ocean (e.g., Pierce et al., 2012; Sen Gupta and McNeil, 2012; WGI AR5 Section 3.8, Table 10.1). With climate change, marine ecosystems are and will be exposed to rising temperature, ocean acidification, expansion of hypoxic zones, and other environmental drivers changing concomitantly.

6.1.1.1. Temperature and Salinity

Over the last 39 years, oceans have warmed at average rates of >0.1°C per decade in the upper 75 m and 0.015°C per decade at 700 m depth (WGI AR5 Section 3.2.2, Figure 3.1). Trends differ regionally, seasonally, and interannually (WGI AR5 Section 2.7; for ocean regions see Section 30.5 in the present volume). Temperature changes are particularly large at El Niño-Southern Oscillation (ENSO) with high (3- to 4-year) and low (5- to 7-year) frequencies, and on multi-decadal scales (>25 years, Figure 6-1). The strongest warming trends are found at high latitudes where most of the inter-decadal variability occurs, while tropical oceans are dominated by interannual frequencies. Global climate models have explored changes in different frequency domains, but their spatial resolution is poor (WGI AR5 Sections 11.3.3, 12.4.7).

Temperature variations are often accompanied by changes in salinity. Increased salinity results from reduced precipitation relative to evaporation, for example, above the thermoclines (layer separating the upper mixed layer from deeper water where temperature and density change rapidly with depth) of subtropical gyres at mid- to low latitudes since 1950 (WGI AR5 Chapter 3). Decreased salinity due to enhanced precipitation relative to evaporation has occurred at some tropical and higher latitudes, exacerbated by sea ice melt (Durack et al., 2012). Both warming and freshening cause enhanced density stratification, a trend projected to continue into the 21st century (WGI AR5 Chapter 3, Section 11.3.3, Figure 12.34; Helm et al., 2010). Mean sea surface temperature in 2090 will be 2.7°C warmer than in 1990 (RCP8.5; WGI AR5 Chapter 12; Bopp et al., 2013).

6.1.1.2. Carbon Dioxide-induced Acidification

Rising carbon dioxide (CO_2) concentrations in air (given as partial pressures, *p*CO2, in µatm) cause increasing upper ocean CO_2 levels (Watson et al., 2009). Starting from a preindustrial value of 280 µatm atmospheric pCO_2 levels will have reached around 500 µatm by 2050 following the Special Report on Emissions Scenarios (SRES; IPCC, 2000) and all RCPs (Moss et al., 2010; Meinshausen et al., 2011). By 2100 values are projected to reach between 420 µatm and 940 µatm depending on the RCP. The rise in pCO_2 causes ocean acidification (OA), measured as a decline in water pH (negative log of proton concentration), accompanied by a fall in both carbonate ion (CO_3^{2-}) concentration and the saturation states (Ω) of various calcium carbonates ($CaCO_3$; Zeebe and Westbroek, 2003; WGI AR5 Section 3.8.2, Box 3.2, Chapter 6, Figure 6.29). Hence, the seawater solubilities of three forms of $CaCO_3$, namely calcite, magnesium-calcite, and aragonite, increase. These minerals are important components of shells and skeletons of many marine organisms (Section 6.3.2).

Ocean acidification occurs on a background of natural temporal and spatial variability of pH, pCO_2, and Ω. In the open ocean, the mean pH (total scale, pH_T) of surface waters presently ranges between 7.8 and 8.4 (WGI AR5 Section 3.8.2). In stratified mid-water layers, largely isolated from gas exchange between surface waters and air, decomposition of organic material leads to lowered oxygen (O_2) and elevated CO_2 levels (Paulmier et al., 2011) associated with lower pH values. The few existing field data of sufficient duration, resolution, and accuracy (WGI AR5 Figure 3.18) show that trends in anthropogenic OA clearly deviate from the envelope of natural variability (Friedrich et al., 2012). OA presently ranges between –0.0013 and –0.0024 pH_T units per year (WGI AR5 Section 3.8.2, Table 3.2, Box 3.2; Dore et al., 2009). Average surface ocean pH has decreased by more than 0.1 units below the preindustrial average of 8.17. By 2100 pH is expected to change by –0.13, –0.22, –0.28, and –0.42 pH_T units, at CO_2 levels of 421, 538, 670, and 936 ppm under RCP2.6, 4.5, 6.0, and 8.5 climate scenarios, respectively (WGI AR5 Figure 6.28). The rate of acidification in surface waters varies regionally and is 50% higher in the northern North Atlantic than in the subtropical Atlantic (Olafsson, 2009). Salinity reduction caused by ice melt or excess precipitation (Jacobs and Giulivi, 2010; Vélez-Belchí et al., 2010) exacerbates OA by diluting the concentrations of substances acting as buffers (Steinacher et al., 2009; Denman et al., 2011). At high sustained CO_2 concentrations the changes in ocean chemistry will take thousands of years to be buffered by the natural dissolution of $CaCO_3$ from sediments and tens to hundreds of thousands of years to be eliminated completely by the weathering of rocks on land (Archer et al., 2009).

6.1.1.3. Hypoxia

The average dissolved oxygen concentration in the ocean is presently 162 µmol kg^{-1} (Sarmiento and Gruber, 2006). Concentrations range from over 500 µmol kg^{-1} in productive Antarctic waters super-saturated with oxygen (Carrillo et al., 2004) to zero in coastal sediments and in permanently anoxic deep layers of isolated water bodies, such as the Black Sea and the Cariaco Basin. Hypoxia results from oxygen depletion in excess of supply as in stratified water bodies (Section 6.1.1.2). Vast Oxygen Minimum Zones (OMZs) exist between less than 100 and more

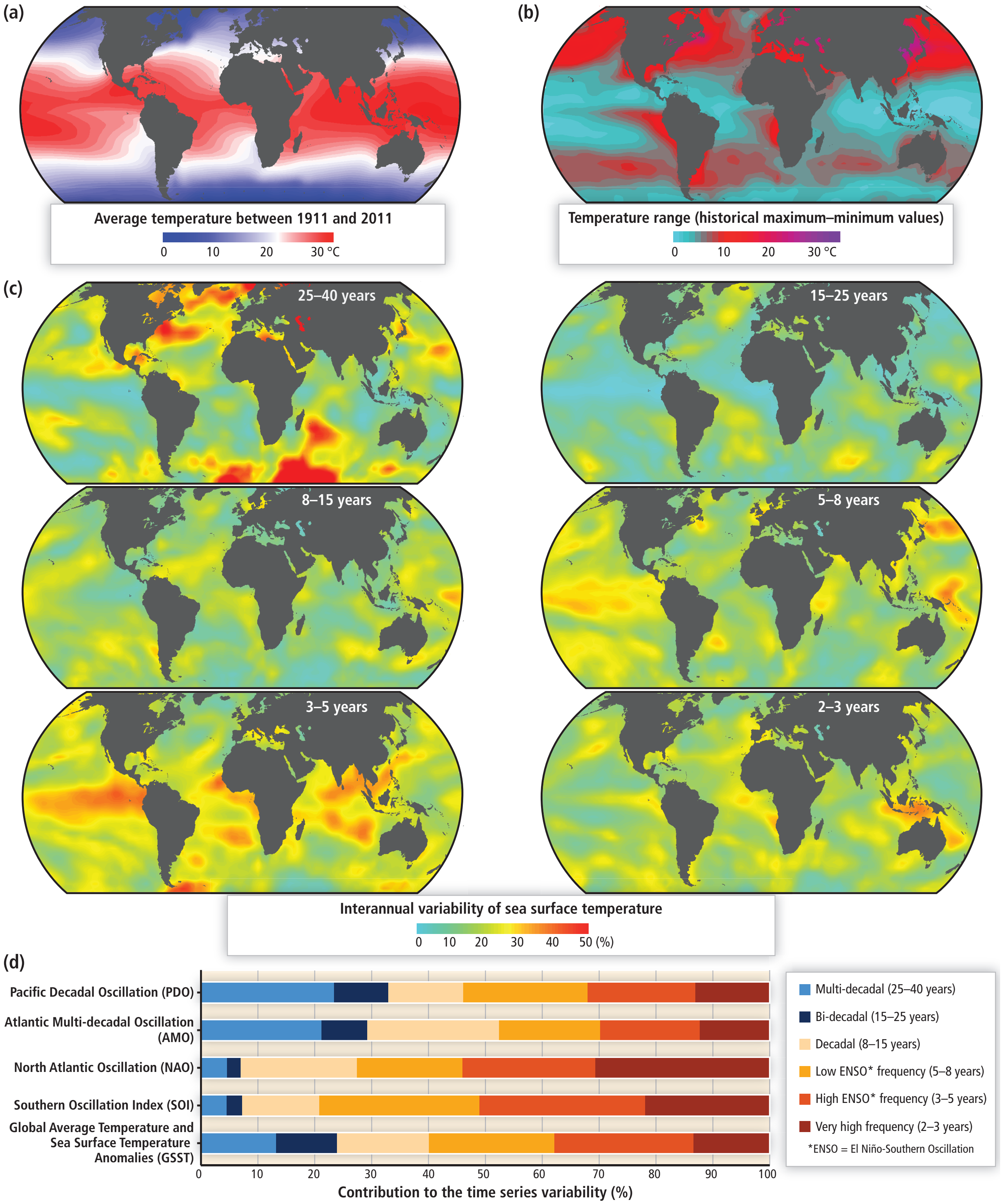


Figure 6-1 | Sea surface temperature variability between 1911 and 2011. (a) The sea surface temperature average for the period. (b) The temperature range calculated as the difference between the maximum and minimum values for each grid component during the century. (c) The spatial distribution of variability by time scales (based on the Extended Reynolds Sea Surface Temperature, NOAA, 2012) corresponds to the multi-decadal (25 to 40 years), bi-decadal (15 to 25 years), decadal (8 to 15 years), low ENSO (El Niño-Southern Oscillation) frequency (5 to 8 years), high ENSO frequency (3 to 5 years), and very high frequency (2 to 3 years) scales. The summed variabilities from the same 2°x2° box in all six maps corresponds to 100% of the time series variability. (d) The spectral density of some of the most widely used climate indices, accumulated in the same frequency windows. The total bar length (100%) corresponds to the cumulative variability of each time series between the 2 and 40 year frequency window. Climate indices were obtained from the NOAA ESRL Physical Sciences Division website.

than 900 m depths in Eastern Atlantic and Pacific tropical oceans. The ecological literature applies the term hypoxia (see Section 6.3.3) to O_2 concentrations below 60 µmol kg^{-1} (estimated at about 5% of global ocean volume; Deutsch et al., 2011). Pacific OMZs regularly reach oxygen levels below 20 µmol kg^{-1} (about 0.8% of global ocean volume; Paulmier and Ruiz-Pino, 2009), lower than Atlantic ones. Suboxic waters at <4.5 µmol O_2 kg^{-1} occupy about 0.03% of the ocean volume, mainly in the northeastern tropical Pacific (Karstensen et al., 2008).

OMZs are naturally present in many habitats including marine sediments, but are also expanding due to anthropogenic influences. Over the past 50 years, open ocean O_2 concentrations have decreased by a mean rate of 0.1 to >0.3 µmol kg^{-1} yr^{-1} (WGI AR5 Section 3.8.3; Stramma et al., 2008). In some OMZs the rate has been much higher due to warming, increased stratification, and rising biological O_2 demand (WGI AR5 Section 3.8.3). Long-term declines in O_2 by about 7 µmol kg^{-1} per decade have been documented at mid-water depths over much of the subarctic North Pacific (Keeling et al., 2010). In coastal regions, extremely hypoxic "dead zones" that exclude animal life, have increased from 42 reported in the 1960s to more than 400 in 2008 and been attributed to high oxygen demand from eutrophication, the local enrichment of nutrients, resulting in organic matter loading and its decay as well as nitrous oxide formation and release (Naqvi et al., 2000; Díaz and Rosenberg, 2008; Zhang et al., 2010).

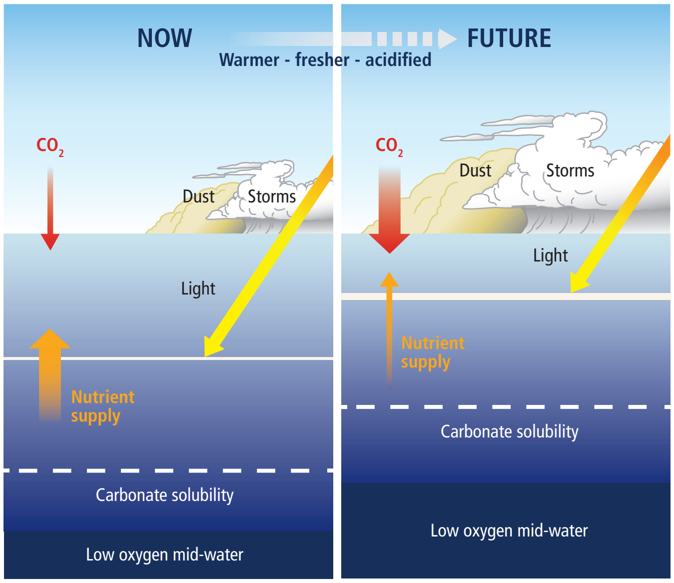


Figure 6-2 | Projected alteration (magnitude and frequency) of oceanic fluxes and atmospheric events due to a changing climate in the coming decades. Ocean properties will be altered from the sunlit surface layer to the mid-water stratum. In the surface ocean, the depth of the mixed layer (solid horizontal line) will shallow resulting in higher mean light levels. Increased density stratification (i.e., a strengthening sea water density gradient represented by the increasing thickness of the solid horizontal line) will reduce the vertical supply of nutrients for photosynthesizing organisms residing in the mixed layer. Anthropogenic CO_2 will acidify, that is, lower the pH of the surface ocean (note this happens in a pH range higher than 7 such that oceans will remain alkaline but less so due to acidification). The penetration of acidified waters to depth will result in a shallower depth (dashed horizontal line) at which $CaCO_3$ structures, such as shells, dissolve. At depth, the location of low-O_2 waters will progressively become shallower. In addition, changes in storm activity and dust deposition will influence ocean physics and chemistry, with consequent effects on ocean biota and hence ecosystems (courtesy of Reusch and Boyd, 2013).

Future warming will *likely* accelerate the spread of hypoxic zones, especially in temperate to sub-polar regions. Most models project decreasing global ocean oxygen contents by 1 to 7% from present-day concentrations in 2100 (Keeling et al., 2010; WGI AR5 Figure 6.30 under RCP8.5), with a mean decline by 3.4% in 2090 compared to the 1990s (Bopp et al., 2013). Warming and freshening of the surface layer will increase stratification and reduce the depth of winter mixing. The evolution of low O_2 zones will be linked to changes in fluvial runoffs (e.g. Milly et al., 2008; see also Section 5.3.4.3), the wind regime (e.g., Vecchi and Soden, 2007), as well as the intensity, duration, and seasonal timing of upwelling events (Snyder et al., 2003; see also Section 30.5.2). The potential contributions of destabilized methane hydrates and bacterial methane oxidation to exacerbate hypoxia and acidification at high latitudes remain to be explored (Westbrook et al., 2009). Currently, there is no consensus on the future volumes of hypoxic and suboxic waters because of large uncertainties in potential biogeochemical effects and in the evolution of tropical ocean dynamics due to both natural and anthropogenic causes (WGI AR5 Section 6.4.5). While volumes with O_2 concentrations <80 µmol kg^{-1} are projected to increase by several percent, suboxic waters <5 µmol O_2 kg^{-1} may undergo a 30% increase by 2100 compared to 2005 (*low confidence*; Bopp et al., 2013).

6.1.1.4. Light and Nutrients

Most models project that the mixed layer at the ocean surface (see Figure 6-2) will become shallower in the coming decades through a strengthening of the vertical density gradient (e.g., Sarmiento et al., 1998; Sallée et al., 2013). Mean light levels encountered by phytoplankton are set by incoming light from solar radiation, the depth of the mixed layer, and the degree to which underwater light is attenuated by living and non-living particles (Kirk, 1994). A shallower mixed layer will *likely* result in the resident phytoplankton receiving higher mean underwater light levels if the organisms are physically mixed through this stratum (Figure 6-2).

Enhanced, seasonally prolonged stratification (Holt et al., 2010), especially in the tropics, the North Atlantic, the Northeast Pacific, and the Arctic (Capotondi et al., 2012), will lead to decreased vertical transport of nutrients to surface waters (Doney, 2010; Figure 6-2). River plumes (Signorini et al., 1999), nutrient accumulation in the pycnocline as reported for North Pacific waters (Whitney, 2011), human-induced eutrophication, enhanced upwelling (Box CC-UP), and tidal mixing and estuarine circulation in coastal oceans could partly compensate for the projected reduction in nutrient supply in the oceans (*limited evidence*, *medium agreement*).

6.1.2. Historical and Paleo-Records

6.1.2.1. Historical Observations

Ocean ecosystems are variable in time and space, and in a non-steady-state, reflected in indices such as the North Atlantic Oscillation (NAO) Index, the Atlantic Multi-decadal Oscillation (AMO), the Arctic Climate Regime Index (ACRI), Pacific Decadal Oscillation (PDO), or the El Niño-Southern Oscillation (ENSO) (WGI AR5 Box 2.5; Figure 6-1; Section 30.5).

The combination of large, global data sets such as Reynolds, National Center for Atmospheric Research (NCAR), International Comprehensive Ocean-Atmosphere Data Set (ICOADS) with multi-decadal time series, for example, near Hawaii (HOT), Bermuda (BATS), the Ligurian Sea (DYFAMED), the Canaries (ESTOC), Kerguelen Island (KERFIX), Hokkaido Island (KNOT), and Taiwan (SEATS) has provided data on the physical and biogeochemical state of the oceans (Karl et al., 2003). These have been augmented by the limited-term, high-resolution programs World Ocean Circulation Experiment (WOCE) and Joint Global Ocean Flux Study (JGOFS).

Historical data sets provide baseline information on ecosystem states and document the responses of biota to both natural variability in the ocean system and surface ocean warming since the 1970s (Figure 6-3; Section 6.3.1). Such data sets are rare and regionally biased. Examples include changes in geographic ranges of plankton and seasonal timing (phenology) of different components of the ecosystem detected by the Continuous Plankton Recorder (CPR: e.g., Edwards et al., 2001; Richardson et al., 2006; Box 6-1) or multi-decadal shifts in pelagic ecosystems (CalCOFI) including higher parts of the food chain such as sardines and anchovies (Brinton and Townsend, 2003; Chavez et al., 2003; Lavaniegos and Ohman, 2003; see also Section 6.3.1) and the skeletal archives of long-lived organisms such as coralline algae (Halfar et al., 2011), bivalves (Schöne et al., 2003), and corals (De'ath et al., 2009).

Systematic, long-term interdisciplinary observations using repeated, highly calibrated measurements at a given field site are required to capture high- and low-frequency events, for example, regime shifts (abrupt changes between contrasting, persistent states of any complex system; deYoung et al., 2008). Direct observations are complemented by satellite remotely sensed data sets. Ocean color data (e.g., Coastal Zone Color Scanner (1978–1986), Sea-Viewing Wide Field-of-View Sensor (SeaWiFS, 1997–2010), and Moderate Resolution Imaging Spectroradiometer (MODIS-AQUA, 2002 to the present); McClain, 2009) provide estimates of chlorophyll concentrations (a proxy for phytoplankton stocks and net primary production (NPP); Sections 6.2.1, 6.3.4; Saba et al., 2011). Total chlorophyll cannot be measured from space; therefore, the near-surface value (approximately one optical depth) is extrapolated to whole water-column chlorophyll based on vertical distribution using region-specific algorithms. Large uncertainties persist, as these estimates reflect both phytoplankton stocks and their physiological status (Dierssen, 2010; Behrenfeld, 2011). The approximately 15-year archived time series of SeaWiFS is too short to reveal trends over time and their causes. It is an example for the general issue that undersampling of ocean phenomena in time and space limits our current ability to assess present states, to distinguish effects of anthropogenic change from natural variability, and to project future changes (Henson et al., 2010; Beaulieu et al., 2013; Box CC-PP).

6.1.2.2. Paleontological Records

Paleontological records in marine sediments provide long-term, low-resolution data on the spatial distributions of organisms and their abundances from all ages and latitudes. This information can be readily related to the concurrent shifts in multiple environmental properties that are also recorded in these sediments. The records provide insights into shifts, expansions, and contractions of biogeographic ranges; species extinctions and emergences; and changes in species abundance, as well as the environmental forcings to which organisms respond. Temporal trends reveal influences of temperature, hypoxia, CO_2, and food availability on organisms and ecosystems (Section 6.1.1; Figure 6-3).

Owing to insufficient resolution, the geological record often does not allow the direct attribution of a biological change to a single driver or the identification of various drivers and their relative importance. Support for projections of future changes in present-day ecosystems and their services is thus limited (*low confidence*; Sections 6.4, 6.5). Nonetheless, information gained from the geological record is invaluable, as both paleo and present climatic shifts share the same combination and sign of environmental changes: increasing atmospheric CO_2 causing warming and CO_2 enrichment in the surface ocean, leading to enhanced stratification of the upper ocean and a decrease in dissolved O_2 (WGI AR5 Chapter 3; Section 5.3). A combination of models (WGI AR5 Chapters 3, 6, 12) and geological data can be used to forecast future impacts on ocean biota (*medium confidence*).

The last glacial-interglacial transition is associated with an average increase in atmospheric CO_2 of approximately 1 µatm per century between 18 and 10 thousand years before present (kyr BP) (WGI AR5 Chapter 5), a significantly slower increase than the approximately 90 µatm in the last century (WGI AR5 Chapters 5, 6). Consequently, the average pH change of 0.002 pH units per century during the glacial-interglacial transition is small relative to the ongoing anthropogenic perturbation of >0.1 pH unit per century (WGI AR5 Section 3.8.2). Overall the upper glacial ocean was more O_2-rich than today's ocean (Jaccard and Galbraith, 2012) and between 0.7°C and 2.7°C colder, with strong regional differences of up to 10°C cooling in the North Atlantic and 2 to 6°C in the Southern Ocean (WGI AR5 Chapter 5, Table 5.2). During warming from the glacial into the interglacial marine plankton such as foraminifera, coccolithophores, diatoms, dinoflagellates, and radiolarians showed marked poleward range expansion (*high confidence*; see WGI AR5 Section 5.7; CLIMAP Project Members, 1976; MARGO Project Members, 2009). Under the lower glacial CO_2 concentrations, calcification in planktonic foraminifera was higher (*limited evidence*, *medium agreement*).

The most prominent abrupt climate change periods in the recent geological record, developing within 10 to 100 years, are associated with Dansgaard-Oeschger (DO) and Heinrich events (WGI AR5 Section 5.7), which occurred repetitively during the last 120 kyr. Whereas the atmospheric changes happened within a few decades, the sea surface temperature in the North Atlantic changed by up to 5°C within decades to centuries (WGI AR5 Section 5.7). Southern Ocean temperature changes were slower (hundreds to thousands of years; Barker et al., 2009). The cold phase of a DO event led to the migration of polar foraminiferal species toward the equator, in the North Atlantic as far south as the Iberian Peninsula (Martrat et al., 2004). Abrupt (approximately 100-year) abundance changes in the Southern Ocean were associated with latitudinal shifts in the Antarctic Circumpolar Current and associated species (Barker et al., 2009) akin to modern changes in plankton range due to warming (Box CC-MB, Box 6-1). During the DO warm phases the Monsoon-driven Arabian Sea upwelling records show enhanced primary

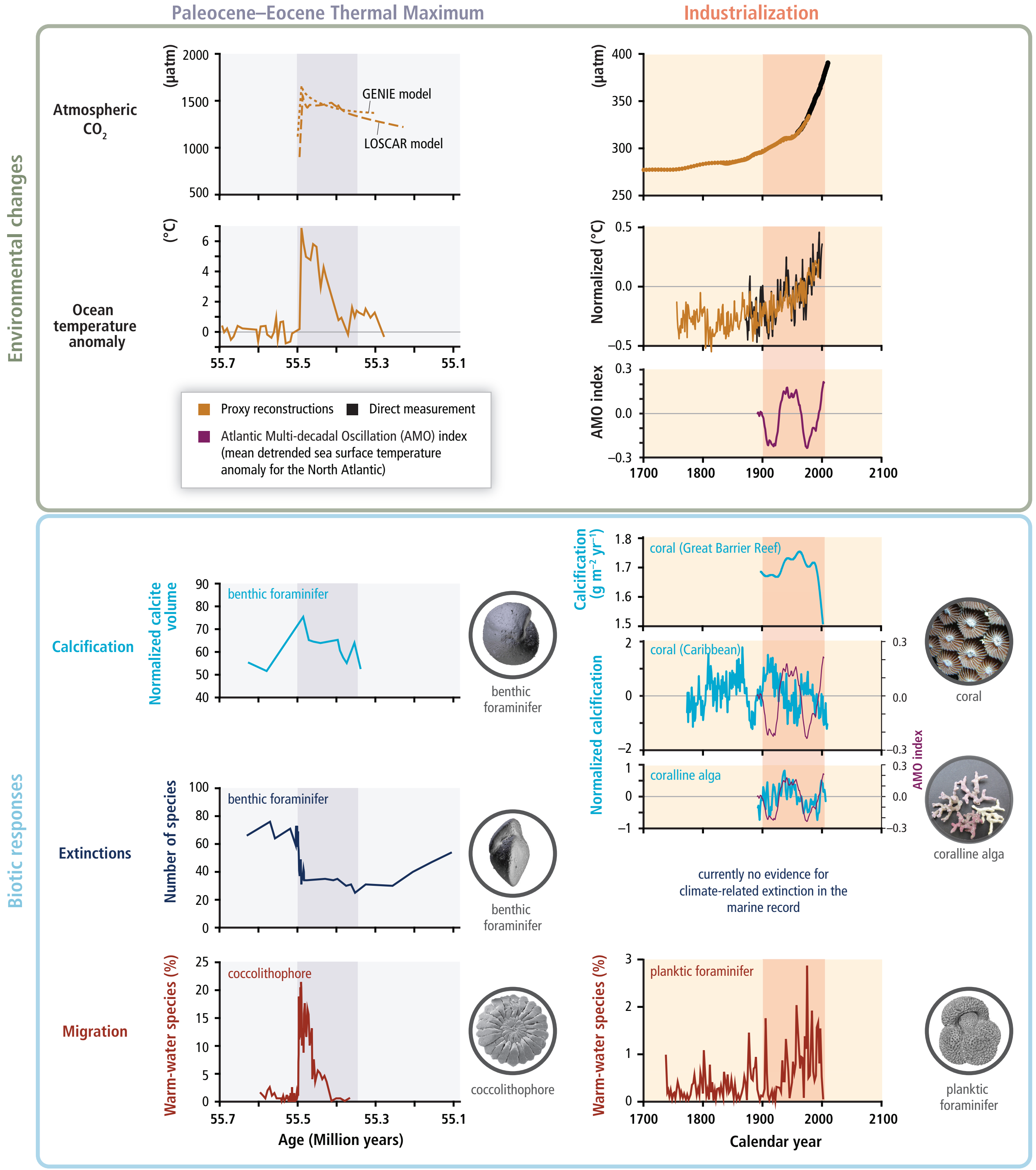


Figure 6-3 | Environmental changes (top) and associated biological responses (bottom) for the Paleocene–Eocene Thermal Maximum (PETM, left) and the industrial era (right). The PETM represents the best geological analog for the future ocean because of its rapid environmental change. Episodes of largest environmental change are indicated with darker bands. Note the different time scale between the two columns. Both time intervals are characterized by rapid warming both on land and in the ocean (modern: Wilson et al., 2006 and PETM: Kennett and Stott, 1991) and increases in CO$_2$ (modern: Etheridge et al. 1996; Keeling et al., 2005 and PETM: Zeebe et al., 2009 (LOSCAR model); Ridgwell and Schmidt, 2010 (Grid Enabled Integrated Earth System Model (GENIE model))). For the recent industrial era, the Atlantic Multi-decadal Oscillation (AMO; see Figure 6-1 and Section 6.1.2.1) is shown to highlight an example of high-frequency sea surface temperature fluctuations (Enfield et al., 2001) and their influence on marine biota. Note the species-specific calcification responses to climate change with decreases, increases, and high variability (coralline alga: Halfar et al., 2011; coral: Vásquez-Bedoya et al., 2012; De'ath et al., 2013; PETM: Foster et al., 2013). While there was extinction during the PETM (Thomas, 2003), there is currently no evidence for climate-related extinction in the marine record. Warming led to migration of warm-water species into previous cold-water habitats (modern: Field et al., 2006; PETM: Bralower, 2002). Pictures are examples of organisms highlighting the processes in each panel, and are not to scale.

and export production, reduced oxygenation, and denitrification, all within approximately 200 years (Higginson et al., 2004).

The last time the atmospheric CO_2 content approached that of today was during the Pliocene warm period (3.3 to 3.0 million years ago (Ma)), with long periods of atmospheric CO_2 levels between 330 and 400 µatm (Pagani et al., 2010; Seki et al., 2010) and equilibrated temperatures approximately 2°C warmer than today (*medium confidence*; Haywood et al., 2009; WGI AR5 Chapter 5). The Mid-Pliocene Warm Period saw a poleward expansion of tropical planktonic foraminifera (*high confidence*; Dowsett, 2007). Coccolithophores (Bown et al., 2004), corals (Jackson and Johnson, 2000), and mollusks (Vermeij and Petuch, 1986) remained unaffected with respect to rates of species extinction or emergences compared to background rates.

Perhaps the best analog for the future ocean is the Paleocene-Eocene Thermal Maximum (PETM, 55.3 Ma). The PETM was an event of warming (Dunkley Jones et al., 2013), and ocean acidification (Zachos et al., 2005) over millennia (Cui et al., 2011; Stassen et al., 2012) with increased runoff and nutrients into the shelf ecosystems. Model simulations for the PETM show 10 times lower rates of CO_2 input and hence ocean acidification compared to today (*medium confidence*; Ridgwell and Schmidt, 2010). Depending on the assumed rate and magnitude of the CO_2 release, models project pH declined by 0.25 to 0.45 units in PETM surface waters and a reduction in surface ocean aragonite saturation from $\Omega = 3$ to $\Omega = 2$ or even as low as 1.5 (Ridgwell and Schmidt, 2010). Warming caused range expansions of warm-water taxa toward higher latitudes (*high confidence*). The composition of plankton assemblages changed both within and between phytoplankton groups (Gibbs et al., 2006; Sluijs and Brinkhuis, 2009), possibly reflecting the warming trend and/or changes in nutrient availability (Sections 6.2.2-3). There was no bias in extinction toward more heavily calcifying species, possibly as slow CO_2 input led to minor surface water acidification. By contrast, benthic foraminifera, the dominant deep water eukaryote, recorded up to 50% extinction (Thomas, 2007). In contrast to sediment dwellers, more mobile pelagic crustaceans (ostracods) did not show any significant change in species composition (Webb et al., 2009). In shallow coastal waters, calcareous algae and corals were replaced by symbiont-bearing benthic foraminifera (*medium confidence*; Scheibner and Speijer, 2008).

The warm climates of the Mesozoic (251 to 65 Ma) led to a number of anoxic events in the oceans (Jenkyns, 2010). In some cases, OMZs expanded vertically, leading to anoxia in upper water layers (Pancost et al., 2004). Some of the Cretaceous oceanic anoxic events were associated with extinctions or increased species turnover (normalized sum of originations and extinctions) of planktonic foraminifera and radiolarians (30%). Such turnover was very small in other groups of organisms (e.g., a maximum of 7% of coccolithophores; Leckie et al., 2002). The attribution of these evolutionary changes to reduced O_2 is tenuous as warming, changes in nutrient supply, and possibly ocean acidification occurred concomitantly (Hönisch et al., 2012).

Global-scale collapse of marine ecosystems is rare, even in the geological record. Some mass extinctions, in particular the Permian Period extinction 251 Ma, have been associated with large-scale inputs of carbon into the atmosphere and ocean, with associated warming and deep-sea O_2 decline (Knoll et al., 2007; Kiessling and Simpson, 2011). The end-Permian mass extinction preferentially affected reef organisms such as corals and sponges resulting in a 4 Myr period without reef builders (Kiessling and Simpson, 2011), and underscores that vulnerabilities differ among organisms depending on anatomy, physiology, and ecology (Knoll and Fischer, 2011). The rates of environmental change and any potential acidification have not yet been accurately constrained for these events.

Of the last 100 Myr, only the last 2 Myr had CO_2 levels of approximately 190 to 280 ppm, comparable to preindustrial values. Values like those predicted for the mid and end of this century can solely be found in the geological record older than 33 Ma, with large uncertainties in the absolute numbers (WGI AR5 Section 5.3; Hönisch et al., 2012). That marine biota thrived throughout high CO_2 times cannot imply that marine organisms will remain unaffected in a future warm, high-CO_2 world. The key environmental issue of the 21st century is one of an unprecedented rate of change, not simply magnitude, of CO_2 levels (Hönisch et al., 2012). The current rate and magnitude of ocean acidification are at least 10 times faster than any event within the last 65 Ma (*high confidence*; Ridgwell and Schmidt, 2010) or even 300 Ma of Earth history (*medium confidence*; Hönisch et al., 2012). The slower events in geological history provide *robust evidence* (*high agreement*) for environmentally mediated changes in biogeographic ranges of fauna and flora, their compositional changes, extinctions, and, to much lesser degree, emergences (*very high confidence*). No past climate change event perfectly parallels future projections of anthropogenic climate change, which is unprecedented in evolutionary history. Existing similarities indicate, however, that future challenges (Sections 6.1.1, 6.3.1-8) may be outside the adaptive capacity of many organisms living in today's oceans (*low* to *medium confidence*).

6.2. Diversity of Ocean Ecosystems and Their Sensitivities to Climate Change

Global-scale observation and modeling studies provide *robust evidence* of present and future climate-mediated alterations of the ocean environment (*high agreement*; Section 6.1.1; WGI AR5 Chapters 3, 6; Bopp et al., 2013), which in turn impact ocean ecosystems (*high confidence*; Boyd and Doney, 2002; Drinkwater et al., 2010; Hoegh-Guldberg and Bruno, 2010). An assessment of present findings and projections requires knowledge of the characteristics of ocean biota and ecosystems and their climate sensitivity.

Life on Earth is diverse as a result of nearly 4 billion years of evolutionary history. Marine microorganisms are the oldest forms of life and the most functionally diverse; multicellular organisms are constrained to limited functional abilities. Knowledge of overarching similarities across the organism domains Archaea, Bacteria, and Eukarya (Woese et al., 1990) or kingdoms Bacteria, Protozoa, Fungi, Plantae, Animalia, and Chromista (Cavalier-Smith, 2004) would facilitate projections of climate impacts. The phylogenetic and metabolic diversity of microbes (i.e., viruses, archaea, bacteria, protists, and microalgae) sustains key ecosystem processes such as primary production, CO_2 fixation and O_2 production, the conversion of nitrogen into ammonia (N_2 fixation), and the use of nitrate, sulfate, CO_2, and metals (iron and manganese) in metabolism instead of O_2 when it is absent. Microbes enhance the horizontal transfer of genetic information between unrelated individuals, thereby enhancing biodiversity (McDaniel et al., 2010). Microbes may respond to climate change by exploiting their large diversity, undergoing species replacements (Karl et al., 2001), and thereby sustain their biogeochemical roles. Species replacements also occur among plants and animals, but in most cases research has focused on their resilience, well-being, abundance, survival, and conservation under climate change (FAQ 6.2).

6.2.1. Pelagic Biomes and Ecosystems

Pelagic organisms are key to biogeochemical processes in the ocean. The base of the marine food web is the photosynthetic fixation of CO_2 by phytoplankton, a process termed (net) primary production (NPP; Box CC-PP). Photosynthesis is controlled by light, temperature, inorganic nutrients (CO_2, nitrate, phosphate, silicate, and trace elements including iron), and the density-dependent stability of the surface mixed-layer depth (MLD) (Section 6.1.1; Figure 6-2; Sverdrup, 1953; González-Taboada and Anadón, 2012). Environmental variability and the displacement of organisms by ocean currents cause variability in phytoplankton productivity, competitiveness, and natural selection (Margalef, 1978) and result in changes in carbon sequestration (Box CC-PP; Figure 6-4). Nutrient limitation leads to a decrease in NPP or chlorophyll levels and a reduction in the amount of energy supplied to higher trophic levels, including fish and invertebrates (*high confidence*; Ware and Thomson, 2005; Brander, 2007), affecting fishery yields (Cheung et al., 2008; Friedland et al., 2012). The wide range of trophic structures in marine food webs and the potentially nonlinear changes in energy transfer under different NPP and temperature scenarios (Stock and Dunne, 2010) hamper accurate projections of changes in higher trophic levels.

6.2.2. Benthic Habitats and Ecosystems

The ocean's primary production is inextricably linked with benthic (sea floor) communities via the biological pump (Figure 6-4), the chemical exchange of nutrients and gases, and the existence of organisms with both pelagic and benthic life history stages. Even in abyssal habitats, a continuous rain of organic detritus serves as the primary source of carbon and energy. Therefore climate impacts on surface marine ecosystems will impact even the deepest benthic communities, even if direct changes to their physical habitat do not occur (Smith et al., 2009).

Benthic organisms living in shallow waters or the intertidal zone (where they encounter temporary exposure to air) are exposed to widely fluctuating and progressively changing means and extremes of environmental variables, such as temperature, oxygen, CO_2, salinity, and sea level (WGI AR5 Chapters 3, 13; Sections 6.3.1-3, 6.3.5). Plants and sessile or slow moving animals may be unable to escape from unfavorable changes except by means of advection of fertilized eggs or planktonic larvae. If climate change harms those species engineering benthic habitats, the entire ecosystem may be impacted. This concerns those ecosystem engineers, which form habitat from the structures they produce (e.g., corals forming skeletons; Section 6.3.1) and those forming habitat through their behavior (e.g., worms reworking and irrigating sediment in a process termed bioturbation). Effects on both types of ecosystem engineers (Sections 6.3.1-8) influence the regeneration of nutrients and affect benthic-pelagic coupling.

6.3. Climate Change Impacts from Organism to Ecosystem

Understanding climate-induced alterations in the functioning of individual organisms, species populations, communities (assemblages of various species), and ecosystems builds on studies in the laboratory, in micro- and mesocosms (closed small- to medium-sized experimental systems approximating natural conditions, holding selected biological communities), and of biota or communities in the field as well as modeling. These data inform us which taxonomic groups in what regions are more susceptible to climate change (Boyd et al., 2011). Empirical studies of marine organism and ecosystem sensitivities have begun identifying the mechanisms and processes linking climate to ecosystem changes (Drinkwater et al., 2010; Ottersen et al., 2010). Changes in ecological community composition, species interactions, and food web dynamics often build on organismal effects elicited by climate forcing (e.g., Section 6.3.1.5; Boyd et al., 2010; Ottersen et al., 2010). The underlying mechanisms respond to climate-related factors in a hierarchy from organism (highest), tissue, cell to molecular (lowest)

Table 6-1 | To assess how a changing climate will alter the ocean's biological pump (Figure 6-4) and determine the resulting biogeochemical feedbacks on global climate, changes in a wide range of processes from cells to ocean basins, and from epipelagic to mesopelagic, must be quantified. This table illustrates the complexity of the integrated knowledge platform needed to provide evidence of these biogeochemical ramifications and thus the present limits to clear conclusions about climate-induced effects on the biological pump (NPP = net primary production; C = carbon; TEP = transparent exopolymer particle; DOM = dissolved organic matter; POM = particulate organic matter).

Alteration of physiological rates	Biogeographical changes/ community shifts	Altered foodweb structure: trophodynamics	Changes to particle dynamics	Biogeochemical changes/ climatic feedbacks
• NPP (Bopp et al., 2002, 2013) • Particle solubilization through bacterial ectoenzymes (Christian and Karl, 1995) • TEP production (Engel et al., 2004) • Microzooplankton grazing rates (Rose et al., 2009)	• Microbial community structure (Giovannoni and Vergin, 2012) • Phytoplankton community structure, e.g., biomes (Boyd and Doney, 2002) • Alteration of zooplankton biomes (Beaugrand et al., 2009) • Faunistic shifts at depth (Jackson and Burd, 2001)	• Altered prey-predator linkages (Lewandowska and Sommer, 2010)	• Faecal pellet geometry (Wilson et al., 2008) • C partitioning between DOM vs. POM, e.g., TEP (Riebesell et al., 2007) • Sinking rates/seawater viscosity (Lam and Bishop, 2008) • Ballasting, e.g., calcite versus opal (Klaas and Archer, 2002)	• Particle flux/C sequestration (Bopp et al., 2002) • Shifts in elemental stoichiometry of planktonic communities (Karl et al., 2003) • Remineralization rate; $[O_2]$, hypoxia; nutrient resupply (Gruber, 2011) • Activity of the microbial loop; vertical carbon export (Grossart et al., 2006; Piontek et al., 2010)

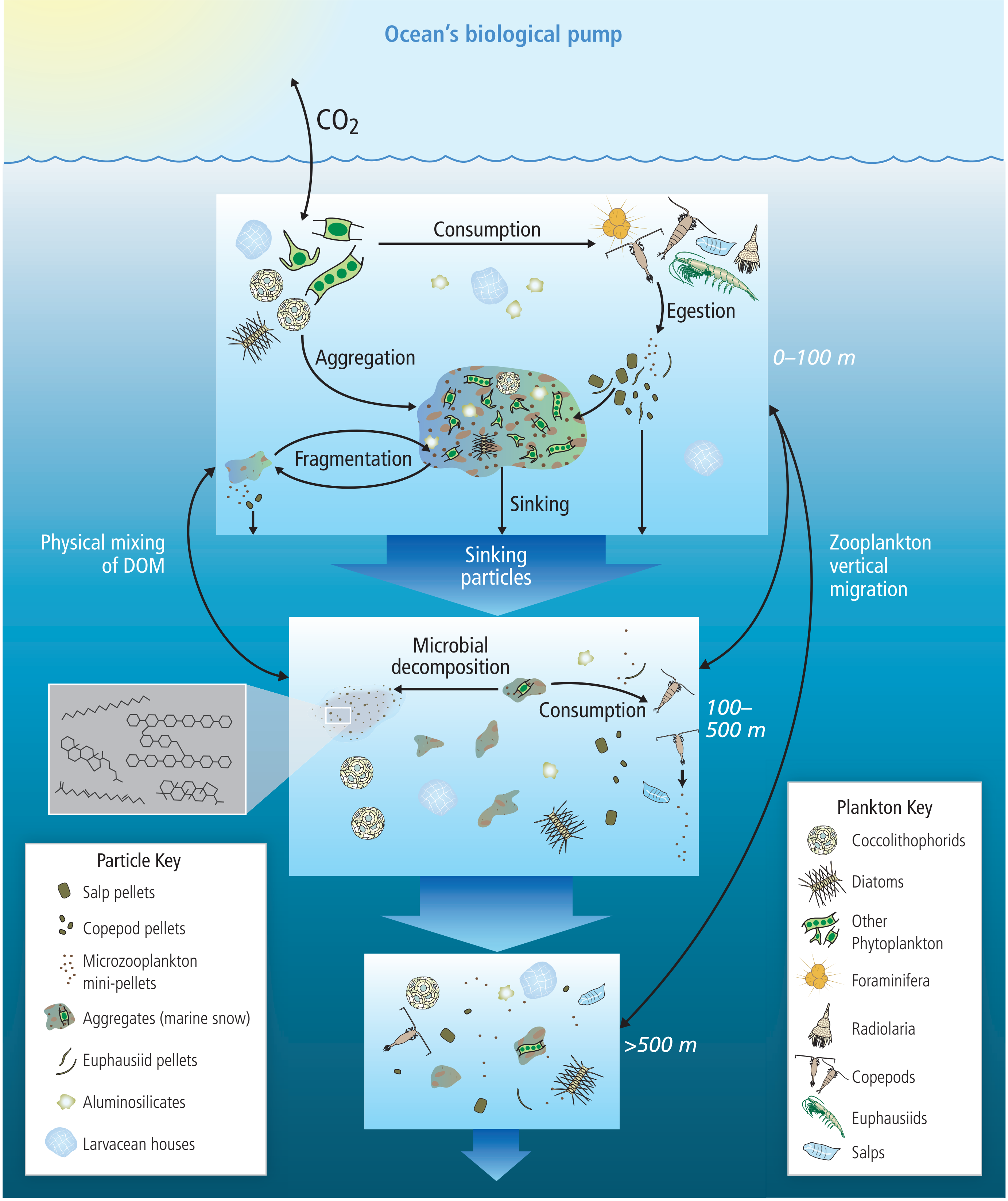


Figure 6-4 | A schematic representation of the ocean's biological pump, which will be influenced by climate change and is a conduit for carbon sequestration. It is difficult to project how the pump might be altered and whether it would represent a positive or negative feedback to climate change through the cumulative effects of affected processes, surface to depth (Table 6-1): shifts in net primary production, floristic and faunistic community composition in the pelagic realm, and in grazing rates; alterations to the ballasting of settling particles and the proportion of net primary production released as dissolved organic matter; modified bacterial enzymatic rates and particle solubilization; faunistic shifts at depth. Note that the relative sizes of the organisms, particles, and particle building blocks are not presented to scale (modified from Buesseler et al. (2008) by J. Cook / WHOI).

Frequently Asked Questions

FAQ 6.2 | What is different about the effects of climate change on the oceans compared to the land, and can we predict the consequences?

The ocean environment is unique in many ways. It offers large-scale aquatic habitats, diverse bottom topography, and a rich diversity of species and ecosystems in water in various climate zones that are found nowhere else.

One of the major differences in terms of the effect of climate change on the oceans compared to land is ocean acidification. Anthropogenic CO_2 enters the ocean and chemical reactions turn some of it to carbonic acid, which acidifies the water. This mirrors what is also happening inside organisms once they take up the additional CO_2. Marine species that are dependent on calcium carbonate ($CaCO_3$), such as shellfish, seastars, and corals, may find it difficult to build their shells and skeletons under ocean acidification. In general, animals living and breathing in water like fish, squid, and mussels have between five and 20 times less CO_2 in their blood than terrestrial animals, so CO_2-enriched water will affect them in different and potentially more dramatic ways than species that breathe in air.

Consider also the unique impacts of climate change on ocean dynamics. The ocean has layers of warmer and colder water, saltier or less saline water, and hence less or more dense water. Warming of the ocean and the addition of more freshwater at the surface through ice melt and higher precipitation increases the formation of more stable layers stratified by density, which leads to less mixing of the deeper, denser, and colder nutrient-rich layers with the less dense nutrient-limited layers near the surface. With less mixing, respiration by organisms in the mid-water layers of stratified oceans will produce oxygen-poor waters, so-called oxygen minimum zones (OMZs). Large, more active fish can't live in these oxygen poor waters, while more simple specialized organisms with a lower need for oxygen will remain, and even thrive in the absence of predation from larger species. Therefore, the community of species living in hypoxic areas will shift.

State-of-the-art ecosystem models build on empirical observations of past climate changes and enable development of estimates of how ocean life may react in the future. One such projection is a large shift in the distribution of commercially important fish species to higher latitudes and reduced harvesting potential in their original areas. But producing detailed projections, for example, what species and how far they will shift, is challenging because of the number and complexity of interactive feedbacks that are involved. At the moment, the uncertainties in modeling and complexities of the ocean system even prevent any quantification of how much of the present changes in the oceans are being caused by anthropogenic climate change or natural climate variability, and how much by other human activities such as fishing, pollution, etc.

It is known, however, that the resilience of marine ecosystems to adjust to climate change impacts is *likely* to be reduced by both the range of factors and their rate of change. The current rate of environmental change is much faster than most climate changes in the Earth's history, so predictions from longer term geological records may not be applicable if the changes occur within a few generations of a species. A species that had more time to adapt in the past may simply not have time to adapt under future climate change.

levels of biological organization (Pörtner, 2002a; Pörtner and Knust, 2007; Raven et al., 2012). Such knowledge aids the interpretation and attribution to climate change of observed effects and is a major asset for projections of future impacts.

The genetic and physiological underpinning of climate sensitivity of organisms sets the boundaries for ecosystem response and provides crucial information on sensitivities, resilience, and the direction and scope of future change. As anthropogenic climate change accelerates, a key issue is whether and how quickly organisms can compensate for effects of individual or multiple drivers, by short-term acclimatization or long-term evolutionary adaptation across generations. Evolutionary adaptation depends on the genetic variation within a population, from which the environment selects the fittest genotypes (Rando and Verstrepen, 2007; Reusch and Wood, 2007). Genetic variation depends on mutation rates, generation time, and population size (Bowler et al., 2010). However, epigenetic mechanisms, such as modifications of the genome by DNA methylation, can also influence fitness and adaptation (Richards, 2006) and can be remarkably rapid as seen in terrestrial ecosystems (Bossdorf et al., 2008). In plants and animals the rate of evolutionary adaptation is constrained by long generation times, but enhanced by high phenotypic variability and high mortality rates among early life stages as a selection pool (e.g., Sunday et al., 2011). The limits to acclimatization or adaptation capacity are presently unknown.

However, mass extinctions occurring during much slower rates of climate change in Earth history (Section 6.1.2) suggest that evolutionary rates in some organisms may not be fast enough to cope.

Comprehensive understanding of climate change effects on ecosystems requires addressing the effects of individual drivers across organism taxa (Sections 6.3.1-4), the integrated action of multiple drivers (Section 6.3.5), the consequences for food webs (Section 6.3.6), and the specific effects on animals breathing in air (Section 6.3.7) and operating at the highest trophic levels.

6.3.1. Temperature Effects

The effects of temperature on ecosystems largely result from organismal responses. This requires that information on organisms' thermal sensitivities, limits, and functional properties is used to assess how temperature changes have affected and will continue to affect species distributions, abundances, diversity, trophic interactions, community assemblages, risks of species extinctions, and ecosystem functioning. Organisms also respond to temperature-driven changes in the physical environment such as stratification, reduced sea ice cover, and freshening. Ambient temperature interacts with other drivers such as ocean acidification and hypoxia (Section 6.3.5). Ambient temperature plays a more limited role for marine mammals and seabirds (Section 6.3.7).

6.3.1.1. Principles

All organisms including marine ones have limited temperature ranges within which they live and function. Organismal performance is related to temperature by curves called thermal reaction norms (Figure 6-5), which *likely* apply across all organisms (Chevin et al., 2010), from viruses (Knies et al., 2006), bacteria (Ratkowsky et al., 1983), and phytoplankton (Eppley, 1972; Thomas et al., 2012) to macroalgae and plants (Bolton and Lüning, 1982; Müller et al., 2009; Vitasse et al., 2010) and animals (Huey and Kingsolver, 1989; Angilletta, 2009). Heat tolerance thresholds differ greatly between organisms and are hypothesized to be lowered by rising organizational complexity and body size (Pörtner, 2002a,b). Maximum heat limits of animals and plants are close to the maximum

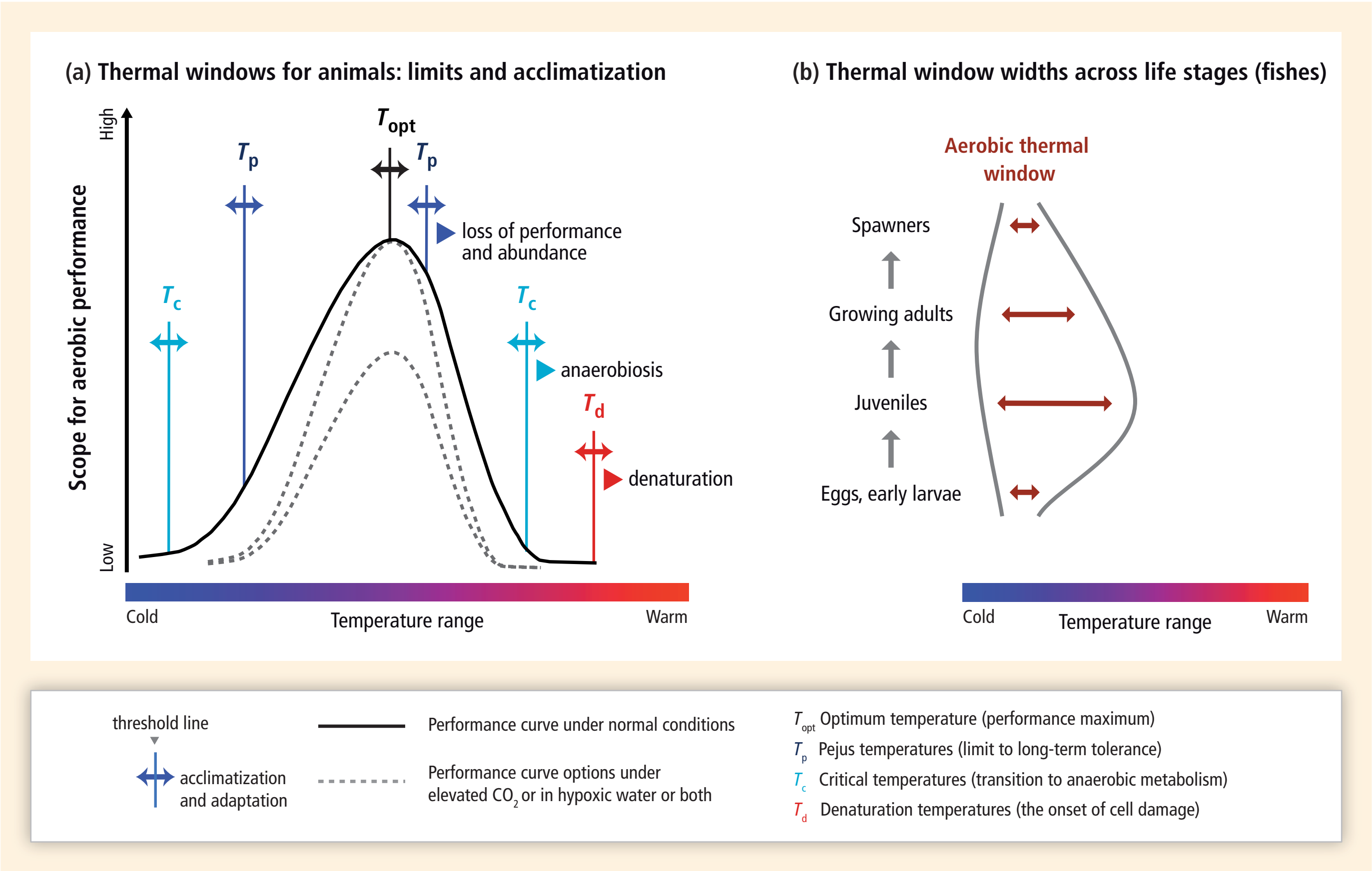


Figure 6-5 | Thermal specialization of an organism explains the why, how, when, and where of climate sensitivity. (a) The thermal tolerance range and performance levels of an organism are described by its performance curve (exemplified for an animal). Each performance (e.g., exercise, growth, reproduction) is maximal at its optimum temperature (T_{opt}), and becomes progressively constrained during cooling or warming. Surpassing the first low- and high-temperature thresholds (T_p; p, pejus: getting worse) means going into time-limited tolerance. Once further cooling or warming surpasses the next low or high thresholds (T_c; c, critical), oxygen availability becomes insufficient and an anaerobic metabolism begins. Denaturation temperatures (T_d) are even more extreme and characterized by the onset of damage to cells and proteins. Horizontal arrows indicate that T_p, T_c, and T_d thresholds of an individual can shift, within limits, between summer and winter (seasonal acclimatization) or when the species adapts to a cooler or warmer climate over generations (evolutionary adaptation). Under elevated CO_2 levels (ocean acidification) and in hypoxic waters performance levels can decrease and thermal windows narrow (dashed gray curves). (b) The width of the thermal range (horizontal arrows) also changes over time when an individual develops from egg to larva to adult and gains weight and size. Blue to red color gradients illustrate the range between cold and warm temperatures (after Pörtner, 2002a, 2012; Pörtner and Farrell, 2008).

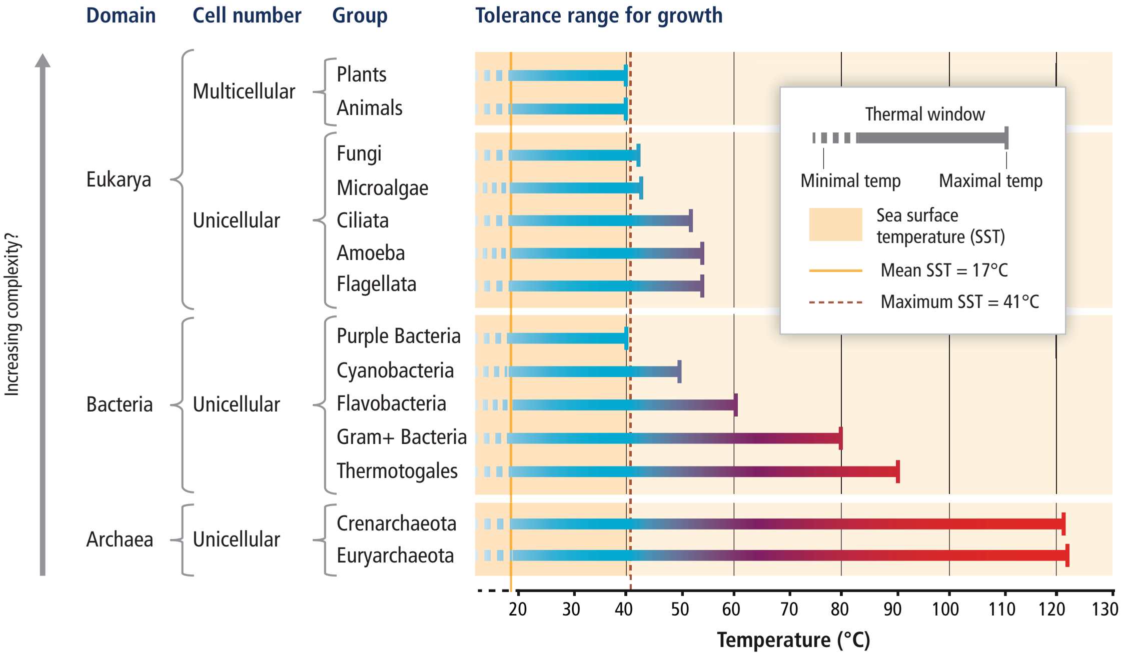


Figure 6-6 | Maximal values of temperature covered by various domains and groups of free-living marine organisms (bacteria to animals; domains and groups modified after Woese et al., 1990). High organizational complexity is hypothesized to be associated with decreasing tolerance to heat and to enable an increase in body size which in turn, decreases heat tolerance further (Sorokin and Kraus, 1962; Chevaldonné et al., 2000; Alker et al., 2001; Baumgartner et al., 2002; Pörtner, 2002a,b; Campbell et al., 2006; De Jonckheere et al., 2009, 2011). In the domain Bacteria, the Thermotogales are less complex and most tolerant to high temperatures (Huber et al., 1986; Tenreiro et al., 1997; Takai et al., 1999; Ventura et al., 2000; Abed et al., 2002). The highest temperature at which growth can occur is 122°C for hydrothermal vent archaea, seen under elevated hydrostatic pressure in laboratory experiments (Kashefi and Lovley, 2003; Takai et al., 2008).

temperature found in the warmest oceans (Figure 6-6). Knowledge of reaction norms, thermal limits, and underlying mechanisms is most advanced in animals (Pörtner et al., 2012; see also Section 6.3.1.4). Their role in underpinning biogeography has not been explored systematically in other organisms (e.g., Green et al., 2008), reducing the confidence level in assessments of thermal impacts. In animals, changes in physiological performances influence growth, body size, behavior, immune defense, feeding, reproductive success, biogeography, phenology, and therefore ecosystem structure and functioning. Shape and width of the curves can shift through acclimatization and evolutionary adaptation (Figure 6-5a) and during life history (Figure 6-5b), with implications for the distribution boundaries of species or populations (Section 6.3.1.5).

For any species, tracking the climate-induced displacement of tolerated ambient temperatures by undergoing shifts in biogeographical ranges to, e.g., higher latitudes during warming (Section 6.3.1.5; Figure 6-7) can be understood as a simple mode of adaptation, implemented through dispersal (e.g., of pelagic life stages), active movements (e.g., of migrating adult fishes), or passive displacement (e.g., of early life stages or plankton with drifting water masses). Conversely, fully completed acclimatization or evolutionary adaptation (Figure 6-5) would involve shifting thermal tolerance ranges and allow species to resist the temperature trend (e.g., warming) and to sustain fitness in their previous habitat.

6.3.1.2. Microbes

Temperature effects on growth, abundance, distribution, phenology, and community structure of highly diverse microbes have large implications for ecosystem functioning (Section 6.3; Box CC-PP). A warming ocean may initially enhance the metabolic rates of microbes (Banse, 1991) and stimulate their overall growth (Bissinger et al., 2008). Data from the Continuous Plankton Recorder (Section 6.1.2) in the Northeast Atlantic confirm that warming from 1960 to 1995 enhanced phytoplankton growth (Edwards et al., 2001). Eventually, with warming, the thermal tolerance of some groups will be challenged (Chevin et al., 2010), leading to the replacement of species. This is reflected in increasing fractions of smaller phytoplankton in warmer relative to colder waters (Morán et al., 2010; Flombaum et al., 2013).

In response to transient warming, phytoplankton distribution in the North Atlantic shifted poleward by hundreds of kilometers per decade since the 1950s. Phenology of plankton in the North Atlantic was also affected, with differences in sensitivity between groups (*high confidence*; Section 6.3.1.5; Box 6-1). Coccolithophore blooms (*Emiliania huxleyi*) in the Bering Sea were reported for the first time during the period 1997–2000, probably in response to a 4°C warming, combined with a shallower mixed layer depth, higher light levels and low zooplankton grazing (Merico et al., 2004). Loss of multi-year Arctic sea ice has had a profound effect on the diversity, structure, and function of the epipelagic microbial assemblage (i.e., found in the layer into which enough light penetrates for photosynthesis) (Comeau et al., 2011), and further warming is likely to have even greater impacts on the food web and on ecosystem services (*medium confidence*). Warming may also have caused the southward range extension of coccolithophores in the Southern Ocean in the 2000s (Cubillos et al., 2007). However, further experimental and field observations (Giovannoni and Vergin, 2012) are required to validate model projections (Taucher and Oschlies, 2011) of differential responses to warming by different microorganisms.

6.3.1.3. Macroalgae and Seagrasses

Macrophytes in coastal waters (Chapter 5) cover 0.6% of the world's marine areas and supply about 2 to 5% of total oceanic production (Smith, 1981; Charpy-Roubaud and Sournia, 1990; Field et al., 1998). They have limited temperature ranges and are sensitive to temperature extremes (*high confidence*), resulting in changes of photosynthesis, growth, reproduction, and survival (following the principles of Figures 6-5, 6-6; and Harley et al., 2012), with consequences for their abundance, distribution, and productivity. Ice retreat in polar areas leads to an expansion of macroalgal distribution, for example, in the Antarctic (Quartino et al., 2013).

Warm- versus cold-water-adapted species may have different sensitivities to warming and show a range of responses in distribution shifts (Lima et al., 2007). Temperate macroalgae with wide windows of thermal tolerance acclimatize by shifting these windows following seasonal temperature changes (Kübler and Davison, 1995). Antarctic and tropical macroalgae are exposed to permanently low or high temperatures, respectively, and have consequently specialized in a limited temperature range, paralleled by a low acclimatization potential (Pakker et al., 1995; Eggert et al., 2006; Gómez et al., 2011). Thus, Antarctic and tropical macroalgae appear to be most vulnerable to warming (*high confidence*; Short and Neckles, 1999). While observations in the tropics indicate that seagrasses tolerate higher temperatures than seaweeds (Campbell et al., 2006), an increase in maximum temperature by >1°C from 1988–1999 to 2002–2006 (Section 30.5.3.1.5) led to increased seagrass shoot mortality in the Mediterranean Sea (Marbà and Duarte, 2010). The molecular basis of acclimatization and evolutionary adaptation, as well as their limitation in relation to the climate regime, require further study in the macrophytes.

6.3.1.4. Animals

The mechanisms shaping the thermal performance curve and, thereby, an animal's thermal niche have been explained by the concept of "oxygen and capacity limited thermal tolerance" (OCLTT), applicable to marine invertebrates and fishes (Pörtner et al., 2010; see also Figure 6-5a, FAQ 6.2). The temperature range at which animals can function best results from optimal oxygen supply at minimal oxygen usage. At temperature extremes, oxygen supply capacity becomes constrained in relation to demand, and metabolism becomes thermally limited. Beyond upper and lower temperature thresholds (T_p, Figure 6-5a), growth, reproduction, and other key functions decrease. These thresholds change during the individual life cycle, and with body size. At large body size, limitations in oxygen supply are exacerbated and heat tolerance limits shift to lower temperatures.

Surpassing species-specific heat tolerance limits (Figure 6-5, T_p) during warming causes a reduction of abundance (Pörtner and Knust, 2007; Katsikatsou et al., 2012), coral losses (Donner et al., 2005), shifts in the seasonal timing of (zooplankton) biomass formation (Mackas et al., 1998; Schlüter et al., 2010), and changes in growth (Lloret and Rätz, 2000; Brunel and Dickey-Collas, 2010). During early life, owing to incomplete development, or as adult spawners, owing to large body size, animals may become more sensitive to warming because of narrower thermal windows (Pörtner et al., 2008). This may cause high vulnerability of winter-spawning Atlantic cod to warming winter to spring temperatures (Table 6-2). In contrast, adult bigeye, bluefin, and skipjack tuna spawn at high temperatures. They need to prevent overheating by moving to cooler (deeper) waters (Lehodey et al., 2011).

Although temperature means are still most commonly used when attributing responses of marine organisms to climate effects, temperature extremes rather than means are most often mediators of effects (e.g., Easterling et al., 2000; Wethey et al., 2011; Wernberg et al., 2013; Figure 6-5). During heat exposure near the borders of the distribution range (including the high intertidal or warming surface waters), reductions in growth, activity, and abundance accompany even small (<0.5°C) shifts in ambient temperature extremes (e.g., Takasuka and Aoki, 2006; Pörtner and Knust, 2007; Nilsson et al., 2009; Neuheimer et al., 2011). Local extinction events follow as a result of mortality or behavioral avoidance of unfavorable thermal environments (Breau et al., 2011). Shifted species distribution ranges follow temperature clines from high to low, usually along latitudes, a lateral gradient at basin scale (Perry et al., 2005; Poloczanska et al., 2013), or a vertical temperature gradient to deeper waters (*high confidence*; Dulvy et al., 2008; Section 6.5.3; see also Figure 6-5b, Box CC-MB).

Adopting OCLTT principles has enabled modeling studies to project climate effects (Section 6.5), and paleo-studies to explain climate-induced mass extinction events and evolutionary patterns in Earth history (Pörtner et al., 2005; Knoll et al., 2007). For example, long-term observations show that warming affects the body size of marine fishes (*medium confidence*). Assessing effects of warming on body size may be complicated by effects on the animal's energy budget, the changing availability and body size of prey species, community structure, species interactions, or effects of fishing (Genner et al., 2010; Cheung et al., 2013a). Below the thermal optimum, warming causes growth and weight-at-age of some juvenile or younger fish populations to increase (e.g., Brunel and Dickey-Collas, 2010; Neuheimer and Grønkjær, 2012). However, OCLTT predicts that small individuals are more heat tolerant than large ones, in line with observations of falling animal body sizes in warming oceans (Box 6-1; e.g., Daufresne et al., 2009). This trend is projected to continue into the 21st century (*medium* to *high confidence*; Cheung et al., 2013a).

Thermal windows of fishes and invertebrates roughly match ambient temperature variability (Figure 6-1) according to climate regime and seasonality (Pörtner and Peck, 2010; Sunday et al., 2012). Sub-Arctic, small, or highly mobile species are eurytherms. They function across a wide temperature range, that is, they have wide thermal windows and distribution ranges, at the expense of higher energetic costs and associated lifestyles (Pörtner, 2002a, 2006). Conversely, high polar species are stenotherms, that is, they have narrow thermal windows and low energy demand lifestyles, making them sensitive to temperature change. In a warming world, polar stenotherms will be marginalized, with no possibility to escape to colder regions (*high confidence*). However, extinction of polar species has not yet been reported. As marine fishes and invertebrates in the Southern Hemisphere are

Table 6-2 | Selected examples of species responses and underlying mechanisms to changing temperature, oxygen level and ocean acidification (OA). References are indicated by superscript numbers and in the footnote.

	Phenomenon	Key drivers	Mechanism/Sensitivity
Biogeography	Northward shift in the distribution of North Sea cod (*Gadus morhua*) stocks between 1977 and 2001. [1,2]	Temperature	Bottlenecks of high sensitivity during early life stages as well as adult spawning stage in winter/early spring.
	Shift from sardines (*Sardinops melanostictus*) to anchovies (*Engraulis japonicus*) in the western North Pacific observed between 1993 and 2003. [3,4]	Temperature	Thermal windows of growth and reproductive output are found at higher temperatures for anchovies than sardines, food preferences of the competing species being similar.
	Variable sensitivity of Pacific tuna species to the availability of dissolved O_2. Bigeye tuna routinely reach depths where ambient O_2 content is below 1.5 ml L^{-1} (≈ 60 µmoles kg^{-1}). [5, 6]	Oxygen	Oxygen transport via hemoglobin is adapted to be highly efficient supporting high metabolic rates as needed during feeding in the OMZ.
	Northward movement of species and the conversion of polar into more temperate and temperate into more subtropical system characteristics in the European Large Marine Ecosystems between 1958–2005. [7, 8]	Warming and current advection	Effects are attributed to climate change but may be influenced by nutrient enrichment and overfishing.
Abundance	Increase in abundance of arctic boreal plankton species, notably the copepods *Calanus hyperboreus*, *Calanus glacialis* and the dinoflagellate *Ceratium arcticum* between 1960 and 2000 in the Newfoundland Shelf, Northwest Atlantic. [9,10]	Temperature	Temperature sensitivity of phyto- and zooplankter resulting from cooling due to increased influx of Arctic water.
	A benthic fish species, the eelpout (*Zoarces viviparus*) at its southern distribution limit, the German Wadden Sea, displayed abundance losses during warming periods and rising summer extreme temperatures between 1993 and 2005, with early disappearance of the largest individuals. [11]	Temperature	Temperature extremes exceed organism's thermal windows, with largest individuals being relatively less tolerant to high temperature than smaller individuals.
	Variable sensitivities to OA within and across animal phyla (Figure 6-10b). [12–21]	Anthropogenic OA, sea water acidification by elevated pCO_2 in OMZs, upwelling areas, involving anthropogenic ocean acidification.	Lowered extracellular (blood plasma) pH causing a lowering of the rates of ion exchange and metabolism in muscle or liver (hepatocytes) of vertebrates and invertebrates. High sensitivity at reduced energy turnover in tissues and/or whole organism by reduced ion exchange, use of more energy efficient transport mechanisms, reduced protein synthesis, enhanced nitrogen release from amino acid catabolism and protein degradation, slower growth.
Phenology	Migration time of pink salmon (*Oncorhynchus gorbuscha*) in Alaska is almost two weeks earlier in 2010s relative to 40 years ago. [22]	Warming	Rapid microevolution for earlier migration timing.
	In the waters around the UK, during a period of warming between 1976 and 2005, the seasonal timing of biological events of all major marine taxonomic groups (plant/phytoplankton, invertebrate and vertebrates) advanced, on average, by 0.31 to 0.43 days $year^{-1}$. [23]	Warming	Sensitivity to seasonal temperature changes as a result of specific thermal windows of different organisms.
Body size and growth	Asymptotic body sizes of different populations of Atlantic cod (*Gadus morhua*) and Atlantic Herring (*Clupea harengus*) are negatively related to temperature. [24, 25]	Warming	At large body size, oxygen supply limitations are exacerbated and the organism reaches its long-term heat tolerance limits at lower temperatures, thus limiting the maximum body size that can be reached.

1. Perry et al. (2005); 2. Pörtner et al. (2008); 3. Takasuka et al. (2007); 4. Takasuka et al. (2008); 5. Lehodey et al. (2011); 6. Seibel (2011); 7. Beaugrand et al. (2009); 8. Philippart et al. (2011); 9. Johns et al. (2001);10. Greene and Pershing (2003); 11. Pörtner and Knust (2007); 12. Reipschläger and Pörtner (1996); 13. Pörtner et al. (2000); 14. Vezzoli et al. (2004); 15. Langenbuch and Pörtner (2003); 16. Fernández-Reiriz et al. (2011); 17. Langenbuch and Pörtner (2002);18. Langenbuch et al. (2006); 19. Michaelidis et al. (2005); 20. Pörtner et al. (1998); 21. Stumpp et al. (2012); 22. Kovach et al. (2012); 23. Thackeray et al. (2010); 24. Taylor (1958); 25. Brunel and Dickey-Collas (2010).

adapted to less variable ocean temperatures than those in the Northern Hemisphere (Jones et al., 1999; Figure 6-1), they may generally be more vulnerable to warming extremes than Northern ones. Tropical species (with thermal windows of intermediate width) live close to the highest temperatures tolerated by marine animals (Figure 6-6). Vulnerability is, therefore, highest for polar stenotherms, similar or lower for tropical, and lowest for temperate species (*high confidence*).

Short-term shifts in thermal thresholds of an individual organism may happen over days and weeks, such as during seasonal acclimatization. Long-term shifts occur over many generations during evolutionary adaptation of a population to cooler or warmer climates (Figure 6-5a; Pörtner, 2006; Pörtner et al., 2008; Eliason et al., 2011). Both acclimatization and adaptation involve adjustments in biochemical characters (membranes, enzymes); however, the capacity to shift those boundaries is limited and depends on the species and the prevailing climate regime (Pörtner et al., 2008, 2012). Ocean acidification, hypoxia, food availability, and stress affect those limits (Section 6.3.5; Figure 6-5a).

Local adaptation may reduce climate vulnerability at the species level, by causing functional and genetic differentiation between populations, thereby enabling the species to cover wider temperature ranges and live in heterogeneous environments. Local adaptation on small spatial scales is particularly strong in intertidal organisms (Kelly et al., 2012). On larger scales, the widening biogeographic and roaming ranges of

Northern Hemisphere eurytherms into Arctic waters (Pörtner et al., 2008) are supported by the differentiation into populations with diverse thermal ranges, combined with high acclimatization capacity. By contrast, such capacity is small in high polar, for example, Antarctic species (Peck et al., 2010). Tropical reef fishes undergo rapid warm acclimation across generations (Donelson et al., 2012) but some may approach animal heat limits. The rates, mechanisms, and limits of thermal acclimatization and evolutionary adaptation are poorly understood (*low confidence*).

6.3.1.4.1. Warm- and cold-water coral communities

Tropical corals live in shallow water and differ from most other animals by hosting dinoflagellates (*Symbiodinium* sp.) in their tissues, which provide the host with organic carbon from photosynthesis and with nitrogen and enable the corals to build and sustain carbonate reefs (Box CC-CR). High light, rapid salinity changes, and small increases in temperature can trigger "coral bleaching", the loss of symbionts and tissue color. In case of warming, early steps involve shifts in the photosynthetic processing of light, generating Reactive O_2 Species (ROS) that may in turn damage the symbionts (Hoegh-Guldberg and Smith, 1989; Glynn and D'Croz, 1990; Jones et al., 1998; Hoegh-Guldberg, 1999). Mass bleaching correlates with small temperature anomalies (+1°C to 2°C of the long-term summer maximum, satellite observations), causing mortalities (Goreau and Hayes, 1994; Strong et al., 2011) and decreasing coral abundance, on average by 1 to 2% per year (*high confidence*; Bruno and Selig, 2007; see also Box CC-CR; Section 30.5.6).

The degree of impact will depend on the coral reefs' adaptability to thermal stress and the interaction of multiple drivers (Meissner et al., 2012; Teneva et al., 2012; see also Box CC-CR). Such capacity is suggested by different heat tolerances among coral genera (Hoegh-Guldberg and Salvat, 1995; Loya et al., 2001), the exchange of genetic clades of *Symbiodinium* with more tolerant varieties (Baker, 2001; Jones et al., 2008), as well as acclimatization phenomena (Howells et al., 2012).

Studies of the thermal sensitivity of deeper-living cold-water corals (without endosymbionts) are scarce. One species, *Lophelia pertusa*, responds to about 3°C warming with a threefold increase in metabolic rate (Dodds et al., 2007), indicating a narrow thermal window in the cold (cf. Pörtner, 2006).

6.3.1.5. Ecosystems

Heat exposure of ecosystem engineers may threaten the existence of a whole ecosystem. During the last warm interglacial period equatorial coral reefs deteriorated and retreated (Kiessling et al., 2012), a finding emphasizing their thermal sensitivity (Veron et al., 2009) and showing that warming oceans can reach temperatures well beyond the upper heat limits of distinct animal groups and marine animals overall (Figure 6-6). In the present-day Great Barrier Reef, a large-scale survey found diverse coral types along a climatic gradient, with no consistent response to climatic drivers (Hughes et al., 2012). However, warm-induced bleaching has contributed to the progressive decrease in live coral cover observed over the last decades (De'ath et al., 2012; see also Box CC-CR; Section 30.5.6).

Within ecosystems, shifting competitive or trophic interactions, differential risks for species extinctions and, thereby, scenarios of community-level responses to temperature change (Urban et al., 2012; Milazzo et al., 2013) can be traced back to changing differences in the performance of participating animal species (Figure 6-7; e.g., Cairns et al., 2008; Harley, 2011; Pörtner, 2012). Knowledge is insufficient to assess interactions of species from different domains, impeding a deeper understanding of shifting distributions, abundances, community assemblages, and food webs in space and time (*low confidence* in current understanding; Parmesan and Matthews, 2005).

For example, in a coastal microcosm (small-scale, simplified experimental ecosystem) resident heterotrophic bacteria were stimulated by warming more than a laboratory-reared phytoplankter (Wohlers-Zöllner et al., 2011). Also, high- to low-latitude transects in both the North and South Atlantic revealed a shift between cold and warm waters, from photo-autotrophs (gaining energy from photosynthesis) to chemo-heterotrophs (Hoppe et al., 2002). Thermal stimulation of bacteria over phytoplankton has biogeochemical implications, for example, microbially mediated CO_2 flow to the atmosphere might increase (Sarmento et al., 2010). The principles and wider applicability of these findings require further investigation (*limited evidence*, *low agreement*; Kirchman et al., 2009).

Observations of shifting distributions and phenologies, reproduction, and range shifts of phytoplankton, zooplankton, other invertebrates, fishes, and seabirds in pelagic and coastal marine ecosystems have at least partly been attributed to temperature-mediated biological responses (*high confidence*; see also Figure 6-8; Box 6-1; Box CC-MB). In the North Atlantic as a key example, many biological events have been occurring earlier in the year (*robust evidence*, *high agreement*; Box 6-1; Section 30.5.1.1.1). Species richness has increased as a result of shifts in ranges and abundances. In the Norwegian and Barents Seas, a time series (1959–2006) of four commercial fish species and their zooplankton prey showed that climate shapes population growth rates through complex influences early in life, including direct temperature effects on growth, indirect effects via the biomass of zooplankton prey, and delayed feedback effects through predators (Stige et al., 2010). Differential species responses to temperature and trophic amplification were demonstrated to modify species interactions at five trophic levels: primary producers (phytoplankton); primary, secondary, and tertiary consumers (zooplankton, fishes, and jellyfishes); and benthic detritivores (echinoderms and bivalves) (Kirby and Beaugrand, 2009). Also, the responses of various plankton functional groups, such as diatoms, dinoflagellates, and copepods, to warming are not synchronous, resulting in predator-prey mismatches that carry over to higher trophic levels (*high confidence*; Edwards and Richardson, 2004; Costello et al., 2006; see also Figure 6-7a; Section 6.3.6). In the intertidal, warming-induced changes in relative species ranges lead to shifts in dominance through competitive interactions and to modifications in predator pressure (Poloczanska et al., 2008; Harley, 2011). Trans-Arctic interchange of species between Atlantic and Pacific has happened repeatedly in warm periods of the Pleistocene (Dodson et al., 2007) and may occur again, now facilitated by ballast transport by enhanced trans-Arctic shipping (*low* to *medium confidence*).

Warming may increase the risk of disease outbreaks or parasite infections, in marine organisms and ecosystems, and ultimately, humans (*medium*

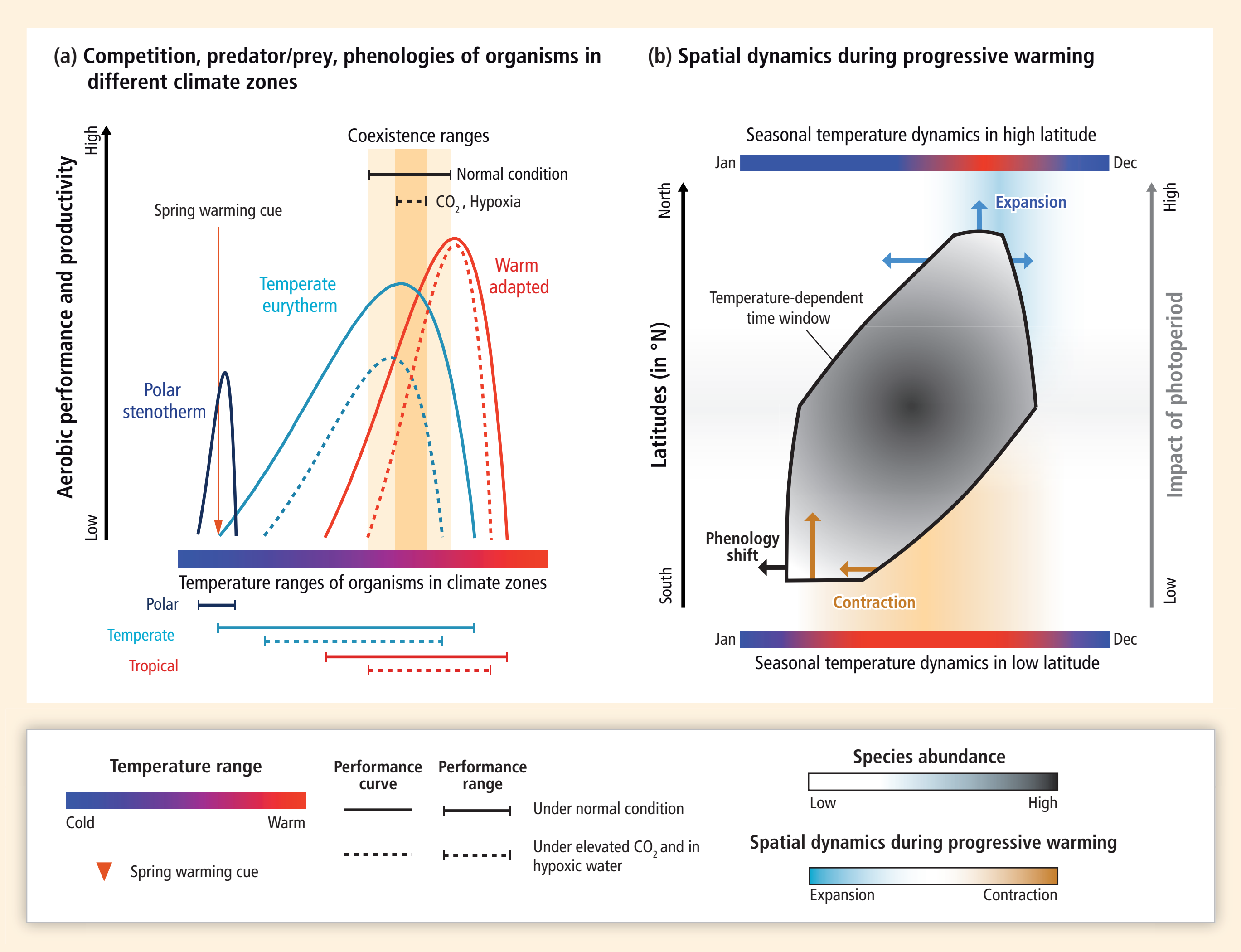


Figure 6-7 | Role of thermal tolerance and performance of organisms at ecosystem level. (a) Thermal tolerance ranges (Figure 6-5) differ between species across polar, temperate, and tropical climate zones, then overlap between coexisting species. Shifting temperatures and specific effects of additional drivers on the respective performance curves (dashed lines) change the fitness of coexisting species relative to each other as well as their temperature range of coexistence (after Pörtner and Farrell, 2008). Warming alters the timing of seasonal activities (e.g., elicited by spring warming cues) to earlier, or can benefit only one of two interacting species (e.g., in predator–prey dynamics or competition), causing shifts in predominance. (b) During climate warming a largely unchanged thermal range of a species causes it to follow its normal temperatures as it moves or is displaced, typically resulting in a poleward shift of the biogeographic range (exemplified for the Northern Hemisphere; modified after Beaugrand, 2009). The polygon delineates the distribution range in space and seasonal time; the level of gray denotes abundance. The Southern time window of tolerated temperatures shifts to earlier and contracts, while the Northern one dilates (indicated by arrows). Species display maximum productivity in low latitude spring, wide seasonal coverage in the center, and a later productivity maximum in the North. The impact of photoperiod (length of daily exposure to light) increases with latitude (gray arrow). Water column characteristics or photoperiod may overrule temperature control in some organisms (e.g., diatoms), limiting northward displacement.

confidence; Altizer et al., 2013; Burge et al., 2014). Some marine pathogens and protist diseases are shifting their distribution poleward as oceans warm (e.g., Baker-Austin et al., 2013; Burge et al., 2014). Climate change may weaken the immune response of hosts, particularly fishes and invertebrates, and increase their susceptibility to disease, as observed during warming in coral reefs of the Pacific and Caribbean (Harvell et al., 2009). Global outbreak frequencies of jellyfish aggregations may follow rising sea surface temperatures (SSTs) (*low confidence*; Mills, 2001; Purcell and Decker, 2005), but evidence is inconclusive. Some studies report an increasing trend (Brotz et al., 2012) and others do not support this view (Condon et al., 2013).

In conclusion, organisms live in limited temperature ranges and are sensitive to temperature extremes (*very high confidence*). Temperature governs the biogeography, diversity, development, reproduction, behavior, and phenology of marine species as well as the composition of communities in both pelagic and benthic systems and the seasonal timing of relevant processes (phenology) (*very high confidence*). Ecosystems functioning at the coldest temperatures and warm adapted ones existing at their upper thermal limits are more sensitive (*medium confidence*).

6.3.2. Carbon Dioxide Effects

Evidence for biological effects of ocean acidification stems from paleo-observations (Section 6.1.2), few observations in the field (Section 6.3.2.5), studies at volcanic CO_2 seeps as natural analogs, and mostly from short- to medium-term (hours to months) experiments in the

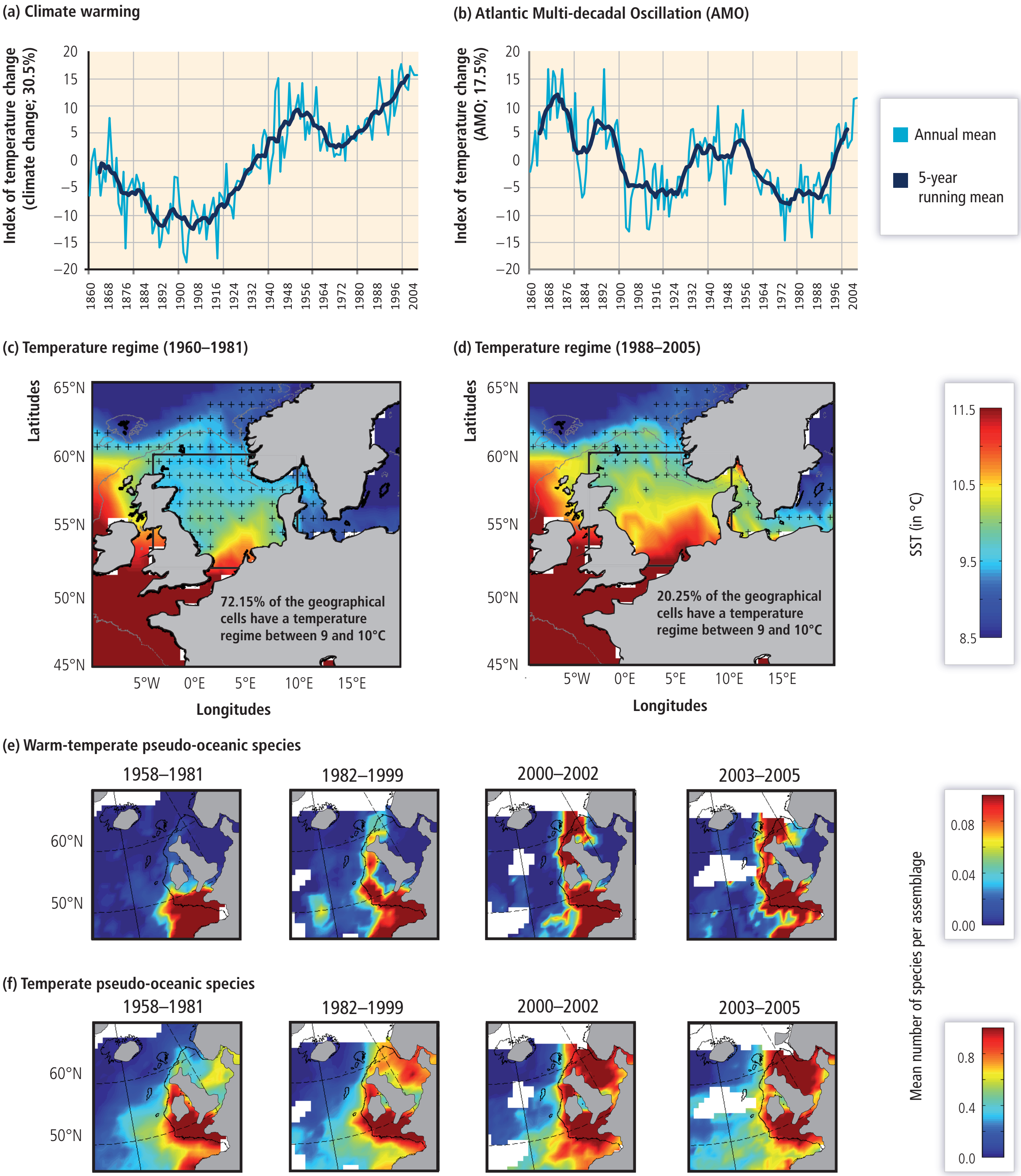


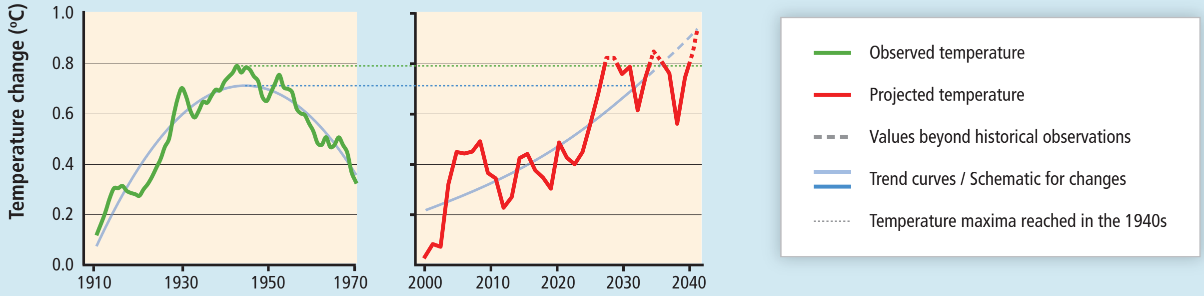
Figure 6-8 | Multi-decadal changes in ecosystem structure in the Northeast Atlantic driven by warming from both anthropogenic climate change and natural climate variability. (a) Index of temperature change over the North Atlantic (31°N to 65°N and 99°W to 11°E) reflecting climate change. This index is the first principal component (i.e., explaining 30.5% of observed variability) based on a principal component analysis (PCA) performed on sea surface temperature. (b) Index of temperature change (17.5% of observed variability) reflecting the Atlantic Multi-decadal Oscillation (AMO). The index is the second principal component. (c, d) Observed mean annual sea surface temperature in the North Sea during 1960–1981 (c) and 1988–2005 (d). The location of the critical thermal boundary (9°C to 10°C) is indicated by "+." (e) Long-term changes in the mean number of warm-temperate pseudo-oceanic species from 1958 to 2005. (f) Long-term changes in the mean number of temperate pseudo-oceanic species from 1958 to 2005. The period 1958–1981 was a period of relative stability and the period 1982–1999 was a period of rapid northward shifts, indicating that the abrupt ecosystem shift observed in the North Sea was part of a large-scale response of the zooplankton biodiversity to warming temperatures (see a–d). Average values are below 1 because they are annual averages. Note that the color bar is 10-fold smaller for warm-temperate pseudo-oceanic species because these species are less frequently observed than their temperate counterparts. Panels (a) and (b) from Edwards et al. (2013), and (c)–(f) from Beaugrand et al. (2008, 2009).

Box 6-1 | An Atlantic Ocean Example:
Long-Term Responses of Pelagic Organisms and Communities to Temperature

Long-term observations (Sections 6.1.2, 30.5.1.1.1) encompassing the pelagic Northeast Atlantic over a 50-year period and longer (Figures 6-8, 6-9) show changes in the seasonal abundance of phytoplankton, rapid northerly displacements of temperate and subtropical zooplankton (e.g., calanoid copepods) and phytoplankton (e.g., dinoflagellates and diatoms), and the resulting changes in the ecosystem functioning and productivity (*high confidence*; Edwards et al., 2001; Beaugrand et al., 2002; Edwards and Richardson, 2004). The range limit of warm water copepods shifted by 10° north since 1960 (Beaugrand et al., 2009), with attendant mismatch in the seasonal timing of trophic levels (predators and prey) and functional groups (Edwards and Richardson, 2004). Modes of climate variability reflected in climate indices like the Northern Hemisphere Temperature (NHT) and the North Atlantic Oscillation (NAO) over multi-decadal periods accompanied these changes (Figure 6-1). In cooler regions, increased phytoplankton activity caused by warming favored growth, resulting in the observed increase in phytoplankton biomass, whereas a decrease in nutrient supply would have prevented growth in warmer regions and caused a decrease in biomass (Richardson and Schoeman, 2004; see also Section 6.3.4). Hinder et al. (2012) attributed a recent decline in North Sea dinoflagellates relative to diatoms to warming, increased summer windiness, and thus water column turbulence. The ecosystem response to natural climate variability in the past provides a glimpse into the climate-induced changes of the near future (Figure 6-9).

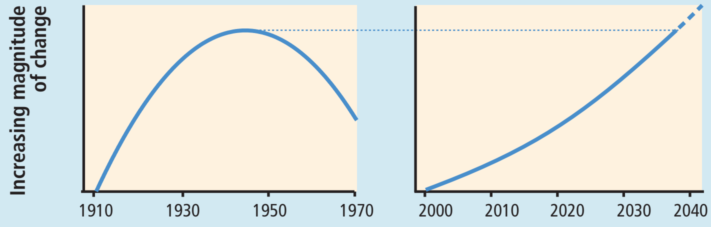
(a) Observed temperature variability in North Atlantic

(b) Projected temperature change (RCP8.5) in North Atlantic



(c) Sketch of observed changes in parameters (see table on the right)

(d) Sketch of projected changes in parameters (see table on the right)



Parameter sketched by curves (c) and (d)	How parameter changes
Biogeography	Poleward expansion and shift
Phenology	Timing variability
Community composition	Species turnover
NPP (mg C m^{-2} day^{-1})	Net primary productivity
Zooplankton (t km^{-2})	Biomass and latitudinal shift
Fish stocks (t km^{-2})	Biomass and latitudinal shift

Figure 6-9 | Schematic depiction of observed effects of approximately 1°C ocean warming in the northern North Atlantic driven by climate variability (a,c) versus effects expected from anthropogenic climate change (b,d). (a) Transient warming and cooling associated with Atlantic Multi-decadal Oscillation (AMO) variability (Drinkwater, 2006), based on the Kola Section temperatures (0 to 200 m; Stations 3 to 7, 71.5° to 72.5°N, 33.5°E) in the Barents Sea obtained from http://www.pinro.ru and filtered using a 20-year running mean. Similar trends occurred across most of the northern North Atlantic although the amplitude and timing of the peaks and troughs varied spatially. (b) Warming driven by climate change for the same region (Representative Concentration Pathway 8.5 (RCP8.5) simulations averaged from Coupled Model Intercomparison Project Phase 5 (CMIP5) models, computed as the mean over the upper 200 m in the grid box (2.5° × 2.5°) centered at 71.25°N and 33.75°E). (c) Warming and subsequent cooling in the northern North Atlantic during the period shown in (a) resulted in complex multi-faceted changes (shown schematically) in net primary production (NPP), zooplankton biomass, and fish stock abundances. There was a general poleward shift and range expansion of many commercial (e.g., Atlantic herring, Atlantic cod, haddock) and non-commercial species, reversed during the subsequent cooling period. Poleward shifts in spawning areas (e.g., Atlantic cod) were also reversed as the waters cooled. Shifts in seasonal timing (phenology) and community composition were influenced by earlier arrivals and later retreat of migratory fish (not shown). For more details see Drinkwater (2006). (d) Projected effects of climate mediated warming on northern sub-polar and polar biota based on model projections of altered NPP (Bopp et al., 2013), and of the range shift of exploited fishes and invertebrates (Cheung et al., 2009, 2013a). The projected trends in (d) will differ with latitude, for example, decreased NPP at lower latitudes and no significant change to NPP in temperate waters (Bopp et al., 2013). Higher NPP supported and is projected to support higher trophic levels at high latitudes (c,d; Section 6.3.4). Note that climate variability will be superimposed on anthropogenic warming (b; see Figures 6-1, 6-8a,b). Dashed lines indicate projected changes to continue beyond the range of historical observations.

Continued next page →

Box 6-1 (continued)

In regions of high vulnerability to climate, mild warming can trigger rapid and substantial ecosystem shifts, offering a way to anticipate future changes (Figure 6-9). In line with the increased understanding of physiology (Section 6.3.1.1), warming in the temperate to polar North Atlantic was paralleled by a reduction in the average body lengths of about 100 copepod species, from 3 to 4 mm to 2 to 3 mm (Beaugrand et al., 2010). Warming also correlated with an increase in species richness among copepods and within the dinoflagellate genus *Ceratium*. In diatoms, which are major contributors to carbon export (Armbrust, 2009), warming and decreasing annual variability in SST resulted in lower diversity, smaller size, and reduced abundance (Beaugrand et al., 2010). Morán et al. (2010) found that temperature alone explained 73% of the variance in the contribution of small cells (picophytoplankton) to total phytoplankton biomass in the eastern and western temperate North Atlantic from –0.6 to 22°C. More recently, Marañón et al. (2012) analyzed data from polar, sub-polar, and tropical regions and suggested that nutrient availability may influence cell size more than temperature.

The ecosystem regime shift observed in North Sea plankton in the late 1980s involved an increase in phytoplankton stocks and changes in species composition and abundance among holozooplankton (animals that are planktonic for their entire lifecycle) (Reid et al., 2001; Kirby and Beaugrand, 2009; Kirby et al., 2009; Lindley et al., 2010). This shift was paralleled by the northward propagation of a critical thermal boundary (CTB, i.e., the boundary of the sub-polar gyre) between the temperate and the polar biomes (Beaugrand et al., 2008; see also Box CC-PP, Figure 1). Warming to above the CTB coincided with pronounced and large-scale variations in phytoplankton productivity, an increase in calanoid copepod diversity (Beaugrand et al., 2008) and herring abundance (Schlüter et al., 2008), a reduction in the mean size of calanoids, and a decrease in the abundance of southern Atlantic cod populations in the North Atlantic Ocean (e.g., the North Sea; Pörtner et al., 2008; Beaugrand et al., 2010). These patterns also extend to the southern North Sea, where elevated salinities and average warming by 1.6°C, both in summer and winter between 1962 and 2007, expanded the time window for growth of microalgae and possibly supported the invasion and increase in numbers of warm-adapted silicified diatoms (Wiltshire et al., 2010). Recent findings indicate a regime shift in the Bay of Biscay and the Celtic and the North Seas in the mid to end 1990s (Luczak et al., 2011). Changing plankton composition and changing abundances of both sardine and anchovies (Raab et al., 2013) paralleled stepwise warming.

Northward range extensions or redistributions in fishes were largest along the European Continental shelf and attributed to regional warming, for example, by 1.0°C from 1977 to 2001 in the North Sea, with winter warming being closely correlated with the shift of Atlantic cod (Perry et al., 2005; see also Section 6.3.1). Similar trends were observed due to warming by 1°C to 2°C in the waters south and west of Iceland during the past 15 years (Valdimarsson et al., 2012). In the Northwest Atlantic Arctic and sub-Arctic, winter and spring warming caused expansion of the area matching the thermal optimum of Atlantic salmon at 4°C to 8°C and caused greater growth (Friedland and Todd, 2012). Pelagic sardines and anchovies entered the North Sea in the early to mid-1990s, after about 40 years of absence, in response to intensified NAO and AMO (Alheit et al., 2012). Red mullet and bass extended into western Norway; Mediterranean and northwest African species extended to the south coast of Portugal (Brander et al., 2003; Beare et al., 2004; Genner et al., 2004; see also Section 30.5.1.1.4).

In the Northwest Atlantic cooling and freshening occurred during the late 1980s to early 1990s and seemed to have the opposite effect, as capelin and their predator, Atlantic cod, shifted farther south (Rose and O'Driscoll, 2002). Between the early 1990s and mid-2000s in the Northwest Atlantic sub-polar gyre, phytoplankton biomass increased, due to warming. At the same time, Arctic copepod species became more abundant, due to increased influx of Arctic water (Head and Pepin, 2010). Although temperatures have risen on the Newfoundland Shelf (Colbourne et al., 2011), capelin and cod remain scarce for reasons probably unrelated to climate (DFO, 2011a,b). Farther south, Arctic freshwater inflows caused freshening and increased stratification of the area around the Gulf of Maine throughout the 1990s, resulting in enhanced phytoplankton abundance, a larger and later fall bloom, increased abundance of small copepods, and a decrease in the large copepod *Calanus finmarchicus* (deYoung et al., 2004; Pershing et al., 2005, 2010). Various fish species showed poleward shifts in distribution (Table 6-2) that were associated with reduced survival of larval cod (Mountain and Kane, 2010) and fewer right whale calves (Greene et al., 2003), but increased herring abundance (Greene and Pershing, 2007).

Frequently Asked Questions

FAQ 6.3 | Why are some marine organisms affected by ocean acidification?

Many marine species, from microscopic plankton to shellfish and coral reef builders, are referred to as calcifiers, species that use solid calcium carbonate ($CaCO_3$) to construct their skeletons or shells. Seawater contains ample calcium but, to use it and turn it into $CaCO_3$, species have to bring it to specific sites in their bodies and raise the alkalinity (lower the acidity) at these sites to values higher than in other parts of the body or in ambient seawater. That takes energy. If high CO_2 levels from outside penetrate the organism and alter internal acidity levels, keeping the alkalinity high takes even more energy. The more energy is needed for calcification, the less is available for other biological processes such as growth or reproduction, reducing the organisms' weight and overall competitiveness and viability.

Exposure of external shells to more acidic water can affect their stability by weakening or actually dissolving carbonate structures. Some of these shells are shielded from direct contact with seawater by a special coating that the animal makes (as is the case in mussels). The increased energy needed for making the shells to begin with impairs the ability of organisms to protect and repair their dissolving shells. Presently, more acidic waters brought up from the deeper ocean to the surface by wind and currents off the Northwest coast of the USA are having this effect on oysters grown in aquaculture.

Ocean acidification affects not only species producing calcified exoskeletons. It affects many more organisms either directly or indirectly and has the potential to disturb food webs and fisheries. Most organisms that have been investigated display greater sensitivity at extreme temperatures so, as ocean temperatures change, those species that are forced to exist at the edges of their thermal ranges will experience stronger effects of acidification.

laboratory or field, exposing organisms to projected future CO_2 levels (Sections 6.3.2.1-4). A surging number of studies is providing evidence that rising CO_2 levels will increasingly affect marine biota and interfere with ecological and biogeochemical processes in the oceans (*high confidence*; FAQs 6.2, 6.3).

6.3.2.1. Principles

The absorption of rising atmospheric CO_2 by oceans and organisms changes carbonate system variables in the water and in organism internal fluids, that is, the relative proportions of CO_2, carbonate, bicarbonate, and hydrogen ions (pH). Internal pH must be tightly controlled, as some processes, such as calcification, release protons thereby affecting pH and as other biochemical processes are pH sensitive. Accumulation of CO_2 and the resulting acidification can also affect a wide range of organismal functions, such as membrane transport, calcification, photosynthesis in plants, neuronal processes in animals, growth, reproductive success, and survival. Effects translate from organism to ecosystem.

The capacity of organisms to resist and compensate for the CO_2-induced acidification of internal fluids depends on acid-base regulation, that is, the capacity of ion exchange to accumulate bicarbonate internally, an aspect unexplored in many phyla (*low* to *medium confidence*; Figure 6-10a; e.g., animals: Heisler, 1986; Claiborne et al., 2002; Pörtner, 2008; phytoplankton: Taylor et al., 2011; see also FAQ 6.3).

In unicellular microbes the regulation of intracellular pH may play a key role in modulating CO_2 responses (Taylor et al., 2011). Findings in invertebrates and fish indicate an additional role for extracellular pH (Figure 6-10a); effective pH values may vary between species. Organisms pre-adapted to elevated CO_2 may minimize the decrease in pH (acidosis). They may also modify their sensitivity such that they respond less or not at all to the acidosis. Recent evidence, however, emphasizes a role for acid-base regulation in a natural low-pH setting. Between two urchin species, only the one successful in maintaining its setpoints of extracellular pH is able to settle close to volcanic CO_2 seeps (Calosi et al., 2013). Compensating for the acidosis may cause increased energy demand and respiration rates. In general, such capacity rises with metabolic energy turnover, for example, it is higher in more active marine animals, such as fishes, cephalopods, and pelagic copepods, and in mobile coastal crabs compared to sessile species (Pörtner et al., 2005, 2011; Ishimatsu et al., 2008; Melzner et al., 2009; Ishimatsu and Dissanayake, 2010; see also Table 6-3). This matches the sensitivity distribution seen among animals at the phylum level (*medium confidence*; Figure 6-9b).

Some species have lower metabolic rates in response to acidosis (Pörtner et al., 1998; Michaelidis et al., 2005; Pörtner, 2008; Liu and He, 2012; Navarro et al., 2013); others display increased energy turnover and food ingestion rates, possibly indicating a capacity to resist acidification effects (Parker et al., 2011; Saba et al., 2012). The effects of the acidosis on various processes relevant to fitness may explain changes in whole-organism energy demand, probably paralleled by modified ion exchange, protein synthesis, and growth and feeding rates. The magnitude of effect depends on the CO_2 concentrations reached (Figure 6-10b).

The internal formation of carbonate from bicarbonate is essential to calcification, which is the formation of solid $CaCO_3$ in internal or external calcified structures, used for defense and structural support. Calcification usually occurs in separate body or cell compartments, where pH and thus CO_3^{2-} concentration and saturation Ω (Section 6.1.1) are maintained at values higher than in other body fluids or ambient water (Taylor et al., 2011; Trotter et al., 2011; McCullough et al., 2012; Venn et al., 2013). CO_2 impedes the formation of carbonate such that calcification rate decreases. It may be maintained by enhanced transport of ions, incurring elevated energetic costs (Figure 6-10).

External carbonate structures like shells rely on ambient seawater being supersaturated with carbonates. Decreasing oceanic carbonate levels reduce the saturation levels (Ω) of calcite or aragonite in the water. Reduction to below unity may lead to the corrosion of carbonate shells (FAQ 6.3). However, many species protect their shells from direct contact with seawater by various types of organic coating (e.g., a periostracum in mollusks and brachiopods, an epicuticle covering the carapace of crustaceans, an epidermis covering the tests of urchins, epithelial tissue covering aragonite in corals, and coralline algae precipitating $CaCO_3$ (mostly Mg-calcite) within their cell wall). A meta-analysis of the effects of ocean acidification on biological processes indicates that reductions in the rate of net calcification (calcification minus dissolution) and survival are the most uniform responses across organisms studied, relative to other, more variable impacts such as reduced growth, development, and abundance (Kroeker et al., 2013; see also Box CC-OA).

Some organisms benefit from elevated CO_2 partial pressures ($p\mathrm{CO}_2$). Photosynthesis and/or nitrogen fixation in selected microorganisms are impacted by OA, but effects are species or taxon specific, possibly depending on how they acquire carbon, that is, the presence and in particular the type, capacity, and energetic costs of carbon-concentrating mechanisms (CCMs; Giordano et al., 2005; Kranz et al., 2011).

A comprehensive picture of responses to CO_2 requires consideration of variable sensitivities between species and life stages and taxon-specific sensitivity distributions, as shown by a meta-analysis of animal data (Wittmann and Pörtner, 2013; see also Figure 6-10b). Echinoderms, bivalves, gastropods, and corals begin to respond negatively at lower CO_2 levels than crustaceans or cephalopods (Figure 6-10b). This sensitivity

Table 6-3 | Tolerances to ocean acidification in marine taxa, assessed from laboratory and field studies of species in the CO_2 partial pressure ($p\mathrm{CO}_2$) range from <650 to >10000 µatm, compared to present day atmospheric levels of 400 µatm. (It should be noted that anthropogenic CO_2 emissions add to the natural variability of CO_2 concentrations in marine environments, which can reach much higher than atmospheric levels.) Variables studied include growth, survival, calcification, metabolic rate, immune response, development, abundance, behavior, and others. Neither all life stages, nor all variables, including the entire range of CO_2 concentrations, were studied in all species. *Confidence* is based on the number of studies, the number of species studied, and the agreement of results within one group. + denotes that possibly more species or strains (genetically distinct populations of the same species) were studied, as only genus or family were specified; beneficial: most species were positively affected; vulnerable: more than 5% of species in a group will be negatively affected by 2100; tolerant: more than 95% of species will not be affected by 2100. RCP 6.0: Representative Concentration Pathway (RCP) with projected atmospheric $p\mathrm{CO}_2$ = 670 µatm; RCP 8.5: $p\mathrm{CO}_2$ = 936 µatm in 2100 (Meinshausen et al., 2011). *Confidence* is limited by the short- to medium-term nature of various studies and the lack of sensitivity estimates on evolutionary time scales, that is, across generations (see separate reference list, Online Supplementary Material). Note that the assessment of variability between species from the same animal phylum has revealed an increase in the fraction of sensitive species with rising CO_2 levels; see Figure 6-10.

Taxon	No. of studies	No. of parameters studied	Total no. of species studied	$p\mathrm{CO}_2$ where the most vulnerable species is negatively affected or investigated $p\mathrm{CO}_2$ range[a] (µatm)	Assessment of tolerance to RCP 6.0 (*confidence*)	Assessment of tolerance to RCP 8.5 (*confidence*)
Cyanobacteria	17	5	9+	180–1250[a]	Beneficial (*low*)	Beneficial (*low*)
Coccolithophores	35	6	7+	740	Tolerant (*low*)	Vulnerable (*medium*)
Diatoms	22	5	28+	150–1500[a]	Tolerant (*low*)	Tolerant (*low*)
Dinoflagellates	12	4	11+	150–1500[a]	Beneficial (*low*)	Tolerant (*low*)
Foraminifers	11	4	22	588	Vulnerable (*low*)	Vulnerable (*medium*)
Seagrasses	6	6	5	300–21000[a]	Beneficial (*medium*)	Beneficial (*low*)
Macroalgae (non-calcifying)	21	5	21+	280–20812[a]	Beneficial (*medium*)	Beneficial (*low*)
Macroalgae (calcifying)	38	10	36+	365	Vulnerable (*medium*)	Vulnerable (*high*)
Warm-water corals	45	13	31	467	Vulnerable (*medium*)	Vulnerable (*high*)
Cold-water corals	10	13	6	445	Vulnerable (*low*)	Vulnerable (*medium*)
Annelids	10	6	17+	1200	Tolerant (*medium*)	Tolerant (*medium*)
Echinoderms	54	14	35	510	Vulnerable (*medium*)	Vulnerable (*high*)
Mollusks (benthic)	72	20	38+	508	Vulnerable (*medium*)	Vulnerable (*high*)
Mollusks (pelagic)	7	8	8	550	Vulnerable (*low*)	Vulnerable (*medium*)
Mollusks (cephalopods)	10	8	5	2200 (850 for trace elements)	Tolerant (*medium*)	Tolerant (*medium*)
Bryozoans	7	3	8+	549	Tolerant (*low*)	Vulnerable (*low*)
Crustaceans	47	27	44+	700	Tolerant (*medium*)	Tolerant (*low*)
Fish[b]	51	16	40	700	Vulnerable (*low*)	Vulnerable (*low*)

[a] Rather than a sensitivity threshold the entire range of investigated $p\mathrm{CO}_2$ values is given for groups of photosynthetic organisms. In all studies photosynthetic rates are stimulated to different, species-specific degrees by elevated $p\mathrm{CO}_2$, indicating low vulnerability. Coccolithophores and calcifying algae are assessed as being more sensitive than other photosynthetic organisms due to reduced calcification and shell dissolution.

[b] Confidence levels for fishes were converted from medium to low, in light of uncertainty on the long-term persistence of behavioral disturbances.

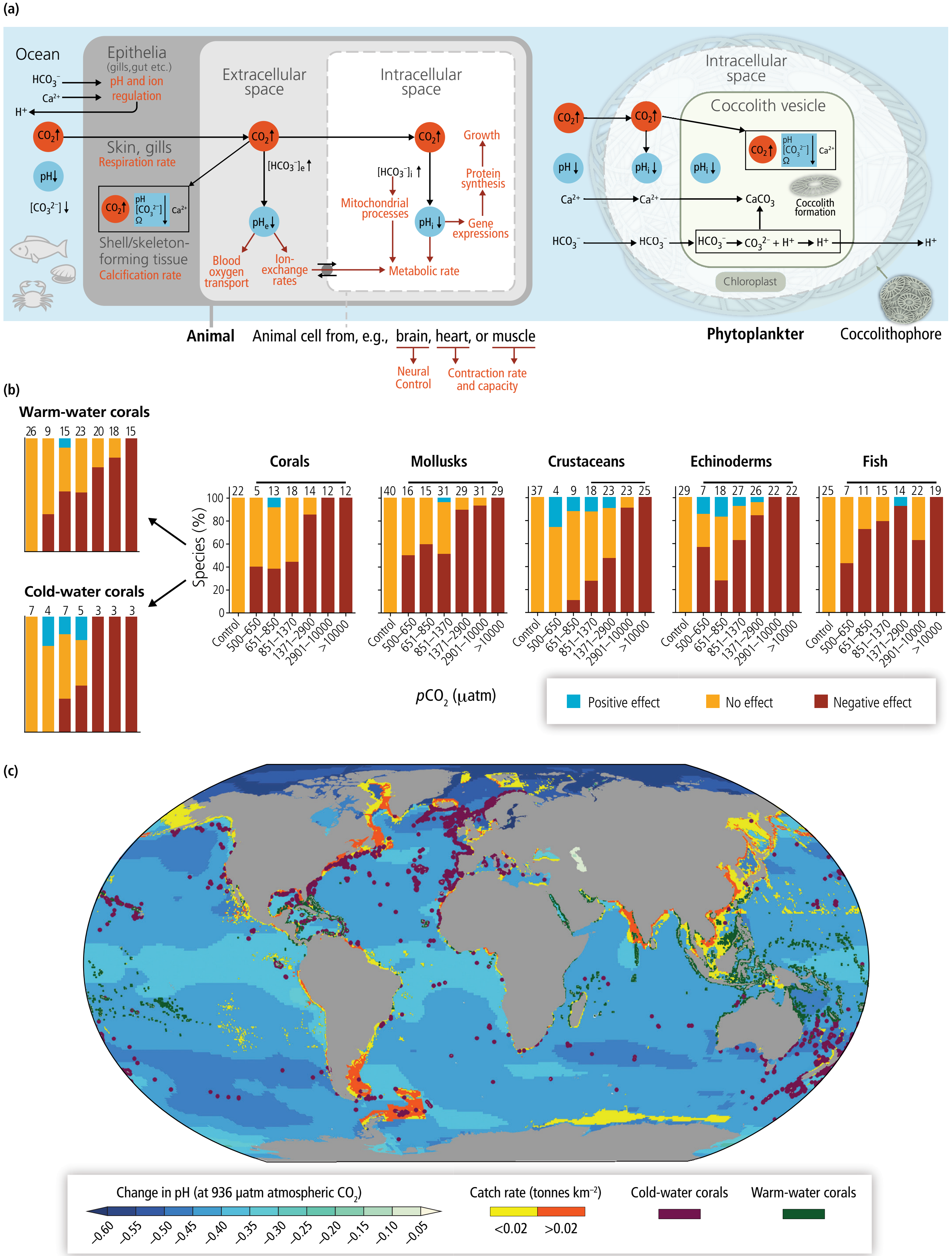

(a)
Ocean
HCO$_3^-$
Ca^{2+}
H+
CO$_2$↑
pH↓
[CO$_3^{2-}$]↓
Epithelia
(gills,gut etc.)
pH and ion regulation
Skin, gills
Respiration rate
Shell/skeleton-forming tissue
Calcification rate
Extracellular space
[HCO$_3^-$]$_e$↑
pH$_e$↓
Blood oxygen transport
Ion-exchange rates
Intracellular space
[HCO$_3^-$]$_i$↑
pH$_i$↓
Growth
Protein synthesis
Gene expressions
Mitochondrial processes
Metabolic rate
Animal
Animal cell from, e.g., brain, heart, or muscle
Neural Control
Contraction rate and capacity
Intracellular space
Coccolith vesicle
CaCO$_3$
Coccolith formation
HCO$_3^-$ → CO$_3^{2-}$ + H+ → H+
Chloroplast
Phytoplankter
Coccolithophore
(b)
Warm-water corals
26 9 15 23 20 18 15
Cold-water corals
7 4 7 5 3 3 3
Corals
22 5 13 18 14 12 12
Mollusks
40 16 15 31 29 31 29
Crustaceans
37 4 9 18 23 23 25
Echinoderms
29 7 18 27 26 22 22
Fish
25 7 11 15 14 22 19
Species (%)
0 20 40 60 80 100
Control 500–650 651–850 851–1370 1371–2900 2901–10000 >10000
pCO$_2$ (µatm)
Positive effect
No effect
Negative effect
(c)
Change in pH (at 936 µatm atmospheric CO$_2$)
–0.60 –0.55 –0.50 –0.45 –0.40 –0.35 –0.30 –0.25 –0.20 –0.15 –0.10 –0.05
Catch rate (tonnes km^{-2})
<0.02 >0.02
Cold-water corals
Warm-water corals

Figure 6-10 | (a) Responses of a schematized marine animal (left) and a phytoplankter (right) to ocean acidification. Effects are mediated via diffusive CO_2 entry (black arrows) into body and cell compartments, resulting in a rise in pCO_2 (highlighted in red), a drop in compartmental pH (highlighted in blue), and their effects (red arrows) on various processes (red text) in tissues and cellular compartments, as well as on calcium carbonate saturation state (Ω) at calcification sites (after Pörtner, 2008; Taylor et al., 2011). Variable sensitivity relates to the degree of pH decline and compensation, depending on the capacity of pH and ion regulation. (b) Distribution of sensitivities across species within animal phyla, under progressively rising water CO_2 levels, as percent of studied cold- and warm-water coral (mostly scleractinia), echinoderm, molluskan, crustacean, and fish species affected negatively, positively, or not at all (for effects considered, see text). As not all life stages, variables, and pCO_2 ranges were covered in all species, two assumptions partially compensate for missing data: 1) Negative effects at low pCO_2 will remain negative at high pCO_2. 2) A positive or neutral outcome at both low and high pCO_2 will be the same at intermediate pCO_2. As responses reported for each species vary for each pCO_2 range, variable species numbers result (on top of columns). The total number of species studied in a group is shown as the number above the control column. The control category corresponds to 380 µatm. For 2100, RCP scenarios falling within each CO_2 partial pressure (pCO_2) category are as follows: RCP4.5 for 500–650 µatm (approximately equivalent to ppm in the atmosphere), RCP6.0 for 651–850 µatm, and RCP8.5 for 851–1370 µatm. By 2150, RCP8.5 falls within the 1371–2900 µatm category. Horizontal lines above columns represent frequency distributions significantly different from controls (Wittmann and Pörtner, 2013). Data for warm- and cold-water corals as in Table 6-3. (c) Areas with reported annual catches of marine calcifiers (crustaceans and mollusks) ≥ 0.005 tonnes km^{-2} depicted on a global map (weighted mean of the orange color area =0.07 tonnes km^{-2}) showing the distribution of ocean acidification in 2100 according to RCP8.5 (WGI AR5 SPM; pH change from 1986–2005 to 2081–2100) as well as the distribution of warm-water (green dots) and cold-water coral communities (purple dots).

pattern resembles the one seen in the Permian mass extinction (Knoll et al., 2007; Knoll and Fischer, 2011). The picture for fishes is less clear, as the present findings of high vulnerability are not met by similar observations in the fossil record. Evolutionary adaptation may thus eliminate or minimize reported effects.

The capacity for pH and ion regulation and other relevant processes can be upregulated by gene expression, as seen in acclimation studies in echinoderm larvae (O'Donnell et al., 2010; Martin et al., 2011) and fishes (Deigweiher et al., 2008; Tseng et al., 2013), in warm-water coral branches (Kaniewska et al., 2012), but not in a study of warm-water coral larvae (Moya et al., 2012). Few studies address whether and to what extent species undergo evolutionary adaptation to high pCO_2, as seen in the coccolithophore *Emiliania huxleyi* over 500 asexual generations (Lohbeck et al., 2012). In organisms with longer generation times, perturbation studies in the laboratory measure tolerance and acclimation, but not adaptation or natural selection. Animal adaptation is accelerated by high functional variability among larvae, enabling selection of resistant genotypes (*low* to *medium confidence*; Sunday et al., 2011; Parker et al., 2012; Pespeni et al., 2013). This may explain the selective mortality seen in Atlantic cod larvae under elevated CO_2 (Frommel et al., 2012). Both acclimatization and adaptation will shift sensitivity thresholds but the capacity and limits of species to acclimatize or adapt remain largely unknown and hence impacts of acute exposures cannot easily be scaled up to effects on the longer, evolutionary time scales of ocean acidification (Wittmann and Pörtner, 2013). Observations in ecosystems characterized by permanently elevated or fluctuating CO_2 levels, such as upwelling areas, OMZs (Section 6.1.1), or seeps, reflect the existence of sensitivity thresholds (*high confidence*; Section 6.3.2.5) but organisms may have evolved a higher resistance to increased CO_2 levels than elsewhere (*low confidence*).

Table 6-3 compiles effects of ocean acidification observed across taxa in laboratory and field experiments. The latter include studies in mesocosms and at natural analogs, submarine CO_2 venting areas at locales such as Ischia, Italy (Hall-Spencer et al., 2008), Papua New Guinea (Fabricius et al., 2011), and Puerto Morelos, Mexico (Crook et al., 2012). It should be noted that anthropogenic CO_2 accumulation according to RCPs adds to the natural variability of CO_2 concentrations in marine environments. Many groups, especially sessile or non-photosynthetic calcifiers, have sensitive species with vulnerability thresholds surpassed under RCP6.0 by 2100 (*low* to *medium confidence*).

Recent meta-analyses also summarize OA effects, two for biogeochemical processes and relative effect sizes (Harvey et al., 2013; Kroeker et al., 2013), one for the distribution of sensitivity between species within major animal phyla and its change depending on ambient pCO_2 (Figure 6-10; Wittmann and Pörtner, 2013). All of these analyses consider the interaction of warming and CO_2 accumulation (Section 6.3.5). Present limitations in understanding the mechanisms of effect and their long-term persistence compounds accurate projections of the long-term effects of OA (*medium confidence*; Wittmann and Pörtner, 2013).

6.3.2.2. Microbes

The physiology of both calcifying (coccolithophores) and non-califying phytoplankton can be influenced by changes in carbonate system variables caused by ocean acidification (Figure 6-10a). Growth and photosynthetic rates of diatoms in laboratory cultures are considered relatively insensitive to elevated CO_2 (Rost et al., 2003; Trimborn et al., 2008). Dinoflagellate sensitivity to elevated CO_2 is poorly studied (Hansen et al., 2007), but in one species carbon fixation rates were enhanced at 750 µatm CO_2 while growth remained unaffected (Fu et al., 2008). Indirect effects of ocean acidification on phytoplankton physiology include altered availability of trace metals needed for many biochemical cycles (Hoffmann et al., 2012).

Harmful algal blooms are a growing problem in coastal waters worldwide (Section 6.4.2.3), and many of the various phytoplankton species that produce bio-accumulated toxins are sensitive to changes in the seawater carbonate buffer system (Hallegraeff, 2010; Fu et al., 2012). For example, the dominance and community structure of harmful bloom dinoflagellates can be profoundly altered by changing pCO_2 (Tatters et al., 2013), and both toxic dinoflagellates and diatoms have been shown to produce higher toxin levels under near-future levels of ocean acidification (Fu et al., 2010; Sun et al., 2011).

Some planktonic N_2-fixing cyanobacteria (diazotrophs), for example, strains (genetically distinct populations of the same species) of offshore cyanobacteria of the genera *Trichodesmium* and *Crocosphaera*, respond to rising CO_2 with increased rates of both carbon and N_2 fixation (Fu et al., 2008; Lomas et al., 2012). In contrast, laboratory studies using the bloom-forming cyanobacteria *Nodularia* (an organism largely found in coastal stratified, eutrophic waters) revealed decreased growth and N_2

fixation under elevated CO_2 conditions (Czerny et al., 2009). The wide range of responses in N_2 fixation (e.g., Hutchins et al., 2007; Levitan et al., 2007; Kranz et al., 2010) may be explained by different CO_2 affinities (i.e., dependences of growth rates on CO_2 concentration) of a range of N_2-fixing cyanobacteria (*Trichodesmium* and *Crocosphaera*) from different oceanic biomes. Some species/strains operate at close to maximum growth rates at present-day oceanic CO_2 levels, whereas others had sub-optimal growth rates under these conditions (Hutchins et al., 2013). To date, the physiological mechanisms underlying these responses remain unknown, especially in open-ocean nitrogen fixers. Cyanobacteria may reallocate energy from their energetically expensive CCMs toward N_2 fixation and the acquisition of growth limiting nutrients (Kranz et al., 2010; Levitan et al., 2010), but evidence for such diversion of energy is lacking. Whether nitrogen fixation will increase with progressive ocean acidification remains to be explored (*low confidence*, *limited* in situ *evidence*, *medium agreement*).

The responses of coccolithophore calcification to OA are species specific and highly variable. The function(s) of calcification are not well understood, making it difficult to evaluate the consequences of lowered calcification (e.g., Rost et al., 2008). Reductions, increases, and unchanged calcification rates (and shell structure) have all been found in different coccolithophore species for RCP8.5 CO_2 conditions projected around 2100 (Riebesell et al., 2000; Zondervan et al., 2001; Langer et al., 2006; Iglesias-Rodriguez et al., 2008). Calcification in coccolithophores is species (Langer et al., 2006) and in *Emiliania huxleyi* even strain specific (Langer et al., 2009, 2011; Hoppe et al., 2011). It thus remains unclear whether OA will result in exoskeletons that are insufficiently calcified for sustained structural support and protection in coccolithophores (*medium evidence*, *low agreement*).

Foraminifera display decreasing calcification and shell weight under elevated CO_2 (Lombard et al., 2010). Changes in historical specimens (Moy et al., 2009; see Section 6.3.2.5.1) and during glacial-interglacial cycles (Barker and Elderfield, 2002) support projections of future reductions in net calcification by foraminifera (*medium* to *high confidence*).

6.3.2.3. Macroalgae and Seagrasses

Primary production, shoot density, reproductive output, and below-ground biomass of seagrasses generally respond positively to elevated pCO_2, indicating CO_2 limitation of their productivity. Such effects were identified in both laboratory and field above 720 to 1800 µatm (*high confidence*; e.g., Palacios and Zimmerman, 2007; Hall-Spencer et al., 2008; Andersson et al., 2011; cf. Section 5.4.2.3). Production, growth, and recruitment of most but not all non-calcifying seaweeds also increased at CO_2 levels from 700 to 900 µatm (RCP8.5; Porzio et al., 2011; Kroeker et al., 2013). Some non-calcifying seaweeds and seagrasses will thus benefit from future ocean acidification (*high confidence*) but OA exposes them to higher than usual grazing as a consequence of losing deterrent phenolic substances (*low confidence*; Arnold et al., 2012).

Calcifying algae (corallines) show complex and species-specific responses of photosynthesis to elevated CO_2, but calcification is impacted once species-specific pCO_2 thresholds are surpassed (*medium confidence*; Anthony et al., 2008; Martin and Gattuso, 2009). At habitat temperature calcification by temperate coralline red and calcareous green algae increased at CO_2 levels up to 900 µatm and decreased only at the highest concentration applied (2850 µatm), but did not fall below rates found at present-day pCO_2 (Ries et al., 2009). During 3 months of exposure, growth of *Lithothamnion glaciale*, a cold-water calcareous red alga, decreased progressively with rising CO_2 levels, and its structural integrity was weakened beyond 590 µatm (Ragazzola et al., 2012), potentially influencing ecosystem function. Some calcifying algae may thus be impacted by future ocean acidification (*medium confidence*).

6.3.2.4. Animals

Studies of marine animals and their life stages show a high diversity and variability of processes affected by ocean acidification. Many variables studied reflect physiological performance (O_2 consumption, exercise, behavior, calcification, growth, immune response, acid-base balance, gene expression, fertilization, sperm motility, developmental time, production of viable offspring, and morphology; Table 6-3; Figure 6-10). In some species growth may be stimulated by OA, in others depressed or unaffected (cf. Gooding et al., 2009; Munday et al., 2009a, 2011a; Dupont et al., 2010). The degree of CO_2-induced acidosis and its compensation by ion exchange may shape sensitivity (Section 6.3.2.1). Full exploitation of the ability to resist pCO_2 increases depends on the availability and high quality of food and the strengthening of fitness (Gooding et al., 2009; Melzner et al., 2011). However, food quality of prey organisms may decrease under elevated pCO_2. For example, slower reproduction and growth of the copepod *Acartia tonsa* under 760 µatm pCO_2 was related to the decreasing quality of its diatom food (Rossoll et al., 2012).

Changes in calcification rates reported from CO_2 manipulation experiments vary widely. Reduced calcification and weakened calcified structures were seen under elevated pCO_2 in corals (see Section 6.3.2.4.2), echinoderms (Kurihara and Shirayama, 2004), mollusks (Gazeau et al., 2013), and larval crustaceans (Arnold et al., 2009; Walther et al., 2011). Some adult limpets and urchins increased calcification rates at pCO_2 from 600 to 900 µatm, before it fell at even higher pCO_2. In some adult crabs, lobsters, and shrimps calcification rates increased further with rising pCO_2 (Ries et al., 2009). Stronger internal structures such as cuttlebones and otoliths resulted from enhanced calcification under elevated pCO_2 in juvenile cuttlefish (cephalopods: Gutowska et al., 2008) and fishes (Checkley, Jr. et al., 2009; Munday et al., 2011b), with unclear impacts on fitness. Energy costs in epithelia or calcification compartments may be enhanced by elevated pCO_2 causing a stimulation of metabolism (Section 6.3.2.1). In some cases, this may indicate imbalances in energy budget rather than increased CO_2 resistance, for example, if costs are down-regulated in muscle or liver. Enhanced calcification can then occur at the expense of growth (*medium confidence*; Wood et al., 2008; Beniash et al., 2010; Thomsen and Melzner, 2010; Parker et al., 2011).

Studies on calcifying zooplankton focused on pteropods (planktonic mollusks with aragonite shells). These form an integral part of the food web, both as grazers and prey, for example, for pink salmon (Armstrong et al., 2005; Hunt et al., 2008). In the Sub-Arctic, the Arctic, and the Southern Ocean, pteropods will reduce calcification in response to OA until at least the end of the century (*medium confidence*; Orr et al., 2005; Comeau et al., 2009; Lischka et al., 2011).

Elevated CO_2 causes behavioral disturbances in fishes (studied mostly in larvae and juveniles; Munday et al., 2010; Ferrari et al., 2011; Domenici et al., 2012; Jutfeld et al., 2013) through neural mechanisms (Nilsson et al., 2012). The long-term persistence and evolutionary relevance of these behavioral effects need further study before general conclusions can be drawn (*low confidence*; Wittmann and Pörtner, 2013; see also Table 6-3).

6.3.2.4.1. Animal life cycles

It is generally held that organisms at early life stages are always more sensitive to environmental stress than adults. In the context of ocean acidification this statement is supported by findings like larval oyster fatalities in aquaculture caused by upwelled CO_2-rich waters (*high confidence*; Barton et al., 2012). A key aspect may also be that larvae growing or developing more slowly under elevated CO_2 as in various groups including fishes (Baumann et al., 2012; see also Section 6.3.2.1) may encounter enhanced mortalities due to prolonged predator exposure. Comparative studies of animal sensitivities to OA over a complete life cycle or during critical transition phases (e.g., fertilization, egg development and hatching, metamorphosis, molting) are scarce and do not support generalized conclusions (*low confidence*).

Effects of elevated CO_2 on one life stage or transition phase may affect or carry over to the next one. Molting success into the final larval stage was reduced in a crab species (Walther et al., 2010). In a sea urchin species, negative impact was found to accumulate during 4 months acclimation of adults reducing reproductive success. This impact was, however, compensated for during extended acclimation of female urchins for 16 months (Dupont et al., 2013). Negative impact was still transferred from urchin larvae to juveniles under elevated pCO_2. Conversely, adult oysters acclimated to high CO_2 acquired resistance which was carried over to their offspring (Parker et al., 2012). More long-term acclimation studies to realistic emission scenarios are needed for generalized conclusions. Furthermore, the preposition that juvenile life stages are always more sensitive than adults needs thorough re-investigation in the context of ocean acidification, especially in the context of the notion that larvae may provide a selection pool for survival of the most suitable phenotypes (*low confidence*; Section 6.3.2.1).

6.3.2.4.2. Warm- and cold-water coral communities

In warm-water reef-building corals, OA causes genus-specific reductions in calcification (Leclercq et al., 2002; Langdon and Atkinson, 2005; Kleypas and Langdon, 2006). Nutrient availability to symbionts may sustain calcification. Heterotrophic feeding by the corals also supports energy-dependent calcification and acid-base regulation, and thus resilience (Edmunds, 2011; Figure 6-10). Females may sacrifice calcification more than males due to energetic trade-offs with reproduction (Holcomb et al., 2012). Warm-water corals are thus sensitive to future OA (*high confidence*; Table 6-3).

The cold-water coral *Lophelia pertusa* shows resilience to ocean acidification. In short-term ship-board incubations pH reductions between 0.15 and 0.3 units (540 and 790 µatm) led to calcification rates reduced by 30 to 56% (Maier et al., 2009), especially in young, fast growing polyps. However, net calcification was maintained at seawater aragonite saturation <1. Exposure to a pCO_2-induced pH reduction by 0.1 units or even to the projected end of century pCO_2 of 930 µatm led to calcification rates being maintained over 6 to 9 months (Form and Riebesell, 2012; Maier et al., 2013). This ability is probably due to a regulated upward shift of pH and carbonate saturation at organismal calcification sites (McCulloch et al., 2012; see also Figure 6-10). Natural distribution of other cold-water species covers wide natural pH gradients in Chilean fjords (*Desmophyllum dianthus*; Jantzen et al., 2013) and ranges into waters with undersaturated carbonates as in Australian waters (four scleractinian corals; Thresher et al., 2011). Pre-adaption to elevated pCO_2 apparently exists; however, species vulnerabilities to further increases in pCO_2 have not been investigated. Again, vulnerability is species specific, colonial scleractinians may be limited to water saturated or near-saturated with aragonite, whereas others are not (Thresher et al., 2011). Conclusions on the relative vulnerability of the group appear premature (Table 6-3). To what extent a further lowering of carbonate saturation values will influence the future distribution of various calcite or aragonite forming cold-water corals is not clear (*low confidence*; Guinotte et al., 2006).

6.3.2.5. Ecosystems

For insight into ecosystem level processes, laboratory studies have been supplemented with experimental studies in large volume mesocosms (i.e., >1000 L) and in the field, and with long-term field observations. Together they inform the debate over the attribution of field observations to ocean acidification.

6.3.2.5.1. Evidence from field observations

Contributions of anthropogenic ocean acidification to climate-induced alterations in the field have rarely been established and are limited to observations in individual species (see also Section 30.5.1.1.3). Shell thinning in modern planktonic foraminifera (collected 1997–2004) in the Southern Ocean compared to those from the Holocene and before was attributed to anthropogenic ocean acidification (Moy et al., 2009). Both anthropogenic OA and the upwelling of CO_2-rich deep waters (Section 30.5.4.1.4) were held responsible for shell thinning in planktonic foraminifera in the Arabian Sea over the last century (de Moel et al., 2009) or in live pteropods collected in 2008 in the Southern Ocean (*medium evidence*, *medium agreement*; Bednaršek et al., 2012). However, no changes were observed in a 57-year record of the composition and abundance of calcifying zooplankton in the increasingly acidified California Current System (Ohman et al., 2009). Possible explanations for the absence of significant responses in some studies include insufficient lengths of time series (Section 6.1.2), organisms being pre-adapted to naturally high CO_2 in upwelling or other systems, linked to a low signal-to-noise ratio, or the difficulty of detecting small OA effects in comparison with larger ecosystem effects of other drivers such as temperature, for example, in calcifying plankton (Beaugrand et al., 2013). Similarly, declines in coral calcification and performance in the field (De'ath et al., 2009) were attributed to thermal extremes, but may also include an as-yet unclear contribution from OA.

6.3.2.5.2. Microbial communities and nutrient cycles

Laboratory experiments, coastal mesocosm studies (Weinbauer et al., 2011), and field experiments (Beman et al., 2011; Law et al., 2012) have yielded various, sometimes conflicting, results on the effects of CO_2 on microbial processes. From a meta-analysis of available data, Liu et al. (2010) conclude that the rates of several microbial processes will be affected by OA, some positively, others negatively. The potential of the microbial community to adapt to ocean acidification and maintain functionality, either by genetic change at the species level or through the replacement of sensitive species or groups at the community level, remains to be explored further. At the present time there is insufficient field-based evidence to conclude that elevated CO_2 will affect natural assemblages of microorganisms (*limited evidence*, *low agreement*) with the possible exception of the negative impact on calcification (Joint et al., 2011).

Experimental studies on OA effects (through reduced pH or increased CO_2) on autotrophic and heterotrophic microbial production have provided inconsistent results. Microbes are characterized by large diversity and broad environmental adaptation, and hence may respond to environmental challenges by exploiting such diversity via species replacements (Krause et al., 2012). This makes it difficult to project the findings of laboratory experiments investigating the response of microbes to OA to the ecosystem level. Relevant variables include cellular elemental stoichiometry (C:N:P ratios; Riebesell, 2004; Fu et al., 2007), rates of CO_2 and N_2 fixation (Riebesell, 2004; Hutchins et al., 2007, 2009), rates of nitrification (Beman et al., 2011), changes in the proportion of dissolved organic carbon (i.e., DOC) to particulate photosynthate produced during carbon fixation (Kim et al., 2011), and the response of viruses (Danovaro et al., 2011).

Field experiments led to the projection that nitrification rates (ammonia oxidation to nitrite and nitrite oxidation to nitrate) of bacteria and archaea will be reduced by 3 to 44% when pH is reduced by 0.05 to 0.14 (Beman et al., 2011), corresponding to a mean rise in CO_2 by approximately 100 µatm. The reported decrease in nitrification occurred regardless of natural pH variability, providing no evidence for acclimation of the nitrifiers to reduced pH, for example, in upwelling areas. Potential changes in microbial cell abundance, possibly as a result of lower cellular nitrification rates, could further decrease the total rate of nitrification.

It remains unclear whether OA has contributed to the systematic changes in phytoplankton abundance and community structure observed over recent decades, which have largely been attributed to warming (Chavez et al., 2011). In natural assemblages from coastal and polar waters, NPP is stimulated by increased CO_2 (*medium confidence*; Riebesell et al., 2008; Tortell et al., 2008). Small differences in CO_2 sensitivity may lead to pronounced shifts in the dominance of species (Tortell et al., 2008; Beaufort et al., 2011). Quantification of the calcite mass of the coccolithophore community in the present ocean and over the last 40 kyr were in large part attributed to shifts between differently calcified species and morphotypes according to carbonate chemistry (Beaufort et al., 2011). The same study, however, also observed heavily calcified *Emiliania huxleyi* morphotypes in upwelling systems characterized by low pH, a finding which highlights the complexity of assemblage-level responses and may indicate pre-adaptation to elevated pCO_2. Owing to the complex response patterns, it is not possible to project ecosystem-level effects from effects on coccolithophore calcification in monospecific culture experiments (*low confidence*). Projections of OA impacts on phytoplankton become even more complicated by synergistic interactions with other drivers (Boyd, 2011; see also Section 6.3.5).

6.3.2.5.3. Macrophytes and macrofauna

Macrofauna and macrophyte communities have been studied in mesocosms and in ecosystems exposed to shifted upwelling regimes or at natural volcanic CO_2 vents (Fabricius et al., 2011; Kroeker et al., 2011). The latter are considered as natural analogs of future ocean acidification. An 8-year trend of (variable) pH decline in upwelled waters along the Northeast Pacific coast was paralleled by shifts in community composition, where shelled species like mussels were replaced by fleshy algae and barnacles (Wootton et al., 2008). Macrofaunal calcifiers at CO_2 vents (Hall-Spencer et al., 2008; Fabricius et al., 2011) and in mesocoms (Christen et al., 2013) display a lowering of species richness. These findings suggest that non-calcifiers increasingly outcompete calcifiers once pH_T decreases to a mean of 7.8 to 7.7 (*medium confidence*). Finally, a loss of calcifiers from mesocosms occurred around 0.5 units below the pH values expected from OA under RCP8.5 by 2100 (*medium confidence*; Christen et al., 2013). At CO_2 seeps, calcitic bryozoans replace coralline algae, which have more soluble high-calcite skeletons (Martin et al., 2008). Seagrasses and non-calcifying algae gain a competitive advantage (Fabricius et al., 2011). Coral communities exposed to high pCO_2 waters (from upwelling or seeps) have lower growth, calcification, and biodiversity (Manzello et al., 2008; Fabricius et al., 2011), resulting in a shift from net accretion to erosion (Box CC-CR). The use of seeps as analogs of future OA is limited as pH variability is high at these sites, such that effective values may be lower than indicated by the average change (Hall-Spencer et al., 2008; Porzio et al., 2011). During periods of high pH at the seeps, they are recolonized by invertebrates and fishes from neighboring areas with normal pH, compromising assessments of long-term sensitivity thresholds. Overall, findings available from mesocosms and natural analogs indicate losses in diversity, biomass, and trophic complexity of benthic marine communities due to elevated CO_2 (*high confidence*) and support the projection of similar shifts in other systems with continued OA (*medium confidence*).

Enhanced freshwater input by poorly buffered rivers or by precipitation, into estuaries, brackish oceans like the Baltic (Section 30.5.3.1.4), and into freshening polar oceans, reduces salinity and alkalinity at rising atmospheric pCO_2 and thereby, alters the carbonate system and enhances OA (Section 6.1.1). Estuaries usually have OMZs, where background pCO_2 is elevated. Its reduction by dilution causes the acidification effect to be somewhat less. Enhanced pH reduction and variability in hyposaline waters may constrain the distribution of sensitive species further (*low confidence*; Miller et al., 2009; Denman et al., 2011).

6.3.2.5.4. Conclusions

Natural analogs and laboratory and mesocosm experiments provide evidence for differential effects of ocean acidification on species and communities. Sensitivity to OA is species specific (*high confidence*); differential sensitivities and associated shifts in performance and distribution will change predator-prey relationships and competitive interactions (*low* to *medium confidence*). OA may stimulate global net primary production (*low confidence*) and nitrogen fixation (*medium confidence*). OA will increase the abundance and primary production of non-calcifying macrophytes, but will be harmful to calcifying algae and many heterotrophs (*medium confidence*). Ecosystems relying on calcified structures and at risk of dissolution under OA include warm-water coral reefs (*high confidence*) and their cold-water equivalent (*medium confidence*). Further studies need to explore how OA may change the composition of communities, impact food webs, and affect higher trophic levels.

6.3.3. Life in Hypoxia and Anoxia

6.3.3.1. Principles

Hypoxia constrains organisms which rely on aerobic metabolism (Section 6.1.1; FAQ 6.2). Below O_2 concentrations of 60 µmol kg^{-1}, commonly termed hypoxic (Section 6.1.1.3), communities undergo species losses and replacements and are transformed into communities with species showing characteristic hypoxia adaptations. However, O_2 can limit animal life at even higher levels, just below air saturation (Gilly et al., 2013). Organisms' tolerance thresholds have been defined by either the critical O_2 partial pressure (P_c) or concentration (O_2crit). Thresholds vary across domains and are highest for large multicellular organisms. Among these, the P_c at rest varies depending on species, body size, and life stage. In animals below the P_c aerobic metabolic rate fails to be maintained and anaerobic metabolism contributes to energy production (Pörtner and Grieshaber, 1993). The critical oxygen threshold is set by the capacity of ventilatory and circulatory systems to supply O_2 and cover demand. The threshold increases once metabolism is stimulated by muscular activity, temperature, or food uptake (Pörtner, 2002a; Ekau et al., 2010; Seibel, 2011; see also Figure 6-11). At extreme temperatures, O_2crit approaches the oxygen content of air-saturated water (Pörtner, 2010; McBryan et al., 2013), indicating high sensitivity to hypoxia in the warmth. Most animals can only sustain anaerobic metabolism temporarily, even if they are energy efficient and survive long periods of anoxia (Grieshaber et al., 1994). Such time-limited tolerance is higher in large than in small individuals or larvae, related to the higher capacity of anaerobic metabolism in large specimens (Gray et al., 2002; Jessen et al., 2009).

6.3.3.2. Microbes

Bacteria and protists consume ambient oxygen down to very low levels in oxygen minimum zones and sustain OMZs by their metabolic diversity (Figure 6-11; WGI AR5 Section 3.8.3). OMZs form habitat for both anaerobic and aerobic microbes that can utilize very low (<1 µmol kg^{-1}) O_2 concentrations (Stolper et al., 2010). Hypoxia is paralleled by elevated pCO_2 and enhanced acidification. Expanding OMZs will select for the proliferation of spezialized microbes (*high confidence*).

6.3.3.3. Animals and Plants

In mesopelagic OMZs, zooplankton also contribute to the development of hypoxia (Robinson et al., 2010; see also FAQ 6.2). During daytime zooplankton congregate at the upper margin of OMZs, where the degradation of organic material causes intensified respiration and oxygen depletion (Bianchi et al., 2013). Animals living permanently in the OMZ still cover virtually all energy demand by aerobic metabolism. This requires special adaptations leading to a reduction in O_2 and energy demand, and the improved ability to use available O_2 efficiently. Enhanced hypoxia tolerance reflected in low O_2crit values is supported by small body size and by cold temperature (Vetter et al., 1994; Pörtner, 2002b; Levin et al., 2009). Accordingly, low O_2 levels support abundant meiofauna (very small fauna, <1 mm) that benefit from abundant food and reduced predation by larger organisms (Levin, 2003). Under suboxia only specialists can survive (Vaquer-Sunyer and Duarte, 2008). Expansion of suboxic and anoxic centres of pelagic OMZs and benthic dead zones will lead to loss of habitat for animal life (*high confidence*).

Large, more active animals such as fishes, crustaceans, and muscular (as opposed to ammoniacal) squids tend to have high O_2 demands associated with high O_2crit thresholds, and are therefore excluded from permanently hypoxic water bodies. However, even in high-activity animal groups some specialists such as Humboldt squid or bigeye tuna have adapted to enter hypoxic environments though only temporarily (Richards et al., 2009; Seibel, 2011). The time-limited tolerance of animals to hypoxia below the O_2crit is maximized by the depression of energy demand, for example, during periods of metabolic arrest (e.g., developmental arrest or diapause of copepods; Auel et al., 2005). Hypoxia-adapted lifeforms will benefit from expanding OMZs (*high confidence*).

There is little information on the hypoxia sensitivity of macrophytes or their O_2crit values. In eelgrass (*Zostera marina*), warming causes the hypoxia threshold to rise due to a strong increase in tissue respiration. Concomitant water or sediment hypoxia can elicit tissue anoxia and sudden die-offs (Raun and Borum, 2013). By contrast, macroalgae attached to rocks rarely encounter anoxia (Raven and Scrimgeour, 1997). Expanding benthic OMZs will constrain the distribution of macrophytes (*medium confidence*).

6.3.3.4. Ecosystems

OMZs, shoaling, and expanding vertically and laterally (Gilly et al., 2013) will cause habitat and abundance losses for intolerant taxa such as mesopelagic (Koslow et al., 2011) and epipelagic fishes with a high O_2 demand (*medium confidence*; Prince et al., 2010; Stramma et al., 2012; see also FAQ 6.2). In line with the distribution of hypoxia sensitivities (Figure 6-11; Sections 6.3.3.1, 6.3.3.3), expanding OMZs will further constrain the distribution of key zooplankton and nekton species and influence their diurnal and ontogenetic vertical migrations (*medium confidence*; Ekau et al., 2010). The composition of microbial and faunal pelagic communities will shift from diverse mid-water assemblages to migrant biota that return to oxygenated surface waters at night (Seibel, 2011). Dissolved O_2, among other factors, plays an important role in shaping large alternating fluctuations of sardine and anchovy abundances, particularly off Peru. Anchovies are not strongly affected by a shallow oxycline (<10 m), while sardines actively avoid such conditions (Bertrand et al., 2010). Where OMZs intersect the continental shelves, groundfishes (McClatchie et al., 2010) and large benthic invertebrates such as crabs display high mortalities (Chan et al., 2008). Susceptibility of early life stages to hypoxia in both pelagic and benthic ecosystems (Ekau et al., 2010) threatens population survival. Effects of hypoxia propagate along the food chain, constraining fish stocks and top predators (*high confidence*; Stramma et al., 2010). Hypoxia reduces biodiversity (Levin et al., 2009; Gooday et al., 2010) and causes the marginalization of calcifiers, due to low metabolic rates and high pCO_2 (*high confidence*; Levin, 2003; Levin et al., 2009).

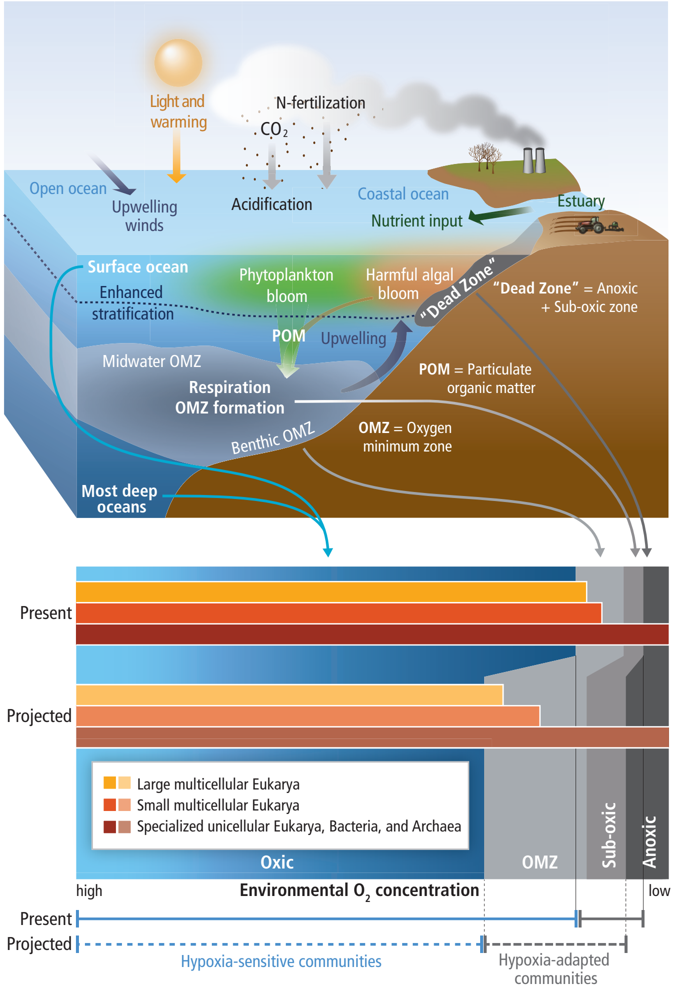


Figure 6-11 | (a) Principal mechanisms underlying the formation of hypoxic conditions and their biological background (modified from Levin et al., 2009; Levin and Sibuet, 2012). The buoyancy flux from fluvial discharges produces sharp density stratification at the base of the freshened layer (also valid for ice melt and high precipitation) near the surface and, hence, vertical mixing is greatly reduced. In consequence, the nutrient inputs from the river and the atmosphere accumulate in a narrow upper layer, leading to blooms of phytoplankton, possibly including harmful algae. The enhancement of oxygen consumption due to aerobic decomposition of sinking particulate organic matter (POM) results in hypoxic conditions of benthic and mid-water oxygen minimum zones (OMZs). Enrichment of nutrients (eutrophication) results in coastal dead zones. In the open oceans, heating of the upper layer increases stratification, while the wind-driven upwelling of hypoxic, nutrient-rich water from deeper layers adds to the formation of the OMZs (Box CC-UP). (b) Distribution of free-living marine organisms (microbes such as archaea, bacteria, protists, small and large multicellular animals, and plants) across the ranges of O_2 concentrations in various water layers. Hypoxia tolerance is enhanced in small compared to large organisms, allowing unicellular species and small animals to thrive in extremely hypoxic habitats. Species richness and body size of animals decrease with falling O_2 levels.

The expansion and enhanced variability of OMZs increases dissimilatory nitrate reduction and anaerobic ammonium oxidation (anammox), both releasing N_2 into the atmosphere, reducing the availability of fixed nitrogen, and limiting oceanic primary productivity (*medium confidence*). Water column denitrification and N_2 fixation are spatially and temporally variable (*limited evidence*, *low confidence*), suggesting that climate effects on these processes are unlikely to operate uniformly (Brandes et al., 2007; Fernandez et al., 2011; Franz et al., 2012).

If O_2 levels decline and OMZs expand, tolerant taxa, such as anaerobic bacteria (Ulloa et al., 2012), gelatinous zooplankton (medusae, ctenophores), selected fishes (gobies, hake), and possibly selected cephalopods (Gilly et al., 2006; Bazzino et al., 2010) will respond with range expansions or population growth. Similar phenomena are expected with intensified upwelling causing extensive mortalities of coastal fishes and invertebrates (Box CC-UP). A community change toward hypoxia-tolerant fauna will occur in mid-water (*high confidence*). The diversity of macroorganisms will decrease and, finally, higher marine organisms will disappear and heterotrophic microorganisms will dominate (*high confidence*). In isolated water bodies such as the Black Sea, warming will lead to the expansion of anoxia and hydrogen sulphide (H_2S) poisoning, reduce pelagic and bottom faunal distributions, and shape trophic relations, energy flows, and productivity (Daskalov, 2003; Fashchuk, 2011).

6.3.4. Mixed Layer Depth and Light Shaping Net Primary Production

The upper ocean is characterized by physical and chemical gradients in the surface mixed layer that influence the magnitude of photosynthetic carbon fixation, often termed net primary production (NPP). The availability of light and nutrients to photoautotrophs sets daily rates of NPP and may be altered directly or indirectly, through changing mixed layer depths, shifts in the circulation regime at different spatial scales, and the physical displacement of organisms (Section 6.1.1.4; Box CC-PP; Figure 6-2). A changing climate will affect mixed layer depth, cloudiness, and/or sea ice areal extent and thickness and thereby modulate NPP (*high confidence*). A stronger vertical density gradient will reduce the communication between the sunlit upper ocean where photosynthesis takes place and the underlying nutrient-rich waters (Figure 6-2). The supplies of plant nutrients (macro-nutrients) such as nitrate, and of micro-nutrients such as iron (Pitchford and Brindley, 1999) vary seasonally (Boyd, 2002) and regionally (Moore et al., 2002), such that NPP may be simultaneously limited (co-limited) by more than one resource (Saito et al., 2008; see also Section 6.3.5).

The changing range and intensity of underwater light will lead to changes in NPP as well as in phytoplankton community composition (Doney, 2006; Boyd et al., 2010). The response of phytoplankton to changing sunlight involves photo-physiological acclimation via changes in cellular chlorophyll, but such acclimation is constrained by unidentified limits (Falkowski and Raven, 1997). A longer growing season, with more sea ice-free days between 1998 and 2009, may have increased NPP in open Arctic waters (Arrigo and van Dijken, 2011; see also Box CC-PP), complemented by massive under-ice blooms as seen in 2011, favored by light that penetrates surface melt ponds and thinner, for example, first-year ice (Arrigo et al., 2012). There are also reports of increased incidences of high phytoplankton stocks, and hence of greater NPP, deeper in the water column (i.e., where it cannot be detected by satellite) during summer in the Arctic, which have implications to assessing changes in NPP from space (Hill et al., 2013). Little is known about shifts from sea ice algae to free-drifting phytoplankton expected with a decrease in sea ice cover and effects of increased light in polar waters in the coming decades (*low confidence*). In the Arctic, summer ice melt led to a rapid export of sea-ice algae to the deep ocean (Boetius et al., 2013). As some krill feed primarily on sea ice algae, it is unclear (*low confidence*) whether they will adapt to feeding mainly on free-drifting phytoplankton (Smetacek and Nichol, 2005).

A range of time series observations, from *in situ* phytoplankton abundances to satellite remote sensing, have been used to assess whether phytoplankton stocks and hence rates of NPP have altered over recent decades. Increases in phytoplankton stocks were found in regions where colder waters had warmed in the Northeast Atlantic, whereas the opposite trend was observed for warm-water regions from a phytoplankton abundance time series (Richardson and Schoeman, 2004). Lower chlorophyll concentrations at warmer SSTs in nutrient-poor low-latitude waters, based on satellite ocean color data, have been interpreted as an effect of increased stratification on phytoplankton stocks. It has thus been suggested that expanding, permanently stratified, low-chlorophyll, tropical regions (WGI AR5 Chapter 3) indicate declining phytoplankton stocks in the warming oligotrophic waters of the North and South Pacific and North and South Atlantic (*limited evidence*, *low agreement* due to methodological uncertainties; Box CC-PP; Polovina et al., 2008; Signorini and McClain, 2012; see also Section 30.5.1.1.2). Furthermore, a transition to conditions favoring increased frequency or even permanence of El Niño in a warmer future (Wara et al., 2005) and further expansion of subtropical ocean gyres (Polovina et al., 2008; see also Section 30.5.6) may lead to lower global ocean NPP (*low* to *medium confidence*).

However, these long-term "blended" projections (i.e., constructing a biomass time series using multiple proxies such as ocean transparency) of a global decrease in phytoplankton biomass (Boyce et al., 2010) have been refuted (Mackas, 2011; McQuatters-Gollop et al., 2011; Rykaczewski and Dunne, 2011). Time series shorter than 20 years do not resolve impacts of bi-decadal variation such as the Pacific Decadal Oscillation or the lunar nodal cycle (e.g., Watanabe et al., 2008; Henson et al., 2010). Analysis of continental shelf ecosystems, including field data in the most productive upwelling areas covering the last 20 years (e.g., Chavez et al., 2011), revealed a large variety of trends at scales of several decades but a general increase in NPP on most shelves (Sherman and Hempel, 2009; Bode et al., 2011), possibly caused by natural climate variability, anthropogenic climate change, and/or anthropogenic eutrophication. Recent field measurements document increasing quantities of both anthropogenic fixed N (Duce et al., 2008) and biologically fixed atmospheric nitrogen (Mouriño-Carballido et al., 2011) entering the open ocean, which could lead to increased NPP especially in warm, stratified tropical and subtropical oceans provided sufficient phosphate and other growth requirements are present (*low confidence*; e.g., Sohm et al., 2011).

For heterotrophs, from bacteria to fish, mammals, and birds, the uptake of organic material as food, ultimately provided by NPP, is central not only to productivity but also for fueling energy-consuming functions including the resistance of organisms to environmental change and pathogens (Sections 6.3.1-2). Any direct influence of climate on the abundance and quality of feed organisms will thus translate to indirect effects on the productivity and well-being of foraging animals (*high confidence*; Figures 6-5a, 6-7a, 6-12).

Overall, pelagic systems respond to climate change by region-specific changes in productivity with the projection of a small net reduction in global ocean NPP by 2100 (*medium confidence*; Box CC-PP). The spatial reorganization of NPP between latitudes affects higher trophic levels by alteration of the composition and functioning of pelagic communities (*medium confidence*).

6.3.5. Concurrent Responses to Multiple Drivers

Climate change alters oceanic properties globally, with concurrent changes in temperature, dissolved CO_2 and O_2, light, and nutrient concentrations (e.g., Sarmiento et al., 1998; Matear and Hirst, 1999; Boyd and Doney, 2002; Ekau et al., 2010; see also Figure 6-2). Additional direct human interventions at regional scale comprise the introduction of non-native species, overfishing, pollution, long-range atmospheric transport of nitrogen, point-source eutrophication, and habitat destruction (Carlton, 2000; Boyd and Hutchins, 2012). Worldwide alterations in marine ecosystems (Pauly et al., 1998; Österblom et al., 2007) have been linked to direct human activities, especially fishing (Frank et al., 2005; deYoung et al., 2008; Casini et al., 2009), but may also be caused to some extent by climate variability and change (Cheung et al., 2013a).

Alteration of each individual property has pronounced effects on organisms from microbes to animals, and hence on ecosystems (Sections 6.3.1-4). The cumulative effects of these factors will result in complex patterns of change, from organismal physiology to the areal extent and boundaries of biogeographic regions (Table 6-4). In many organisms, effects of ocean acidification interact with those of other key drivers such as temperature and hypoxia (Boyd, 2011; Gruber, 2011; Pörtner, 2012) and translate from molecular to ecosystem level impacts. In phytoplankton, low light (Zondervan et al., 2002) or nitrogen limitation (Sciandra et al., 2003) limit beneficial OA effects on photosynthesis and have a strong negative effect on plankton calcification (Rokitta and Rost, 2012). Nutrients and light support functional adjustments to OA through gene expression changes (Dyhrman et al., 2006; Richier et al., 2009).

Similar to today, paleo-events such as the Palaeocene-Eocene Boundary demonstrate concurrent warming, enhanced stratification of the oceans, deoxygenation of deeper waters, and OA, albeit at a rate more than 10 times slower than today's rate (Section 6.1.2). Both the complexity of paleo-ecosystem changes and the complexity of present effects confound the clear attribution of biological trends to individual drivers (Parmesan et al., 2011). For warming and hypoxia, changes are accelerated by effects of shifting seasonal or even diurnal extremes and their frequency on organisms and ecosystems (*medium evidence*, *medium agreement*) (e.g., Pörtner and Knust, 2007; Díaz and Rosenberg, 2008). This may also apply to effects of anthropogenic OA (*limited evidence*, *low agreement*).

6.3.5.1. Principles

Effects of various climate drivers on ocean ecosystems are intertwined and effects may be exacerbated by responses of biota. For example, warming reduces O_2 solubility and enhances biotic O_2 demand, which exacerbates hypoxia, produces CO_2, and causes acidification (Millero, 1995; Brewer and Peltzer, 2009). Drivers act with either additive, synergistic (i.e., amplification of) or antagonistic (i.e., diminution of) effects. A meta-analysis of 171 experimental studies that exposed marine systems to two or more drivers identified cumulative effects that were additive (26%), synergistic (36%), or antagonistic (38%) (Crain et al., 2008). Effects range from direct impacts of ocean warming on organismal physiology (Pörtner and Knust, 2007) to ocean acidification acting together with warming, for example, on coccolithophore calcite production and abundances (Feng et al., 2009), or with hypoxia and/or salinity changes (Table 6-4). Interactions of predominantly temperature, ocean acidification, and hypoxia have *likely* been involved in climate-driven evolutionary crises during Earth history (Pörtner et al., 2005; see also Section 6.1.2).

Effects on individual organisms may also reflect intertwined impacts of ocean warming, acidification, and hypoxia, which may operate through interrelated functional principles (Pörtner, 2012). Such knowledge helps to reconcile apparently contrasting findings. For example, warming toward the thermal optimum (Figure 6-5a) stimulates resistance to OA; CO_2-induced disturbances of growth and calcification were reversed by concomitant warming (Findlay et al., 2010; Sheppard-Brennand et al., 2010; Walther et al., 2011). Warming to above optimum temperatures, however, constrains performance and exacerbates sensitivity to hypoxia and/or elevated CO_2 (Figure 6-5, e.g., via decreased calcification; Rodolfo-Metalpa et al., 2011). Both hypoxia and/or elevated CO_2 in turn enhance heat sensitivity, as seen for CO_2 in crustaceans (via decreased heat limits: Walther et al., 2009; Findlay et al., 2010), coral reef fishes (via reduced performance: Munday et al., 2009b), and corals (via decreased calcification and CO_2-enhanced bleaching: Reynaud et al., 2003; Anthony et al., 2008). This translates into a narrowing of the thermal niche (Walther et al., 2009; see also Figure 6-5), which will shrink biogeographic ranges, affect species interactions, and shift phenologies (Figure 6-7a). Hence, extreme warming and hypoxia exacerbate CO_2 effects and vice versa (*medium confidence*). Such principles need to be reconfirmed across organism taxa (Pörtner, 2012).

Differences in organism adaptation to a climate zone's characteristic temperatures, temperature variability, oxygen content, and ocean chemistry may shape vulnerability to climate change. In high polar species evolutionary cold adaptation enhances vulnerability to warming

Table 6-4 | Potential interactions between modes of anthropogenic forcing (environmental; foodwebs; harvesting) on different levels of biological organisation. These interactions, from simple to complex, are illustrated with examples from the published literature. Unknown denotes no published information is available for each of these categories. NA denotes not applicable for this category.

Biological organization studied at ecosystem level	Anthropogenic forcing			
	Single environmental driver	Multiple environmental drivers	Fishing/foodwebs	Fishing/climate change
Individuals	Lab experiments and field observations show that warming alters organismal physiology and thereby growth (Pörtner and Knust, 2007). 🟧🟦⬜	Shipboard manipulation experiment addressing interactive effects of temperature and CO_2 on coccolithophore calcification (Feng et al., 2009). 🟧⬜⬜	NA	Unknown
Population	Physiological effects of warming change population abundance *in situ* (Pörtner and Knust, 2007). 🟧🟦⬜ Lab cultures show how altered pH elicits different responses of coccolithophore species (Langer et al., 2006). 🟧⬜⬜	Lab cultures show differential responses of cyanobacterial groups to temperature and CO_2 (Fu et al., 2007). 🟧⬜⬜	Altered maturation age and growth rate of populations due to fishing (Fairweather et al., 2006; Hseih et al., 2006). ⬜🟦⬜	Interactive effects on cod populations of fishing and alteration of salinity (Lindegren et al., 2010). ⬜⬜⬛
Ecosystem	Mesocosm experiments simulating the effect of individual drivers (e.g., ocean acidification effects on benthos: Christen et al., 2013; and on pelagic communities: Riebesell et al., 2013). 🟧🟦⬛	Mesocosm experiments studying differential effects of light and temperature, on copepods versus diatoms (Lewandowska and Sommer, 2010). 🟧⬜⬜	Effects of fishing on ecosystem structure — trophic cascades (Frank et al., 2005). ⬜🟦⬜	Interplay of fishing and climate pressures on ecosystems promotes lower trophic levels (Kirby et al., 2009); ⬜🟦⬜ enhances diversity loss in benthic communities (Griffith et al., 2011). ⬜⬜⬛
Biome	Time series observations on warming and geographical shifts of zooplankton biomes (Beaugrand et al., 2009). ⬜🟦⬜	Unknown	Unknown	Unknown

Approaches: 🟧 = Experiments (lab or field) 🟦 = Observations ⬛ = Modeling ⬜ = Not applied (*medium confidence*). In OMZs, marine sediments, and in polar waters (due to high solubility in the cold), CO_2 levels are elevated and adaptation may reduce sensitivity and reliance on calcified structures (Clark et al., 2009; Walther et al., 2011; Maas et al., 2012). The observed shift from "overcalcified" to "weakly calcified" coccolithophores *Emiliania huxleyi* in cold waters may reflect a related shift in ecotype dominance (*limited evidence*, *medium agreement*; Cubillos et al., 2007).

Despite such potential adaptation, polar calcifiers exposed to higher CO_2 and lower carbonate saturation levels have been hypothesized to be highly sensitive to further CO_2 accumulation (*limited evidence*, *high agreement*; Orr et al., 2005). Here it appears relevant that cold temperature reduces energy demand and thereby lowers resistance to ocean acidification. Both energy demand and resistance are higher in eurytherms than in high polar and deep sea stenotherms (*limited evidence*, *medium agreement*; Pörtner, 2006; e.g., crustaceans: Pane and Barry, 2007: cf. Whiteley, 2011). In turn, tropical species may be more sensitive than temperate zone species (Pörtner et al., 2011). This rough differentiation of sensitivity is complicated by the local adaptation of populations from within-species genetic variability (*low confidence*).

Temperature influences hypoxia sensitivity (Section 6.3.3). Warming causes the minimum tolerated O_2 level to rise, enhancing vulnerability (*high confidence*). Conversely, hypoxia enhances vulnerability to warming in animals. This may occur fastest in warm oceans, where metabolic rates are higher and animals live closer to upper thermal limits (*medium confidence*; Pörtner, 2010). However, evolutionary adaptation has led to high hypoxia tolerance (low P_c or O_2crit values) in some warm-adapted coral reef fishes. Further warming then causes a rise in P_c which cannot be compensated for (Nilsson et al., 2010). Limits to hypoxia adaptation coincide with upper thermal limits (*medium confidence*).

Complexity in responses rises with the number of drivers involved. Enhanced river runoff and increased precipitation cause a shift from marine to more brackish and even freshwater communities, with unclear consequences for effects of other drivers. Falling primary production reduces resilience of higher trophic levels (Kirby and Beaugrand, 2009; Stock et al., 2011). The introduction of non-indigenous species, when supported by climate-induced shifts in interactions, may promote the displacement of ecotypes and shifts in ecosystem functioning, for example, in the Mediterranean Sea (Occhipinti-Ambrogi, 2007; Coll et al., 2010).

6.3.5.2. Microbes

Both synergistic and antagonistic effects of multiple drivers on microbial biota in the surface ocean have been observed in manipulation or modeling experiments (Folt et al., 1999; Boyd et al., 2010; Gruber, 2011). The productivity of many microbes was simultaneously limited by, for example, availability of nitrate and phosphate, cobalt and iron (Saito et al., 2002; Bertrand et al., 2007), or iron and light (Boyd et al., 2010; see also Section 6.2.2). Warming and high CO_2 synergistically enhanced photo-physiological rates of the cyanobacterium *Synechococcus*, whereas the cyanobacterial group *Prochlorococcus* showed no change (Fu et al., 2007). The magnitude of CO_2 effects on growth, fixation rates, or elemental ratios within single species is often strongly modulated by nutrient availability and light conditions (e.g., Sciandra et al., 2003; Zondervan et al., 2002; Kranz et al., 2010). Such differences cause floristic shifts in phytoplankton with the potential to restructure predator-prey interactions (Table 6-4).

Co-limiting factors vary by group, such as nitrogen fixers (e.g., Hutchins et al., 2007; Kranz et al., 2010), diatoms (Boyd et al., 2010), and coccolithophores (e.g., Feng et al., 2009; Rokitta and Rost, 2012). This limits the ability to project climate change effects (Boyd et al., 2010). The most reliable projections at ocean basin scale come from modeling, which mainly points to synergistic effects, such as those of elevated CO_2, hypoxia, and warming. For example, OA is projected to alter sinking particles (C:N ratio and/or reduced calcite content and slower sinking) with a consequent knock-on effect on water column O_2 demand already stimulated by warming, thereby causing expansion of OMZs (Gruber, 2011).

6.3.5.3. Animals and Plants

High oxygen availability alleviates thermal stress as seen in fish and mollusks (Mark et al., 2002; Pörtner et al., 2006). Conversely, hypoxia reduces heat tolerance (Section 6.3.5.1), but acclimation to hypoxia compensates for this and increases thermal tolerance (Burleson and Silva, 2011), for example, by enhancing blood pigment content or reducing energy demand. Tolerances to hypoxia and to high temperature may positively correlate in some fishes, indicating potential for adaptive evolution under climate change (*low confidence*; McBryan et al., 2013).

As a consequence of hypoxia narrowing thermal ranges (Section 6.3.5.1), combined warming and expanding hypoxia may cause mid-water mesopelagic and demersal fish stocks to decline at rates much quicker than anticipated in the California Current Ecosystem (McClatchie et al., 2010; Koslow et al., 2011). In benthic fauna, warming will also increase vulnerability to hypoxia. Experiments showed a rise in lethal oxygen concentrations by 25% and thereby reducing survival by 36% at 4°C warmer temperatures (Vaquer-Sunyer and Duarte, 2011). Hence, warming is expected to expand the area of ecosystems affected by hypoxia even if oxygen concentrations remain unchanged (*high confidence*). Under combined hypoxia and warming, CO_2 can extend short-term passive tolerance (despite constraining long-term tolerance). It facilitates a reduction in energy demand (Reipschläger et al., 1997; Pörtner et al., 2000), thereby extending survival of transient extremes of temperatures or hypoxia (*medium confidence*).

In macroalgae (non-calcifying) light availability modulates the response to elevated pCO_2 and temperature levels (Russell et al., 2011; Sarker et al., 2013). In warm-water corals, warming acting synergistically with CO_2 reduces calcification and increases sensitivity to bleaching (*high confidence*; Anthony et al., 2008). Combined warming and OA following SRES B1 (≈RCP4.5, reduced emission) and A1FI (≈RCP8.5, business-as-usual) scenarios in mesocosms caused losses of symbionts and corals, and a nocturnal decalcification of the reef community in summer. Present-day conditions already imply reduced resilience to episodic extreme events such as cyclones (Dove et al., 2013; see also Box CC-CR).

6.3.5.4. Ecosystems

The cumulative impacts of climate change drivers underlie alterations of species interactions and ecosystem structure and functioning, including changes in trophodynamics and the physical and chemical characteristics of habitats (*high confidence*). These effects combine with more indirect effects, such as shifts in stratification and productivity, expanding oxygen minimum zones, and the changing composition and biomass of food (partly resulting from direct effects on prey organisms) (*high confidence*). These complexities reduce the precision and reliability of quantitative projections (Section 6.5), including uncertainties concerning shifts in upwelling and their future role in global primary production and the development of fish stocks (Box CC-UP).

At the level of animal communities, effects of various drivers remain largely unexplored, some are highly complex. For example, the net eastward shift of Pacific skipjack tuna between 1980 and 2009 was linked to the shifting aggregation of macrozooplankton and micronekton, involving complex interactions of climate variability (due to ENSO; Section 30.5.2), warming ocean surface, shallowing mixed layer depth relative to the position of the warm pool, and the convergence of the pool with the Pacific Equatorial Divergence Province (Lehodey et al., 2011; see also Section 30.5.6.1.1). Interactive drivers will affect the relative performance of interacting species, thereby shifting species ranges, interactions, and food webs (*medium confidence*; Figure 6-7a). Adaptation to various climate zones modifies the roles of light and temperature in seasonalities and species interactions (Bradshaw and Holzapfel, 2010). Moderate hypoxia expansion in warming seas, for example, as the stratified central North Sea (Queste et al., 2013) may well influence the degree of temperature-induced species displacements (Figure 6-7b).

Impacts of climate change on benthic ecosystem engineers can also profoundly alter ecosystems. Tropical corals respond to ocean warming and acidification by increased bleaching, impeded calcification rates, and increased incidence of disease (*high confidence*; Veron et al., 2009; Veron, 2011; see also Sections 6.3.1-2, 30.5.6; Box CC-CR). In coral reefs under multiple stressors, differentiation of these large-scale phenomena into species-specific sensitivities is highly uncertain as trend data are virtually nonexistent (Brainard et al., 2011). Little is known about impacts on deep-water or cold-water corals and sponges, tropical calcified algae, bryozoans, sponges, and tube-forming serpulid worms (Wood, 1999). The reliance of all of these on surface productivity makes them vulnerable to any alteration in food supply. Projected severe stress from increased temperature, hypoxia, and ocean acidification will cause reduced performance and increasing mortality in ecosystem engineers (*high confidence*), and a deterioration of habitat characteristics for other organisms (*medium* to *low confidence*).

As a corollary, shifts in the geographical distributions of marine species (e.g., to higher latitudes or deeper waters; Figure 6-7b; Section 6.5.2) cause changes in community composition and interactions (Harley, 2011; Simpson et al., 2011; Hazen et al., 2013). Some species may gain predominance and abundance from fitness benefits (Figure 6-7) while others become less competitive or easier prey (Occhipinti-Ambrogi, 2007). Thereby, climate change will reassemble communities and affect biodiversity, with differences over time and between biomes and latitudes (*high confidence*; Parmesan and Matthews, 2005; Sala and Knowlton, 2006; Cheung et al., 2009; Parmesan et al., 2011; see also Box CC-PP; Section 6.5).

6.3.6. Food Web Consequences

Community reassembly under climate change involves a change in species composition and strongly alters food web structure, for example, causing shifts in trophic pathways (Kirby and Beaugrand, 2009; Moloney et al., 2011; see also Figure 6-12), some of which are irreversible (Jarre and Shannon, 2010). Through trophic cascades (Cury et al., 2003; Luczak et al., 2011), climate affects predation, competition, and food availability (e.g., via changes in NPP; Figure 6-12; Utne-Palm et al., 2010), including fish stocks (Parsons and Lear, 2001; Brown et al., 2010). Trophic amplification then drives an ecosystem towards a new stable structure or regime, which may be difficult to reverse (Folke et al., 2004). Warming may result in consumer control of food web structure because respiration of heterotrophic zooplankton and bacteria increases more strongly with warming than does photosynthesis of autotrophic phytoplankton (*medium confidence*; O'Connor et al., 2009).

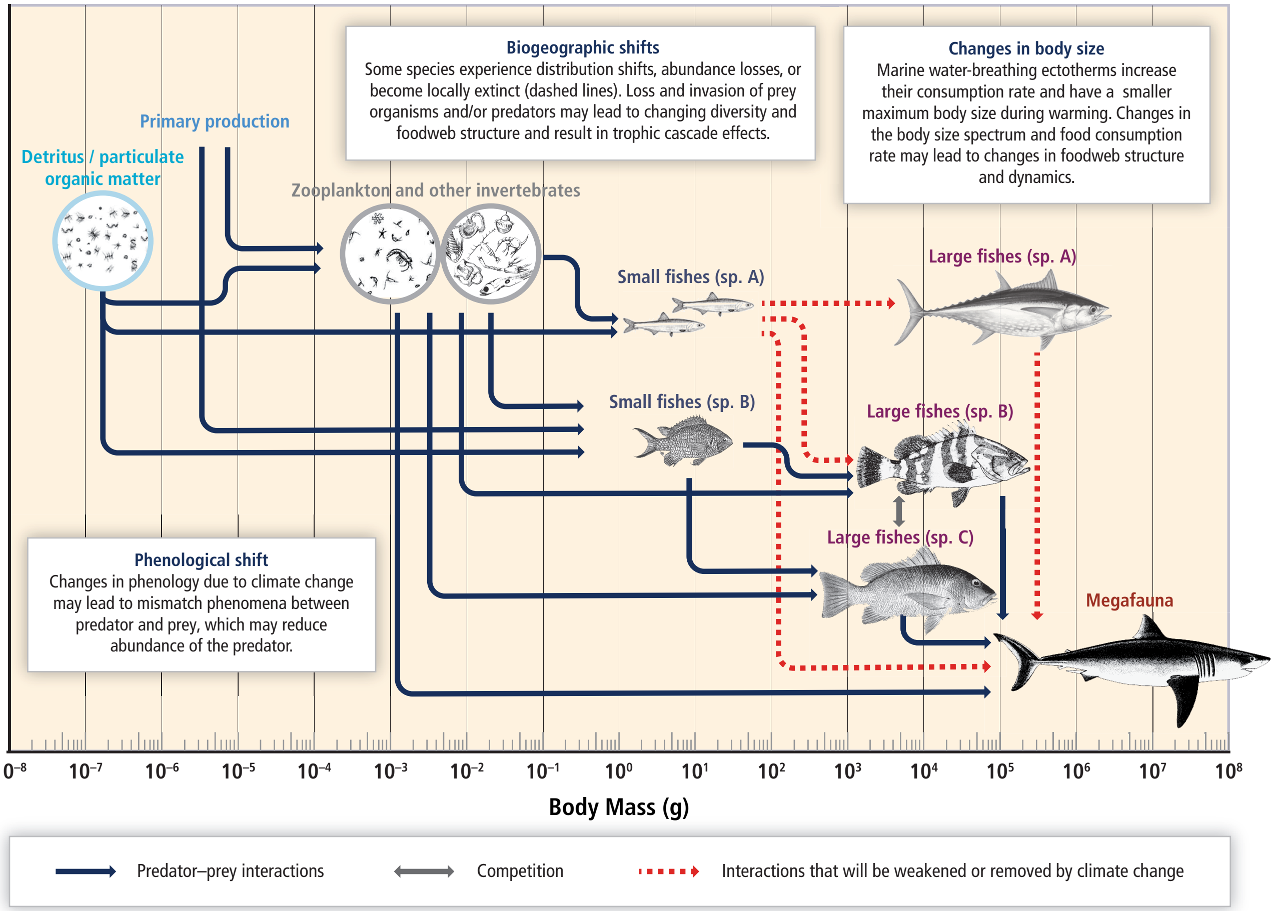
Many impacts of climate change on food webs resemble those caused by fishing, pollution, eutrophication, and associated hypoxia (Section 6.3.3), and habitat change (Brander, 2007); unambiguous attribution to climate remains difficult (*low* to *medium confidence*; Parmesan et al., 2011). Some of these factors also affect food web responses to climate change. Fishing truncates the age and size structure of populations, making them more dependent on annual recruitment and reducing their ability to buffer environmental fluctuations (Genner et al., 2010; Planque et al., 2010; Botsford et al., 2011; see also Figure 6-12). Both adult and larval fishes show greater variability in abundance in exploited compared to unexploited populations (Hsieh et al., 2008). Warming, acidification, and removal of top or competing predators may all contribute to large fluctuations in gelatinous plankton (e.g., jellyfish) populations (*low confidence*; Molinero et al., 2005; Richardson and Gibbons, 2008; Richardson et al., 2009; Condon et al., 2012).

Analyzing impacts on key species provides insight into how individual components of a food web will respond to perturbations. However, projections of future states must include the complex food web interactions that influence the species and system-level responses, which affect stability and resilience of the overall ecosystem (Neutel et al., 2007; Dunne and Williams, 2009; Romanuk et al., 2009). There is no single approach currently available that includes the complex links within and among ecosystems, biogeochemistry, and climate as needed for projections of future states of marine food webs (Fulton, 2011; Moloney et al., 2011). In conclusion, there is *low confidence* in the quantitative projections of such changes (for further discussion see Section 6.5).

6.3.7. Marine Reptiles, Mammals, and Birds

6.3.7.1. Principles

Marine reptiles (turtles, snakes, crocodiles), mammals, and seabirds breathe air but live mostly in water; some shift or expand their ranges



Biogeographic shifts
Some species experience distribution shifts, abundance losses, or become locally extinct (dashed lines). Loss and invasion of prey organisms and/or predators may lead to changing diversity and foodweb structure and result in trophic cascade effects.

Changes in body size
Marine water-breathing ectotherms increase their consumption rate and have a smaller maximum body size during warming. Changes in the body size spectrum and food consumption rate may lead to changes in foodweb structure and dynamics.

Phenological shift
Changes in phenology due to climate change may lead to mismatch phenomena between predator and prey, which may reduce abundance of the predator.

Figure 6-12 | Schematic diagram of expected responses to climate change in a marine food web. A coupled pelagic and benthic food web is structured by the body size spectrum of species. Combined warming, hypoxia, and ocean acidification reduce body size, shift biogeographies, change species composition and abundance, and reconfigure trophic linkages and interaction dynamics. Fishing generally removes large-bodied species and truncates the body-size spectrum of the community. This confounds the detection and attribution of food web responses to climate change. Arrows represent species interactions (e.g., between predator and prey or competitors for food or space). Broken lines reflect the potential loss of populations and trophic linkages due to climate change.

as a result of climate warming. The body temperature of ectothermic reptiles is set by ambient conditions; only at large body size may their body store heat and its temperature be higher than ambient. Reptiles are thus more responsive to temperature than homeothermic seabirds and marine mammals (McMahon and Hays, 2006), which regulate their body temperature by adjusting metabolic heat production and insulation from the environment, a trait beneficial especially in the cold. Various degrees of body core insulation in mammals and birds constrain their distribution to either warmer or colder waters (by poor or high insulation, respectively). However, large body sizes enable some aquatic air breathers to travel across the widest temperature ranges possible in some of the largest migrations on Earth.

Changes in water chemistry and hypoxia have minimal direct influences on the air-breathing vertebrates, reflecting their large independence from physical and chemical drivers in the oceans. There is evidence for increased sound propagation in a CO_2-enriched ocean, but no evidence yet for any effect on biota (Ilyina et al., 2010). If habitat structures offering retreat or ambush disappear, this will increase the energetic costs of life. Warming waters increase the cost of pursuit-diving as prey fishes increase swimming velocity. The predation success of such mammals (e.g., sea lions) and seabirds (e.g., penguins, cormorants) is thus constrained to waters ≤20°C (Cairns et al., 2008), a trend that extrapolates into the future (*low* to *medium confidence*). As prey distributions shift, foragers tied to land between trips may be constrained by the physiological costs of finding prey (Péron et al., 2012; Hazen et al., 2013). If food items are only found in thermally restricted areas or move to greater depths, mammals and birds may become constrained to certain distribution ranges or to the physiological limits of their diving ability (McIntyre et al., 2011). Conversely, hypoxic habitat compression for fishes may facilitate foraging opportunities for their air-breathing predators (Hazen et al., 2009). Accordingly, many air-breathers encounter changing habitat and food availability with climate change (*high confidence*).

6.3.7.2. Field Observations

Some species of seabirds, marine mammals, and sea turtles have responded to the anomalous ocean climate of the 20th century (*high confidence*; Hughes, 2000). There is insufficient information to assess effects on sea snakes or crocodiles. Poleward distribution shifts of turtles consistent with recent warming have been recorded in almost all marine groups. Decadal-scale climate fluctuations affect their recruitment success and nesting abundance (Van Houtan and Halley, 2011), with an inverse correlation between warming and abundance in various species and regions (Balazs and Chaloupka, 2004; Chaloupka et al., 2008; Mazaris et al., 2009). Extreme weather causes nest flooding, considerably reducing hatching success (Van Houtan and Bass, 2007); projected sea level rise (WGI AR5 Chapter 13) will exacerbate such impact. Those with high fidelity to nesting and foraging sites (Cuevas et al., 2008) are impacted more than those capable of changing those sites (Fish et al., 2009; Hawkes et al., 2009). Continued warming, modulated by changing rainfall (Santidrián Tomillo et al., 2012), may skew turtle sex ratios toward females, increase egg and hatchling mortality (Fuentes et al., 2009), cause earlier onset of nesting (Pike et al., 2006; Mazaris et al., 2008), decrease nesting populations (Chaloupka et al., 2008), and shift dietary breadths (Hawkes et al., 2009), leading to projected recruitment declines (e.g., leatherback turtles; Saba et al., 2012). Vulnerability due to shifting sex ratio alone remains unclear, as nesting beaches have persisted with low production of male hatchlings over decades or longer (*low confidence*; Godfrey et al., 1999; Broderick et al., 2000; Hays et al., 2003). The absence of sea turtles in certain regions may be best explained by the temporal unavailability of food resources or strong thermoclines restricting their bottom foraging abilities (Braun-McNeill et al., 2008; Gardner et al., 2008).

Seabird range modifications probably caused by climate change were recorded in polar areas and the temperate zone of the North Atlantic (Grémillet and Boulinier, 2009). Temperate species have shifted their ranges to higher latitudes in both hemispheres (Bunce et al., 2002; Robinson et al., 2005; La Sorte and Jetz, 2010). Some species, like the king penguin, follow shifting foraging zones (Péron et al., 2012); others, such as the emperor penguin, are affected by changing habitat structure (sea ice; Jenouvrier et al., 2012). Warming causes many bird species to breed earlier (Sydeman and Bograd, 2009). High-latitude, cool-water species undergo extended breeding seasons (Chambers et al., 2011). There is often no agreement, whether changes reflect solely ocean warming, or a combination of factors, such as fishing pressure on seabirds' prey species, sea level rise, and pollution (Galbraith et al., 2005; Votier et al., 2005; Heath et al., 2009). Most shifts in range and seasonal activity involve shifts in trophic relationships (*medium confidence*). Seabirds with narrow geographic domains are expected to be more susceptible to climate change (Chambers et al., 2005; Grémillet and Boulinier, 2009), even leading to local extinctions (e.g., the Galápagos penguin: Vargas et al., 2007; or the marbled murrelet: Becker et al., 2007).

The distribution, phenology, and migratory timing of marine mammals are also shaped by predator-prey dynamics and climate impacts on specific habitats (Calambokidis et al., 2009; Salvadeo et al., 2011). Some marine mammals, that is, dolphin, porpoise, and whale species, shift their distribution poleward to follow the movement of their prey (*medium confidence*; Springer et al., 1999; MacLeod et al., 2005; Simmonds and Isaac, 2007; Salvadeo et al., 2010). As in birds, vulnerability to climate change is high for marine mammals with narrow geographic ranges and high habitat dependence. For example, the critically endangered vaquita, endemic to the Northern Gulf of California, cannot move north because of the land barrier (MacLeod, 2009). The polar bear (Laidre et al., 2008; Rode et al., 2012) and the walrus depend on sea ice as a platform for hunting, resting, and giving birth. For polar bears, access to prey such as ringed seals has been disrupted by the later formation and earlier breakup of sea ice in the eastern Canadian Arctic. Seasonal migrants into the Arctic (fin, minke, gray, killer, humpback whales) may increasingly compete with species adapted to operate in habitat with sea ice (some seals, narwhal, bowhead whale, beluga). Both may benefit from the net loss of sea ice, which will offer them better access to foraging in a pelagic-dominated ecosystem (Moore and Huntington, 2008).

6.3.8. Summary and Conclusions

An organism's capacity to perform, but also its access to food energy fueling that performance, shape its sensitivity to climate change (*high confidence*). Extreme temperatures surpassing the fringes of the thermal envelope cause local abundance losses, extinction, and shifts in temperature-dependent distribution ranges (*high confidence*; Section 6.3.1).

Some climate change effects detected in the field can be attributed to temperature, but few allow clear attribution to other drivers (Sections 6.3.1-5, 6.6). In fishes and invertebrates, specialization in regional climate regimes co-defines sensitivity to warming, acidification, and hypoxia (*high confidence*; Section 6.3.5). In marine mammals, birds, and ectothermic reptiles, changes in life history and population dynamics have often not been directly attributed to climate drivers (*low confidence*), but rather to the availability of habitat and food (*high confidence*; Section 6.3.7).

Natural climatic variability (Figure 6-1) and anthropogenic change, with a strong role of warming, cause large-scale changes in biogeography, abundance, diversity, community composition, and structure of marine species (*very high confidence*; Section 6.3.1). Warming reduces body size (*medium confidence*; Section 6.3.1). Differential species responses modify their interactions across trophic levels through trophic amplification (*medium* to *high confidence*; Section 6.3.6).

Some tropical species and ecosystems exist close to upper thermal limits placing them among the marine ecosystems most affected by climate change (*high confidence*; Section 6.3.1). Corals and coral reefs are primary examples. However, other factors change concomitantly, such that quantifying ecosystem changes attributable to warming or other drivers has not always been possible (Section 6.3.5).

Under future climate change ocean acidification will affect marine organisms and ecosystems for centuries (*high confidence*; Sections 6.3.2, 6.3.5). To date, very few ecosystem-level changes in the field have been attributed to anthropogenic or local ocean acidification (*medium confidence*; Section 6.3.2). Concomitant trends of warming, O_2 depletion, OA, and other drivers prevent clear attribution to OA (Section 6.3.5).

Elevated CO_2 levels stimulate primary production of some macroalgae and seagrass species (*high confidence*), causing them to be more competitive than calcifying organisms (*medium confidence*; Section 6.3.2). High sensitivities to OA are associated with low capacities to maintain pH in internal fluids (*high confidence*). Calcification rates in sensitive invertebrates, including corals, echinoderms, and mollusks, decrease under OA, especially if combined with temperature extremes

Frequently Asked Questions

FAQ 6.4 | What changes in marine ecosystems are likely because of climate change?

There is general consensus among scientists that climate change significantly affects marine ecosystems and may have profound impacts on future ocean biodiversity. Recent changes in the distribution of species as well as species richness within some marine communities and the structure of those communities have been attributed to ocean warming. Projected changes in physical and biogeochemical drivers such as temperature, CO_2 content and acidification, oxygen levels, the availability of nutrients, and the amount of ocean covered by ice will affect marine life.

Overall, climate change will lead to large-scale shifts in the patterns of marine productivity, biodiversity, community composition, and ecosystem structure. Regional extinction of species that are sensitive to climate change will lead to a decrease in species richness. In particular, the impacts of climate change on vulnerable organisms such as warm-water corals are expected to affect associated ecosystems, such as coral reef communities.

Ocean primary production of the phytoplankton at the base of the marine food chain is expected to change but the global patterns of these changes are difficult to project. Existing projections suggest an increase in primary production at high latitudes such as the Arctic and Southern Oceans (because the amount of sunlight available for photosynthesis of phytoplankton goes up as the amount of water covered by ice decreases). Decreases are projected for ocean primary production in the tropics and at mid-latitudes because of reduced nutrient supply. Alteration of the biology, distribution, and seasonal activity of marine organisms will disturb food web interactions such as the grazing of copepods (tiny crustaceans) on planktonic algae, another important foundational level of the marine food chain. Increasing temperature, nutrient fluctuations, and human-induced eutrophication may support the development of harmful algal blooms in coastal areas. Similar effects are expected in upwelling areas where wind and currents bring colder and nutrient-rich water to the surface. Climate change may also cause shifts in the distribution and abundance of pathogens such as those that cause cholera.

Most climate change scenarios foresee a shift or expansion of the ranges of many species of plankton, fish, and invertebrates toward higher latitudes, by tens of kilometers per decade, contributing to changes in species richness and altered community composition. Organisms less likely to shift to higher latitudes because they are more tolerant of the direct effects of climate change or less mobile may also be affected because climate change will alter the existing food webs on which they depend.

In polar areas, populations of species of invertebrates and fish adapted to colder waters may decline as they have no place to go. Some of those species may face local extinction. Some species in semi-enclosed seas such as the Wadden Sea and the Mediterranean Sea also face higher risk of local extinction because land boundaries around those bodies of water will make it difficult for those species to move laterally to escape waters that may be too warm.

(*high confidence*; Section 6.3.5). Thresholds beyond which effects occur can be quantified only with *low confidence*; there are differential sensitivities and thresholds between taxa and species (*high confidence*; Section 6.3.2).

Expansion of oxygen minimum zones leads to community shifts clearly attributable to extreme hypoxia (*high confidence*; Section 6.3.3). Gradual effects of a progressive decline in ocean O_2 levels on communities have not been sufficiently explored.

In general, community reassembly with new species coming in will occur in the transition to future climates (*medium confidence*) and lead to new ecosystem states (*low confidence*; Section 6.3.6). Climate change interacts with top-down human interferences, such as fisheries or other forms of harvesting, which accelerate impacts (*medium confidence*). Nonlinearities challenge the projection of marine ecosystem trajectories (FAQ 6.4).

In microbes, a conceptual foundation suitable to support an integrated understanding of climate impacts on individual species and communities is lacking. Specific physiological responses, such as in primary production, N_2 fixation, or calcification, can be attributed to multiple environmental drivers associated with climate change (*high confidence*; Sections 6.3.1-5).

6.4. Human Activities in Marine Ecosystems: Adaptation Benefits and Threats

Human societies benefit from resources and processes supplied by marine ecosystems, so-called ecosystem services. Attributing and projecting ecosystem changes and their effects on human communities caused by climate change including ocean acidification is challenging. Insufficient observations compound an understanding of long-term changes and the definition of baseline conditions. Some of the challenges are related to the difficulty of projecting how human communities will adapt to changing marine ecosystem benefits.

6.4.1. Ecosystem Services

Marine ecosystem services (e.g., Chapter 5) include products (food, fuel, biochemical resources), climate regulation and biogeochemical processes (CO_2 uptake, carbon storage, microbial water purification), coastal protection, provision of space and waterways for maritime transport, cultural services (recreational and spiritual opportunities, aesthetic enjoyment), and functions supporting all other ecosystem services (nutrient cycling, photosynthesis, habitat creation). Most components of the marine environment contribute to more than one major category of ecosystem service: for example, ocean primary productivity is classified as a supporting service, but it affects provisioning services via changes in fisheries, generation of fossil fuel resources, regulating services via the global carbon cycle and climate regulation, and cultural services via the enjoyment of a healthy ecosystem. Rarely has economic damage of climate change to a whole ecosystem been evaluated and projected. The projected loss of tropical reef cover due to ocean acidification under SRES A1 and B2 scenarios will cause damages of US$870 and 528 billion (year 2000 value) by 2100, respectively (cost rising with parallel economic growth; Brander et al., 2012; see also Box CC-OA). Such loss is felt most strongly in the respective regions.

6.4.1.1. Food from the Sea

Fisheries provide 3 billion people with almost 20% of their average per capita intake of animal protein (FAO, 2012a), 400 million depend critically on fish for their food (Garcia and Rosenberg, 2010). Total world marine capture fisheries catches stabilized in the mid-1990s at about 90 million tons per year. Marine aquaculture of primarily mollusks and crustaceans contributes more than 63 million tons annually to seafood production, mostly concentrated in coastal areas (FAO, 2012b). The growth of aquaculture has decelerated, but is still considered a development opportunity and a strong need in regions such as Africa and Latin America (Section 7.4.2.2).

Climate-induced shifts in ecosystems and fisheries production will create significant challenges to sustainability and management (Section 7.5.1.1.3), particularly for countries with fewer resources and lower adaptive capacity, including many low-latitude and small island nations (*high confidence*; Allison et al., 2009; Worm et al., 2009; Cooley et al., 2012; see also Sections 7.2.1.2, 7.4.2.1, 30.6.2; WGIII AR5 Section 2.1). Vulnerability will be exacerbated by increases in the frequency and severity of extreme events (e.g., floods or storms) damaging infrastructure, homes, health, livelihoods, or non-marine food security (Kovats et al., 2003; Rosegrant and Cline, 2003; Adger et al., 2005; Haines et al., 2006).

The projected trends in fish stocks will widen the disparity in food security between developing and developed nations. Fish migrations due to warming (Section 6.3.1) have already shifted the composition of fisheries catches (Pinsky and Fogarty, 2012; Cheung et al., 2013a) and altered stock distributions (Sabatés et al., 2006). Further warming may be beneficial for fisheries productivity in some regions such as the North Atlantic, because of the poleward shift of exploited species and changes in primary productivity (Arnason, 2007; Stenevik and Sundby, 2007; Cheung et al., 2010; see also Box 6-1; Section 30.5.1.1.1), or for some Pacific Islands due to the eastward redistribution of tuna stocks (Lehodey, 2000; Lehodey et al., 2011). Resulting changes in accessibility and fishing operations costs are projected to straddle economic zones, perturb international fishery agreements, and cause excessive exploitation (Hannesson, 2007; Sumaila et al., 2011; see also Sections 7.3.2.4, 7.4.2; WGIII AR5 Section 4.3.7).

Invertebrate fisheries and aquaculture appear very vulnerable to the impacts of ocean acidification (Barton et al., 2012; see also Box CC-OA; Figure 6-10). This concerns especially shelled mollusks, with a substantial decline in their global production projected between 2020 and 2060 under the SRES A2 business-as-usual scenario (Cooley and Doney, 2009; Cooley et al., 2012). Effects on calcifying plankton will propagate through the food web, making estimates of economic impact on fish catch by OA difficult, also due to complex interactions with other stressors like warming and fisheries management (Griffith et al., 2012; Branch et al., 2013). Model projections suggest a potential loss of up to 13% (SRES A1FI scenario) to annual total fishery value in the USA, or globally more than US$100 billion annually by 2100 (Cooley and Doney, 2009; Narita et al., 2012). Vulnerability differs highly between nations according to the contribution of such fisheries to their economy (Cooley et al., 2012; see also Sections 7.3.2.4, 7.4.2). These projections are sensitive to the projected vulnerabilities of the organisms to ocean acidification (*medium confidence*; Section 6.3.2).

Fishing reduces abundances at high trophic levels, but increases abundances at mid-trophic levels. It reduces species numbers, simplifies ecosystem structure, and increases ecosystem sensitivity to climate change (Perry et al., 2010). Exploitation of fish stocks and the alteration of their demography, population dynamics, and life history traits (Petitgas et al., 2006; Perry et al., 2010; Planque et al., 2010) can reduce the capacity of fish populations to buffer changes in climate variability (Ottersen et al., 2006; Genner et al., 2010), and increase variability in population size. Interactions between warming, OA, and human activities such as fishing may thus exacerbate climate impacts on a wide range of ocean processes and services, including marine fisheries (*medium confidence*; Tables 6-4, 6-6; Section 30.6.2).

A 2°C global temperature increase by 2050 is estimated to cause global losses in landed value of US$17 to 41 billion annually (in 2005 value), with an estimated cost of adaptation for the fisheries of US$7 to 30 billion annually over a 40-year time frame between 2010 and 2050. The largest loss in landed value is projected to occur in East Asia and the Pacific (*low confidence*; Sumaila and Cheung, 2010). Overall impacts and the regional manifestations will partially depend on the flexibility and response capacities of food production systems (Elmqvist et al., 2003; Planque et al., 2011a).

Specific implications for the fishing industry are still poorly known, as future projections of shifts in primary production and knock-on effects through food webs and into fisheries remain uncertain (*low confidence* in effects of changing NPP; Planque et al., 2011b; Stock et al., 2011).

6.4.1.2. Other Provisioning Services

Reductions in marine biodiversity due to climate change and other anthropogenic stressors (Tittensor et al., 2010), such as OA (CBD, 2009) and pollution, might reduce the discovery of genetic resources from marine species useful in pharmaceutical, aquaculture, agriculture, and other industries (Arrieta et al., 2010), leading to a loss of option value from marine ecosystems. Climate change increases the demand for marine renewable energy such as wind and wave power, though with potential ecosystem impacts of their infrastructure (Section 6.4.2).

6.4.1.3. Climate Regulation and Extreme Events

The effect of climate change on marine biota will alter their contribution to climate regulation, that is, the maintenance of the chemical composition and physical processes in the atmosphere and oceans (*high confidence*; Beaumont et al., 2007). Regulatory mechanisms in which organisms (especially phytoplankton) play a key role, include control of the level of atmospheric CO_2 through the balance between photosynthesis and respiration (Johnson et al., 2010), and through the biological and alkalinity pump (Falkowski, 1997; Feely et al., 2008). They also include the modulation of further greenhouse gases such as nitrous oxide (N_2O; Jin and Gruber, 2003; Law, 2008; see also Section 6.1.1.3), and the modulation of other climatically reactive gases such as dimethylsulfide (DMS; Vogt et al., 2008). A projected decrease in global ocean NPP (Section 6.5.1) may result in decreased export of biogenic carbon to the deep ocean (Bopp et al., 2002; Boyd and Doney, 2002; Hashioka and Yamanaka, 2007). A positive feedback on climate change may result; however, many of the factors controlling the pump are poorly understood (Figure 6-4; WGI AR5 Chapter 6).

Coastal marine ecosystems reduce the effects of floods and storm surges which account for most of the natural disasters affecting people in coastal regions (IPCC, 2012a). Empirical and modeling studies show that coral reefs contribute to buffering the impact of tsunamis (Fernando et al., 2005; Gravelle and Mimura, 2008; see also Sections 5.4.2.4, 30.5; Box CC-CR). Experiments and models indicate that warming and OA slow coral growth by nearly 50% by 2050 (Box CC-CR; Section 5.4.2.4), making some islands and coastal areas more vulnerable to tsunamis, storm surges, wave energy, and coastal erosion (*high confidence*). Wetlands and mangroves provide biologically diverse buffer zones (Section 5.4.2.3). The combined impacts of climate change, pollution, deoxygenation, and other overlapping stressors, on mangroves and wetlands have not been determined (Cooley et al., 2009; Cooley, 2012). Some of these stressors enhance each other's effects in coastal systems (Feely et al., 2010; Cai et al., 2011; Howarth et al., 2011).

6.4.1.4. Cultural Services

Cultural services encompass a wide array of services with marine biodiversity as a core component supporting recreation and tourism as the economically most relevant. Tropical coral reefs and their enormous biodiversity sustain substantial tourist industries, presently with global annual net benefits of about US$9.6 billion (Cesar et al., 2003; see also Box CC-CR; Section 30.6.2.2). If reef services degrade, coastal visitors might choose alternative attractions (UNWTO, 2008). Increased travel to see disappearing ecosystem types (e.g., Antarctica: Liggett et al., 2011) or in previously inhospitable areas or seasons (Amelung et al., 2007; Moore, 2010) create new pressures and are unsustainable as the locations of key attractors shift (e.g., cetaceans: Lambert et al., 2010; Salvadeo et al., 2013).

Climate change may endanger harvests of marine species with spiritual and aesthetic importance to indigenous cultures, raising ethical questions about cultural preservation (e.g., Nuttall, 1998). In coastal communities, losing the aesthetic values of marine ecosystems may harm local economies: better water quality and fewer harmful algal blooms are related to higher shellfish landings and real estate prices (Jin et al., 2008).

Some heritage benefits of preserving marine ecosystems consist of the economic value of a healthy, diverse ecosystem to future generations. Any climate-related biodiversity loss or pollution of marine ecosystems would decrease the bank of resources for future opportunities. For example, the research and conservation value of coral reef biodiversity and its non-use value are estimated together at US$5.5 billion annually (Cesar et al., 2003). As with spiritual and aesthetic benefits, maintaining heritage benefits under climate change poses challenges for managers concerning equity and ethics as well as multigenerational (and possibly multi-cultural) ethical questions.

6.4.1.5. Supporting Services

Fully identifying the services supporting other ecosystem benefits is virtually impossible, as they are diverse in nature and scale. Ecosystem engineers play an important role in these services. Damage to calcifying algae and corals will reduce habitat for other species (Section 6.3.5), biodiversity, cultural and leisure values, and their climate regulation capacity.

Waterways for shipping are expected to change in the next several decades (*very high confidence*; Chapter 28; Section 30.6.2.3). Reductions in Arctic sea ice allow new trade routes such as the Northwest Passage (Wilson et al., 2004; Granier et al., 2006), enabling economically viable trans-Arctic shipping, and access to regional resources for exploitation and tourism. This development would increase emission of greenhouse gases and other pollutants (Lauer et al., 2009; Corbett et al., 2010), and facilitate the invasion of non-indigenous species carried on hulls and in ballast waters (Lewis et al., 2004).

6.4.2. Management-Related Adaptations and Risks

6.4.2.1. Ecosystem Management

A changing climate will have both positive and negative consequences for managing ocean resources (*high confidence*) (Eide and Heen, 2002;

Eide, 2007; see also Section 6.4.1). Ecosystem-based management (EBM, an approach recognizing all, including human interactions, within an ecosystem) or the ecosystem approach (EA, a strategy for the integrated management of living resources promoting both conservation and sustainable use) are increasingly adopted globally (FAO, 2003) to deal with the multitude of human pressures on marine ecosystems (Sherman et al., 2005; Hoel, 2009). Extended EBM addresses changes driven by climate and human activities, considering that diverse drivers will interact and confound each other (Planque et al., 2010; Eero et al., 2011; see also Section 6.3.5). Human activities will undermine resilience to other, including climate, impacts or undermine the effectiveness of mitigation and adaptation measures, by increasing variability (thereby reducing predictability), and limiting scope for adaptation (*high confidence*; e.g., Hughes, 2004; Sissener and Bjørndal, 2005; Eero et al., 2011). Thus, managing ecosystems under climate change increases the resilience of ecosystems and adaptive capacity of management systems through reducing other human perturbations (e.g., overfishing) (Brander, 2008; see also Section 7.5.1.1.3). Managing ecosystems also reduces the consequences of ocean acidification until CO_2 emission reduction becomes effective (Rau et al., 2012; Billé et al., 2013; McLeod et al., 2013; see also Box CC-OA). Ecosystem resilience is enhanced by reducing regional eutrophication (Falkenberg et al., 2013), or in aquaculture by avoiding acidified water (Barton et al., 2012) and by selecting and cultivating pre-adapted strains (Parker et al., 2012).

However, effects of climate change cannot be reversed by reducing the impacts of non-climatic drivers, emphasizing the need for adaptive management. Increased variability of ecosystem responses to climate change and the low predictability of some biological responses undermine the effectiveness of management and conservation measures. A particular risk is that climate change may contribute to large-scale ecosystem regime shifts (Section 6.3.1.5; Box 6-1). Detecting and forecasting such shifts from time series of environmental and biological data (Carpenter and Brock, 2006; deYoung et al., 2008), is constrained by an insufficient number of observations and limited quantitative understanding (Section 6.1.2). Biogeographic shifts challenge spatial management (Box CC-MB; Sections 6.3.1, 6.5), which is a fundamental part of EBM (Douvere, 2008), and demand that "fixed in law forever" site-attached zoning to protect specific species may need to become more flexible to maintain the original objectives as species move or community structures shift (*high confidence*; Soto, 2001; Hawkins, 2012).

6.4.2.2. Geoengineering Approaches

Geoengineering approaches to mitigate climate change and its effects, include Solar Radiation Management (SRM) and Carbon Dioxide Removal (CDR; see Table 6-5; IPCC, 2012b). SRM aims to reduce warming by increasing albedo, for example, via stratospheric injection of sulfate aerosol (Crutzen, 2006). SRM may affect marine ecosystems through changes in precipitation. With continued CO_2 emissions it leaves ocean acidification largely unabated as it cannot mitigate rising atmospheric CO_2 concentrations (Vaughan and Lenton, 2011; Williamson and Turley, 2012). Termination of SRM after its implementation involves the risk of rapid climate change and more severe effects on ecosystems (Russell et al., 2012).

Proposed CDR techniques include both ocean- and land-based approaches (Vaughan and Lenton, 2011; see also Section 30.6.4). CO_2 storage in geological reservoirs may occur beneath the seafloor, for example, in porous marine aquifers, and includes the risk of CO_2 leakage to the marine environment. Proposals to directly or indirectly sequester CO_2 into the ocean (Caldeira et al., 2005; Boyd, 2008; Shepherd et al., 2009; see also Table 6-5; WGIII AR5 Section 7.5.5) include, among others, the use of ocean fertilization techniques by nutrient addition, the direct storage of biomass in the deep ocean, the addition of alkalinity for build-up of dissolved inorganic carbon (DIC; i.e., carbonate), and the direct CO_2 injection into the deep ocean (Williamson et al., 2012). All of these approaches have potentially negative consequences for marine ecosystems.

Ocean fertilization by adding iron to high-nutrient low-chlorophyll (HNLC) oceanic waters could increase productivity and the net export of organic material to the deep ocean and its consecutive decomposition, causing deep-water accumulation of CO_2. Fertilization would affect all major marine biogeochemical cycles of the ocean with unclear side effects that could include the formation of methane (CH_4) and N_2O (Law, 2008) or the stimulation of harmful algal blooms (Trick et al., 2010). The enhanced NPP would add more carbon to the base of food webs (de Baar et al., 2005) and stimulate growth, for example, of deep-sea benthos (Wolff et al., 2011). Any regional increase in organic material (through fertilization or intentional storage of biomass) would cause enhanced O_2 demand and deep-water O_2 depletion (Sarmiento et al., 2010; Table 6-5), increasing the level and extent of hypoxia and associated impacts on marine ecosystems (Sections 6.3.3, 6.3.5, 30.5.7). The synergistic effects of CO_2-induced acidification will exacerbate the biological impacts (*high confidence*).

Neutralizing the acidifying water by the addition of alkalinity, for example, calcium oxide, would require large-scale terrestrial mining with associated consequences (Caldeira et al., 2005). The biological effects of increased concentrations of Ca^{2+} ions and dissolved inorganic carbon remain insufficiently explored. Direct injection of CO_2 or its localized disposal in the ocean (e.g., as a lake in a deep-sea valley) causes locally highly increased CO_2 and acidification effects on deep-sea organisms (*high confidence*; Caldeira et al., 2005; see also Section 6.3.3.4). In contrast to long-term ocean fertilization or storage of biomass, this technique leaves the oxygen inventory of the deep ocean untouched (*limited evidence*, *medium agreement*; Pörtner et al., 2005).

The knowledge base on the implementation of SRM and CDR techniques and associated risks is presently insufficient. Comparative assessments suggest that the main ocean-related geoengineering approaches are very costly and have large environmental footprints (*high confidence*; Boyd, 2008; Vaughan and Lenton, 2011; Russell et al., 2012).

6.4.2.3. Health Issues

Human health and near-shore ecosystems may be directly impacted by climate change effects on harmful algal blooms (HABs; Edwards et al., 2006; see also Section 30.6.3) or disease vectors. Planktonic time-series archives and nearshore sediment cores containing HAB cysts have revealed few examples of strong linkages between altered HABs and

Table 6-5 | Challenges for the oceans that will arise from the employment of a range of geoengineering methods (SRM = solar radiation management; CDR = carbon dioxide removal).

Topic	Brief description	Challenge and impact	References
Solar radiation management techniques	Deflection of approximately 1.8% of sunlight, by various techniques, is able to offset the global mean temperature effects of a doubling of atmospheric CO_2 content from pre-industrial values.	Will leave ocean acidification unabated (*high confidence*). Response of primary production to light reduction unclear.	Crutzen (2006); Caldeira and Wood (2008)
Ocean storage by direct injection	Capture of CO_2 post-combustion from mainly coastal power plants, followed by injection of liquid CO_2 by pipeline or from a ship into the deep ocean.	Will add to ocean acidification and create localized harm to marine life (*high confidence*). Quantities will be small relative to the atmospheric invasion signal. CO_2 injected will dissolve and be transported by ocean circulation with eventual surface exposure.	Caldeira et al. (2005)
Sub-sea geologic storage	Capture of CO_2 from extracted gas or from post-combustion followed by well injection into a porous submarine aquifer beneath impermeable geologic strata.	Extensive experience in place from the Norwegian Sleipner field activity in the North Sea. No evidence of ocean impact from leakage to date.	Benson et al. (2005)
Ocean fertilization	Spreading of trace amounts of reduced iron over very large areas of the surface ocean where excess nutrients occur. Overcoming the local iron deficiency creates extensive phytoplankton blooms drawing down sea surface pCO_2. Fertilization can also be carried out by using direct or indirect (ocean pipes) addition of macronutrients to oceanic regions where they are depleted.	Much of the exported organic matter is remineralized at shallow depths, creating local oxygen stress and shallow CO_2 enrichment and methane and N_2O production. These effects are temporary and the effective retention time is short. If sustained, reduced surface ocean and increased deep ocean acidification. O_2 loss in ocean interior (*medium confidence*).	de Baar et al. (1995); de Baar et al. (2005); Pörtner et al. (2005); Boyd et al. (2007); Buesseler et al. (2008); Law (2008); Cao and Caldeira (2010)
Artificial upwelling or downwelling	Ocean fertilization by bringing nutrient rich deep water (from 200 to 1000 m) to the surface. Downwelling occurs in parallel, transporting physically dissolved CO_2 into the deep ocean.	Deep water contains high levels of CO_2, which if released counteracts the binding of CO_2 by fertilization. No evidence available.	Lovelock and Rapley (2007); Oschlies et al. (2010)
Sequestration of organic carbon	Storage of terrestrial biomass in the coastal or deep ocean.	Physical impact, regional loss of oxygen, CO_2 accumulation and acidification during degradation; increases in methane, N_2O, and H_2S. No evidence available.	Metzger and Benford (2001); Strand and Benford (2009)
Carbonate neutralization	Dissolution of power plant flue gas into sea water yielding an acidic solution that is neutralized by addition of crushed limestone. The resulting bicarbonate-rich fluid is discharged to the ocean.	Involves the transport and crushing to fine scale of large quantities of limestone and the processing of very large quantities of sea water. Environmental impact issues not yet explored.	Rau (2011)
Accelerated olivine weathering	Uses wind powered electrochemical processes to remove HCl from the ocean and neutralizes the acid with silicate minerals such as olivine for disposal. The net result is to add alkalinity to the ocean akin to natural silicate weathering processes.	Complex system as yet untested in pilot processes. Involves mining and crushing large quantities of silicate minerals. Very long time scale consequences uncertain.	House et al. (2007); Köhler et al. (2010)

climate fluctuations (Dale et al., 2006; see also Section 30.5.3.1.2). HABs can be stimulated by warming, nutrient fluctuations in upwelling areas, eutrophication in coastal areas, and enhanced surface stratification (*medium confidence*). Species-specific responses involve shifts in seasonal cycles and blooms (Johns et al., 2003). Ocean acidification may exacerbate the toxicity of species in coastal oceans under nutrient-limited conditions (Tatters et al., 2012; Sun et al., 2011). Suitable adaptation measures include appropriate monitoring of biotoxin problems (Hallegraeff, 2010).

Continued warming of tropical and temperate coastal habitats, excessive nutrient loading leading to phytoplankton and zooplankton blooms, and sea water inundation due to sea level rise are all projected to exacerbate the expansion and threat of cholera (*medium confidence*; see also Sections 11.5.2.1, 30.6.3), although attribution to climate change is confounded by climate variability and non-climate drivers (Lafferty, 2009; Dobson, 2009).

Cholera and its pathogen, the marine bacterium, *Vibrio cholera*, have been widely studied. The pathogen associates with marine organisms, especially chitinized zooplankton (Vezzulli et al., 2010). Where cholera is endemic (e.g., India, Bangladesh, Latin America), outbreaks correlate with warming and high zooplankton abundance (Lobitz et al., 2000; Lipp et al., 2002). Based on an 18-year climate record for Bangladesh, Pascual et al. (2000) reported cholera outbreaks at ENSO events, and the recent reappearance of cholera in Peru has also been linked to the intense 1991–1992 ENSO (Lipp et al., 2002). An increase in sustained maximum temperatures of the Baltic Sea (Section 30.5.3.1.4) has been related to an increase in reported *Vibrio* infections; highest human mortality rates were associated with *V. vulnificus* infections (Baker-Austin et al., 2013). Continued warming of tropical and temperate coastal habitats, excessive nutrient loading leading to phytoplankton and zooplankton blooms, and seawater inundation due to sea level rise are all projected to exacerbate the expansion and threat of cholera (*medium confidence*).

Ciguatera poisoning may occur when people consume fish, mainly from tropical reefs, that have ciguatoxins from the epiphytic dinoflagellate *Gambierdiscus* sp. Historical records show significant correlations between ciguatera poisoning and sea surface temperature in South Pacific nations (Hales et al., 1999). However, the relationship is nonlinear and dependent on the thermal window of the specific dinoflagellate (Llewellyn, 2010). This casts doubt on the accuracy of projected increases in ciguatera poisoning using linear extrapolations from observations (*low confidence*).

6.4.3. Conclusions

Human societies benefit from and depend on marine ecosystem services, including the provisioning of food and other goods, regulation of climate and extreme events, and cultural and supporting services (Section 6.4.1). Attributing and projecting climate-change-mediated shifts in these services remains a challenge, due to the intrinsic difficulty of assessments, lack of baseline and long time series data, and confounding human impacts. However, empirical and modeling studies indicate that climate change impacts on marine ecosystems lead to changes in provisioning, regulating, and supporting services (*high confidence*), as well as cultural services (*limited evidence*, *medium agreement*).

Food production from the sea is facing diverse stressors (Section 6.4.1.1), such as overfishing and habitat degradation, which interact with climate change phenomena, including warming (Section 6.3.1), ocean acidification (Section 6.3.2), and hypoxia (Section 6.3.3). Projections of impacts on capture fisheries are constrained by uncertainties in marine primary production (*medium evidence*, *medium agreement*; Section 6.5.1). Negative effects are projected to be most significant in developing nations in tropical regions (*high confidence*). Nations at higher latitudes may even benefit from climate change effects on ocean ecosystems, at least initially (Section 6.5.3).

Climate change effects on biota will alter their climate regulation through mechanisms such as carbonate production, the biological pump, the balance between photosynthesis and respiration, and the modulation of greenhouse gases (*high confidence*; Section 6.4.1.3). However, projections of the direction and magnitudes of feedbacks are at an early stage (*low confidence*).

Future management of ecosystems and fisheries might have to aim for increasing ecosystem resilience to climate change, for example, through reductions of other human perturbations (Section 6.4.2.1). Active ocean geoengineering strategies to ameliorate climate change may prove detrimental to the functioning of ecosystems, which highlights the need for further research and careful governance (Section 6.4.2.2). There is limited understanding of how harmful algal blooms and pathogens affecting human health will respond to climate change (Section 6.4.2.3; *medium* to *low confidence*).

6.5. Projections of Future Climate Change Impacts through Modeling Approaches

A range of models explore climate change effects on marine biota, from primary producers to higher trophic levels, and test hypotheses about responses of marine species, food webs, and ecosystems (Rose et al., 2010; Fulton et al., 2011; Stock et al., 2011; see also FAQ 6.2). Both empirical and mechanistic approaches are used over a range of temporal and spatial scales (Barange et al., 2010; Stock et al., 2011). There is an increasing need for upscaling from molecular and physiological to ecosystem level (e.g., Le Quesne and Pinnegar, 2012). Uncertainty in projections of changes in marine ecosystems is partly contingent on the level of confidence in climatic and oceanographic projections (Section 6.1.1; WGI AR5 Section 9.8). Models are currently useful for developing scenarios of directional changes in net primary productivity, species distributions, community structure, and trophic dynamics of marine ecosystems, as well as their implications for ecosystem goods and services under climate change. However, specific quantitative projections by these models remain imprecise (*low confidence*; Hannah et al., 2010; Rose et al., 2010; Stock et al., 2011; FAQ 6.4).

Earth System Models couple atmosphere, cryosphere, and hydrosphere (including the oceans), as well as climate and carbon cycles, and project changes in ocean biogeochemistry under a range of CO_2 emission scenarios (WGI AR5 Chapter 6). Models focusing on population and species level responses comprise models of population dynamics, models of species distribution, and models which explicitly link effects of changes in ocean physics and chemistry to changes in interactions between species at different trophic levels, or human activities such as fishing and aquaculture (Rose et al., 2010).

6.5.1. Oceanic Primary Production

Climate-induced effects on global ocean NPP comprise changes in its long-term average, seasonal timing, and peak amplitude (Henson et al., 2013). The magnitude, direction, and pattern of projected changes vary with differences in model structure and parameterization (Box CC-PP; Figure 6-13). Unknown accuracy of current NPP observations further increases the uncertainty of projections, as does the incomplete understanding of effects of multiple drivers on NPP (Sections 6.3.1-5, 6.4). Global coupled climate-ocean biogeochemical Earth System Models (WGI AR5 Chapter 6) project an increase in NPP at high latitudes but a decrease in permanently stratified oceans at mid-latitudes, in the tropics (west tropical Pacific, tropical Indian Ocean, tropical Atlantic), and in the North Atlantic (*medium confidence*; Steinacher et al., 2010; Bopp et al., 2013) (Figure 6-13). The overall result is a reduction in global mean NPP under all RCP scenarios (*medium confidence* in the direction of projected trends, *low confidence* in the magnitude of change).

6.5.2. Higher Trophic Levels

Projected future changes in temperature and other physical and chemical oceanographic factors are expected to affect the distribution and abundance of marine fishes and invertebrates, as elaborated by species distribution models. Limits of distribution ranges of 1066 exploited species are projected to undergo shifts by a median of around 50 km per decade to higher latitudes by 2050 relative to 2000 under the SRES A1B (≈RCP6.0) scenario (Cheung et al., 2009). Some species shift toward the equator following a regional temperature gradient (Burrows et al., 2011; Cheung et al., 2013b; Pinsky et al., 2013). The rate of range shifts is projected to be three times higher for pelagic than for demersal fishes (Cheung et al., 2009), the latter shifting at a rate of around 27 to 36 km per decade (Cheung et al., 2013b). However, the expansion of hypoxic waters may have a greater impact than warming on demersal fishes (Koslow et al., 2011). As a result of distribution shifts, high-latitude regions (the Arctic, Southern Ocean) are projected to have high rates of species invasions. Intermediate latitudes are expected to undergo both invasions and local extinctions. High rates of local extinction are projected for the tropics and semi-enclosed seas (e.g., Mediterranean Sea, Persian Gulf). In addition, the future productivity and distribution

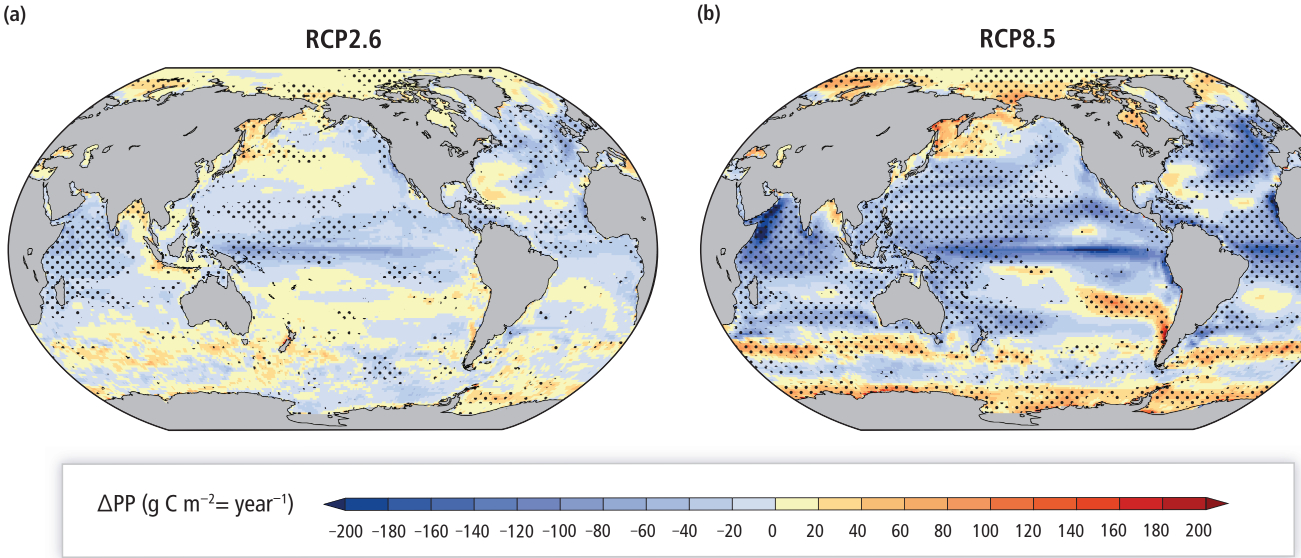


Figure 6-13 | Multi-model annual mean changes of projected vertically integrated net primary production (small and large phytoplankton) under the low-emission scenario Representative Concentration Pathway 2.6 (RCP2.6) (a) and the high-emission scenario RCP8.5 (b) for the period 2090 to 2099 relative to 1990 to 1999 (after Bopp et al., 2013). To indicate consistency in the sign of change, regions are stippled where 80% of the 10 models from the Coupled Model Intercomparison Project Phase 5 (Bopp et al. 2013) agree on the sign of change.

of higher trophic level organisms are projected to change due to changes in primary productivity (Section 6.3.6). For example, the migration route of Pacific sardine is projected to shift because of changes in primary productivity and food availability (Ito et al., 2010). The global pattern of distribution shifts is generally consistent with regional-scale projections and past observations (e.g., Lenoir et al., 2011; Cheung et al., 2013a). However, detailed quantitative projections are sensitive to model structure and assumptions (Hare et al., 2012; Jones et al., 2013) and responses of specific populations may differ from average species responses (Hazen et al., 2013).

Coral reefs are projected to undergo long-term degradation by 2020 to 2100 relative to the 2000s under RCP2.6, 4.5, and 8.5 or their equivalents (Section 30.5.6). Reefs projected to be threatened most by bleaching under the SRES A1B scenario by 2100 include the Central and Western Equatorial Pacific, Coral Triangle, and parts of Micronesia and Melanesia (Teneva et al., 2012). These projections assume that coral bleaching occurs when SST exceeds a certain threshold, and that there is limited potential to shift such threshold by adaptation. Reef degradation will impact ecosystem services (Hoegh-Guldberg, 2011; see also Section 6.4; Box CC-CR).

Some groups of marine air-breathing fauna are projected to shift in distribution and abundance (Section 6.3.7). Cetacean richness will increase above 40° latitude in both hemispheres, while at lower latitudes both pinniped and cetacean richness are projected to decrease by 2040–2049 relative to 1990–1999 under the SRES A1B scenario (Kaschner et al., 2011). Using SST as a predictor, the distribution of loggerhead turtles is projected to expand poleward in the Atlantic Ocean and to gain habitat in the Mediterranean Sea by 2070–2089 relative to 1970–1989 (Witt et al., 2010). Leatherback turtle may decrease in abundance at a rate of 7% per decade because of reduced hatching success with warming following the SRES A2 scenario (Saba et al., 2012). Abundances of some seabirds such as European breeding seabirds (Huntley et al., 2007), Cassin's auklet in the California Current Ecosystem, or emperor penguin in Antarctica are projected to decline because of climate-induced changes in oceanographic conditions, such as temperature and upwelling intensity (Wolf et al., 2010; see also Box CC-UP), or summer sea ice conditions (Jenouvrier et al., 2012). The diversity of megafaunal responses to climate change will have cascading ecosystem impacts, and will affect ecosystem services such as tourism (*high confidence*; Sections 6.3.7, 6.4.1).

6.5.3. Ecosystems and Fisheries

One of the most direct impacts of climate change on marine ecosystem services is through fisheries (Sections 6.4.1, 7.2.1.2, 7.3.2.4, 7.4.2). Projected climate impacts on fisheries are based on recruitment, growth, mortality, abundance, and distribution of fish stocks as well as changes in ocean NPP (Cheung et al., 2008), evaluated from chlorophyll concentration and other variables such as sea surface temperature (Campbell et al., 2002). Friedland et al. (2012) suggested that chlorophyll concentration, indicating both phytoplankton production and biomass, is a better predictor of the fishery yield in large marine ecosystems than NPP. While the principle holds that catch potential is dependent on energy from primary production, quantitative projections of catch potential are limited by residual uncertainty on the best possible indicators of primary production and biomass.

Assuming that the potential fish catch is proportional to NPP, the fish catch in the North Pacific Ocean subtropical biome is projected to increase by 26% through expansion of the biome, while catches in the temperate and equatorial biomes may decrease by 38 and 15%, respectively, through contraction of the biomes by 2100 relative to 2000 under the SRES A2 (RCP6.0 to 8.5) scenario (Polovina et al., 2011). Changes in phytoplankton size structure are projected to affect fisheries catch potential (Cheung et al., 2011), resulting in a 0 up to 75.8% decrease in the potential catch of large fishes in the central North Pacific

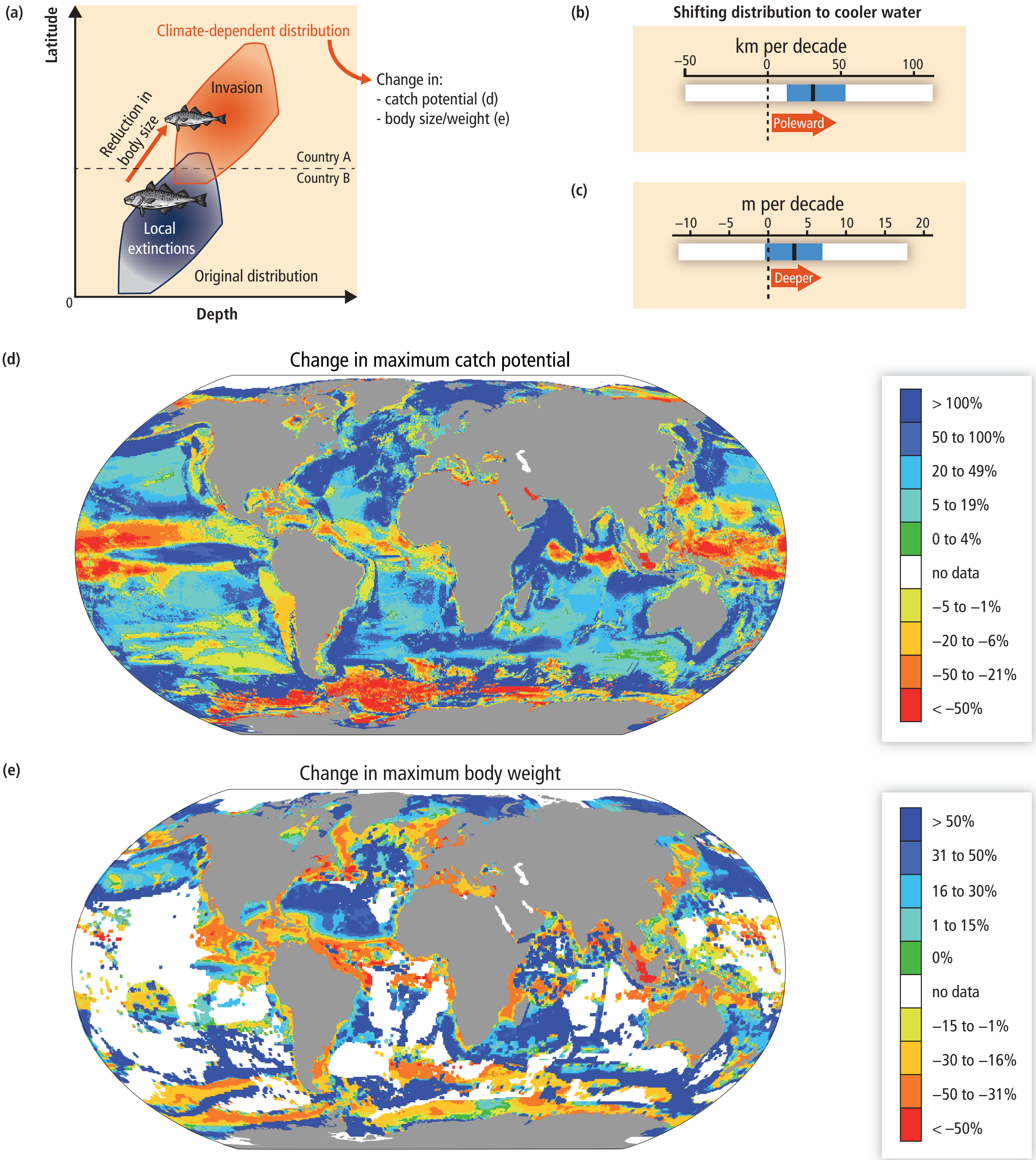


Figure 6-14 | Climate change effects on the biogeography, body size, and fisheries' catch potential of marine fishes and invertebrates. (a) Shifts in distribution range and reduction in body size of exploited fish driven by projected warming, oxygen depletion, and sea ice retreat (cf. Figure 6-7). Whenever the shift in distribution does not fully compensate for warming and hypoxia, the result will be a decrease in body size. Shifts in (b) latitudinal and (c) depth distribution of 610 exploited demersal fishes are projected to have a median (central line of the box) of 31 km per decade and 3.3 m per decade, respectively, with variation between species (box boundary: 25th and 75th percentiles) from 1991–2010 to 2041–2060 under the SRES A2 (between RCP6.0 and 8.5) scenario (Cheung et al., 2011, 2013b). (d) Combining species' range shifts with projected changes in net primary production leads to a projected global redistribution of maximum catch potential. (Analysis includes approximately 1000 species of exploited fishes and invertebrates, under warming by 2°C according to SRES A1B (≈RCP6.0), comparing the 10-year averages 2001–2010 and 2051–2060; redrawn from Cheung et al., 2010.). (e) Changes in species distribution and individual growth are projected to lead to reduced maximum body size of fish communities at a certain site. The analysis includes 610 species of marine fishes, from 1991–2010 to 2041–2060 under SRES A2 (approximately RCP6.0 to 8.5; Cheung et al., 2013b), without analysis of potential impacts of overfishing or ocean acidification. Key assumptions of the projections are that current distribution ranges reflect the preferences and tolerances of species for temperature and other environmental conditions and that these preferences and tolerances do not change over time. Catch potential is determined by species range and net primary production. Growth and maximum body size of fishes are a function of temperature and ambient oxygen level.

and increases of up to 43% in the California Current region over the 21st century under the SRES A2 scenario (Woodworth-Jefcoats et al., 2013). Globally, climate change is projected to cause a large-scale redistribution of global catch potential, with an average 30 to 70% increase in yield at high latitudes and up to 89% in some regions, after 2°C warming from preindustrial periods following SRES A1B (≈RCP6.0) (Cheung et al., 2010; Blanchard et al., 2012; see also Figure 6-14). Redistribution between areas, with average catch potential remaining unchanged, will occur at mid latitudes. A 40 to 60% drop will occur in the tropics and in Antarctica by the 2050s relative to the 2000s (*medium confidence* for direction of trends in fisheries yields, *low confidence* for the magnitude of change). This highlights high vulnerabilities in the economies of tropical coastal countries (Allison et al., 2009; see also Section 6.4).

Fisheries targeting specific species may show more complex responses to climate change. For example, driven by changes in temperature and primary production, catches of skipjack and bigeye tuna in the south Pacific are projected to increase by 2035 relative to 1980–2000 under the SRES B1 and A2 scenario, but for 2100, skipjack tuna catch is projected to decrease under the A2 scenario, while bigyeye tuna catch decreases under both A2 and B1 scenarios (Lehodey et al., 2011). Regionally, tuna catches in the Western Pacific are projected to decrease, while those in the Eastern Pacific will increase (Lehodey et al., 2011). Mollusk fisheries under ocean acidification is discussed under Section 6.4.1.

Identifying responses to climate change is complicated by species interactions and multiple stressors. Major marine habitats and biodiversity hotspots are projected to encounter cumulative impact from changes in temperature, pH, oxygen, and primary productivity by the end of the 21st century (RCP4.5 and 8.5) (Mora et al., 2013). Acidification and hypoxia will reduce maximum catch potential over 50 years from about 2000 onward in both the North Atlantic and Northeast Pacific (Ainsworth et al., 2011; Cheung et al., 2011). Changes in O_2 content as well as warming will drive a global decrease of community-averaged maximum body size of 14 to 24% of exploited demersal marine fishes by 2050 relative to the 2000s under the SRES A2 (RCP6.0 to 8.5) scenario (Cheung et al., 2013b; see also Figure 6-14). The decrease in maximum body size may affect natural mortality rates and trophic interactions, and reduce yield-per-recruit and thus potential catch. Responses of exploited marine species and their fisheries may interact with other human stressors such as overfishing, exacerbating their impacts (e.g., Lindegren et al., 2010; Ainsworth et al., 2011). Through species shifts climate change may also cause overlap of habitats of species targeted by fishing with habitat of threatened species, potentially increasing the chances of the latter of being caught as bycatch (Jones et al., 2013). Moreover, differences in vulnerability and adaptive capacity of species to changing environmental and ecosystem conditions will affect the responses of fisheries to climate change (e.g., Le Borgne et al., 2011; Griffith et al., 2011).

The complex and nonlinear interactions and responses of both biophysical and socioeconomic systems to climate change may lead to changes that have a low probability of occurrence based on empirical data (Doak et al., 2008). The risk of such low-probability but potentially high-impact events may be underestimated in existing model projections (Williams and Jackson, 2007; Lindenmayer et al., 2010). Projected changes in the distribution and production potential of fisheries resources are expected to affect economics, human livelihood, and food security (Allison et al., 2009; Sumaila and Cheung, 2010; *low confidence* in the magnitude and direction of the projected socioeconomic impacts).

6.5.4. Conclusions

Modeling projects that the distribution of invertebrates, fishes, and some marine mammals, birds, and reptiles will shift further under most emission scenarios, with rates and directions of shifts consistent with those observed in the last century (*high confidence*; Sections 6.3.1-7). These projections are valid for those species that adapt not at all or incompletely to warmer temperatures and the associated ecosystem changes, as indicated by present trends (Section 6.3.1; Box CC-MB). For non-adapting species rates of shift will thus increase with increasing rates of warming and higher emission scenarios (*high confidence*), unless the shift is blocked by geographic or other barriers (e.g., light regime; Figure 6-7). The average shift in distribution will continue to be poleward at large spatial scales (*high confidence*; Section 6.5.2; Box CC-MB). Species richness and the abundance of warm-water species will increase at high latitudes (*high confidence*) and decrease in the tropics (*medium confidence*; Section 6.5.2). Projections for individual species and populations are more variable and sensitive to model parameters.

Maximum fisheries catch potential is projected to increase at high and decrease at low latitudes by 2050 under SRES B1 (≈RCP4.5) and A1B (≈RCP6.0) climate scenarios (*medium confidence*; Section 6.5.3). Quantifying such projections is constrained by uncertainties in projected primary production rates (Sections 6.3.4, 6.5.1), biological responses such as species interactions (Section 6.3.6), and in projected effects of multiple climate drivers and human activities (*low confidence*; Section 6.3.5).

Models that integrate climate and ocean changes with biological responses and interactions, and with current human activities, have led to agreement on species and food web responses to climate change (Section 6.5.3). However, most of these models do not include trophic interactions. They insufficiently consider physiological principles and none include evolutionary adaptations that affect responses of biota to physical and chemical changes.

Projections of ocean biogeochemistry represent the open oceans rather well, but coastal and shelf regions only poorly. From a global perspective, open ocean NPP will decrease moderately by 2100 under both medium (SRES B1 or RCP4.5) and high emission scenarios (*medium confidence*; A2 or RCP6.0 to 8.5; Sections 6.3.4, 6.5.1), paralleled by an increase in NPP at high latitudes and a decrease in the tropics (*medium confidence*; Sections 6.3.4, 6.5.1; Box CC-PP).

Overall, the projected responses of marine organisms and ecosystems to climate change include changes in primary productivity (*medium confidence*), species' life history (*medium confidence*), distribution, abundance, and diversity across marine food webs (*high confidence*) in a time frame of 20 to 80 years from 2010, with substantially larger long-term (end of 21st century) responses under high emission scenarios (*high confidence*). These changes will be largest under business-as-usual scenarios (RCP8.5) and increase the vulnerability of human societies, by affecting income, employment, and food security through their effects on fisheries, tourism, and regulatory services such as coastal protection (*medium confidence*; Section 6.4.1.3; Box CC-CR).

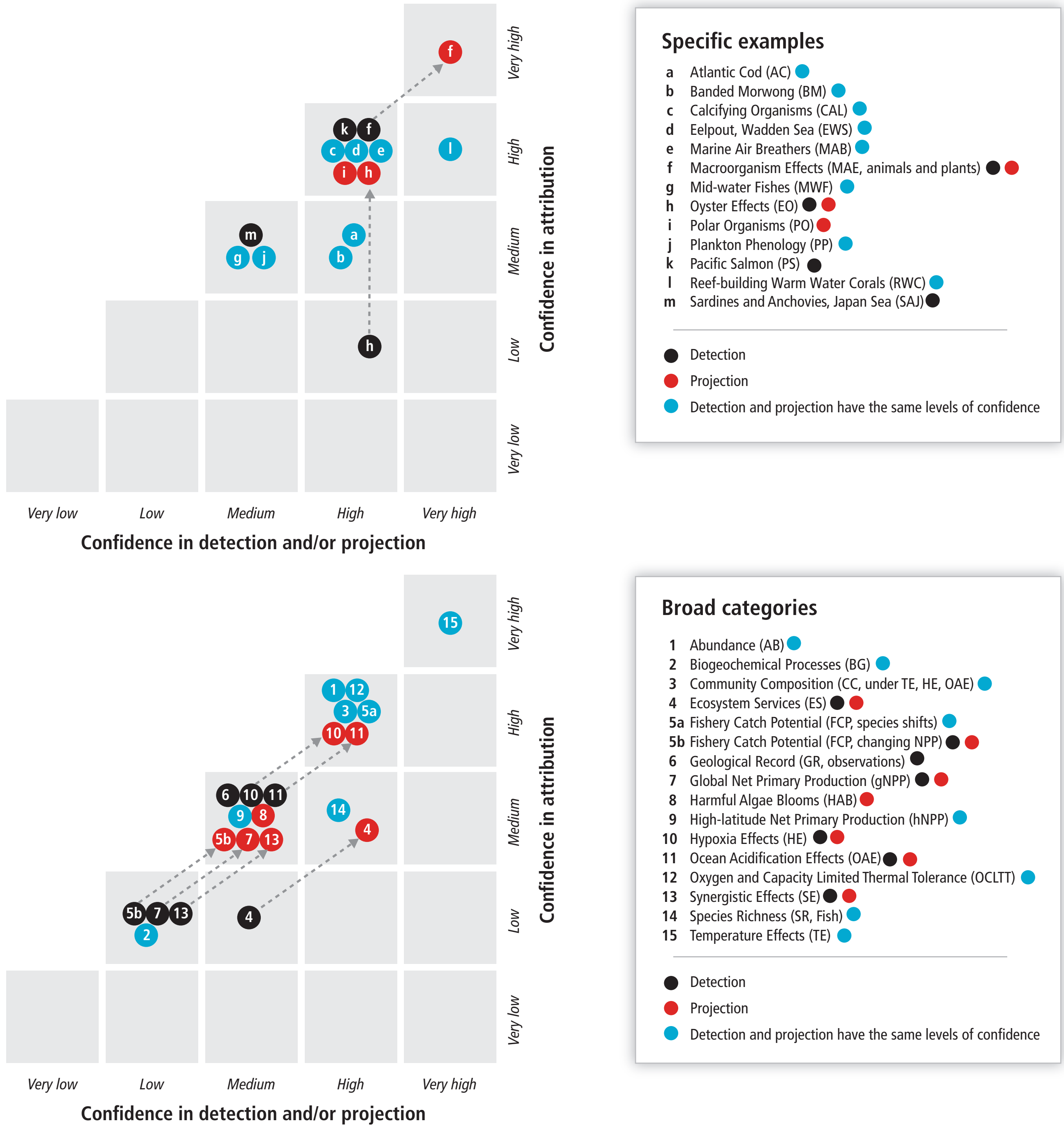


Figure 6-15 | Overview of the levels of confidence in detection, as well as in projection, of climate change effects on ocean systems, in relation to the levels of confidence in attributing these effects to the respective climate forcings. Case studies, processes, and concepts relevant in assessing the effects of climate change are represented by their acronyms in both text and figure. While confidence in the presence of effects is often high, the direct attribution to one driver in field experiments is difficult, as drivers are often highly correlated with each other (e.g., warming with changes in stratification, hence reduced nutrient supply). Some climate change impacts have been condensed into broad categories to avoid overpopulating the figures (e.g., Bio-Geochemical processes, BG). Note that the term "attribution" is used for both present-day detections in the field and future projections, the latter including qualitative and quantitative extrapolations and simulations of future conditions from fundamental principles, experiments, and models. Firm knowledge from experiments (field, laboratory, and modeling) simulating future conditions enhances the respective confidence levels to those for detection or projection. The empirical observations resulting from those experiments are directly attributable to the respective drivers. Confidence in attribution is enhanced if these experiments identify the underlying mechanisms and their responses. See text for the discussion of depicted examples and categories. Confidence assignments focus on the nature and size of effects, not on model capacity to quantify their magnitude reliably.

6.6. Chapter Conclusions and Key Uncertainties

This section provides an overview of confidence levels in the detection and projection of climate change effects on ocean systems, and of confidence levels in their attribution to different forcings. It distinguishes between effects previously observed and those projected, and considers confidence in the knowledge of underlying principles as discussed in this chapter. While the anthropogenic signal is conspicuous in the oceans (Section 6.1.1), clear attribution to anthropogenic influences on climate is not always possible in individual case studies, owing to the inherent variability of the system (Figure 6-15; acronyms of relevant processes, capitalized, link between text and figure).

Present-day observations and those from the Geological Record (GR; Figure 6-15) show similar signs of response to environmental changes, for example, warming at high CO_2 levels, and similar ecological consequences in the ocean (*robust evidence*, *medium agreement*; *medium confidence*). However, the ongoing rate of anthropogenic CO_2 release and hence ocean acidification is unprecedented in the last 65 Ma (*high confidence*) and probably the last 300 Ma (Section 6.1.2).

6.6.1. Key Risks Related to Climate Change: Constraints on Ecosystem Services

Empirical studies provide evidence that climate change has impacted marine ecosystems (*high confidence*; FAQ 6.4; Table 6-6) and has caused changes in provisioning, regulating, and supportive Ecosystem Services (ES; *medium confidence*). Climate change may also have affected cultural services (*limited evidence*, *medium agreement*) but attribution of impacts to these services remains a challenge (*low confidence*), owing to the intrinsic difficulties of assessing these services, the lack of long time-series data, and confounding human impacts. In light of available understanding of cause and effect of climate change impacts on marine ecosystems (*high confidence*), future climate change will affect some ecosystem services (*high confidence* in projection, *medium confidence* in attribution). Projected changes in the availability of marine resources and ecosystem services are expected to affect economics, human livelihood, and food security. Vulnerability is highest for the national economies of tropical coastal countries (*high confidence*).

6.6.1.1. Redistribution and Constraints on Microbial Functions and Primary Productivity

Laboratory and mesocosm studies have identified various microbially mediated processes responding to climate-induced changes in light, nutrient supply, temperature, CO_2, and hypoxia (*high confidence*). Such processes include nitrogen fixation and the nitrogen cycle, carbon sequestration and export production, calcification, respiration, O_2 production, climate-feedback by dimethylsulfide (DMS) production, and nutrient recycling. However, changes in these Bio-Geochemical processes (BG) in the field are difficult to detect, project, and attribute to climate change (*low confidence*; Sections 6.3.1-5).

The trends in net primary production recently reported for much of the low-latitude ocean using satellite observations differ considerably from those few long-term direct estimates of NPP at oceanic time series sites (Sections 6.1.2, 6.3.4). Increased NPP at high latitudes (hNPP, detected and attributable to climate change with *medium confidence*; Section 6.3.4; Box CC-PP) are indicated by satellite images (*medium confidence*) and due to reduction and thinning of sea ice. Trends in NPP will be strengthened with further warming (*medium confidence*). Modeling projects that global NPP (gNPP) will decrease by 2100 under RCP scenarios (*medium confidence*; Section 6.5.1; Box CC-PP).

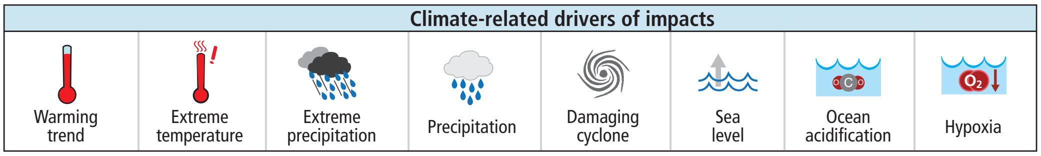
6.6.1.2. Warming-Induced Species Redistribution, Loss of Biodiversity, and Fisheries Catch Potential

Long-term observations show variability in oceanographic conditions with a key role of temperature and changing oceanographic regimes causing observed changes in ecosystem structure and fish stocks (*very high confidence*; cf. Section 30.7.1.1). Temperature Effects (TE) reflect the differential specialization of all life forms in limited ambient temperature ranges (*very high confidence*). Temperature exerts strong MAcroorganism Effects (MAE), that is, on animals and plants. Warming is presently causing and will cause species displacements and largely poleward shifts in biogeographic distribution of zooplankton and fishes, paralleled by altered seasonal activity, species abundance, migration, and body size (*high* to *very high confidence*; Section 6.3.1), and leading to shifts in Community Composition (CC; *high confidence*; Box 6-1). Causes and effects are understood for fishes and most invertebrates via their Oxygen and Capacity Limited Thermal Tolerance (OCLTT; *robust evidence*, *medium agreement*; *high confidence*; Section 6.3.1). Such knowledge supports projections into the future (*high confidence*; Section 6.5), which are influenced by the limited potential of organisms to adapt. Alterations in species ABundance (AB) result when organisms encounter shifting mean and extreme temperatures (*high confidence* in detection and attribution). Such trends will be exacerbated during future warming (*high confidence*; Section 6.5.1).

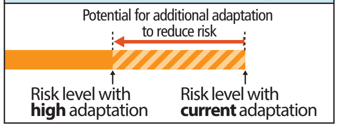
Among prominent examples, warming has caused and will cause northward shift and expansion of the geographic distribution of North Atlantic Cod (AC; *high confidence* in detection or projection, *medium confidence* in detection or projection and attribution; Section 6.3.1) and shifting growth patterns in relation to the distribution of Banded Morwong around New Zealand (BM; *high confidence* in detection or projection, *medium confidence* in detection or projection and attribution). Warming has shifted dominant species from Sardines to Anchovies in the Sea of Japan (SAJ; *medium confidence* in detection, *medium confidence* in detection and attribution; Sections 6.3.1, 6.3.6). Warming extremes have reduced and will further reduce the abundance of Eelpout in the Wadden Sea (EWS; *high confidence* in detection or projection, *high confidence* in detection or projection and attribution; Section 6.3.1). Extreme warming events increase mortalities of Pacific Salmon during spawning migrations (PS; *high confidence* in detection, *high confidence* in detection and attribution; Section 6.3.1) in Fraser River, Canada. At temperate and high latitudes, communities display increasing fish Species Richness (SR) resulting from latitudinal shifts of species and attributed to warming and loss of sea ice, although the relative contributions of regional climate variation and long-term global trends have not been quantified (*high confidence* in detection, *medium confidence* in detection and attribution; Sections 6.3.1, 6.5.2). Latitudinal species shifts are

Table 6-6 | Coastal and oceanic key risks from climate change and the potential for risk reduction through mitigation and adaptation. Key risks are identified based on assessment of the literature and expert judgments made by authors of the various WGII AR5 chapters, with supporting evaluation of evidence and agreement in the referenced chapter sections. Each key risk is characterized as *very low*, *low*, *medium*, *high*, or *very high*. Risk levels are presented for the near-term era of committed climate change (here, for 2030–2040), in which projected levels of global mean temperature increase do not diverge substantially across emissions scenarios. Risk levels are also presented for the longer-term era of climate options (here, for 2080–2100), for global mean temperature increase of 2°C and 4°C above pre-industrial levels. For each time frame, risk levels are estimated for the current state of adaptation and for a hypothetical highly adapted state. As the assessment considers potential impacts on different physical, biological, and human systems, risk levels should not necessarily be used to evaluate relative risk across key risks. Relevant climate variables are indicated by symbols. Acronyms for oceans sub-regions are as follows: HLSBS = High-Latitude Spring Bloom Systems; EUS = Equatorial Upwelling Systems; SES = Semi-Enclosed Seas; CBS = Coastal Boundary Systems; EBUE = Eastern Boundary Upwelling Ecosystems; STG = Sub-Tropical Gyres, DS = Deep Sea (>1000 m).

Climate-related drivers of impacts



Level of risk & potential for adaptation



Risks to ecosystems and adaptation options

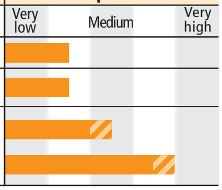
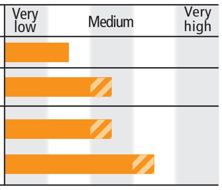
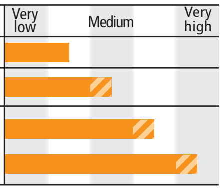
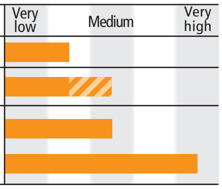
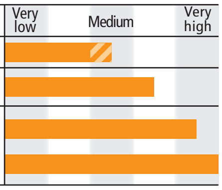
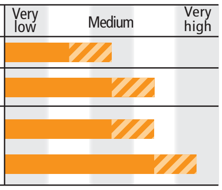
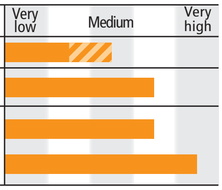
Key risk	Adaptation issues & prospects	Climatic drivers	Timeframe	Risk & potential for adaptation
Changes in ecosystem productivity associated with the redistribution and loss of net primary productivity in open oceans. (*medium confidence*) [6.5.1, 6.3.4, 30.5.1-2, Box CC-PP]	Adaptation options are limited to the translocation of industrial fishing activities due to regional decreases (low latitude) versus increases (high latitude) in productivity, or to the expansion of aquaculture.	Warming trend Extreme temperature	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
Distributional shift in fish and invertebrate species, fall in fisheries catch potential at low latitudes, e.g., in EUS, CBS, and STG regions. (*high confidence*) [6.3.1, 6.5.2-3, 30.5.1-4, 30.6.2, Box CC-MB]	Evolutionary adaptation potential of fish and invertebrate species to warming is limited as indicated by their changes in distribution to maintain temperatures. Human adaptation options involve the large-scale translocation of industrial fishing activities following the regional decreases (low latitude) versus (possibly transient) increases (high latitude) in catch potential as well as deploying flexible management that can react to variability and change. Further options include improving fish resilience to thermal stress by reducing other stressors such as pollution and eutrophication, the expansion of sustainable aquaculture and development of alternative livelihoods in some regions.	Warming trend Extreme temperature	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
High mortalities and loss of habitat to larger fauna including commercial species due to hypoxia expansion and effects. (*high confidence*) [6.3.3, 30.5.3-5]	Human adaptation options involve the large-scale translocation of industrial fishing activities as a consequence of the hypoxia-induced decreases in biodiversity and fisheries catch of pelagic fish and squid. Special fisheries may benefit (Humboldt squid). Reducing the amount of organic carbon running off coastlines by controlling nutrients and pollution running off agricultural areas can reduce microbial activity and consequently limit the extent of the oxygen drawdown and the formation of coastal dead zones.	Warming trend Hypoxia Precipitation Extreme precipitation	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
Ocean acidification: Reduced growth and survival of commercially valuable shellfish and other calcifiers, e.g., reef building corals, calcareous red algae. (*high confidence*) [5.3.3.5, 6.1.1, 6.3.2, 6.4.1.1, 30.3.2.2, Box CC-OA]	Evidence for differential resistance and evolutionary adaptation of some species exists but is likely limited by the CO_2 concentrations and high temperatures reached; adaptation options include the shift to exploiting more resilient species or the protection of habitats with low natural CO_2 levels, as well as the reduction of other stresses, mainly pollution and limiting pressures from tourism and fishing.	Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
Reduced biodiversity, fisheries abundance and coastal protection by coral reefs due to heat-induced mass coral bleaching and mortality increases, exacerbated by ocean acidification, e.g., in CBS, SES, and STG regions. (*high confidence*) [5.4.2.4, 6.3.1, 6.4.2, 30.3.1.1, 30.3.2.2, 30.5.3-6, Box CC-CR]	Evidence of rapid evolution by corals is very limited or nonexistent. Some corals may migrate to higher latitudes. However, the movement of entire reef systems is unlikely given estimates that they need to move at the speed of 10 – 20 km yr^{-1}. Human adaptation options are limited to reducing other stresses, mainly enhancing water quality and limiting pressures from tourism and fishing. This option will delay the impacts of climate change by a few decades but is likely to disappear as thermal stress increases.	Warming trend Damaging cyclone Extreme temperature Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
Coastal inundation and habitat loss due to sea level rise, extreme events, changes in precipitation, and reduced ecological resilience, e.g., in CBS and STG subregions. (*medium to high confidence*) [5.4.2.3-7, 5.5.2, 5.5.4, 30.5.6, Box CC-CR]	Options to maintain ecosystem integrity are limited to the reduction of other stresses, mainly pollution and limiting pressures from tourism, fishing, physical destruction, and unsustainable aquaculture; reducing deforestation and increasing reforestation of river catchments and coastal areas to retain sediments and nutrients; increased mangrove, coral reef, and seagrass protection and restoration to protect numerous ecosystem goods and services such as coastal protection, tourist value, and fish habitat.	Warming trend Sea level Ocean acidification Extreme precipitation Damaging cyclone	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high
Marine biodiversity loss with high rate of climate change. (*medium confidence*) [6.3.1-3, 6.4.1.2-3, Table 30.4, Box CC-MB]	Adaptation options are limited to the reduction of other stresses, mainly to reducing pollution and to limiting pressures from tourism and fishing.	Warming trend Extreme temperature Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	 Very low / Medium / Very high

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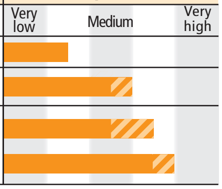
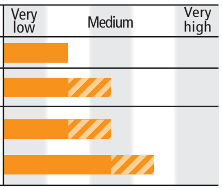
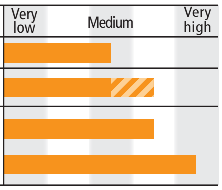
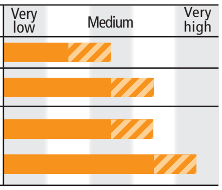
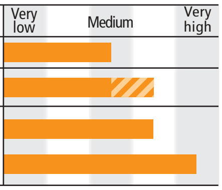
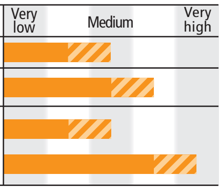
Table 6-6 (continued)

Risks to fisheries				
Key risk	**Adaptation issues & prospects**	**Climatic drivers**	**Timeframe**	**Risk & potential for adaptation**
Decreased production of global shellfish fisheries. (*high confidence*) [6.3.2, 6.3.5, 6.4.1.1, 30.5.5, 30.6.2.1, Box CC-OA]	Effective shift to alternative livelihoods, changes in food consumption patterns, and adjustment of (global) markets.	🌡 🌡! O C O	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	
Global redistribution and decrease of low-latitude fisheries yields are paralleled by a global trend to catches having smaller fishes. (*medium confidence*) [6.3.1, 6.4.1, 6.5.3, 30.5.4, 30.5.6, 30.6.2]	Increasing coastal poverty at low latitudes as fisheries becomes smaller – partially compensated by the growth of aquaculture and marine spatial planning, as well as enhanced industrialized fishing efforts.	🌡	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	
Redistribution of catch potential of large pelagic-highly migratory fish resources, such as tropical Pacific tuna fisheries. (*high confidence*) [6.3.1, 6.4.3, Table 30.4]	International fisheries agreements and instruments, such as the tuna commissions, may have limited success in establishing sustainable fisheries yields.	🌡	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	
Variability of small pelagic fishes in Eastern Boundary Upwelling systems is becoming more extreme at interannual to multi-decadal scales, making industry and management decisions more uncertain. (*medium confidence*) [6.3.2, 6.3.3, 30.5.5, Box CC-UP]	Development of new and specific management tools and models may have limited success to sustain yields. Reduction in fishing intensity increases resilience of the fisheries.	O_2↓ 🌡 O C O	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	
Decrease in catch and species diversity of fisheries in tropical coral reefs, exacerbated by interactions with other human drivers such as eutrophication and habitat destruction. (*high confidence*) [6.4.1, 30.5.3-4, 30.5.6, Table 30-4, Box CC-CR]	Restoration of overexploited fisheries and reduction of other stressors on coral reefs delay ecosystem changes. Human adaptation includes the usage of alternative livelihoods and food sources (e.g., coastal aquaculture).	🌡 O C O	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	
Current spatial management units, especially the MPAs, may fail in the future due to shifts in species distribution and community structure. (*high confidence*) [6.3.1, 6.4.2.1, 30.5.1, Box CC-MB]	Continuous revision and shifts of MPA borders, and of MPA goals and performance.	🌡 O_2↓ 🌡! O C O	Present Near term (2030 – 2040) Long term 2°C (2080 – 2100) 4°C	

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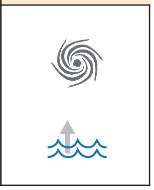
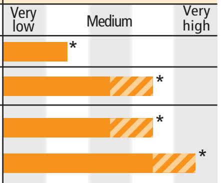
projected to continue in the 21st century under all IPCC emission scenarios (*high confidence*; Sections 6.3.1, 6.3.5, 6.3.7, 6.4.1, 6.5.2).

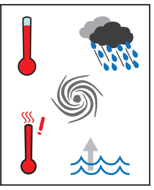
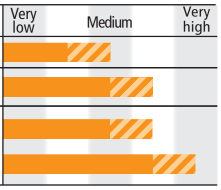
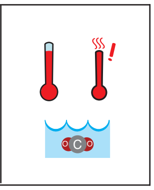
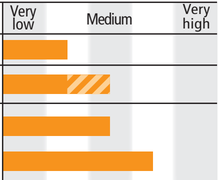
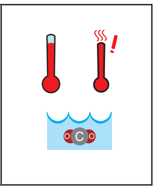
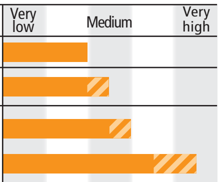
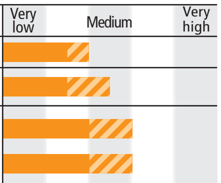
Climate-induced regime shifts and regional changes in Plankton Phenology (PP; *medium confidence*) have caused and will cause changes in food composition and availability to animals. Species shifts and changing species composition lead to changes in Fishery Catch Potential (FCP; *high confidence*; 5a in Figure 6-15), partly attributable to climate change (*high confidence*) and to sustained fishing pressure (Section 6.5.3). Fisheries Catch Potentials (FCP) will be redistributed, decrease at low latitudes, and increase at high latitudes (*high confidence*; 5a in Figure 6-15). These trends will possibly be strengthened by the projected decrease in NPP at low latitudes and increase in NPP at high latitudes (*medium confidence*; Sections 6.5.2-3; 5b in Figure 6-15). Polar Organisms (PO) that are unable to migrate to cooler waters, and to acclimatize or to adapt to warming, will become marginalized, contributing to the projected high species turnover in polar areas (*high confidence*; Sections 6.3.1, 6.5.2).

Detected effects on Marine Air Breathers (MAB: mammals, seabirds, and reptiles) include changing abundances and phenology, shifts in species distribution, and in sea turtle sex ratios (*high confidence*), all of which are partly attributed to climate change (*high confidence*). However, few effects are directly linked to climate drivers (e.g., temperature-driven turtle sex ratio); most effects are due to shifts in habitat structure (e.g., loss of sea ice), changing availability of prey organisms, or changes in

Table 6-6 (continued)

Risks to humans and infrastructure				
Key risk	**Adaptation issues & prospects**	**Climatic drivers**	**Timeframe**	**Risk & potential for adaptation**
Coastal socioeconomic security. (*high confidence*) [5.5.2, 5.5.4, 30.6.5, 30.7.1, Table 30-4]	Human adaptation options involve (1) protection using coastal defences (e.g. seawalls) and soft measures (e.g., mangrove replanting and enhancing coral growth); (2) accommodation to allow continued occupation of coastal areas by making changes to human activities and infrastructure; and (3) managed retreat as a last viable option. Options vary from large-scale engineering works to smaller scale community projects. Options are available under the more traditional CZM (coastal zone management) framework but increasingly under DRR (disaster risk reduction) and CCA (climate change adaptation) frameworks.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

**High confidence* in existence of adaptation measures, *Low confidence* in magnitude of risk reduction

Key risk	**Adaptation issues & prospects**	**Climatic drivers**	**Timeframe**	**Risk & potential for adaptation**
Reduced livelihoods and increased poverty. (*medium confidence*) [6.4.1-2, 30.6.2, 30.6.5]	Human adaptation options involve the large-scale translocation of industrial fishing activities following the regional decreases (low latitude) versus increases (high latitude) in catch potential and shifts in biodiversity. Artisanal local fisheries are extremely limited in their adaptation options by available financial resources and technical capacities, except for their potential shift to other species of interest.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts due to increased frequency of harmful algal blooms (*medium confidence*) [6.4.2.3, 30.6.3]	Adaptation options include improved monitoring and early warning system, reduction of stresses favoring harmful algal blooms, mainly pollution and eutrophication, as well as the avoidance of contaminated areas and fisheries products.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts on marine resources threatening regional security as territorial disputes and food security challenges increase (*limited evidence, medium agreement*) [AR5 SREX, 30.6.5, 30.7.2, 12.4-6, 29.3]	Decrease in marine resources, movements of fish stocks and opening of new seaways , and impacts of extreme events coupled with increasing populations will increase the potential for conflict in some regions, drive potential migration of people, and increase humanitarian crises.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts on shipping and infrastructure for energy and mineral extraction increases as storm intensity and wave height increase in some regions (e.g., high latitudes) (*high confidence*) [AR5 SREX, 30.6.2.3-4, 30.6.5, 29.3]	Adaptation options are to limit activities to particular times of the year and/or develop strategies to decrease the vulnerability of structures and operations.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

foraging efficiency, in both mammals (polar bears, walruses) and birds (penguins, albatrosses). Such trends will be exacerbated by future warming (*high confidence*; Sections 6.3.7, 6.5.2).

6.6.1.3. Expanding Hypoxia Affecting Marine Resources

Hypoxic zones in marine sediments and pelagic OMZs will continue to expand in the future, owing to climate-induced warming trends (Section 6.1.1). Local and regional Hypoxia Effects (HE) have been observed (*medium confidence*) and will be exacerbated in the future (*high confidence*; Section 6.3.3) causing habitat loss for groundfishes and pelagic predators and affecting the distribution of key zooplankton and nekton species (*medium confidence*). Progressive hypoxia is causing shifts in community composition toward hypoxia-tolerant species, excluding calcifiers due to elevated pCO_2 (*high confidence*), benefiting specialized microbes, and leading to reduced biodiversity and the loss of higher life forms (*high confidence*; Section 6.3.3). Loss of deep habitat and biomass of Mid-Water Fishes (MWF; Section 6.3.3; *medium confidence* in detection) off California is also attributed to hypoxia (*medium confidence*). These trends will continue into the future (*medium confidence*).

6.6.1.4. Constraints on Marine Calcifiers and Associated Fisheries and Aquaculture due to Ocean Acidification

Ocean acidification will exert negative effects on species and whole ecosystems and their services, especially those relying on carbonate structures such as warm-water coral reefs (*high confidence*; cf. Section 30.7.1.2). Presently, only a small number of field observations have detected Ocean Acidification Effects (OAE; *medium confidence*), but experiments and natural analogs support reliable but qualitative projections and attribution (*high confidence*). A specific glimpse into the future of anthropogenic OA is provided by negative Effects of upwelled CO_2-rich waters on Pacific Oysters (EO) introduced to aquaculture along the North American west coast (*high confidence* in detection, *low confidence* in attribution to anthropogenic causes). Findings in experimental laboratory and field studies as well as at natural analogs support attribution of projected effects to future CO_2

concentrations (*medium confidence*), with species-specific sensitivities across phyla (*high confidence*). Projected effects are most harmful to strong CALcifiers (CAL; *high confidence*), for example, some echinoderms, bivalves, gastropods, warm-water corals, and crustose algae, and less harmful to some crustaceans and, possibly, fishes. Projections from experimental studies and observations at natural analogs indicate shifts in Community Composition (CC) to more active animals and from calcifiers (CAL) to non-calcifiers in all organism groups (*high confidence* in both projection and attribution to increased CO_2; Section 6.3.2; Table 6-3).

6.6.1.5. Interactions of Climate-Related Drivers Exacerbating Impacts on Organisms, Ecosystems, and Their Services

Climate change involves interactions of temperature with other climate-related drivers and their effects (ocean acidification, hypoxia, freshening, nutrient supply, organism shifts resulting in changing interactions between species, changes in habitat structure, e.g., loss of sea ice). Strong interactions with other human impacts like eutrophication, fishing, and other forms of harvesting accelerate and amplify climate-induced changes (*high confidence*; Section 6.3.5, 30.7.1.1). Harmful algal blooms (HAB) will be stimulated by warming, nutrient fluctuations in upwelling areas, eutrophication in coastal areas (Table 6-6), ocean acidification, and enhanced surface stratification (*medium confidence*). Synergistic Effects (SE) will be exacerbated in the future (*medium confidence*), but have not yet been clearly detected and attributed in the field (*low confidence*). For projected future effects, attribution of observed impacts to such synergisms is supported by experimental evidence, especially in animals and plants (*medium confidence*).

Increased bleaching and decreased calcification displayed by several Reef-building Warm-water Corals (RWC; *very high confidence*) over the last 3 decades are attributed to the ongoing warming trend, and the associated rise in extreme temperature events and amplitudes (*high confidence*; Sections 6.3.1, 30.5.6; Box CC-CR). Such trends will be exacerbated by future warming and synergistic effects (*high confidence*; cf. Section 30.5.4.2), with some amelioration by latitudinal shifts and evolutionary adaptation (Section 6.3.1; *low confidence*). Ocean acidification will have an increasing influence on reefs (*high confidence*), as indicated by similar phenomena during mass extinctions in Earth history.

6.6.2. Key Uncertainties

Key uncertainties result from insufficient knowledge of ocean systems. International organizations (both inter- and non-governmental) have the opportunity to play a key role in coordinating research concepts and approaches, working toward a coherent picture of climate change effects on the global ocean. Countries around the world have limited capacity and infrastructure to study the ocean's response to climate change. Long-term observational time series are especially lacking, in both quantity and quality. Research has provided valuable insights, but a unifying approach addressing principles across organism domains and ecosystems is still missing. Processes investigated so far differ largely by study organisms (plants, animals, phytoplankton, and bacteria) and by level of organization (ecosystem, whole organism, tissue, cell, molecular). Especially for microbes, available data are patchy and reported trends are often in different directions, partly due to different experimental protocols and/or over-reliance on species or strains of microbes that are readily culturable, and hence have been used for decades in laboratory research. The knowledge base of climate impacts on species, strains, or communities in the field is insufficient. Scaling from physiological studies on individual species to ecosystem changes has been successful in individual cases but has not been widely implemented, for example, to shifts in species interactions or food webs. An integrated framework of climate sensitivity at the ecosystem level that considers multiple drivers and their interactive effects needs to be developed further. This includes an in depth understanding of ecosystem structure (physical and biological) and functioning, of ecosystem complexity and species interactions, and of the resulting implications for biogeochemical processes. For all climate drivers, especially ocean warming, acidification, and hypoxia, studies integrating mechanistic knowledge and evolutionary adaptation over generations are needed. Research should also cover various climate zones and biomes. Laboratory and modeling experiments are needed to test hypotheses building on long-term field observations and observations at natural or paleo-analogs. Models should better integrate observations and mechanism-based understanding, and better project future interactions between human and natural systems in a changing climate.

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6

Ocean Systems
Supplementary Material

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The Ocean

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Frequently Asked Questions

Executive Summary

The Ocean plays a central role in Earth's climate and has absorbed 93% of the extra energy from the enhanced greenhouse effect and approximately 30% of anthropogenic carbon dioxide (CO_2) from the atmosphere. Regional responses are addressed here by dividing the Ocean into seven sub-regions: High-Latitude Spring Bloom Systems (HLSBS), Eastern Boundary Upwelling Ecosystems (EBUE), Coastal Boundary Systems (CBS), Equatorial Upwelling Systems (EUS), Subtropical Gyres (STG), Semi-Enclosed Seas (SES), and the Deep Sea (DS; >1000 m). An eighth region, Polar Seas, is dealt with by Chapter 28. {Figure 30-1; WGI AR5 6.3.1; WGI AR5 Boxes 3.1, 3.8}

Global average sea surface temperatures have increased since both the beginning of the 20th century and the 1950s (certain). The average sea surface temperature (SST) of the Indian, Atlantic, and Pacific Oceans has increased by 0.65°C, 0.41°C, and 0.31°C, respectively, over the period 1950–2009 (*very likely*, *p*-value ≤ 0.05). Changes in the surface temperatures of the ocean basins are consistent with temperature trends simulated by ocean-atmosphere models with anthropogenic greenhouse gas (GHG) forcing over the past century (*high confidence*). Sub-regions within the Ocean also show *robust evidence* of change, with the influence of long-term patterns of variability (e.g., Pacific Decadal Oscillation (PDO); Atlantic Multi-decadal Oscillation (AMO)) contributing to variability at regional scales, and making changes due to climate change harder to distinguish and attribute. {30.3.1; Figure 30-2e-g; Table 30-1; WGI AR5 2.4.2-3, 3.2, 10.4.1, 14}

Uptake of CO_2 has decreased ocean pH (approximately 0.1 unit over 100 years), fundamentally changing ocean carbonate chemistry in all ocean sub-regions, particularly at high latitudes (*high confidence*). The current rate of ocean acidification is unprecedented within the last 65 Ma (*high confidence*), if not the last 300 Ma (*medium confidence*). Warming temperatures, and declining pH and carbonate ion concentrations, represent risks to the productivity of fisheries and aquaculture, and the security of regional livelihoods given the direct and indirect effects of these variables on physiological processes (e.g., skeleton formation, gas exchange, reproduction, growth, and neural function) and ecosystem processes (e.g., primary productivity, reef building and erosion) (*high confidence*). {6.1.2, 6.2-3, 30.3.2, 30.6; WGI AR5 3.8.2; WGI AR5 Boxes 3.2, 5.3.1}

Regional changes observed in winds, surface salinity, stratification, ocean currents, nutrient availability, and oxygen depth profile in many regions may be a result of anthropogenic GHG emissions (*low* to *medium confidence*). Marine organisms and ecosystems are *likely* to change in response to these regional changes, although evidence is limited and responses uncertain. {6.2-3, 30.3, 30.5; WGI AR5 2.7, 3.3-8, 10.4.2, 10.4.4}

Most, if not all, of the Ocean will continue to warm and acidify, although the rates will vary regionally (*high confidence*). Differences between Representative Concentration Pathways (RCPs) are *very likely* to be minimal until 2040 (*high confidence*). Projected temperatures of the surface layers of the Ocean, however, diverge as the 21st century unfolds and will be 1°C to 3°C higher by 2100 under RCP8.5 than RCP2.6 across most ocean sub-regions. The projected changes in ocean temperature pose serious risks and vulnerabilities to ocean ecosystems and dependent human communities (*robust evidence*, *high agreement*; *high confidence*). {6.5, 30.3.1-2, 30.7.1; Figure 30-2e-g; Table 30-3; WGI AR5 11.3.3, 12.4.7; WGI AR5 Box 1.1}

Rapid changes in physical and chemical conditions within ocean sub-regions have already affected the distribution and abundance of marine organisms and ecosystems. Responses of species and ecosystems to climate change have been observed from every ocean sub-region (*high confidence*). Marine organisms are moving to higher latitudes, consistent with warming trends (*high confidence*), with fish and zooplankton migrating at the fastest rates, particularly in HLSBS regions. Changes to sea temperature have also altered the phenology, or timing of key life-history events such as plankton blooms, and migratory patterns and spawning in fish and invertebrates, over recent decades (*medium confidence*). There is *medium* to *high agreement* that these changes pose significant uncertainties and risks to fisheries, aquaculture, and other coastal activities. Ocean acidification maybe driving similar changes (*low confidence*), although there is *limited evidence* and *low agreement* at present. The associated risks will intensify as ocean warming and acidification continue. {6.3-4, 30.4-5; Table 30-3; Box CC-MB}

Regional risks and vulnerabilities to ocean warming and acidification can be compounded by non-climate related stressors such as pollution, nutrient runoff from land, and over-exploitation of marine resources, as well as natural climate variability (*high confidence*). These influences confound the detection and attribution of the impacts of climate change and ocean acidification on ecosystems yet may also represent opportunities for reducing risks through management strategies aimed at reducing their influence, especially in CBS, SES, and HLSBS. {5.3.4, 18.3.3-4, 30.1.2, 30.5-6}

Recent changes to wind and ocean mixing within the highly productive HLSBS, EBUE, and EUS are *likely* to influence energy transfer to higher trophic levels and microbial processes. There is, however, *limited evidence* and *low agreement* on the direction and magnitude of these changes and their relationship to ocean warming and acidification (*low confidence*). In cases where Net Primary Productivity (NPP) increases or is not consumed (e.g., Benguela EBUE, *low confidence*), the increased transfer of organic carbon to deep regions can stimulate microbial respiration and reduce O_2 levels (*medium confidence*). Oxygen concentrations are also declining in the tropical Pacific, Atlantic, and Indian Oceans (particularly EUS) due to reduced O_2 solubility at higher temperatures, and changes in ocean ventilation and circulation. {6.3.3, 30.3, 30.5.1-2, 30.5.5; Box CC-PP; WGI AR5 3.8.3}

Global warming will result in more frequent extreme events and greater associated risks to ocean ecosystems (*high confidence*). In some cases (e.g., mass coral bleaching and mortality), projected increases will eliminate ecosystems, and increase risks and vulnerabilities to coastal livelihoods and food security (e.g., CBS in Southeast Asia; SES, CBS, and STG in the Indo-Pacific) (*medium* to *high confidence*). Reducing stressors not related to climate change represents an opportunity to strengthen the ecological resilience within these regions, which may help them survive some projected changes in ocean temperature and chemistry. {5.4, 30.5.3-4, 30.5.6, 30.6.1; Figure 30-4; Box CC-CR; IPCC, 2012}

The highly productive HLSBS in the Northeastern Atlantic has changed in response to warming (*medium evidence*, *high agreement*), with a range of consequences for fisheries. These ecosystems are responding to recent warming, with the greatest changes being observed since the late 1970s in the phenology, distribution, and abundance of plankton assemblages, and the reorganization of fish assemblages (*high confidence*). There is *medium confidence* that these changes will have both positive and negative implications depending on the particular HLSBS fishery and the time frame. {6.4.1.1, 6.5.3, 30.5.1, 30.6.2.1; Boxes CC-MB, 6-1}

EUS, which support highly productive fisheries off equatorial Africa and South America, have warmed over the past 60 years (Pacific EUS: 0.43°C, Atlantic EUS: 0.54°C; *very likely*, p-value ≤ 0.05). Although warming is consistent with changes in upwelling intensity, there is *low confidence* in our understanding of how EUS will change, especially in how El Niño-Southern Oscillation (ENSO) and other patterns of variability will interact in a warmer world. The risk, however, of changes to upwelling increases with average global temperature, posing significant uncertainties for dependent ecosystems, communities, and fisheries. {30.5.2; WGI AR5 14.4}

The surface waters of the SES show significant warming from 1982 and most CBS show significant warming since 1950. Warming of the Mediterranean has led to the recent spread of tropical species invading from the Atlantic and Indian Oceans. Projected warming increases the risk of greater thermal stratification in some regions, which can lead to reduced O_2 ventilation and the formation of additional hypoxic zones, especially in the Baltic and Black Seas (*medium confidence*). In some CBS, such as the East China Sea and Gulf of Mexico, these changes are further influenced by the contribution of nutrients from coastal pollution contributing to the expansion of hypoxic (low O_2) zones. These changes are *likely* to influence regional ecosystems as well as dependent industries such as fisheries and tourism, although there is *low confidence* in the understanding of potential changes and impacts. {5.3.4.3, 30.5.3-4; Table 30-1}

Coral reefs within CBS, SES, and STG are rapidly declining as a result of local stressors (i.e., coastal pollution, overexploitation) and climate change (*high confidence*). Elevated sea temperatures drive impacts such as mass coral bleaching and mortality (*very high confidence*), with an analysis of the Coupled Model Intercomparison Project Phase 5 (CMIP5) ensemble projecting the loss of coral reefs from most sites globally by 2050 under mid to high rates of ocean warming (*very likely*). {29.3.1.2, 30.5.3-4, 30.5.6; Figure 30-10; Box CC-CR}

The productive EBUE and EUS involve upwelling waters that are naturally high in CO_2 concentrations and low in pH, and hence are potentially vulnerable to ocean warming and acidification (*medium confidence*). There is *limited evidence* and *low agreement* as to how upwelling systems are *likely* to change (*low confidence*). Declining O_2 and shoaling of the aragonite saturation horizon through ocean acidification increase the risk of upwelling water being low in pH and O_2, with impacts on coastal ecosystems and fisheries, as has been seen already (e.g., California Current EBUE). These risks and uncertainties are *likely* to involve significant challenges for fisheries and associated livelihoods along the west coasts of South America, Africa, and North America (*low* to *medium confidence*). {22.3.2.3, 30.3.2.2, 30.5.2, 30.5.5; Boxes CC-UP, CC-PP}

Chlorophyll concentrations measured by satellites have decreased in the STG of the North Pacific, Indian, and North Atlantic Oceans by 9%, 12%, and 11%, respectively, over and above the inherent seasonal and interannual variability from 1998 to 2010 (*high confidence*; *p*-value $\leq$ 0.05). Significant warming over this period has resulted in increased water column stratification, reduced mixed layer depth, and possibly decreases in nutrient availability and ecosystem productivity (*limited evidence*, *medium agreement*). The short time frame of these studies against well-established patterns of long-term variability leads to the conclusion that these changes are *about as likely as not* due to climate change. {6.3.4, 30.5.6; Table 30-1; Box CC-PP; WGI AR5 3.8.4}

The world's most abundant yet difficult to access habitat, the DS, is changing (*limited evidence*, *medium agreement*), with warming between 700 and 2000 m from 1957 to 2010 *likely* to involve a significant anthropogenic signal (*medium confidence*). Decreased primary productivity of surface waters (e.g., STG) is *likely* to reduce the availability of organic carbon to DS ecosystems. Understanding of the risks of climate change and ocean acidification to the DS is important given the size of the DS region but is limited (*low confidence*). {30.5.7; Figure 30-2; WGI AR5 3.2.4; WGI AR5 Figures 3.2, 3.9}

Changes to surface wind and waves, sea level, and storm intensity will increase the vulnerability of ocean-based industries such as shipping, energy, and mineral extraction (*medium confidence*). Risks to equipment and people may be reduced through the design and use of ocean-based infrastructure, together with the evolution of policy (*medium agreement*). Risks and uncertainties will increase with further climate change. New opportunities as well as risks for shipping, energy, and mineral extraction, and international issues over access and vulnerability, may accompany warming waters, particularly at high latitudes. {10.2.2, 10.4.4, 28.2.6, 28.3.4, 30.3.1, 30.6.2; IPCC, 2012}

Changes to ocean temperature, chemistry, and other factors are generating new challenges for fisheries, as well as benefits (*high agreement*). Climate change is a risk to the sustainability of capture fisheries and aquaculture development, adding to the threats of over-fishing and other non-climate stressors. In EUS and STG, shifts in the distribution and abundance of large pelagic fish stocks will have the potential to create "winners" and "losers" among island nations and economies. There has been a boost in fish stocks of high-latitude fisheries in the HLSBS of the North Pacific and North Atlantic, partly as a result of 30 years of increase in temperature. This is *very likely* to continue, although some fish stocks will eventually decline. A number of practical adaptation options and supporting international policies can minimize the risks and maximize the opportunities. {7.4.2, 7.5.1.1.2, 29.4, 30.6-7}

Adaptation strategies for ocean regions beyond coastal waters are generally poorly developed but will benefit from international legislation and expert networks, as well as marine spatial planning (*high agreement*). Fisheries and aquaculture industries with high technology and/or large investments, as well as marine shipping and oil and gas industries, have high capacities for adaptation due to greater development of environmental monitoring, modeling, and resource assessments. For smaller scale fisheries and developing nations, building social resilience, alternative livelihoods, and occupational flexibility represent important strategies for reducing the vulnerability of ocean-dependent human communities. Building strategies that include climate forecasting and early-warning systems can reduce impacts of warming and ocean acidification in the short term. Overall, there is a strong need to develop ecosystem-based monitoring and adaptation strategies to mitigate rapidly growing risks and uncertainties to the coastal and oceanic industries, communities, and nations (*high agreement*). {7.5.1.1, 30.6}

Significant opportunity exists within the Ocean and its sub-regions for reducing the CO_2 flux to the atmosphere (*limited evidence*, *medium agreement*). Ecosystems such as mangroves, seagrass, and salt marsh offer important carbon storage and sequestration opportunities (e.g., Blue Carbon; *limited evidence*, *medium agreement*). Blue Carbon strategies can also be justified in terms of the ecosystem services provided by coastal vegetated habitats such as protection against coastal erosion and storm damage, and maintenance of habitats for fisheries species. Sequestration of anthropogenic CO_2 into deep ocean areas still faces considerable hurdles with respect to the expense, legality, and vulnerability of storage sites and infrastructure. There are also significant opportunities with the Ocean for the development of offshore renewable energy such as wind and tidal power. {5.5.7, 30.6.1, 30.6.4}

International frameworks for collaboration and decision making are critically important for coordinating policy that will enable mitigation and adaptation by the Ocean sectors to global climate change (e.g., United Nations Convention on the Law of the Sea (UNCLOS)). These international frameworks offer an opportunity to solve problems collectively, including improving fisheries management across national borders (e.g., reducing illegal, unreported, and unregulated (IUU) fishing), responding to extreme events, and strengthening international food security. Given the importance of the Ocean to all countries, there is a need for the international community to progress rapidly to a "whole of ocean" strategy for responding to the risks and challenges posed by anthropogenic ocean warming and acidification. {30.7.2}

30.1. Introduction

The Ocean exerts a profound influence as part of the Earth, interacting with its atmosphere, cryosphere, land, and biosphere to produce planetary conditions. It also directly influences human welfare through the provision and transport of food and resources, as well as by providing cultural and economic benefits. The Ocean also contributes to human welfare indirectly through the regulation of atmospheric gas content and the distribution of heat and water across the planet. This chapter examines the extent to which regional changes to the Ocean can be accurately detected and attributed to anthropogenic climate change and ocean acidification, building on the conclusions of Chapter 6, which focuses on the marine physiological and ecological responses to climate change and ocean acidification. Detailed assessment of the role of recent physical and chemical changes within the Ocean to anthropogenic climate change is provided in WGI AR5 (particularly Chapters 2, 3, 13, and 14). In this chapter, impacts, risks, and vulnerabilities associated with climate change and ocean acidification are assessed for seven ocean sub-regions, and the expected consequences and adaptation options for key ocean-based sectors are discussed. Polar oceans (defined by the presence of sea ice in the north and by the Polar Front in the south) are considered in Chapter 28.

Given that climate change affects coastal and low-lying sub-regions of multiple nations, detailed discussion of potential risks and consequences for these regions occurs in the relevant chapters of this report (e.g., Chapters 5 and 29, as well as other regional sections).

30.1.1. Major Sub-regions within the Ocean

The Ocean represents a vast region that stretches from the high tide mark to the deepest oceanic trench (11,030 m) and occupies 71% of the Earth's surface. The total volume of the Ocean is approximately 1.3 billion km^3, with approximately 72% of this volume being below 1000 m (Deep Sea (DS); Section 30.5.7). There are considerable challenges in assessing the regional impacts of climate change on the Ocean. Devising an appropriate structure to explore the influence of climate change across the entire Ocean region and the broad diversity of life forms and habitats is challenging. Longhurst (1998) identified more than 50 distinct ecological provinces in the Ocean, defined by physical characteristics and the structure and function of phytoplankton communities. Longhurst's scheme, however, yields far more sub-regions than could be sensibly discussed in the space allocated within AR5. Consequently, comparable principles were used with a division of the non-polar ocean into seven larger sub-regions similar to Barber (1988). It is recognized that these sub-regions do not always match physical-chemical patterns or specific geographies, and that they interact strongly with terrestrial regions through weather systems and the exchange of materials. Different ocean sub-regions may also have substantially different primary productivities and fishery catch. Notably, more than 80% of fishery catch is associated with three ocean sub-regions: Northern Hemisphere High-Latitude Spring Bloom Systems (HLSBS), Coastal Boundary Systems (CBS), and Eastern Boundary Upwelling Ecosystems (EBUE; Table SM30-1, Figure 30-1). The DS (>1000 m) is included as a separate category that overlaps with the six other ocean sub-regions dealt with in this chapter.

30.1.2. Detection and Attribution of Climate Change and Ocean Acidification in Ocean Sub-regions

The central goal of this chapter is to assess the recent literature on the Ocean as a region for changes that can be attributed to climate change and/or ocean acidification. Detailed assessments of recent physical and chemical changes in the Ocean are outlined in WGI AR5 Chapters 2, 3, 6, 10, 13, and 14. The detection and attribution of climate change and ocean acidification on marine organisms and ecosystems is addressed in Chapter 6. This chapter draws on these chapters to investigate regional changes in the physical, chemical, ecological, and socioeconomic aspects of the Ocean and the extent to which they can be attributed to climate change and ocean acidification.

Generally, successful attribution to climate change occurs when the full range of possible forcing factors is considered and those related to climate change are found to be the most probable explanation for the detected change in question (Section 18.2.1.1). Comparing detected changes with the expectations of well-established scientific evidence also plays a central role in the successful attribution of detected changes. This was attempted for seven sub-regions of the Ocean. There are a number of general limitations to the detection and attribution of impacts to climate change and ocean acidification that are discussed elsewhere (Section 18.2.1) along with challenges (Section 18.2.2). Different approaches and "best practice" guidelines are discussed in WGI AR5 Chapters 10 and 18, as well as in several other places (Hegerl et al., 2007, 2010; Stott et al., 2010). The fragmentary nature of ocean observing, structural uncertainty in model simulations, the influence of long-term variability, and confounding factors unrelated to climate change (e.g., pollution, introduced species, over-exploitation of fisheries) represent major challenges (Halpern et al., 2008; Hoegh-Guldberg et al., 2011b; Parmesan et al., 2011). Different factors may also interact synergistically or antagonistically with each other and climate change, further challenging the process of detection and attribution (Hegerl et al., 2007, 2010).

30.2. Major Conclusions from Previous Assessments

An integrated assessment of the impacts of climate change and ocean acidification on the Ocean as a region was not included in recent IPCC assessments, although a chapter devoted to the Ocean in the Second Assessment Report (SAR) did "attempt to assess the impacts of projected regional and global climate changes on the oceans" (Ittekkot et al., 1996). The fact that assessments for ocean and coastal systems are spread throughout previous IPCC assessment reports reduces the opportunity for synthesizing the detection and attribution of climate change and ocean acidification across the physical, chemical, ecological, and socioeconomic components of the Ocean and its sub-regions. The IPCC Fourth Assessment Report (AR4) concluded, however, that, while terrestrial sub-regions are warming faster than the oceans, "Observations since 1961 show that the average temperature of the global ocean has increased to depths of at least 3000 m and that the ocean has been taking up over 80% of the heat being added to the climate system" (AR4 Synthesis Report, p. 30). AR4 also concluded that sea levels had risen due to the thermal expansion of the Ocean but recognized that

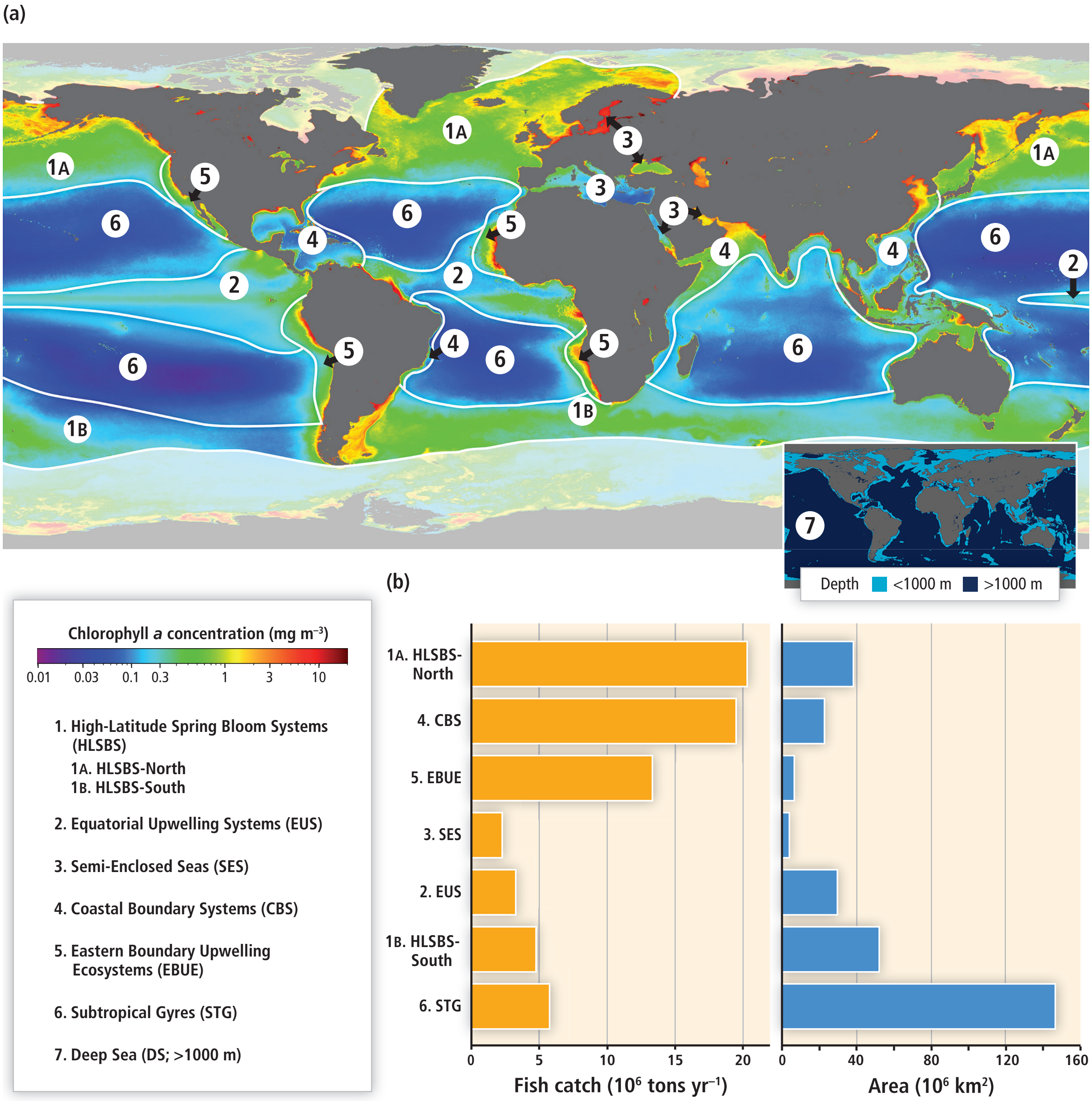


Figure 30-1 | (a) Separation of the world's oceans into seven major sub-regions (excluding an eighth area, Polar Oceans, which is considered in Chapter 28; white shaded area). The chlorophyll-*a* signal measured by SeaWiFS and averaged over the period from Sep 4, 1997 to 30 Nov 2010 (NASA) provides a proxy for differences in marine productivity (with the caveats provided in Box CC-PP). Ecosystem structure and functioning, as well as key oceanographic features, provided the basis for separating the Ocean into the sub-regions shown. The map insert shows the distribution of Deep Sea (DS) habitat (>1000 m; Bathypelagic and Abyssopelagic habitats combined). (b) Relationship between fish catch and area for each ocean subregion. Left panel: average fish catch (as millions tons yr^{-1}) for the period 1970–2006. Right panel: surface area (millions km^2). The top three bars (subregions HLSBS-North, CBS, and EBUE) cover 19% of the world oceans' area and provide 76% of the world's fish catches. Values for fish catch, area, and primary productivity of the ocean sub-regions are listed in Table SM30-1.

our understanding of the dynamics of glaciers and ice sheets was "too limited to assess their likelihood or provide a best estimate or an upper boundary for sea level rise" (WGI AR4 SPM). Changes to ocean temperature and density have been identified as having the potential to alter large-scale ocean circulation. AR4 concluded that, with respect to the Meridional Overturning Circulation (MOC), "it is *very likely* that up to the end of the 20th century the MOC was changing significantly at interannual to decadal time scales" (WGI AR4 Box 5.1, p. 397), despite limited evidence of a slowing MOC.

According to AR4, "Sea-level rise over the last 100 to150 years is probably contributing to coastal erosion in many places," including the east coast of the United States and the United Kingdom (WGII AR4 Section 1.3.3.1, p. 92). The AR4 assessment was *virtually certain* that rising atmospheric carbon dioxide (CO_2) had changed carbonate chemistry of the ocean (i.e., buffering capacity, carbonate and bicarbonate concentrations), and that a decrease in surface pH of 0.1 had occurred over the global ocean (calculated from the uptake of anthropogenic CO_2 between 1750 and 1994; Sabine et al., 2004; Raven et al., 2005; WGI AR4 Section 5.4.2.3; WGI AR4 Table 7.3). Large-scale changes in ocean salinity were also observed from 1955 to 1998 and were "characterized by a global freshening in sub-polar latitudes and salinification of shallower parts of the tropical and subtropical oceans" (WGI AR4 Chapter 5 ES, p. 387). In this case, freshening was observed in the Pacific, with increased salinity being observed in the Atlantic and Indian Oceans (WGI AR4 Sections 5.3.2-5). These changes in surface salinity were qualitatively consistent with expected changes to surface freshwater flux. Freshening of mid- and high-latitude waters together with increased salinity at low latitudes were seen as evidence "of changes in precipitation and evaporation over the oceans" (WGI AR4 SPM, p. 7).

Substantial evidence presented in AR4 indicated that changing ocean conditions have extensively influenced marine ecosystems (WGII AR4 Table 1.5). AR4 noted that there is an "accumulating body of evidence to suggest that many marine ecosystems, including managed fisheries, are responding to changes in regional climate caused predominately by warming of air and sea surface temperatures (SST) and to a lesser extent by modification of precipitation regimes and wind patterns" (WGII AR4 Section 1.3.4.2, p. 94). Observed changes in marine ecosystems and managed fisheries reported within AR4 included changes to plankton community structure and productivity, the phenology and biogeography of coastal species, intertidal communities on rocky shores, kelp forests, and the distribution of pathogens and invasive species. Changes were also observed in coral reefs (primarily increased mass coral bleaching and mortality) and migratory patterns and trophic interactions of marine birds, reptiles, and mammals, as well as of a range of other marine organisms and ecosystems (WGII AR4 Table 1.5), although a separate exercise in detection and attribution of changes due to climate change (as done for terrestrial studies) was not done as part of AR4.

30.3. Recent Changes and Projections of Future Ocean Conditions

Evidence that increasing concentrations of atmospheric CO_2 have resulted in the warming and acidification of the upper layers of the Ocean has strengthened since AR4. Understanding the full suite of physical and chemical changes to the Ocean is critical to the interpretation of the past and future responses of marine organisms and ecosystems, especially with respect to the implications for coastal and low-lying areas.

30.3.1. Physical Changes

30.3.1.1. Heat Content and Temperature

The Ocean has absorbed 93% of the extra heat arising from the enhanced greenhouse effect (1971–2010), with most of the warming (64%) occurring in the upper (0 to 700 m) ocean (1971–2010; WGI AR5 Section 3.2.3, Figure 3.2, Box 3.1). It is certain that global average SSTs have increased since the beginning of the 20th century, with improvements and growth of data sets and archives, and the understanding of errors and biases since AR4 (WGI AR5 Section 2.4.2). It is *virtually certain* that the upper ocean (0 to 700 m depth) has warmed from 1971 to 2010 (Figure 30-2a), while it is *likely* that the surface layers of the Ocean have warmed from the 1870s to 1971. Rates of increase in temperature are highest near the surface of the Ocean (>0.1°C per decade in the upper 75 m from 1971 to 2010) decreasing with depth (0.015°C per decade at 700 m; Figure 30-2b,c). It is *very likely* that the intensification of this warming near the surface has increased thermal stratification of the upper ocean by about 4% between 0 and 200 m depth from 1971 to 2010 in all parts of the ocean north of 40°S. It is *likely* that the Ocean has warmed between 700 and 2000 m from 1957 to 2010, with the warming signal becoming less apparent or non-existent at deeper depths (WGI AR5 Sections 3.2.1-3, Figures 3.1, 3.2, 3.9). These changes include a significant anthropogenic signal (*virtually certain*; Gleckler et al., 2012; Pierce et al., 2012), with the surface waters of all three ocean basins warming at different rates that exceed those expected if there were no changes to greenhouse gas (GHG) forcing over the past century (Figure 30-2e,f,g). In this respect, the observed record also falls within the range of historical model outputs that include increases in the concentration of GHGs as opposed to models that do not (Figure 30-2e,f,g).

Data archives such as Hadley Centre Interpolated SST 1.1 (HadISST1.1) contain SSTs reconstructed from a range of sources, allowing an opportunity to explore mean monthly, gridded, global SST from 1870 to the present (Rayner et al., 2003). The published HadISST1.1 data set (higher temporal and spatial resolution than HadSST3) was used to explore trends in historic SST within the sub-regions of the Ocean (Figure 30-1a; see definition of regions in Figure SM30-1 and Table SM30-2, column 1). The median SST for 1871–1995 from the Comprehensive Ocean-Atmosphere Data Set (COADS) were merged with data from the UK Met Office Marine Data Bank (MDB) to produce monthly globally complete fields of SST on a 1° latitude-longitude SST grid from 1870 to the present.

The surface layers of the three ocean basins have warmed (p-value ≤ 0.05, *very likely*), with the Indian Ocean (0.11°C per decade) warming faster than the Atlantic (0.07°C per decade) and Pacific (0.05°C per decade) Oceans (*high confidence*; Table 30-1). This is consistent with the depth-averaged (0 to 700 m) temperature trend observed from 1971 to 2010 (Figure 30-2a).

While some regions (e.g., North Pacific) did not show a clear warming trend, most regions showed either significant warming in the average temperature, or significant warming in either/or the warmest and coolest months of the year, over the period 1950–2009 (HadISST1.1 data; Table 30-1). Trends in SST show considerable sub-regional variability (Table 30-1; Figure 30-2a). Notably, the average temperature of most HLSBS did not increase significantly from 1950 to 2009 (except in the Indian Ocean; Table 30-1) yet the temperatures of the warmest month (North and South Atlantic, and Southeastern Pacific) and of the coolest month (North and South Atlantic, and South Pacific) showed significant upward trends over this period (p-value ≤ 0.05; Table 30-1).

The two EUS warmed from 1950 to 2009 (Pacific EUS: 0.07°C per decade, Atlantic EUS: 0.09°C per decade; Table 30-1). The average monthly SST of the SES did not warm significantly, although the temperature of the coolest month increased significantly within the Baltic Sea (0.35°C per decade or 2.11°C from 1950 to 2009), as did the temperatures of the warmest months in the Black (0.14°C per decade or 0.83°C from 1950 to 2009), Mediterranean (0.11°C per decade or 0.66°C from 1950 to 2009), and Red (0.05°C per decade or 0.28°C from 1950 to 2009) Seas over the period 1950–2009 (*very likely*; Table 30-1). Studies over shorter periods (e.g., 1982–2006; Belkin, 2009) report significant increases in average SST of the Baltic (1.35°C), Black (0.96°C), Red (0.74°C), and Mediterranean (0.71°C) Seas. Such studies

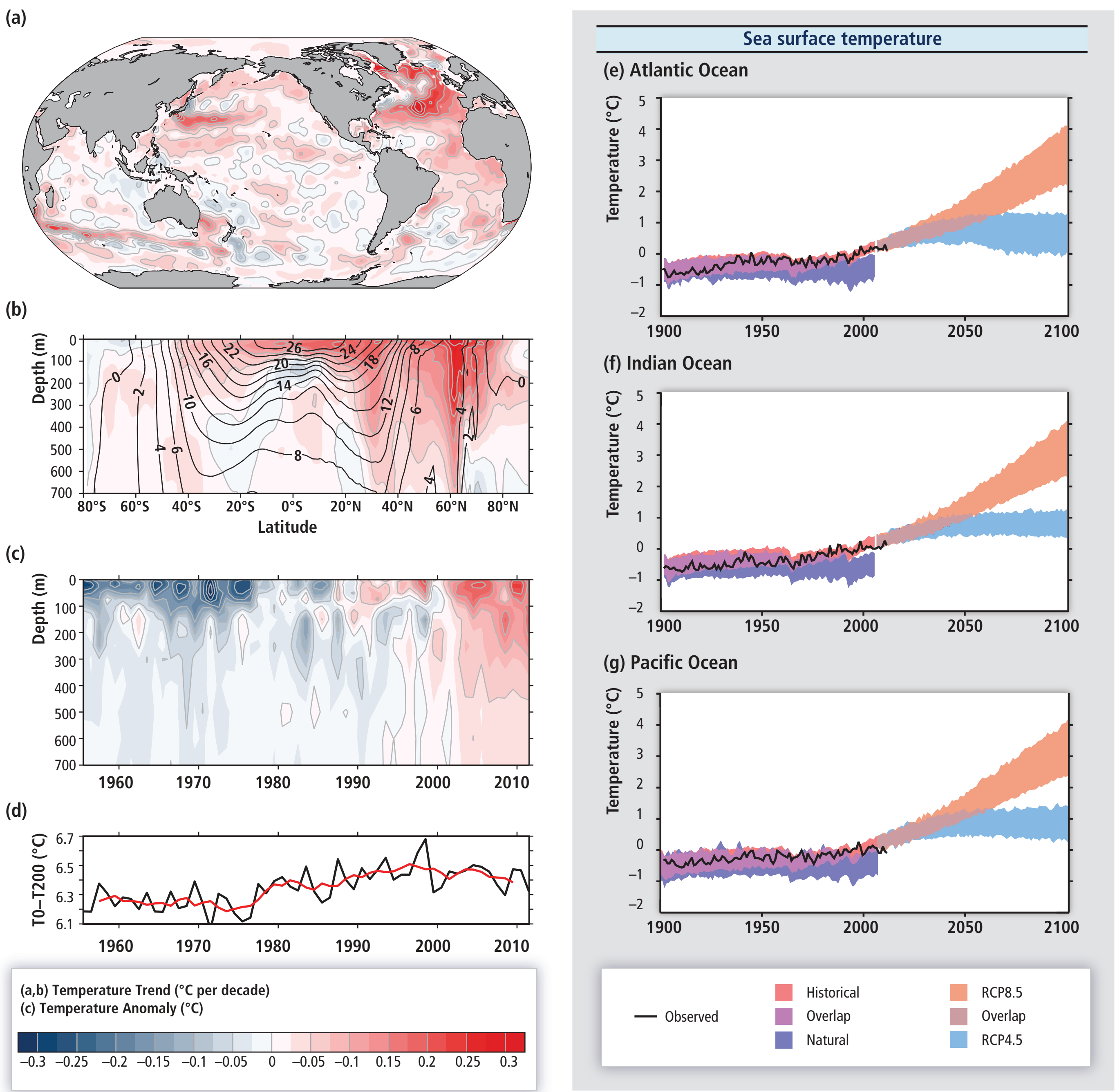


Figure 30-2 | (a) Depth-averaged 0 to 700 m temperature trend for 1971–2010 (longitude vs. latitude, colors and gray contours in degrees Celsius per decade). (b) Zonally averaged temperature trends (latitude vs. depth, colors and gray contours in degrees Celsius per decade) for 1971–2010, with zonally averaged mean temperature over plotted (black contours in degrees Celsius). (c) Globally averaged temperature anomaly (time vs. depth, colors and gray contours in degrees Celsius) relative to the 1971–2010 mean. (d) Globally averaged temperature difference between the Ocean surface and 200 m depth (black: annual values; red: 5-year running mean). [(a–d) from WGI AR5 Figure 3.1] (e)–(g) Observed and simulated variations in past and projected future annual average sea surface temperature over three ocean basins (excluding regions within 300 km of the coast). The black line shows estimates from Hadley Centre Interpolated sea surface temperature 1.1 (HadISST1.1) observational measurements. Shading denotes the 5th to 95th percentile range of climate model simulations driven with "historical" changes in anthropogenic and natural drivers (62 simulations), historical changes in "natural" drivers only (25), and the Representative Concentration Pathways (RCPs; blue: RCP4.5; orange: RCP8.5). Data are anomalies from the 1986–2006 average of the HadISST1.1 data (for the HadISST1.1 time series) or of the corresponding historical all-forcing simulations. Further details are given in Panels (a)-(d) originally presented in WGI AR5 Fig 3.1 and Box 21-2.

Table 30-1 | Regional changes in sea surface temperature (SST) over the period 1950–2009 using the ocean regionalization specified in Figure 30-1(a) (for further details on regions defined for analysis, see Figure SM30-1 and Table SM30-2, column 1). A linear regression was fitted to the average of all 1×1 degree monthly SST data extracted from the Hadley Centre HadISST1.1 data set (Rayner et al., 2003) for each sub-region over the period 1950–2009. All SST values less than –1.8°C, together with all SST pixels that were flagged as being sea ice, were reset to the freezing point of seawater (–1.8°C) to reflect the sea temperature under the ice. Separate analyses were also done to explore trends in the temperatures extracted from the coldest-ranked and the warmest-ranked month of each year (Table SM30-2). The table includes the slope of the regression (°C per decade), the *p*-value for the slope being different from zero and the total change over 60 years (i.e., the slope of linear regression multiplied by six decades) for each category. The *p*-values that exceed 0.05 plus the associated slope and change values have an orange background, denoting the lower statistical confidence in the slope being different from zero (no slope). Note that changes with higher *p*-values may still describe informative trends although the level of confidence that the slope is different from zero is lower.

Sub-region	Area	Regression slope			Total change over 60 years			*p*-value, slope different from zero		
		°C per decade (coolest month)	°C per decade (all months)	°C per decade (warmest month)	Change over 60 years (coolest month)	Change over 60 years (all months)	Change over 60 years (warmest month)	°C per decade (coolest month)	°C per decade (all months)	°C per decade (warmest month)
1. High-Latitude Spring Bloom Systems (HLSBS)	Indian Ocean	0.056	0.087	0.145	0.336	0.522	0.870	0.000	0.003	0.000
	North Atlantic Ocean	0.054	0.073	0.116	0.324	0.438	0.696	0.001	0.15	0.000
	South Atlantic Ocean	0.087	0.063	0.097	0.522	0.378	0.582	0.000	0.098	0.000
	North Pacific Ocean (west)	0.052	0.071	0.013	0.312	0.426	0.078	0.52	0.403	0.462
	North Pacific Ocean (east)	0.016	0.04	0.016	0.096	0.24	0.096	0.643	0.53	0.444
	North Pacific Ocean	0.033	0.055	0.015	0.198	0.33	0.09	0.284	0.456	0.319
	South Pacific Ocean (west)	0.043	0.017	0.044	0.258	0.102	0.264	0.016	0.652	0.147
	South Pacific Ocean (east)	0.047	0.031	0.052	0.282	0.186	0.312	0.000	0.396	0.003
	South Pacific Ocean	0.046	0.027	0.050	0.276	0.162	0.300	0.000	0.467	0.000
2. Equatorial Upwelling Systems (EUS)	Atlantic Equatorial Upwelling	0.101	0.090	0.079	0.606	0.540	0.474	0.000	0.000	0.000
	Pacific Equatorial Upwelling	0.079	0.071	0.065	0.474	0.426	0.39	0.096	0.001	0.071
3. Semi-Enclosed Seas (SES)	Arabian Gulf	0.027	0.099	0.042	0.162	0.594	0.252	0.577	0.305	0.282
	Baltic Sea	0.352	0.165	0.06	2.112	0.99	0.36	0.000	0.155	0.299
	Black Sea	–0.004	0.053	0.139	–0.024	0.318	0.834	0.943	0.683	0.009
	Mediterranean Sea	0.035	0.084	0.110	0.21	0.504	0.660	0.083	0.32	0.006
	Red Sea	0.033	0.07	0.047	0.198	0.42	0.282	0.203	0.138	0.042
4. Coastal Boundary Systems (CBS)	Atlantic Ocean (west)	0.137	0.123	0.127	0.822	0.738	0.762	0.000	0.000	0.000
	Caribbean Sea/Gulf of Mexico	0.023	0.024	0.019	0.138	0.144	0.114	0.193	0.498	0.281
	Indian Ocean (west)	0.097	0.100	0.096	0.582	0.600	0.576	0.000	0.000	0.000
	Indian Ocean (east)	0.099	0.092	0.080	0.594	0.552	0.480	0.000	0.000	0.000
	Indian Ocean (east), Southeast Asia, Pacific Ocean (west)	0.144	0.134	0.107	0.864	0.804	0.642	0.000	0.000	0.000
5. Eastern Boundary Upwelling Ecosystems (EBUE)	Benguela Current	0.062	0.032	0.002	0.372	0.192	0.012	0.012	0.437	0.958
	California Current	0.117	0.122	0.076	0.702	0.732	0.456	0.026	0.011	0.125
	Canary Current	0.054	0.089	0.106	0.324	0.534	0.636	0.166	0.014	0.000
	Humboldt Current	0.051	0.059	0.104	0.306	0.354	0.624	0.285	0.205	0.013
6. Subtropical Gyres (STG)	Indian Ocean	0.141	0.112	0.103	0.846	0.672	0.618	0.000	0.000	0.000
	North Atlantic Ocean	0.042	0.046	0.029	0.252	0.276	0.174	0.048	0.276	0.038
	South Atlantic Ocean	0.079	0.083	0.098	0.474	0.498	0.588	0.000	0.017	0.000
	North Pacific Ocean (west)	0.065	0.071	0.059	0.390	0.426	0.354	0.000	0.018	0.000
	North Pacific Ocean (east)	0.008	0.042	0.051	0.048	0.252	0.306	0.617	0.133	0.014
	North Pacific Ocean	0.034	0.055	0.051	0.204	0.33	0.306	0.001	0.053	0.000
	South Pacific Ocean (west)	0.060	0.076	0.092	0.360	0.456	0.552	0.002	0.000	0.000
	South Pacific Ocean (east)	0.055	0.056	0.088	0.330	0.336	0.528	0.000	0.058	0.000
	South Pacific Ocean	0.056	0.060	0.089	0.336	0.360	0.534	0.000	0.027	0.000

Continued next page →

Table 30-1 (continued)

	Sub-region	Regression slope			Total change over 60 years			*p*-value, slope different from zero		
		°C per decade (coolest month)	°C per decade (all months)	°C per decade (warmest month)	Change over 60 years (coolest month)	Change over 60 years (all months)	Change over 60 years (warmest month)	°C per decade (coolest month)	°C per decade (all months)	°C per decade (warmest month)
Coral Reef Provinces; see Figure 30-4(b)	Caribbean Sea/Gulf of Mexico	0.026	0.024	0.023	0.156	0.144	0.138	0.107	0.382	0.203
	Coral Triangle and Southeast Asia	0.137	0.131	0.098	0.822	0.786	0.588	0.000	0.000	0.000
	Indian Ocean (east)	0.081	0.097	0.116	0.486	0.582	0.696	0.000	0.000	0.000
	Indian Ocean (west)	0.091	0.100	0.102	0.546	0.600	0.612	0.000	0.000	0.000
	Pacific Ocean (east)	0.079	0.094	0.101	0.474	0.564	0.606	0.106	0.000	0.023
	Pacific Ocean (west)	0.072	0.073	0.073	0.432	0.438	0.438	0.000	0.000	0.000
Basin Scale	North Atlantic Ocean	0.045	0.061	0.090	0.270	0.366	0.540	0.002	0.198	0.000
	South Atlantic Ocean	0.076	0.074	0.101	0.456	0.444	0.606	0.000	0.041	0.000
	Atlantic Ocean	0.060	0.068	0.091	0.360	0.408	0.546	0.000	0.000	0.000
	North Pacific Ocean	0.030	0.052	0.046	0.180	0.312	0.276	0.000	0.248	0.006
	South Pacific Ocean	0.055	0.048	0.075	0.330	0.288	0.450	0.000	0.115	0.000
	Pacific Ocean	0.043	0.052	0.046	0.258	0.312	0.276	0.000	0.000	0.006
	Indian Ocean	0.130	0.108	0.106	0.780	0.648	0.636	0.000	0.000	0.000

are complicated by the influence of patterns of long-term variability and by the small size and land-locked nature of SES. Coastal Boundary Systems (except the Caribbean and Gulf of Mexico) all showed highly significant (*p*-value ≤ 0.05) warming (0.09°C to 0.13°C per decade; Table 30-1). Among the EBUE, the Canary and Californian Current regions exhibited a significant rate of change in the average SST (0.09°C per decade and 0.12°C per decade, respectively; *p*-value ≤ 0.05), while the Benguela and Humboldt Currents did not show significant temperature changes from 1950 to 2009 (*p*-value ≤ 0.05; Table 30-1). There was some variability between EUBEs in terms of the behavior of the coolest and warmest months. The temperature of the coolest month increased significantly from 1950 to 2009 in the case of the Benguela and California Currents (0.06°C per decade and 0.12°C per decade, respectively; p-value ≤ 0.05), while there was a significant increase in the temperature of the warmest month in the case of the Canary and Humboldt Currents (0.11°C per decade and 0.10°C per decade, respectively; Table 30-1).

The average temperature of STG showed complex patterns with increasing temperatures (1950–2009) in the Indian, South Atlantic, and South Pacific Oceans (*very likely*; 0.11°C, 0.08°C, and 0.06°C per decade, respectively; *p*-value ≤ 0.05), but not in the North Atlantic or North Pacific Ocean (*p*-value ≤ 0.05). These rates are half the value reported over shorter periods (e.g., 1998–2010; Table 1 in Signorini and McClain, 2012) and based on NOAA_OI_SST_V2 data. Given the sensitivity of coral reefs to temperature (Eakin et al., 2010; Strong et al., 2011; Lough, 2012; Box CC-CR), trends in key coral reef regions were also examined using the World Resources Institute's Reefs at Risk report (www.wri.org) to identify HadISST1.1 grid cells containing coral reefs (Figure 30-4b). Grouping the results into six major coral reef regions, coral reef waters (with the notable exception of the Gulf of Mexico and Caribbean) were found to show strong increases in average temperature (0.07°C to 0.13°C per decade) as well as the temperature of the coolest (0.07°C to 0.14°C decade) and warmest months (*very likely*) (0.07°C to 0.12°C per decade; Table 30-1). These trends in temperature have resulted in an absolute increase in sea temperature of 0.44°C to 0.79°C from 1950 to 2009.

Given the essential role that temperature plays in the biology and ecology of marine organisms (Box CC-MB; Sections 6.2-3; Pörtner, 2002; Poloczanska et al., 2013), the speed of isotherm migration ultimately determines the speed at which populations must either move, adapt, or acclimate to changing sea temperatures (Pörtner, 2002; Burrows et al., 2011; Hoegh-Guldberg, 2012). Burrows et al. (2011) calculated the rate at which isotherms are migrating as the ratio of the rate of SST change (°C yr^{-1}) to the spatial gradient of temperature (°C km^{-1}) over the period 1960–2009 (Figure 30-3). Although many of these temperature trajectories are toward the polar regions, some are not and are influenced by features such as coastlines. This analysis and others (e.g., North Atlantic; González-Taboada and Anadón, 2012) reveals that isotherms in the Ocean are moving at high velocities (to over 200 km per decade), especially at low latitudes (*high confidence*; Figure 30-3). Other sub-regions showed smaller velocities with contracting isotherms (cooling) in some areas (e.g., the Central and North Pacific, and Atlantic Oceans; Figure 30-3). There are also changes in the timing of seasonal temperatures in both spring and fall/autumn (Burrows et al., 2011; Poloczanska et al., 2013), which, together with other variables (e.g., light, food availability, geography), are *likely* to affect biological processes such as the migration of species to higher latitudes, and the timing and synchrony of reproductive and other seasonal behaviors.

Excursions of sea temperature above long-term summer temperature maxima (or below long-term temperature minima) significantly affect marine organisms and ecosystems (Hoegh-Guldberg, 1999; Bensoussan et al., 2010; Crisci et al., 2011; Harley, 2011). Consequently, calculating heat stress as a function of exposure time and size of a particular temperature anomaly is useful in understanding recent changes to

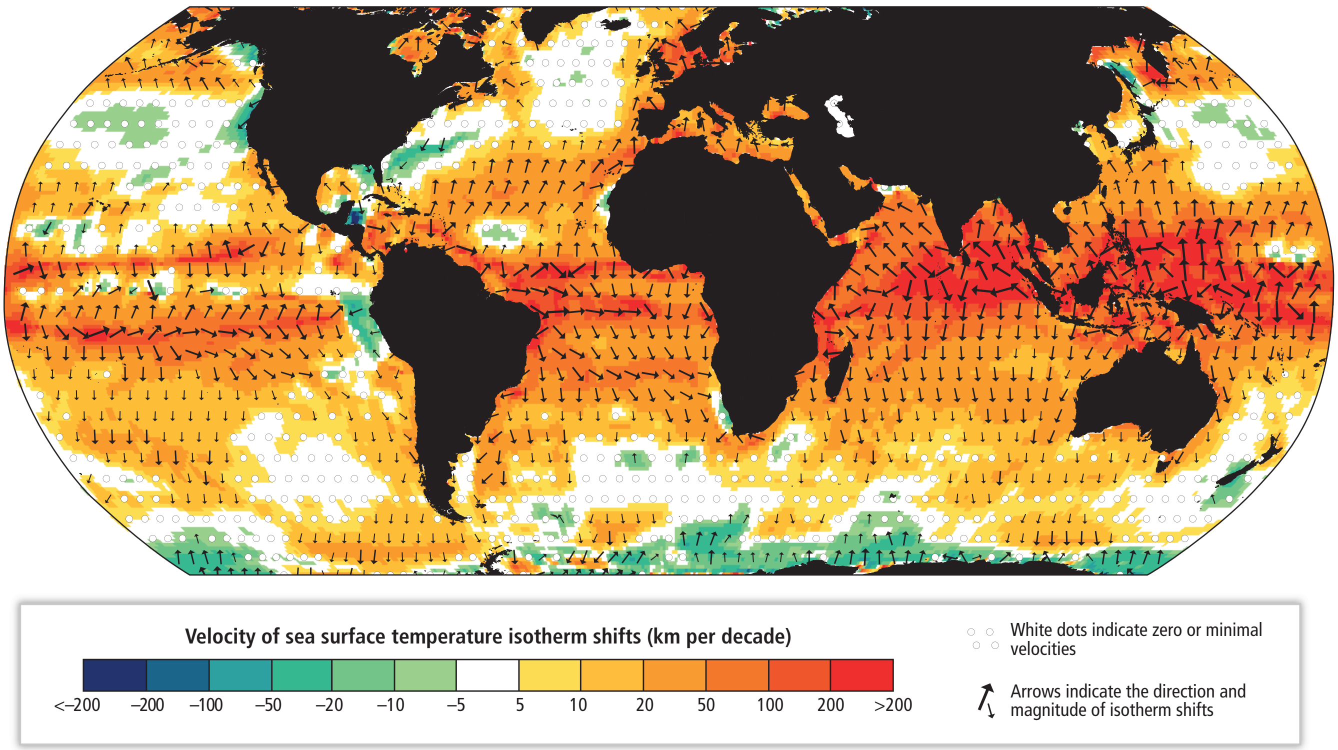


Figure 30-3 | Velocity at which sea surface temperature (SST) isotherms shifted (km per decade) over the period 1960–2009 calculated using Hadley Centre Interpolated sea surface temperature 1.1 (HadISST1.1), with arrows indicating the direction and magnitude of shifts. Velocity of climate change is obtained by dividing the temperature trend in °C per decade by the local spatial gradient °C km^{-1}. The direction of movement of SST isotherms are denoted by the direction of the spatial gradient and the sign of the temperature trend: toward locally cooler areas with a local warming trend or toward locally warmer areas where temperatures are cooling. Adapted from Burrows et al., 2011.

organisms and ecosystems (e.g., coral reefs and thermal anomalies; Strong et al., 2011). The total heat stress accumulated over the period 1981–2010 was calculated using the methodology of Donner et al. (2007) and a reference climatology based on 1985–2000 in which the highest monthly SST was used to define the thermal threshold, above which accumulated thermal stress was calculated as "exposure time multiplied by stress" or Degree Heating Months (DHM) as the running total over 4 consecutive months. While most sub-regions of the Ocean experienced an accumulation of heat stress (relative to a climatology based on the period 1985–2000), equatorial and high-latitude sub-regions in the Pacific and Atlantic Oceans have the greatest levels of accumulated heat stress (Figure 30-4a). These are areas rich in thermally sensitive coral reefs (Figure 30-4b; Strong et al., 2011). There was also a higher proportion of years that have had at least one stress event (DHM > 1) in the last 30 years (1981–2010, Figure 30-4c) than in the preceding 30 years (1951–1980; Figure 30-4c,d).

The three ocean basins will continue warming under moderate (RCP4.5) to high (RCP8.5) emission trajectories (*high confidence*) and will only stabilize over the second half of the century in the case of low range scenarios such as RCP2.6 (Figure 30-2e,f,g; WGI AR5 AI.4–AI.8). Projected changes were also examined for specific ocean sub-regions using ensemble averages from Atmosphere-Ocean General Circulation Models (AOGCM) simulations available in the Coupled Model Intercomparison Project Phase 5 (CMIP5) archive (Table SM30-3) for the four scenarios of the future (RCP2.6, RCP4.5, RCP6.0, and RCP8.5; van Vuuren et al., 2011). Ensemble averages for each RCP are based on simulations from 10 to 16 individual models (Table SM30-3). The subset of CMIP5 models were chosen because each has historic runs enabling the derivation of the maximum monthly mean (MMM) climatology from 1985 to 2000, ensuring that all anomalies were comparable across time periods and across RCPs (Figure 30-10). Model hind-cast changes matched those observed for ocean sub-regions for the period 1980–2009 (HadISST1.1; Figure 30-2), with the model ensemble slightly overestimating the extent of change across the different ocean sub-regions (slope of observed/model = 0.81, $r^2 = 0.76$, p-value ≤ 0.001). In this way, the absolute amount of change projected to occur in the ocean sub-regions was calculated for near-term (2010–2039) and long-term (2070–2099) periods (Table SM30-4). In the near term, changes in the temperature projected for the surface layers of the Ocean were largely indistinguishable between the different RCP scenarios owing to the similarity in forcing up to 2040. By the end of the century, however, SSTs across the ocean sub-regions were 1.8°C to 3.3°C higher under RCP8.5 than those projected to occur under RCP2.6 (Table SM30-4; Figure 30-2e,f,g). The implications of these projected changes on the structure and function of oceanic systems are discussed below.

30.3.1.2. Sea Level

The rate of sea level rise (SLR) since the mid-19th century has been larger than the mean rate during the previous two millennia (*high confidence*). Over the period 1901–2010, global mean sea level (GMSL) rose by 0.19 (0.17 to 0.21) m (WGI AR5 Figure SPM.3; WGI AR5 Sections 3.7, 5.6, 13.2). It is *very likely* that the mean rate of global averaged SLR was 1.7 (1.5 to 1.9) mm yr^{-1} between 1901 and 2010, 2.0 (1.7 to 2.3) mm yr^{-1}

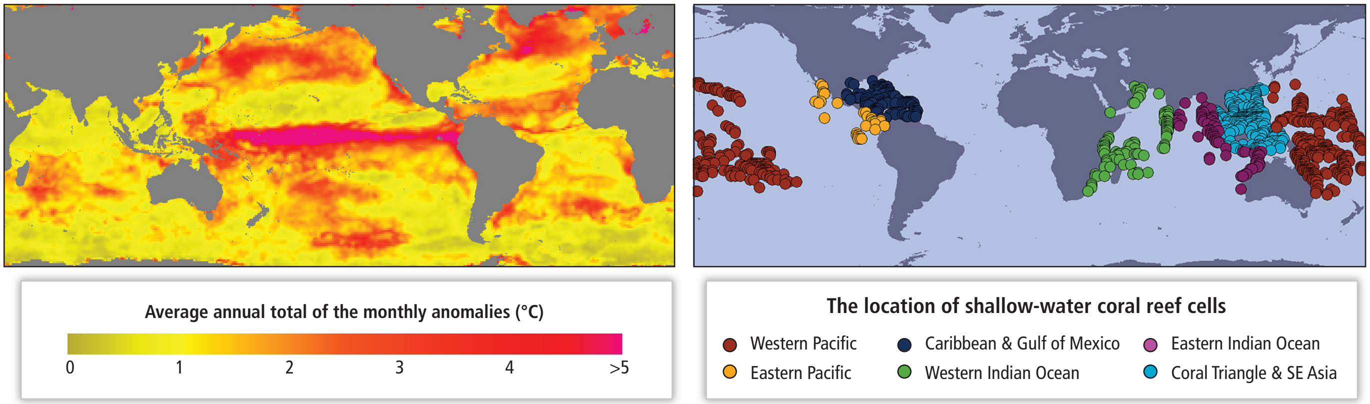
between 1971 and 2010, and 3.2 (2.8 to 3.6) mm yr^{-1} between 1993 and 2010 (WGI AR5 SPM, Section 3.7). These observations are consistent with thermal expansion of the Ocean due to warming plus the addition of water from loss of mass by melting glaciers and ice sheets. Current rates of SLR vary geographically, and can be higher or lower than the GMSL for several decades at time due to fluctuations in natural variability and ocean circulation (Figure 30-5). For example, rates of SLR are up to three times higher than the GMSL in the Western Pacific and Southeast Asian region, and decreasing in many parts of the Eastern Pacific for the period 1993–2012 as measured by satellite altimetry (Figure 30-5; WGI AR5 Section 13.6.5).

SLR under increasing atmospheric GHG concentrations will continue for hundreds of years, with the extent and rate of the increase in GMSL being dependent on the emission scenario. Central to this analysis is the millennial-scale commitment to further SLR that is *likely* to arise from the loss of mass of the Greenland and Antarctic ice sheets (WGI AR5 Section 13.5.4, Figure 13.13). SLR is *very likely* to increase during the 21st century relative to the period 1971–2010 due to increased ocean warming and the continued contribution of water from loss of mass from glaciers and ice sheets. There is *medium confidence* that median SLR by 2081–2100 relative to 1986–2005 will be (5 to 95% range of process-based models): 0.44 m for RCP2.6, 0.53 m for RCP4.5, 0.55 m for RCP6.0, and 0.74 m for RCP8.5. Higher values of SLR are possible but are not backed by sufficient evidence to enable reliable estimates of the probability of specific outcomes. Many semi-empirical model projections of GMSL rise are higher than process-based model projections (up to about twice as large), but there is no consensus in the scientific community about their reliability and there is thus *low confidence* in their projections (WGI AR5 Sections 13.5.2, 13.5.3, Table 13.6, Figure 13.12).

It is considered *very likely* that increases in sea level will result in greater levels of coastal flooding and more frequent extremes by 2050 (WGI AR5 Section 13.7.2; IPCC, 2012). It is *about as likely as not* that the frequency of the most intense storms will increase in some ocean basins,

(a) Total thermal stress for the period 1981–2010

(b) Coral reef provinces and locations



(c) Proportion of years with thermal stress (1951–1980)

(d) Proportion of years with thermal stress (1981–2010)

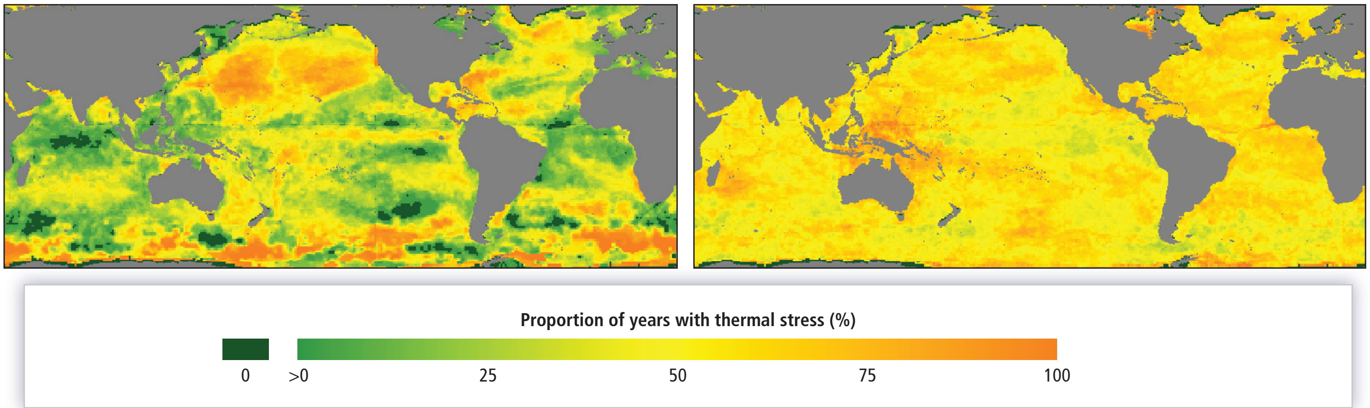


Figure 30-4 | Recent changes in thermal stress calculated using Hadley Centre Interpolated sea surface temperature data (HadISST1.1). A monthly climatology was created by averaging the HadISST monthly SST values over a reference period of 1985–2000 to create 12 averages, one for each month of the year. The Maximum Monthly Mean climatology was created by selecting the hottest month for each pixel. Anomalies were then created by subtracting this value from each sea surface temperature value, but allowing values to be recorded only if they were greater than zero (Donner et al., 2007). Two measures of the change in thermal stress were calculated as a result: The total thermal stress for the period 1981–2010, calculated by summing all monthly thermal anomalies for each grid cell (a); and the proportion of years with thermal stress, which is defined as any year that has a thermal anomaly, for the periods 1951–1980 (c) and 1981–2010 (d). The location of coral reef grid cells used in Table 30-1 and for comparison to regional heat stress is depicted in (b). Each dot is positioned over a 1 × 1 degree grid cell within which lies at least one carbonate coral reef. The latitude and longitude of each reef is derived from data provided by the World Resources Institute's Reefs at Risk report (http://www.wri.org). The six regions are as follows: red—Western Pacific Ocean; yellow—Eastern Pacific Ocean; dark blue—Caribbean and Gulf of Mexico; green—Western Indian Ocean; purple—Eastern Indian Ocean; and light blue—Coral Triangle and Southeast Asia.

although there is *medium agreement* that the global frequency of tropical cyclones is *likely* to decrease or remain constant (WGI AR5 Sections 14.6, 14.8). Although understanding of associated risks is relatively undeveloped, coastal and low-lying areas, particularly in southern Asia, as well as the Pacific Ocean and North Atlantic regions, face increased flood risk (Sections 5.3.3.2, 8.2.3.3, 9.3.4.3). Future impacts of SLR include increasing penetration of storm surges into coastal areas and changing patterns of shoreline erosion (Section 5.3), as well as the inundation of coastal aquifers by saltwater (Sections 5.4.2.5, 29.3.2). Regionally, some natural ecosystems may reduce in extent (e.g., mangroves), although examples of habitat expansion have been reported (Brown et al., 2011). Overall, changes to sea level are *very likely* to modify coastal ecosystems such as beaches, salt marshes, coral reefs, and mangroves (Section 5.4.2; Box CC-CR), especially where rates of sea level rise are highest (e.g., Southeast Asia and the Western Pacific).

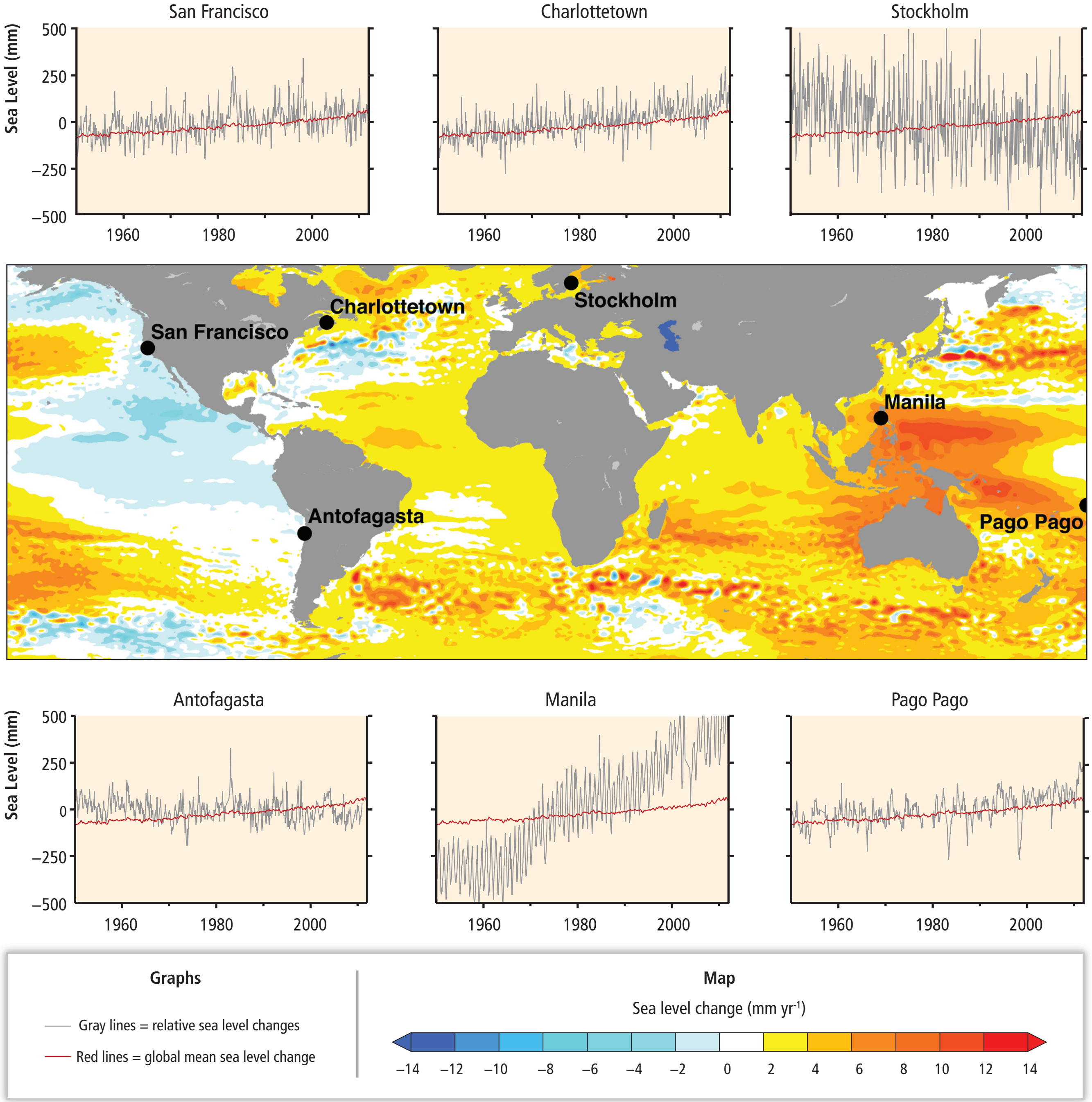


Figure 30-5 | Map of the rate of change in sea surface height (geocentric sea level) for the period 1993–2012 derived from satellite altimetry. Also shown are relative sea level changes (gray lines) from selected tide gauge stations for the period 1950–2012. For comparison, an estimate of global mean sea level change is shown (red lines) with each tide gauge time series. The relatively large short-term oscillations in local sea level (gray lines) are due to the natural climate variability and ocean circulation. For example, the large regular deviations at Pago Pago are associated with the El Niño-Southern Oscillation. Figure originally presented in WGI AR5 FAQ 13.1, Figure 1.

30.3.1.3. Ocean Circulation, Surface Wind, and Waves

Circulation of atmosphere and ocean (and their interactions) drives much of the chemical, physical, and biological characteristics of the Ocean, shaping phenomena such as ocean ventilation, coastal upwelling, primary production, and biogeochemical cycling. Critical factors for transporting nutrients from deep waters to the marine primary producers in the upper layers of the ocean include wind-driven mixing and upwelling.

There has been a poleward movement of circulation features, including a widening of the tropical belt, contraction of the northern polar vortex, and a shift of storm tracks and jet streams to higher latitudes (*medium confidence*; WGI AR5 Sections 2.7.5-6, 2.7.8; WGI AR5 Box 2.5). Long-term patterns of variability (years to decades) continue to prevent robust conclusions regarding long-term changes in atmospheric circulation and winds in many cases (WGI AR5 Section 2.7.5). There is *high confidence*, however, that the increase in northern mid-latitude westerly winds from the 1950s to the 1990s, and the weakening of the Pacific Walker Circulation from the late 19th century to the 1990s, have been largely offset by recent changes (WGI AR5 Sections 2.7.5, 2.7.8; WGI AR5 Box 2.5). Wind stress has increased since the early 1980s over the Southern Ocean (*medium confidence*; WGI AR5 Section 3.4.4), and tropical Pacific since 1990 (*medium confidence*), while zonal mean wind stress may have declined by 7% in the equatorial Pacific from 1862–1990 due to weakening of the tropical Walker Circulation (*medium confidence*; WGI AR5 Section 3.4.4; Vecchi et al., 2006). For example, it is *very likely* that the subtropical gyres of the major ocean basins have expanded and strengthened since 1993. However, the short-term nature of observing means that these changes are *as likely as not* to be due to decadal variability and/or due to longer term trends in wind forcing associated with climate change (WGI AR5 Section 3.6). Other evidence of changes in ocean circulation is limited to relatively short-term records that suffer from low temporal and spatial coverage. Therefore, there is *very low confidence* that multi-decadal trends in ocean circulation can be separated from decadal variability (WGI AR5 Section 3.6.6). There is no evidence of a long-term trend in large-scale currents such as the Atlantic Meridional Overturning Circulation (AMOC), Indonesian Throughflow (ITF), the Antarctic Circumpolar Current (ACC), or the transport of water between the Atlantic Ocean and Nordic Seas (WGI AR5 Section 3.6; WGI AR5 Figures 3.10, 3.11).

Wind speeds may have increased within the regions of EBUE (*low confidence* in attribution to climate; e.g., California Current, WGI AR5 Section 2.7.2). Changing wind regimes have the potential to influence mixed layer depth (MLD) and upwelling intensity in highly productive sub-regions of the world's oceans, although there is *low agreement* as to whether or not upwelling will intensify or not under rapid climate change (Bakun, 1990; Bakun et al., 2010; Box CC-UP).

Surface waves are influenced by wind stress, although understanding trends remains a challenge due to limited data. There is *medium confidence* that significant wave height (SWH) has increased since the mid-1950s over much of the North Atlantic north of 45°N, with typical winter season trends of up to 20 cm per decade (WGI AR5 Section 3.4.5). There is *low confidence* in the current understanding of how SWH will change over the coming decades and century for most of the Ocean. It remains an important knowledge gap (WGI AR5 Section 3.4).

30.3.1.4. Solar Insolation and Clouds

Solar insolation plays a crucially important role in the biology of many marine organisms, not only as a source of energy for photosynthesis but also as a potential co-stressor in the photic zone (with temperature), as is seen during mass coral bleaching and mortality events (e.g., Hoegh-Guldberg, 1999). Global surface solar insolation (from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Reanalysis Project; Kalnay et al., 1996) decreased by 4.3 W m^{-2} per decade from the 1950s until 1991, after which it increased at 3.3 W m^{-2} per decade until 1999 (Ohmura, 2009; Wild, 2009), matching a broad suite of evidence from many land-based sites (WGI AR5 Section 2.3.3). Although there is consistency between independent data sets for particular regions, there is substantial ambiguity and therefore *low confidence* in observations of global-scale cloud variability and trends (WGI AR5 Section 2.5.6). There is also *low confidence* in projections of how cloudiness, solar insolation, and precipitation will change as the planet warms due to the large interannual and decadal variability (El Niño-Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO)), short observation time series, and uneven spatial sampling, particularly in the early record (before 1950; WGI AR5 Section 2.5.8).

30.3.1.5. Storm Systems

As agents of water column mixing, storms (from small atmospheric disturbances to intense tropical cyclones) can remix nutrients from deeper areas into the photic zone of the Ocean, stimulating productivity. Storms can also reduce local sea temperatures and associated stress by remixing heat into the deeper layers of the Ocean (Carrigan and Puotinen, 2011). Large storms can destroy coastal infrastructure and coastal habitats such as coral reefs and mangrove forests, which can take decades to recover (Lotze et al., 2011; De'ath et al., 2012).

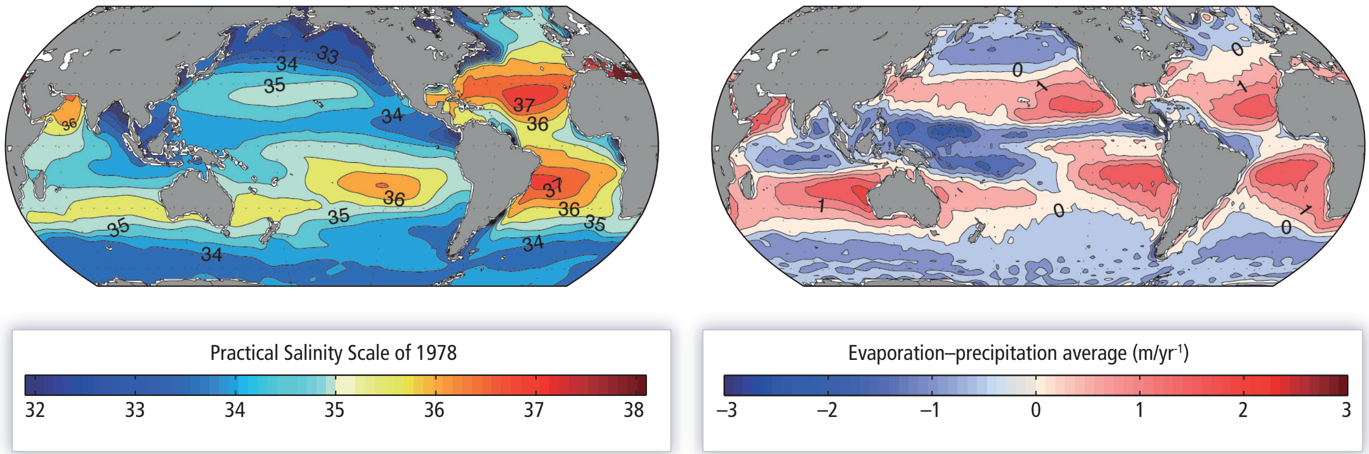
Although there is *low confidence* for long-term trends in tropical cyclone activity globally (largely due to the lack of reliable long-term data sets), it is *virtually certain* that the frequency and intensity of the strongest tropical cyclones in the North Atlantic have increased since the 1970s (WGI AR5 Section 2.6.3). There is *medium agreement* that the frequency of the most intense cyclones in the Atlantic has increased since 1987 (WGI AR5 Section 2.6.3) and *robust evidence* of inter-decadal changes in the storm track activity within the North Pacific and North Atlantic (Lee et al., 2012). It is also *likely* that there has been a decrease in the number of land-falling tropical cyclones along the East Australian coast since the 19th century (WGI AR5 Section 2.6.3; Callaghan and Power, 2011). It is *likely* that these patterns are influenced by interannual variability such as ENSO, with land-falling tropical cyclones being twice as common in La Niña versus El Niño years (*high confidence*; Callaghan and Power, 2011). There has been an increase in the number of intense wintertime extratropical cyclone systems since the 1950s in the North Pacific. Similar trends have been reported for the Asian region, although analyses are limited in terms of the spatial and temporal coverage of reliable records (WGI AR5 Section 2.6.4). There is *low confidence*, however, in large-scale trends in storminess or storminess proxies over the last century owing to the lack of long-term data and inconsistencies between studies (WGI AR5 Section 2.6.4).

30.3.1.6. Thermal Stratification

As heat has accumulated in the Ocean there has been a 4% increase in thermal stratification of the upper layers in most ocean regions (0 to 200 m, 40-year record) north of 40°S (WGI AR5 Section 3.2.2). Increasing thermal stratification has reduced ocean ventilation and the depth of mixing in many ocean sub-regions (*medium confidence*; WGI AR5 Section 3.8.3). This in turn reduces the availability of inorganic nutrients and consequently primary productivity (*medium confidence*; Section 6.3.4). In the STG, which dominate the three major ocean basins (Section 30.5.6), satellite-derived estimates of surface chlorophyll and primary production decreased between 1999 and 2007 (Box CC-PP). In contrast, however, *in situ* observations at fixed stations in the North Pacific and North Atlantic Oceans (Hawaii Ocean Time-series (HOT) and Bermuda Atlantic Time-series Study (BATS)) showed increases in nutrient and chlorophyll levels and primary production over the same period, suggesting that other processes (e.g., ENSO, PDO, North Atlantic Oscillation (NAO), winds, eddies, advection) can counteract broad-scale trends at local scales (Box CC-PP). The continued warming of the surface layers of the Ocean will *very likely* further enhance stratification and potentially limit the nutrient supply to the euphotic zone in some areas. The response of upwelling to global warming is *likely* to vary between regions and represents a complex interplay between local and global variables and processes (Box CC-UW).

(a) Climatological-mean sea surface salinity (1955–2005)

(b) Annual mean evaporation–precipitation (1950–2000)



(c) The 58-year (2008 minus 1950) sea surface salinity change

(d) The 30-year (2003–2007 average centered at 2005, minus the 1960–1989 average centered at 1975) sea surface salinity difference

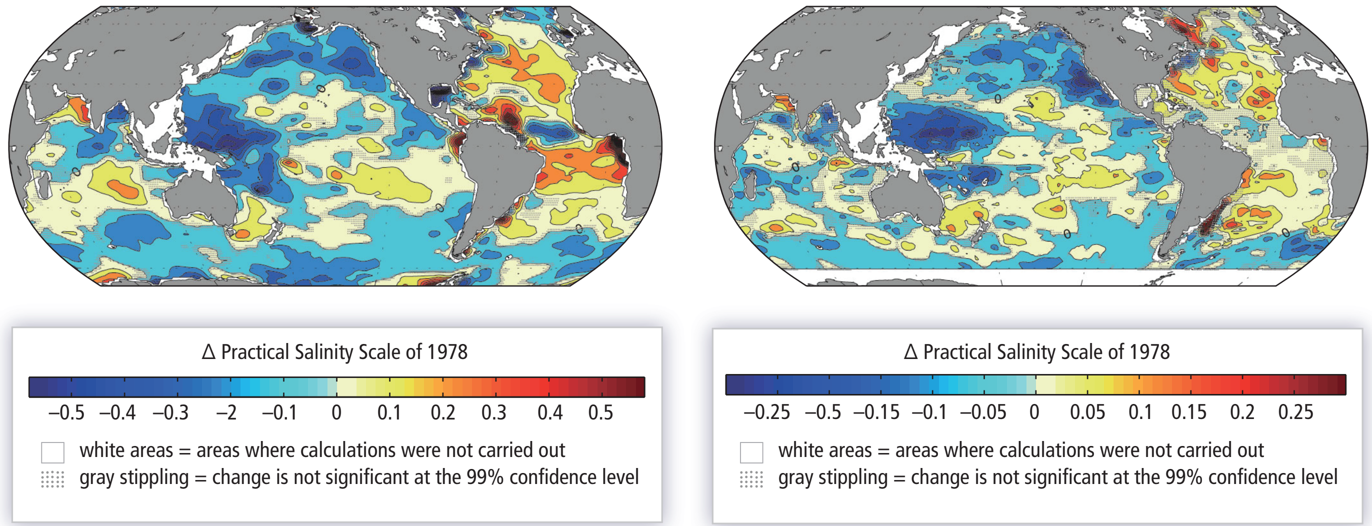


Figure 30-6 | (a) The 1955–2005 climatological-mean sea surface salinity (Antonov et al., 2010) color contoured at 0.5 Practical Salinity Scale 1978 (PSS78) intervals (black lines). (b) Annual mean evaporation-precipitation averaged over the period 1950–2000 (National Centers for Environmental Prediction (NCEP)) color contoured at 0.5 m yr–1 intervals (black lines). (c) The 58-year (2008 minus 1950) sea surface salinity change derived from the linear trend (PSS78), with seasonal and El Niño-Southern Oscillation (ENSO) signals removed (Durack and Wijffels, 2010) color contoured at 0.116 PSS78 intervals (black lines). (d) The 30-year (2003–2007 average centered at 2005, minus the 1960–1989 average centered at 1975) sea surface salinity difference (PSS78) (Hosoda et al., 2009) color contoured at 0.06 PSS78 intervals (black lines). Contour intervals in (c) and (d) are chosen so that the trends can be easily compared, given the different time intervals in the two analyzes. White areas in (c) and (d) are marginal seas where the calculations are not carried out. Regions where the change is not significant at the 99% confidence level are stippled in gray. Figure originally presented as WGI AR5 Figure 3.4. All salinity values quoted in the chapter are expressed on the Practical Salinity Scale 1978 (PSS78) (Lewis and Fofonoff, 1979).

30.3.2. Chemical Changes

30.3.2.1. Surface Salinity

The global water cycle is dominated by evaporation and precipitation occurring over ocean regions, with surface ocean salinity varying with temperature, solar radiation, cloud cover, and ocean circulation (Deser et al., 2004). Changes in salinity influence stratification of water masses and circulation. Ocean salinity varies regionally (Figure 30-6a) and is a function of the balance between evaporation and precipitation (Durack and Wijffels, 2010; WGI AR5 Section 3.3). Evaporation-dominated regions (Figure 30-6b) such as the STG and Atlantic and Western Indian Oceans (WGI AR5 Section 3.3.3) have elevated salinity, while areas of high precipitation such as the North Pacific, northeastern Indian Ocean, Southeast Asia, and the eastern Pacific have relatively low salinities (WGI AR5 Section 3.3.3; Figure 30-6a). It is *likely* that large-scale trends in salinity have also occurred in the Ocean interior, deriving from changes to salinity at the surface and subsequent subduction (WGI AR5 Section 3.3).

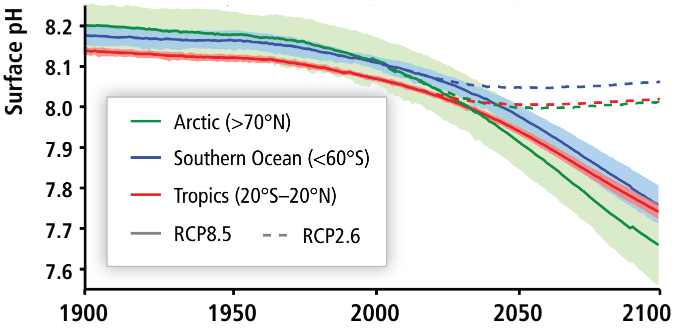
Salinity trends are consistent with the amplification of the global hydrological cycle (Durack et al., 2012; Pierce et al., 2012), a consequence of a warmer atmosphere *very likely* producing the observed trend in greater precipitation, evaporation, atmospheric moisture (Figure 30-6b), and extreme events (WGI AR5 Sections 2.6.2.1, 3.3.4; IPCC, 2012). Spatial patterns in salinity and evaporation-precipitation are correlated, providing indirect evidence that these processes have been enhanced since the 1950s (WGI AR5 Sections 3.3.2-4; WGI AR5 Figures 3.4, 3.5, 3.20d; WGI AR5 FAQ 3.3). These trends in salinity are *very likely* to have a discernible contribution from anthropogenic climate change (WGI AR5 Section 10.4.2). The combined changes in surface salinity and temperature are consistent with changes expected due to anthropogenic forcing of the climate system and are inconsistent with the effects of natural climate variability, either internal to the climate system (e.g., ENSO, PDO; Figure 30-6c,d) or external to it (e.g., solar forcing or volcanic eruptions; Pierce et al., 2012). There is *high confidence* between climate models that the observed trends in ocean salinity will continue as average global temperature increases (Durack and Wijffels, 2010; Terray et al., 2012). Ramifications of these changes are largely unknown but are of interest given the role of ocean salinity and temperature in fundamental processes such as the AMOC.

30.3.2.2. Ocean Acidification

The Ocean has absorbed approximately 30% of atmospheric CO_2 from human activities, resulting in decreased ocean pH and carbonate ion concentrations, and increased bicarbonate ion concentrations (Box CC-OA; WGI AR5 Box 3.2; WGI AR5 Figure SM30-2). The chemical response to increased CO_2 dissolving into the Ocean from the atmosphere is known with *very high confidence* (WGI AR5 Section 6.4.4). Factors such as temperature, biological processes, and sea ice (WGI AR5 Section 6.4) play significant roles in determining the saturation state of seawater for polymorphs (i.e., different crystalline forms) of calcium carbonate. Consequently, pH and the solubility of aragonite and calcite are naturally lower at high latitudes and in upwelling areas (e.g., California Current EBUE), where organisms and ecosystems may be relatively more exposed to ocean acidification as a result (Feely et al., 2012; Gruber et al., 2012; Figures 30-7a,b, SM30-2). Aragonite and calcite concentrations vary with depth, with under-saturation occurring at deeper depths in the Atlantic (calcite: 3500 to 4500 m, aragonite: 400 to 3000 m) as opposed to the Pacific and Indian Oceans (calcite: 100 to 3000 m, aragonite: 100 to 1200 m; Feely et al., 2004, 2009; Orr et al., 2005; Figure 30-8).

Surface ocean pH has decreased by approximately 0.1 pH units since the beginning of the Industrial Revolution (*high confidence*) (Figure 30-7a; WGI AR5 Section 3.8.2; WGI AR5 Box 3.2), with pH decreasing at the rate of –0.0013 and 0.0024 pH units yr^{-1} (WGI AR5 Section 3.8.2; WGI AR5 Table 3.2). The presence of anthropogenic CO_2 diminishes with

(a) Surface pH



(b) Change in surface pH in 2090s from 1990s (RCP8.5)

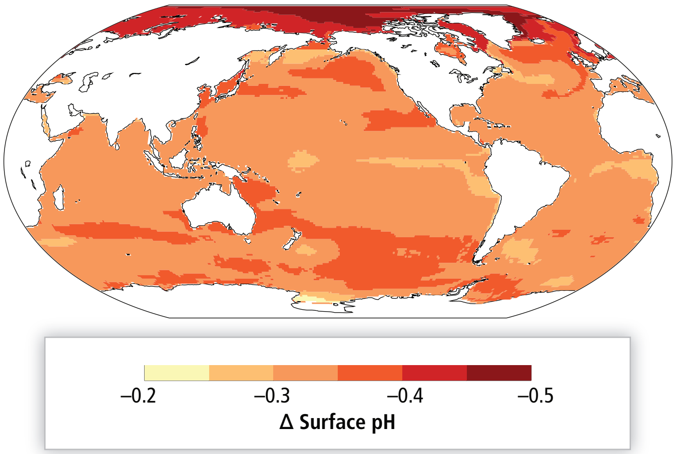


Figure 30-7 | Projected ocean acidification from 11 Coupled Model Intercomparison Project Phase 5 (CMIP5) Earth System models under RCP8.5 (other Representative Concentration Pathway (RCP) scenarios have also been run with the CMIP5 Models): (a) Time series of surface pH shown as the mean (solid line) and range of models (shaded area), given as area-weighted averages over the Arctic Ocean (green), the tropical oceans (red), and the Southern Ocean (blue). (b) Maps of the median model's change in surface pH from 1990s. Panel (a) also includes mean model results from RCP2.6 (dashed lines). Over most of the Ocean, gridded data products of carbonate system variables are used to correct each model for its present-day bias by subtracting the model-data difference at each grid cell following (Orr et al., 2005). Where gridded data products are unavailable (Arctic Ocean, all marginal seas and the Ocean near Indonesia), the results are shown without bias correction. The bias correction reduces the range of model projections by up to a factor of four; for example, in panel (a) compare the large range of model projections for the Arctic (without bias correction) to the smaller range in the Southern Ocean (with bias correction). Figure originally presented in WGI AR5 Figure 6.28.

depth. The saturation horizons of both polymorphs of calcium carbonate, however, are shoaling rapidly (1 to 2 m yr^{-1}, and up to 5 m yr^{-1} in regions such as the California Current (Orr et al., 2005; Feely et al., 2012). Further increases in atmospheric CO_2 are *virtually certain* to further acidify the Ocean and change its carbonate chemistry (Figures SM30-2, 30-7, 30-8). Doubling atmospheric CO_2 (~RCP4.5; Rogelj et al., 2012) will decrease ocean pH by another 0.1 unit and decrease carbonate ion concentrations by approximately 100 µmol kg^{-1} in tropical oceans (Figure 30-8a) from the present-day average of 250 µmol kg^{-1} (*high confidence*). Projected changes for the open Ocean by 2100 (Figures 30-7, 30-8) range from a pH change of –0.14 unit with RCP2.6 (421 ppm CO_2, +1°C, 22% reduction of carbonate ion concentration) to a pH change of –0.43 unit with RCP8.5

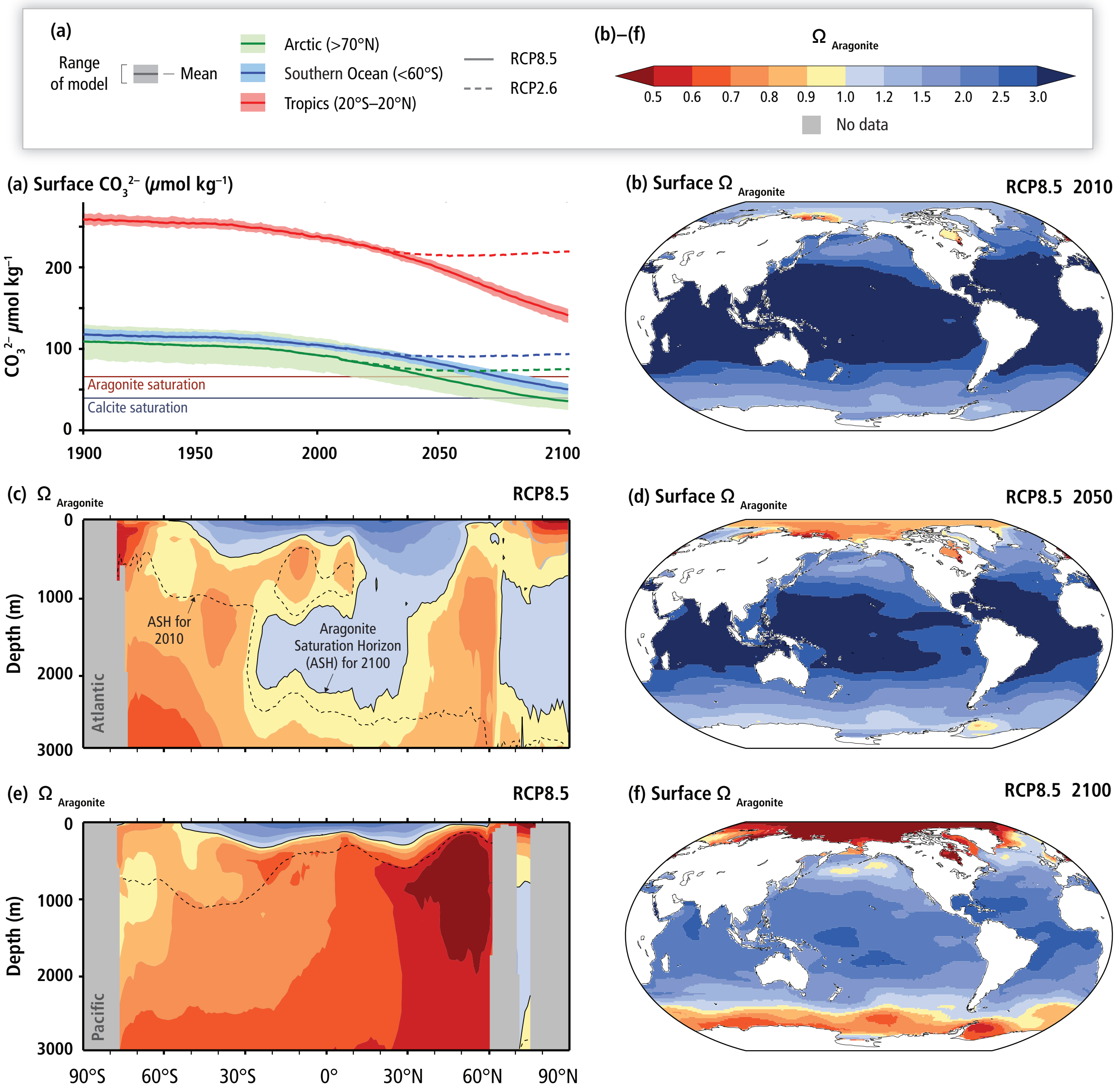


Figure 30-8 | Projected aragonite saturation state ($\Omega_{Aragonite}$) from 11 Coupled Model Intercomparison Project Phase 5 (CMIP5) Earth System Models under Representative Concentration Pathway 8.5 (RCP8.5) scenario. (a) Time series of surface carbonate ion (CO_3^{2-}) concentration shown as the mean (solid line) and range of models (shaded area), given as area-weighted averages over the Arctic Ocean (green), the tropical oceans (red), and the Southern Ocean (blue); maps of the median model's surface $\Omega_{Aragonite}$ in (b) 2010, (d) 2050, and (f) 2100; and zonal mean sections (latitude vs. depth) of $\Omega_{Aragonite}$ in 2100 over (c) the Atlantic Ocean and (e) the Pacific Ocean, while the ASH (Aragonite Saturation Horizon) is shown for 2010 (dotted line) and 2100 (solid line). Panel (a) also includes mean model results from RCP2.6 (dashed lines). As for Figure 30-7, gridded data products of carbonate system variables (Key et al., 2004) are used to correct each model for its present-day bias by subtracting the model-data difference at each grid cell following Orr et al. (2005). Where gridded data products are unavailable (Arctic Ocean, all marginal seas, and the Ocean near Indonesia), results are shown without bias correction. Figure originally presented in WGI AR5 Figure 6.29.

(936 ppm CO_2, +3.7°C, 56% reduction of carbonate ion concentration). The saturation horizons will also become significantly shallower in all oceans (with the aragonite saturation horizon between 0 and 1500 m in the Atlantic Ocean and 0 and 600 m (poles vs. equator) in the Pacific Ocean; Sabine et al., 2004; Orr et al., 2005; WGI AR5 Section 6.4.4; WGI AR5 Figure 6.28). Trends toward under-saturation of aragonite and calcite will also partly depend on ocean temperature, with surface polar waters expected to become seasonally under-saturated with respect to aragonite and calcite within a couple of decades (Figure 30-8c,d,e,f; Box CC-OA; McNeil and Matear, 2008).

Overall, observations from a wide range of laboratory, mesocosm, and field studies reveal that marine macro-organisms and ocean processes are sensitive to the levels of ocean acidification projected under elevated atmospheric CO_2 (*high confidence*; Box CC-OA, Section 6.3.2; Munday et al., 2009; Kroeker et al., 2013). Ecosystems that are characterized by high rates of calcium carbonate deposition (e.g., coral reefs, calcareous plankton communities) are sensitive to decreases in the saturation states of aragonite and calcite (*high confidence*). These changes are *very likely* to have broad consequences such as the loss of three-dimensional coral reef frameworks (Hoegh-Guldberg et al., 2007; Manzello et al., 2008; Fabricius et al., 2011; Andersson and Gledhill, 2013; Dove et al., 2013) and restructuring of food webs at relatively small (~50 ppm) additional increases in atmospheric CO_2. Projected shoaling of the aragonite and calcite saturation horizons are *likely* to impact deep water (100 to 2000 m) communities of scleractinian corals and other benthic organisms as atmospheric CO_2 increases (Orr et al., 2005; Guinotte et al., 2006; WGI AR5 Section 6.4.4), although studies from the Mediterranean and seamounts off southwest Australia report that some deep water corals may be less sensitive (Thresher et al., 2011; Maier et al., 2013). Organisms are also sensitive to changes in pH with respect to physiological processes such as respiration and neural function (Section 6.3.2). Owing to the relatively short history, yet growing effort, to understand the implications of rapid changes in pH and ocean carbonate chemistry, there are a growing number of organisms and processes reported to be sensitive. The impact of ocean acidification on marine organisms and ecosystems continues to raise serious scientific concern, especially given that the current rate of ocean acidification (at least 10 to 100 times faster than the recent glacial transitions (Caldeira and Wickett, 2003; Hoegh-Guldberg et al., 2007)) is unprecedented within the last 65 Ma (*high confidence*; Ridgwell and Schmidt, 2010) and possibly 300 Ma of Earth history (*medium confidence*; Hönisch et al., 2012; Section 6.1.2).

30.3.2.3. Oxygen Concentration

Dissolved O_2 is a major determinant of the distribution and abundance of marine organisms (Section 6.3.3). Oxygen concentrations vary across ocean basins and are lower in the eastern Pacific and Atlantic basins, and northern Indian Ocean (Figure 30-9b; Section 6.1.1.3). In contrast, some of the highest concentrations of O_2 are associated with cooler high-latitude waters (Figure 30-9b). There is *high agreement* among analyses providing *medium confidence* that O_2 concentrations have decreased in the upper layers of the Ocean since the 1960s, particularly in the equatorial Pacific and Atlantic Oceans (WGI AR5 Section 3.8.3; WGI AR5 Figure 3.20). A formal fingerprint analysis undertaken by Andrews et al. (2013) concluded that recent decreases in oceanic O_2 are due to external influences (*very likely*). Conversely, O_2 has increased in the North and South Pacific, North Atlantic, and Indian Oceans, and is consistent with greater mixing and ventilation due to strengthening wind systems (WGI AR5 Section 3.8.3). The reduction in O_2 concentration in some areas of the Ocean is consistent with that expected from higher ocean temperatures and a reduction in mixing (increasing stratification) (WGI AR5 Section 3.8.3). Analysis of ocean O_2 trends over time (Helm et al., 2011b) reveals that the decline in O_2 solubility with increased temperature is responsible for no more than 15% of the observed change. The remaining 85%, consequently, is associated with increased deep-sea microbial respiration and reduced O_2 supply due to increased ocean stratification (WGI AR5 Section 6.1.1.3). In coastal areas, eutrophication can lead to increased transport of organic carbon into adjacent ocean habitats where microbial metabolism is stimulated, resulting in a rapid drawdown of O_2 (Weeks et al., 2002; Rabalais et al., 2009; Bakun et al., 2010).

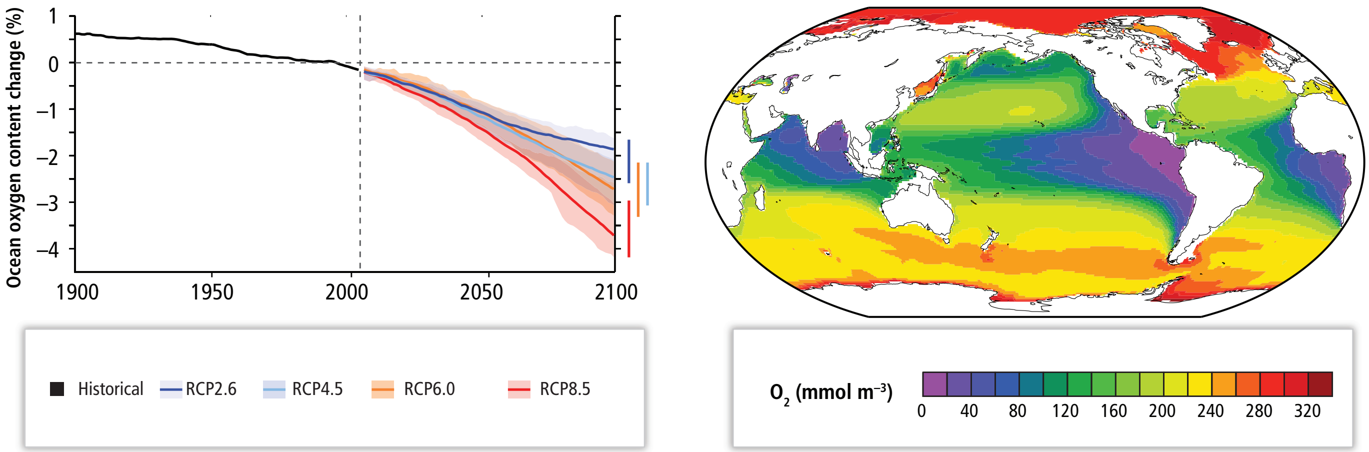
The development of hypoxic conditions (defined as O_2 concentrations below ~60 µmol kg^{-1}) over recent decades has been documented across

Frequently Asked Questions

FAQ 30.1 | Can we reverse the impacts of climate change on the ocean?

In less than 150 years, greenhouse gas (GHG) emissions have resulted in such major physical and chemical changes in our oceans that it will take thousands of years to reverse them. There are a number of reasons for this. Given its large mass and high heat capacity, the ability of the Ocean to absorb heat is 1000 times larger than that of the atmosphere. The Ocean has absorbed at least nine-tenths of the Earth's heat gain between 1971 and 2010. To reverse that heating, the warmer upper layers of the Ocean have to mix with the colder deeper layers. That mixing can take as much as 1000 years. This means it will take centuries to millennia for deep ocean temperatures to warm in response to today's surface conditions, and at least as long for ocean warming to reverse after atmospheric GHG concentrations decrease (*virtually certain*). But climate change-caused alteration of basic conditions in the Ocean is not just about temperature. The Ocean becomes more acidic as more carbon dioxide (CO_2) enters it and will take tens of thousands of years to reverse these profound changes to the carbonate chemistry of the ocean (*virtually certain*). These enormous physical and chemical changes are producing sweeping and profound changes in marine ecosystems. Large and abrupt changes to these ecosystems are unlikely to be reversible in the short to medium term (*high confidence*).

(a) Ocean oxygen content change (%) **(b) Oxygen concentrations in the 1990s (200–600 m)**



(c) 2090s, changes from 1990s (RCP2.6) **(d) 2090s, changes from 1990s (RCP8.5)**

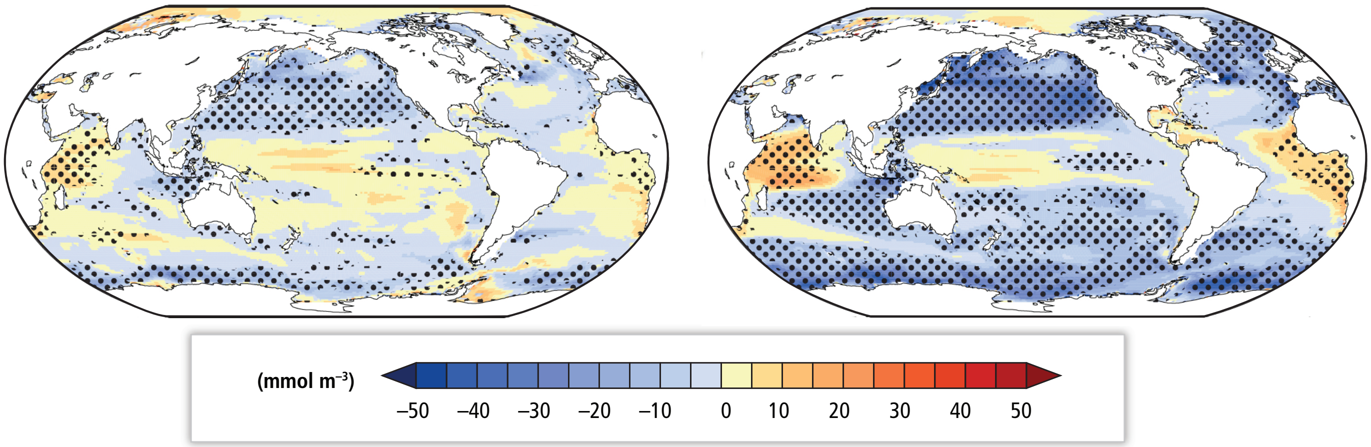


Figure 30-9 | (a) Simulated changes in dissolved O_2 (mean and model range as shading) relative to 1990s for Representative Concentration Pathway 2.6 (RCP2.6), RCP4.5, RCP6.0, and RCP8.5. (b) Multi-model mean dissolved O_2 (mmol m^{-3}) in the main thermocline (200 to 600 m depth average) for the 1990s, and changes in the 2090s relative to 1990s for RCP2.6 (c) and RCP8.5 (d). To indicate consistency in the sign of change, regions are stippled when at least 80% of models agree on the sign of the mean change. These diagnostics are detailed in Cocco et al. (2013) in a previous model intercomparison using the Special Report on Emission Scenarios (SRES)-A2 scenario and have been applied to Coupled Model Intercomparison Project Phase 5 (CMIP5) models here. Models used: Community Earth System Model 1–Biogeochemical (CESM1-BGC), Geophysical Fluid Dynamics Laboratory–Earth System Model 2G (GFDL-ESM2G), Geophysical Fluid Dynamics Laboratory–Earth System Model 2M (GFDL-ESM2M), Hadley Centre Global Environmental Model 2–Earth System (HadGEM2-ES), Institute Pierre Simon Laplace–Coupled Model 5A–Low Resolution (IPSL-CM5A-LR), Institute Pierre Simon Laplace–Coupled Model 5A–Medium Resolution (IPSL-CM5A-MR), Max Planck Institute–Earth System Model–Low Resolution (MPI-ESM-LR), Max Planck Institute–Earth System Model–Medium Resolution (MPI-ESM-MR), Norwegian Earth System Model 1 (Emissions capable) (NorESM1). Figure originally presented in WGI AR5 Figure 6.30.

a wide array of ocean sub-regions including some SES (e.g., Black and Baltic Seas), the Arabian Sea, and the California, Humboldt, and Benguela Current systems, where eruptions of hypoxic, sulfide-laden water have also occurred in some cases (Weeks et al., 2002). Localized, seasonal hypoxic "dead zones" have emerged in economically valuable coastal areas such as the Gulf of Mexico (Turner et al., 2008; Rabalais et al., 2010), the Baltic Sea (Conley et al., 2009), and the Black Sea (Kideys, 2002; Ukrainskii and Popov, 2009) in connection with nutrient fluxes from land. Over a vast region of the eastern Pacific stretching from southern Chile to the Aleutian Islands, the minimum O_2 threshold (less than 2 mg l^{-1} or ~60 µmol^{-1}) is found at 300 m depth and upwelling of increasingly hypoxic waters is well documented (Karstensen et al., 2008). Hypoxic waters in the northern Arabian Sea and Bay of Bengal are located close to continental shelf areas. Long-term measurements reveal that O_2 concentrations are declining in these waters, with *medium evidence* that economically significant mesopelagic fish populations are being threatened by a reduction in suitable habitat as respiratory stress increases (Koslow et al., 2011). It should be noted that hypoxia profiles based on a critical threshold of 60 µmol kg^{-1} can convey an overly simplistic message given that critical concentrations of O_2 in this regard are very much species, size, temperature, and life history stage specific. This variability in sensitivity is, however, a critical determinant for any attempt to understand how ecosystems will respond to changing future O_2 levels (Section 6.3.3).

There is *high agreement* among modeling studies that O_2 concentrations will continue to decrease in most parts of the Ocean due to the effect of temperature on O_2 solubility, microbial respiration rates, ocean ventilation, and ocean stratification (Figure 30-9c,d; WGI AR5 Table 6.14; Andrews et al., 2013), with implications for nutrient and carbon cycling, ocean productivity, marine habitats, and ecosystem structure (Section 6.3.5). The outcomes of these global changes are *very likely* to be influenced by regional differences in variables such as wind stress, coastal processes, and the supply of organic matter.

30.4. Global Patterns in the Response of Marine Organisms to Climate Change and Ocean Acidification

Given the close relationship between organisms and ecosystems with the physical and chemical elements of the environment, changes are expected in the distribution and abundance of marine organisms in response to ocean warming and acidification (Section 6.3; Boxes CC-MB, CC-OA). Our understanding of the relationship between ocean warming and acidification reveals that relatively small changes in temperature and other variables can result in often large biological responses that range from simple linear trends to more complex non-linear outcomes. There has been an increase in studies that focus on the influence and consequences of climate change for marine ecosystems since AR4 (Hoegh-Guldberg and Bruno, 2010; Poloczanska et al., 2013), representing an opportunity to examine, and potentially attribute, detected changes within the Ocean to climate change.

Evidence of global and regional responses of marine organisms to recent climate change has been shown through assessments of multiple studies focused on single species, populations, and ecosystems (Tasker, 2008; Thackeray et al., 2010; Przeslawski et al., 2012; Poloczanska et al., 2013). The most comprehensive assessment, in terms of geographic spread and number of observed responses, is that of Poloczanska et al. (2013). This study reveals a coherent pattern in observed responses of ocean life to recent climate change across regions and taxonomic groups, with 81% of responses by organisms and ecosystems being consistent with expected changes to recent climate change (*high confidence*; Box CC-MB). On average, spring events in the Ocean have advanced by 4.4 ± 0.7 days per decade (mean ± SE) and the leading edges of species' distributions have extended (generally poleward) by 72.0 ± 0.35 km per decade. Values were calculated from data series ranging from the 1920s to 2010, although all series included data after 1990. The fastest range shifts generally occurred in regions of high thermal velocity (the speed and direction at which isotherms move (Burrows et al., 2011; Section 30.3.1.1)). Subsequently, Pinsky et al. (2013), using a database of 360 fish and invertebrate species and species groups from coastal waters around North America, showed differences in the speed and directions that species shift can be explained by differences in local climate velocities (Box CC-MB).

30.5. Regional Impacts, Risks, and Vulnerabilities: Present and Future

This section explores the impacts, risks, and vulnerabilities of climate change for the seven sub-regions within the Ocean. There is considerable variability from region to region, especially in the extent and interaction of climate change and non-climate change stressors. Although the latter may complicate attribution attempts in many sub-regions, interactions between the two groups of stressors may also represent opportunities to reduce the overall effects on marine organisms and processes of the environmental changes being driven by climate change (including ocean acidification) (Crain et al., 2008; Griffith et al., 2012).

30.5.1. High-Latitude Spring Bloom Systems

High-Latitude Spring Bloom Systems (HLSBSs) stretch from 35°N to the edge of the winter sea ice (and from 35°S to the polar front) and provide 36% of world's fish catch (Figure 30-1b). Although much of the North Pacific is iron limited (Martin and Fitzwater, 1988) and lacks a classical spring bloom (McAllister et al., 1960), strong seasonal variability of primary productivity is pronounced at all high latitudes because of seasonally varying photoperiod and water column stability (Racault et al., 2012). Efficient transfer of marine primary and secondary production to higher trophic levels, including commercial fish species, is influenced by the magnitude as well as the spatial and temporal synchrony between successive trophic production peaks (Hjort, 1914; Cushing, 1990; Beaugrand et al., 2003; Beaugrand and Reid, 2003).

Frequently Asked Questions

FAQ 30.2 | Does slower warming in the Ocean mean less impact on plants and animals?

The greater thermal inertia of the Ocean means that temperature anomalies and extremes are lower than those seen on land. This does not necessarily mean that impacts of ocean warming are less for the ocean than for land. A large body of evidence reveals that small amounts of warming in the Ocean can have large effects on ocean ecosystems. For example, relatively small increases in sea temperature (as little as 1°C to 2°C) can cause mass coral bleaching and mortality across hundreds of square kilometers of coral reef (*high confidence*). Other analyses have revealed that increased temperatures are spreading rapidly across the world's oceans (measured as the movement of bands of equal water temperature or isotherms). This rate of warming presents challenges to organisms and ecosystems as they try to migrate to cooler regions as the Ocean continues to warm. Rapid environmental change also poses steep challenges to evolutionary processes, especially where relatively long-lived organisms such as corals and fish are concerned (*high confidence*).

30.5.1.1. Observed Changes and Potential Impacts

30.5.1.1.1. North Atlantic

The average temperature of the surface waters of the North Atlantic HLSBS has warmed by 0.07°C per decade, resulting in an increase in sea temperature of 0.44°C between 1950 and 2009 (*likely*) (p-value = 0.15; Table 30-1). Over the same period, both winter and summer temperatures have increased significantly (0.05°C per decade and 0.12°C per decade respectively, p-value $\leq$ 0.05). Since the 1970s, the Atlantic Ocean has warmed more than any other ocean basin (0.3°C per decade; Figure 30-2a; WGI AR5 Section 3.2.2), with greatest warming rates over European continental shelf areas such as the southern North Sea, the Gulf Stream front, the sub-polar gyres, and the Labrador Sea (MacKenzie and Schiedek, 2007a,b; Levitus et al., 2009; Lee et al., 2011; González-Taboada and Anadón, 2012). Basin-wide warming in the North Atlantic since the mid-1990s has been driven by global warming and the current warm phase of the Atlantic Multi-decadal Oscillation (AMO) (Wang and Dong, 2010; WGI AR5 Section 14.7.6).

The North Atlantic is one of the most intensively fished ocean sub-regions. The major areas for harvesting marine living resources span the eastern North American, European, and Icelandic shelves (Livingston and Tjelmeland, 2000). In addition, the deep regions of the Nordic Seas and the Irminger Sea contain large populations of pelagic fish such as herring, blue whiting, and mackerel and mesopelagic fish such as pearlsides and redfish. The region covers a wide latitudinal range from 35°N to 80°N and, hence, a large span in thermal habitats. This is reflected in the latitudinal gradient from subtropical/temperate species along the southern fringe to boreal/arctic species along the northern fringe.

Climate change is *virtually certain* to drive major changes to the northern fringes of the Atlantic HLSBS by 2100. For the Barents Sea region, which borders the HLSBS and Arctic regions, modeling projections from 1995 to 2060 (SRES B2 scenario) gave an increase in phytoplankton production of 8%, an increase in Atlantic zooplankton production of 20%, and a decrease of Arctic zooplankton production of 50% (Ellingsen et al., 2008). These changes result in a total increase in zooplankton production in the HLSBS section of the Barents Sea and a decrease in the Arctic section. Together with poleward shifts of fish species, a substantial increase in fish biomass and catch is also *very likely* at the northern fringes of the HLSBS (Cheung et al., 2011). However, for some species such as capelin, which feeds in summer at the ice edge and spawns in spring at the southern Atlantic Norwegian/Murman coast of the Barents Sea, the continuous temperature increase is *very likely* to cause discontinuous changes in conditions. The limited migration potential for this small pelagic fish is also *likely* to drive an eastward shift in spawning areas to new spawning grounds along the Novaja Semlja coast (Huse and Ellingsen, 2008).

Observations of fish and other species moving to higher latitudes (Beare et al., 2005; Perry et al., 2005; Collie et al., 2008; Lucey and Nye, 2010) within the North Atlantic HLSBS are consistent with results of modeling exercises (Stenevik and Sundby, 2007; Cheung et al., 2011). Examples from the Barents (Section 28.2.2.1), Nordic, and North Seas (Box 6-1; Section 23.4.6) show how warming from the early 1980s influenced North Atlantic ecosystems, where substantial biological impacts such as large-scale modification of the phenology, abundance, and distribution of plankton assemblages and reorganization of fish assemblages have been observed (Beaugrand et al., 2002; Edwards, 2004; Edwards and Richardson, 2004; Tasker, 2008; Nye et al., 2009; Head and Pepin, 2010; Simpson et al., 2011). The ranges of some cold-water zooplankton assemblages in the northeast Atlantic have contracted towards the Arctic since 1958, and have been replaced by warm-water zooplankton assemblages (specifically copepods) (*high confidence*), which moved up to 1000 km northward (Beaugrand et al., 2002; Beaugrand, 2009). Although changes to surface circulation may have played a role (Reid et al., 2001), the primary driver of the shift was shown to be regional warming (Beaugrand et al., 2002; Beaugrand, 2004). Reorganization of zooplankton communities and an observed decline in mean size has implications for energy transfer to higher trophic levels including commercial fish stocks (Beaugrand et al., 2003; Kirby and Beaugrand, 2009; Lindley et al., 2010; Section 23.4.6). Warm-water species of fish have increased in abundance on both sides of the North Atlantic (*medium confidence*; Beare et al., 2005; Collie et al., 2008; Genner et al., 2010; Hermant et al., 2010; Lucey and Nye, 2010; Simpson et al., 2011). The diversity of zooplankton and fish has increased as more diverse warm-water assemblages extend northward in response to changing environmental conditions (*high confidence*; Kane, 2007; Hiddink and ter Hofstede, 2008; Beaugrand, 2009; Mountain and Kane, 2010; ter Hofstede et al., 2010; Box 6-1; Section 23.4.6).

The past decade has been the warmest decade ever recorded in the Barents Sea, resulting in large populations of krill shrimp and pelagic and demersal fish stocks linked to the Atlantic and boreal ecosystem of the Barents Sea (*high confidence*; Johannesen et al., 2012; Section 28.2.2.1). Recruitment to boreal fish stocks such as cod, haddock, and herring has increased (Eriksen et al., 2012). The relatively warm Atlantic waters have advanced northward and eastward (Årthun et al., 2012) and sea ice has retreated along with the Arctic water masses. As a result, boreal euphausiids, which are mainly confined to Atlantic water, have increased in biomass and distribution (Dalpadado et al., 2012), enhancing growth of young cod *Gadus morhua* (boreal) as well as the more Arctic (arcto-boreal) capelin (*Mallotus villosus*). The abundance of amphipods of more Arctic origin has decreased, resulting in poorer feeding conditions for polar zooplankton predators such as polar cod (*Boreogadus saida*). Blue whiting (*Micromesistius poutassou*), which spawns west of the British Isles and feeds on zooplankton in the Norwegian Sea during the summer, extended their summer feeding distribution into the Barents Sea during the recent warm period.

The Norwegian Sea is one of the two core regions for the herbivore copepod *Calanus finmarchicus*, an important prey species for pelagic fish and early life stages of all fish around the rim of this high-latitude sea including the North Sea and the Barents Sea (Sundby, 2000). *C. finmarchicus* is the main food item for some of the world's largest fish stocks such as the Norwegian spring-spawning herring (*Clupea harengus*), blue whiting (*M. poutassou*), and northeast Atlantic mackerel (*Scomber scombrus*). These stocks have increased considerably during the recent warming that started in the early 1980s (Huse et al., 2012). The individual size of herring has also increased, enabling longer feeding migrations to utilize boreal zooplankton occurring closer to distant Arctic water masses. Mackerel (*Scomber scombrus*) has advanced northward and westward into Icelandic waters (Astthorsson et al., 2012) and was even observed in East Greenland water in summer 2013 (Nøttestad et al., 2013). Since 2004, the sum of spawning stock biomass of the three pelagic fish species (herring, blue whiting, and mackerel) leveled out at around 16 million tonnes.

Observed changes in the phenology of plankton groups in the North Sea over the past 50 years are driven by climate forcing, in particular regional warming (*high confidence*; Edwards and Richardson, 2004; Wiltshire and Manly, 2004; Wiltshire et al., 2008; Lindley et al., 2010; Lindley and Kirby, 2010; Schluter et al., 2010), although responses are species-specific with substantial variation within functional groups (Edwards and Richardson, 2004; Box 6-1). For example, the peak maximum abundance of the copepod *C. finmarchicus* advanced by 10 days from the 1960s to the 2000s, but its warm-water equivalent, *C. helgolandicus*, did not advance (Bonnet et al., 2005). In the North Sea, bottom temperatures in winter have warmed by 1.6°C (1980–2004; Dulvy et al., 2008). The whole demersal fish community shifted deeper by 3.6 m per decade over the period 1980–2004, although mean latitude of the whole community did not show net displacement (Dulvy et al., 2008). Within the community, cool-water specialists generally shifted northward while abundant warm-water species shifted southward, reflecting winter warming of the southern North Sea. The cold winter temperatures of the shallow regions of the southern North Sea have acted to exclude species with warm-water affinities. Trawl survey data from the rapidly warming southern North Sea suggests waves of immigration by southern species such as red mullet (*Mullus surmuletus*), anchovy (*Engraulis encrasicholus*), and sardines (*Sardina pilchardus*), linked to increasing population sizes and warming temperatures (Beare et al., 2004, 2005).

In the northeast Atlantic, range expansions and contractions linked to changing climate have also been observed in benthic crustaceans, bivalves, gastropods, and polychaetes (*medium confidence*; Mieszkowska et al., 2007; Beukema et al., 2009; Berke et al., 2010). For example, the southern range limit of the common intertidal barnacle, *Semibalanus balanoides*, contracted northward along European coastlines at a rate of 15 to 50 km per decade since 1872, and its retreat is attributed to reproductive failure as winter temperatures warm (Southward et al., 2005; Wethey and Woodin, 2008). *Chthamalus montagui*, its warm-water competitor, increased in abundance to occupy the niche vacated by *S. balanoides* (*high confidence*; Southward et al., 1995; Poloczanska et al., 2008).

Many of the longest and most comprehensive time series used to investigate the ecological consequences of climate fluctuations and fishing, that span periods of cooling and warming over the past century, are from the northeast Atlantic (Toresen and Østvedt, 2000; Southward et al., 2005; Sundby and Nakken, 2008; Edwards et al., 2010; Poloczanska et al., 2013). Meta-analysis of 288 long-term data sets (spanning up to 90 years) of zooplankton, benthic invertebrates, fish, and seabirds from the OSPAR Commission Maritime Area in the North-east Atlantic showed widespread changes in distribution, abundance, and seasonality that were consistent (77%) with expectations from enhanced greenhouse warming (Tasker, 2008). The study brought together evidence of changes in ocean climate and ecological responses across a range of species that encompassed both exploited and unexploited species from a variety of information types including peer-reviewed reports from International Council for the Exploration of the Sea (ICES) Working Groups. In particular, observations indicated poleward shifts in zooplankton communities, increasing abundance of fish species in the northern part of their ranges and decreases in southern parts, and the expansion of benthic species into more northerly or less coastal areas (*high confidence*).

The major portion of the literature on the influence of climate change on the North Atlantic region covers time spans that are longer than for most other sub-regions of the Ocean. Even here, however, the bulk of the literature is limited to the last 30 to 50 years. The few publications covering the first half of the 20th century represent an important longer term perspective on the influence of climate change (Toresen and Østvedt, 2000; Drinkwater, 2006; Sundby and Nakken, 2008; Bañón, 2009; Astthorsson et al., 2012). For example, distinct changes in fauna were associated with a pronounced warming period over 1920–1940 (Wood and Overland, 2010), when fish and other fauna shifted northward (Iversen, 1934; Southward et al., 2005; Drinkwater, 2006; Hátún et al., 2009). The major lesson from these reports is that a rapid large-scale temperature increase occurred in the high-latitude North Atlantic between the 1920s and 1940s, with basin-scale consequences for marine ecosystems that are comparable to warming and observed impacts over the last 30 years. The former event was of great concern within the scientific community, particularly during the late 1940s and early 1950s (Iversen, 1934; Tåning, 1949, 1953; Southward, 1980). However, with the subsequent long-term cooling in the 1970s, discussion around climate responses was discontinued (Southward, 1980). The centennial-long perspective indicates that multi-decadal variability has played a major role in changes observed over the past 30 years. The 150-year instrumental record shows distinct warm phases of the AMO during approximately 1930–1965 and from 1995, and cool phases between approximately 1900–1930 and 1960–1995 (WGI AR5 Section 14.7.6). However, it is *virtually certain* that the enhanced warming in recent decades cannot be explained without external forcing (WGI AR5 Section 10.3.1.1.3). Understanding the changes in inter-decadal variability over the next century is particularly important. The current warm phase of the AMO is *likely* to terminate in the next few decades, leading to a cooling influence in the North Atlantic and potentially offsetting some of the effects of global warming (WGI AR5 Sections 11.3.2.4.1, 14.7.6). Over the transition period, the climate of the North Atlantic is *likely* to change more rapidly than during previous transitions since 1900.

30.5.1.1.2. North Pacific

Sub-decadal variability in the North Pacific HLSBS is dominated by ENSO (Trenberth, 1990; WGI AR5 Section 14.4). Unlike the North Atlantic HLSBS, the North Pacific HLSBS does not show any significant trends in temperature over time, *very likely* as a consequence of climate variability influences on long-term warming patterns (1950–2009; Table 30-1). Decadal and longer periods of variability in the North Pacific are reflected in the principal mode, the Pacific Decadal Oscillation (PDO; WGI AR5 Section 14.7.3), with periodicities in SST of both 15 to 25 years and 50 to 70 years (Minobe, 1997; Mantua and Hare, 2002). Further modes of climate variability include the North Pacific Gyre Oscillation (NPGO; Di Lorenzo et al., 2008; Chhak et al., 2009). The PDO exhibits SST anomalies of one sign along the eastern boundary and the opposite sign in western and central Pacific. The PDO has been reported to have an anthropogenic component (Bonfils and Santer, 2011) but confidence in this is *very low* (*limited evidence*, *low agreement*; WGI AR5 Section 10.3.3). The interplay of the phases of these modes of variability has strong influence on high-latitude Pacific ecosystems (*very high confidence*). In the space of 3 years, the eastern North Pacific fluctuated from one of the warmest years in the past century (2005) to one of the coldest (2008) (McKinnell et al., 2010; McKinnell and Dagg, 2010). This rapid change was accompanied by large changes in primary productivity, zooplankton communities, and fish and seabird populations (McKinnell et al., 2010; McKinnell and Dagg, 2010; Batten and Walne, 2011; Bi et al., 2011; Keister et al., 2011).

Climate transitions among phases of variability tend to be characterized by abrupt reorganization of the ecosystems as dynamic trophic relationships among species alter (Hunt et al., 2002; Peterson and Schwing, 2003; Litzow and Ciannelli, 2007; Litzow et al., 2008; Alheit, 2009). Periods of broad-scale environmental change were observed across high-latitude ecosystems in the North Pacific HLSBS (eastern Bering Sea and Gulf of Alaska) during 1976–1978, 1987–1989, and 1998–1999. These periods were associated with regime shifts in foraging fish that occurred in 1979–1982, 1988–1992, and 1998–2001. The changes indicate how basin-scale variability such as the PDO can manifest across distinct ecosystems (Overland et al., 2008; Link et al., 2009a,b). Phenological shifts observed in the zooplankton communities of the North Pacific were *very likely* in response to decadal climate variability, with distinct changes noted after the climate shifts of the 1970s and 1990s (Mackas et al., 1998; Peterson and Schwing, 2003; Chiba et al., 2006). Modeling evidence suggests a weak shift in PDO toward more occurrences of the negative phase but the credibility of projections remains uncertain (WGI AR5 Section 14.7.3). It is *about as likely as not* that the PDO will change its form or behavior in the future (WGI AR5 Section 14.7.3).

The Kuroshio-Oyashio Extension (KOE) in the northwest Pacific displays pronounced decadal-scale variability (Yatsu et al., 2008; Sugisaki et al., 2010). "Warm periods" in the mid-1970s and late 1980s were accompanied by dramatic changes in pelagic ecosystems and sardine and anchovy stocks (Chiba et al., 2008; Yatsu et al., 2008). Observations and climate model simulations indicate that global warming is *likely* to further alter the dynamics of the Kuroshio Current and the KOE over the coming century (McPhaden and Zhang, 2002; Sakamoto et al., 2005; Wu et al., 2012; Zhang et al., 2014). Alteration of the KOE will alter the timing, magnitude, and structure of spring blooms in the western Pacific and have implications for pelagic fish recruitment, production, and biogeochemical cycles (Ito et al., 2004; Hashioka et al., 2009; Yatsu et al., 2013).

Commercial catches of salmon species in the North Pacific HLSBS follow decadal fluctuations in climate (Hare and Mantua, 2000; Mantua and Hare, 2002). Catches peaked in the warm periods of the 1930s–1940s and 1990s–2000s, with 2009 yielding the highest catch to date, and warming trends are *about as likely as not* to have contributed to recent peaks in some sub-regions (Morita et al., 2006; Irvine and Fukuwaka, 2011). Poleward range shifts of some large pelagic fish in the western North Pacific, such as yellowtail *Seriola quinqueradiata* and Spanish mackerel *Scomberomorus niphonius*, were attributed, in part, to regional warming (*high confidence*) and these two species are projected to shift 39 to 71 km poleward from the 2000s to 2030s under SRES A1B (Tian et al., 2012; Jung et al., 2014). Anticipating ecological responses to future anthropogenic climate change also requires evaluation of the role of changes to climate beyond warming per se. For example, declining sea level pressure in the North Pacific is *likely* influenced by anthropogenic forcing (Gillett et al., 2003; Gillett and Stott, 2009; WGI AR5 Section 10.3.3.4) and sea level pressure in turn is related to atmospheric climate parameters (e.g., turbulent mixing via wind stress) that regulate commercially significant fish populations (Wilderbuer et al., 2002).

The northern fringe of the Bering Sea is among the most productive of marine sub-regions and includes the world's largest single-species fishery, walleye pollock (*Theragra chalcogramma*; Hunt et al., 2010). This region underwent major changes in recent decades as a result of climate variability, climate change, and fishing impacts (Litzow et al., 2008; Mueter and Litzow, 2008; Jin et al., 2009; Hunt et al., 2010; Section 28.2.2.1). Seasonal sea ice cover declined since the 1990s (to 2006), although there is no linear trend between 1953 and 2006, and the initiation of spring ice retreat over the southeastern Bering Sea shelf started to occur earlier (Wang et al., 2007a). Concurrent with the retreat of the "cold pool," an area of reduced water temperature (<2°C) on the northern Bering Sea shelf that is formed as a consequence of sea ice and is maintained over summer (Hunt et al., 2010), bottom trawl surveys of fish and invertebrates show a significant community-wide northward distribution shift and a colonization of the former cold pool areas by sub-Arctic fauna (*high confidence*; Wang et al., 2006a; Mueter and Litzow, 2008).

Over a vast region of the eastern Pacific stretching from southern Chile to the Aleutian Islands, waters low in dissolved O_2 (Oxygen Minimum Zone (OMZ)) are found at 300 m depth (Karstensen et al., 2008). Sporadic upwelling of these low-O_2 waters along the continental shelf is well documented, where biological respiration can further reduce dissolved O_2 levels and result in hypoxic or anoxic conditions that lead to mortality of coastal fishes and invertebrates (Grantham et al., 2004; Chan et al., 2008). The magnitude and severity of seasonal hypoxic conditions in shallow-shelf waters of the eastern North Pacific HLSBS increased in recent decades (Bograd et al., 2008; Chan et al., 2008). In addition, minimum pH values in the water column usually occur near the depths of the OMZ (WGI AR5 Box 3.2). A shoaling of the aragonite saturation horizon has *likely* resulted in low-aragonite conditions within the density layers being upwelled on the shelf of the west coast of the USA, increasing the risk of seasonally upwelled water being relatively acidified (Feely et al., 2008) with observed impacts on Pacific oyster (*Crassostrea gigas*) hatcheries (Barton et al., 2012). In the time period 1991–2006, reductions in pH in the North Pacific between 800 and ~100 m were attributed in approximately equal measure to anthropogenic and natural variations (Byrne et al., 2010; WGI AR5 Section 3.8.2; WGI AR5 Figure 3.19).

30.5.1.1.3. Southern Hemisphere

The seasonal peaks in phytoplankton productivity in the Southern Hemisphere are much less pronounced and are of smaller magnitude than those at Northern Hemisphere high latitudes (Yoder et al., 1993). The Southern Hemisphere HLSBS is broadly bounded by the subtropical front and the sub-Antarctic front. Associated with the subtropical front is intense biological activity of bloom-forming coccolithophores (phytoplankton) (Brown and Yoder, 1994). The calcifying plankton assemblages play a key role in carbon cycles in the region and the transport of carbon to deep ocean sediments. The coccolithophore, *Emiliania huxleyi*, extended its range south of 60° in the southwest Pacific (141°E to 145°E) over the 2 decades since 1983 (Cubillos et al., 2007). Although the drivers for this range extension are not clear, it was proposed that the extension is facilitated by surface warming or changes in the abundance of grazing zooplankton.

Large regions of the sub-Antarctic surface waters are *likely* to become undersaturated with respect to aragonite during winter by 2030, which will impact calcifying plankton and Southern Ocean ecosystems (McNeil and Matear, 2008; Bednaršek et al., 2012; Section 28.2.2.2). Shell weights of the modern foraminifer, *Globigerina bulloides*, in the sediments of the sub-Antarctic region of the HLSBS south of Australia were observed to be 30 to 35% lower than those from sediment cores representing preindustrial periods, consistent with a recent decline in pH (Moy et al., 2009). Examination of the pteropod, *Limacina helicina antarctica*, captured from polar waters further south shows severe levels of shell dissolution consistent with the shoaling of the aragonite saturation horizon and indicates that the impact of ocean acidification is already occurring (Bednaršek et al., 2012).

While the South Pacific HLSBS has not shown warming overall, both the warmest and coolest months show a slight, but significant, increase over time (both 0.05°C per decade from 1950 to 2009, p-value ≤ 0.05; Table 30-1), although some areas within this sub-region have warmed. For example, the western Tasman Sea has shown enhanced warming since 1900 as compared to average global trends (*high confidence*). This has been driven by changes in large-scale wind-forcing leading to a southward expansion of the South Pacific STG and intensification of the southward-flowing East Australian Current (EAC; Cai, 2006; Hill et al., 2008; Wu et al., 2012; WGI AR5 Section 3.6.2). Model simulations suggest both stratospheric ozone depletion and greenhouse forcing contribute to the observed trend in wind stress (Cai and Cowan, 2007). Coinciding with this warming and intensified EAC is the observation that a number of benthic invertebrates, fish, and zooplankton are now found further south than they were in the mid-20th century (Ling, 2008; Pitt et al., 2010; Last et al., 2011). Warming facilitated the establishment of the grazing urchin, *Centrostephanus rodgersii*, in eastern Tasmania during the late 1970s (*high confidence*), which has resulted in deleterious effects on macroalgal beds (Ling, 2008; Ling et al., 2008, 2009; Banks et al., 2010).

30.5.1.2. Key Risks and Vulnerabilities

Projected changes to the temperature of surface waters match those of the past 50 years, with average sea temperatures in the HLSBS regions projected to increase by 0.35°C to 1.17°C in the near term (2010–2039) and by 1.70°C to 4.84°C over the long term (2010–2099) under the "business as usual" (BAU) RCP8.5 scenario (Table SM30-4). Under the lower case scenario considered here (RCP2.6), projected rates of regional warming are much lower (0.12°C to 0.79°C) in the near term, with slight cooling for some regions in the long term (–0.16°C to 1.46°C). Risks to HLSBS from warming of surface waters include changes to primary production and carbon cycling, and the reorganization of ecosystems in response to warmer and more acidified oceans. Both primary production and the timing of the spring bloom in HLSBS are very sensitive to environmental change. Latitudinal shifts in the distribution of phyto- and zooplankton communities will alter seasonality, community composition, and bloom dynamics (Beaugrand, 2009; Ito et al., 2010; Shoji et al., 2011). Alteration of the structure and composition of plankton communities can propagate through high-latitude food webs due to tight trophic linkages (Edwards and Richardson, 2004; Beaugrand et al., 2010; Beaugrand and Kirby, 2010). Mechanisms are complex, and tend to be non-linear, with impacts on ecosystems, fisheries, and biogeochemical cycles being hard to project with any certainty (Box CC-PP). A reorganization of commercial fish stocks, with attendant social and economic disruption, is a key risk of ongoing climate change in HLSBS sub-regions. AR4 reported that the productivity of some marine fisheries is *likely* to increase in the North Atlantic (WGII AR4 Sections 10.4.1, 12.4.7). A large number of publications since then has substantially extended documentation of these trends and has begun to elucidate the nuances in how marine ecosystems and organisms respond (Sumaila et al., 2011).

An additional risk exists for sub-polar areas from the loss of seasonal sea ice. Decreases in seasonal sea ice are *very likely* to lead to increases in the length of the growth season and the intensity of the light available to fuel phytoplankton growth and, hence, enhance primary production and attending modifications of ecosystem structure (Arrigo et al., 2008). In the long term, however, primary production may decrease due to the reduced supply of nutrients to the surface layers (Box CC-PP). The decline in Arctic sea ice will open ecological dispersal pathways, as well as new shipping routes (Section 30.6.2.3), between the North Atlantic and the North Pacific; large numbers of the Pacific diatom, *Neodenticula seminae*, were found in the North Atlantic in 1999 (Reid et al., 2007).

HLSBSs are also vulnerable to rapid changes in the carbonate chemistry of ocean waters. Ocean acidification will produce additional and large-scale challenges. There is *medium agreement* that calcifying organisms in these regions will be negatively affected by ocean acidification, with substantial impacts on higher trophic levels, although there is *limited evidence* at this point.

30.5.2. Equatorial Upwelling Systems

The largest upwelling systems are found in the equatorial regions of the eastern Pacific and Atlantic Oceans (Figure 30-1a). Equatorial Upwelling Systems (EUS) produce highly productive "cold tongues" that stretch westward across equatorial areas, which is different from other upwelling systems (e.g., EBUE; Section 30.5.5). The associated upwelling is a consequence of the Earth's rotation and easterly (westward) winds and currents, which drive water northward and southward at the northern and southern edges of these sub-regions. As result, cold, nutrient-rich, and high CO_2/low pH waters are transported from the deeper layers of the Ocean to the surface, driving high levels of primary productivity that support 4.7% of total global fisheries productivity (Table SM30-1; Figure 30-1b). Interannual modes of variability (e.g., ENSO; WGI AR5 Section 14.4) dominate EUS, particularly in the Pacific (Barber et al., 1994; McCarthy et al., 1996; Signorini et al., 1999; Le Borgne et al., 2002;

Christian and Murtugudde, 2003; Mestas-Nuñez and Miller, 2006; Pennington et al., 2006; Wang et al., 2006b). Upwelling of the Pacific EUS declines during El Niño events, when the trade winds weaken, or even reverse, and is strengthened during La Niña events. ENSO periodicity controls primary productivity and consequently has a strong influence over associated fisheries production (Mestas-Nuñez and Miller, 2006). The Intertropical Convergence Zone (ITCZ; WGI AR5 Section 14.3.1.1), an important determinant of regional ocean temperature, is located at the edges of the Indian and Pacific equatorial upwelling zone and influences a range of variables including productivity, fisheries, and precipitation. The EUS are also affected by inter-decadal variability (e.g., Inter-decadal Pacific Oscillation (IPO); Power et al., 1999; WGI AR5 Section 14.3).

30.5.2.1. Observed Changes and Potential Impacts

The average sea temperature associated with the EUS has increased significantly (p-value ≤ 0.05), by 0.43°C and 0.54°C from 1950 to 2009 in the Pacific and Atlantic EUS, respectively (Table 30-1). In the Pacific, regional variability in SST trends is driven by the temporal patterns in ENSO and the more frequent El Niño Modoki or Central Pacific El Niño events in recent decades (*high confidence*; Ashok et al., 2007; Yu and Kao, 2007; Lee and McPhaden, 2010; WGI AR5 Section 14.2.4.4). The faster warming of the Atlantic EUS is *likely* to be associated with a weakening of upwelling (Tokinaga and Xie, 2011). SLR in the eastern equatorial Pacific has been decreasing by up to -10 mm yr^{-1} since 1993 (Church et al., 2006; Figure 30-5).

Coral reefs in the EUS of the eastern Pacific (e.g., Galápagos and Cocos Islands) have relatively low species diversity and poorly developed carbonate reef frameworks, due to the low pH and aragonite saturation of upwelling waters (*high confidence*; Glynn, 2001; Manzello et al., 2008; Manzello, 2010). Prolonged periods of elevated temperature associated with El Niño have negatively affected corals, kelps, and associated organisms, and resulted in several possible local extinctions (*high confidence*; Glynn, 2011). Since 1985, coral reefs from west of South America to the Gilbert Islands of Kiribati have experienced the highest levels of thermal stress relative to other areas (Donner et al., 2010). In 1982/1983, mass coral bleaching and mortality affected most of the reef systems within the eastern equatorial Pacific (Glynn, 1984; Baker et al., 2008). Subsequent canonical El Niño and Central Pacific El Niño events in 1997/1998, 2002/2003, 2004/2005, and 2009/2010 (WGI AR5 Section 14.4.2; WGI AR5 Figure 14.13) triggered mass coral bleaching by adding to the background increases in sea temperatures (*high confidence*; Donner et al., 2010; Obura and Mangubhai, 2011; Vargas-Ángel et al., 2011). In some locations, impacts of El Niño have also interacted with other anthropogenic changes, such as those arising from changes to fishing pressure (Edgar et al., 2010), further complicating the attribution of recent ecological changes to climate change.

30.5.2.2. Key Risks and Vulnerabilities

Climate models indicate that ENSO is *virtually certain* to continue to be a major driver of oceanic variability over the coming century, although not all models can accurately replicate its behavior (WGI AR5 Section 9.5.3). Superposition of a warming ocean on future ENSO activity (possibly modified in frequency and intensity) is *likely* to result in oceanic conditions that are different from those experienced during past El Niño and La Niña events (Power and Smith, 2007). Temperatures within EUS sub-regions are projected to continue to warm significantly (p-value $\leq$ 0.05). Under RCP8.5, SST of the Atlantic EUS is projected to increase by 0.81°C over 2010–2039 and 2.56°C over 2010–2099, with similar increases projected for the Pacific EUS (Table SM30-4). Differences between RCPs for the two EUS become clear beyond mid-century, with warming of SST over 2010–2099 being 0.43°C and 0.46°C under RCP2.6, and 3.01°C and 3.03°C under RCP8.5, for Pacific and Atlantic EUS respectively (Table SM30-4). These projected increases in sea temperature will increase heat stress and ultimately irreversibly degrade marine ecosystems such as coral reefs (*very likely*). Further increases in atmospheric CO_2 will cause additional decrease in pH and aragonite saturation of surface waters (adding to the low pH and aragonite saturation of upwelling conditions), with significant differences between emission trajectories by the middle of the century. These changes in ocean carbonate chemistry are *very likely* to negatively affect some

Frequently Asked Questions

FAQ 30.3 | How will marine primary productivity change with ocean warming and acidification?

Drifting microscopic plants known as phytoplankton are the dominant marine primary producers at the base of the marine food chain. Their photosynthetic activity is critically important to life in general. It provides oxygen, supports marine food webs, and influences global biogeochemical cycles. Changes in marine primary productivity in response to climate change remain the single biggest uncertainty in predicting the magnitude and direction of future changes in fisheries and marine ecosystems (*low confidence*). Changes have been reported to a range of different ocean systems (e.g., High-Latitude Spring Bloom Systems, Subtropical Gyre Systems, Equatorial Upwelling Systems, and Eastern Boundary Upwelling Ecosystems), some of which are consistent with changes in ocean temperature, mixing, and circulation. However, direct attribution of these changes to climate change is made difficult by long-term patterns of variability that influence productivity of different parts of the Ocean (e.g., Pacific Decadal Oscillation). Given the importance of this question for ocean ecosystems and fisheries, longer time series studies for understanding how these systems are changing as a result of climate change are a priority (*high agreement*).

marine calcifiers, although many of the species from this region are adapted to the low aragonite and calcite saturation states that result from equatorial upwelling, albeit with much lower rates of calcification (Manzello, 2010; Friedrich et al., 2012). A substantial risk exists with respect to the synergistic interactions between sea temperature and declining pH, especially as to how they influence a large number of key biological processes (Box CC-OA).

There is *low confidence* in the current understanding of how (or if) climate change will influence the behavior of ENSO and other long-term climate patterns (Collins et al., 2010; WGI AR5 Section 12.4.4.2). There is also low agreement between different CMIP5 General Circulation Models (GCMs) on how ocean warming will affect ENSO, with no significant change to ENSO amplitude in half of the models examined, and both increasing and decreasing activity in others (Guilyardi et al., 2012). These differences appear to be a consequence of the delicate balance within ENSO between dampening and amplifying feedbacks, and the different emphasis given to these processes within the different GCMs (Collins et al., 2010). Other studies have looked at the interaction between the STG and EUS, and warming of surface waters in the Pacific, with at least one study projecting the possible expansion of the STG at the expense of the EUS (Polovina et al., 2011). In the latter case, the area of equatorial upwelling within the North Pacific would decrease by 28%, and primary production and fish catch by 15%, by 2100. Many of the projected changes imply additional consequences for pelagic fisheries resulting from the migration of fish stocks deriving from changing distribution of particular sea temperatures (Lehodey et al., 2006, 2008, 2011; Cheung et al., 2010; Sumaila et al., 2011; Bell et al., 2013b). These projections suggest that fisheries within EUS will experience increased vulnerability as a result of climate change (*low confidence*).

30.5.3. Semi-Enclosed Seas

Semi-Enclosed Seas (SES) represent a subset of ocean sub-regions that are largely land locked and consequently heavily influenced by surrounding landscapes and local climates (Healy and Harada, 1991). In most cases, they support small but regionally significant fisheries (3.3% of global production; Table SM30-1; Figure 30-1b) and opportunities for other industries such as tourism. Five SES (all over 200,000 km^2 with single entrances <120 km wide) are considered here. This particular geography results in reduced circulation and exchange with ocean waters, and jurisdictions for these water bodies that are shared by two or more neighboring states. In many cases, the small volume and disconnected nature of SES (relative to coastal and oceanic environments) makes them highly vulnerable to both local and global stressors, especially with respect to the much reduced options for the migration of organisms as conditions change.

30.5.3.1. Observed Changes and Potential Impacts

30.5.3.1.1. Arabian Gulf

The Arabian Gulf (also referred to as the Persian Gulf), along with the Red Sea, is the world's warmest sea, with both extreme negative and positive temperature excursions (annual temperature range of 12°C to 35°C). Like other SES, the Arabian Gulf is particularly vulnerable to changing environmental conditions as a result of its landlocked nature. Trends in SST were not significant over the period 1950–2009 (Table 30-1), which is probably due to long-term variability, and a consequence of regional and abrupt changes that occurred in the late 1980s (Conversi et al., 2010). In keeping with this, recent (1985–2002) localized analyses (e.g., Kuwait Bay) show strong and significant warming trends (based in this case on Advanced Very High Resolution Radiometer (AVHRR) National Oceanic and Atmospheric Administration (NOAA) satellite data) of 0.6°C per decade (Al-Rashidi et al., 2009). There is *limited evidence* and *low agreement* as to how this variability influences marine ecosystems and human activities within the Arabian Gulf, although impacts on some ecosystem components (e.g., coral reefs) have been defined to some extent. The mass coral bleaching and mortality that occurred in 1996 and 1998 were a direct result of the sensitivity of reef-building corals to unusually elevated sea temperatures (*high confidence*; Riegl, 2002, 2003; Box CC-CR). These changes to coral reefs have resulted in a loss of fish species that feed on coral-associated invertebrates while herbivores and planktivorous fish abundances have increased (*medium confidence*; Riegl, 2002). Despite coral ecosystems in this sub-region being adapted to some of the highest temperatures in shallow seas on Earth, anthropogenic climate change is driving higher frequencies and intensities of mass coral bleaching and mortality (Riegl et al., 2011). Other biological changes (e.g., harmful algal blooms and fish kills; Heil et al., 2001) have been associated with the increasing sea temperatures of the Arabian Gulf, although attribution to increasing temperatures as opposed to other factors (e.g., water quality) is limited (Bauman et al., 2010).

30.5.3.1.2. Red Sea

Few studies have focused on attributing recent changes in Red Sea ecosystems to climate change (including ocean acidification). The Red Sea warmed by 0.74°C from 1982 to 2006 (Belkin, 2009), although trends in the average SST, however, are not significant from 1950 to 2009 (p-value > 0.05; Table 30-1) owing to a high degree of variability involved when longer periods were examined (supplementary material in Belkin, 2009). The temperature of the warmest month of the year, however, showed a significant increase over the 60-year period (0.05°C per decade; Table 30-1). Regional trends within the Red Sea may also differ, with at least one other study reporting higher rates of warming for the central Red Sea (1.46°C, relative to 1950–1997 NOAA Extended Reconstructed SST (ERSST) v3b climatology; Cantin et al., 2010).

Long-term monitoring of coral community structure and size over 20 years shows that average colony size of corals has declined (*high confidence*) and species' latitudinal limits may have changed (*medium confidence*). The decline in average colony size is ascribed to heat-mediated bleaching as well as increases in coral diseases and crown of thorns starfish (*Acanthaster* sp.) predation (Riegl et al., 2012). The patterns of this decline correlate well with the pattern of recent heating in the Red Sea (Raitsos et al., 2011), with the biggest changes being seen in the southern part of the Red Sea. Skeletal growth of the long-lived massive coral *Diploastrea heliopora* has declined significantly, *very likely* due to warming temperatures (*medium confidence*; p-value ≥ 0.05; Cantin et al., 2010).

Cantin et al. (2010) proposed that the massive coral *Diploastrea helipora* will cease to grow in the central Red Sea by 2070 under SRES A1B and A2 (*medium confidence*), although this may not hold for other coral species. For example, an increase in linear extension of *Porites* corals, beginning in the 1980s, was recorded in the northern Red Sea (Heiss, 1996), where temperatures have increased by 0.74ºC from 1982 to 2006 (Belkin, 2009), suggesting that these corals were living in sub-optimal conditions (cooler waters). They may therefore benefit from elevated temperature before reaching their thermal threshold, at which point growth rates would be predicted to decline, as they are doing in other oceans. Riegl and Piller (2003) concluded that coral habitats at moderate depths in the Red Sea might provide important refugia from some aspects of climate change in the future (*limited evidence*). Silverman et al. (2007) quantified the sensitivity of net coral reef ecosystem calcification to changes in carbonate chemistry (pH, aragonite saturation). Their results demonstrate a strong negative effect of ocean acidification on ecosystem-scale calcification and decalcification, and show that small changes in carbonate dissolution could have large-scale implications for the long-term persistence of carbonate coral reef systems within the Red Sea (Silverman et al., 2007, 2009).

30.5.3.1.3. Black Sea

The temperature of the surface waters of the Black Sea increased by 0.96ºC from 1982 to 2006 (Belkin, 2009), which is consistent with other studies (*high confidence*; Buongiorno Nardelli et al., 2010; Bozkurt and Sen, 2011). As with other SES (i.e., Arabian Gulf and Baltic, Mediterranean, and Red Seas), longer data sets do not reveal a significant trend due to large-scale variability prior to 1982, which may be due to the influence of AMO, NAO, and other long-term sources of variability (Table 30-1; supplementary material in Belkin, 2009). Buongiorno Nardelli et al. (2010) observed that short-term SST variability (week-month) is strongly influenced by interactions with the overlying atmosphere, which itself is strongly influenced by the surrounding land temperatures. As with the Mediterranean and Red Seas, however, a significant upward trend in the temperature is recorded in the warmest month of the year over the period 1950–2009 (Table 30-1). Freshwater discharge from rivers draining into the Black Sea has remained more or less constant since the early 1960s (Ludwig et al., 2009). Increasing water temperature has steadily eliminated the Cold Intermediate Layer (CIL; temperatures below 8°C) throughout the Black Sea basin over 1991–2003 (*high confidence*; Oguz et al., 2003). Reduced water column mixing and upwelling during warmer winter periods has reduced the supply of nutrients to the upper layers of the Black Sea (Oguz et al., 2003) and expanded areas of low O_2 in the deeper parts of the Black Sea, which is the world's largest anoxic marine basin (*high confidence*; Murray et al., 1989). These changes coincided with the collapse of fish stocks and the invasion by the ctenophore, *Mnemiopsis leidyi*, in the 1980s (Oguz et al., 2008), while inputs of nutrients such as phosphate from the Danube River has decreased strongly since 1992–1993 (Oguz and Velikova, 2010). Environmental perturbations explain the declining levels of primary productivity, phytoplankton, bacterioplankton, and fish stocks in the Black Sea from the mid-1990s (Yunev et al., 2007; Oguz and Velikova, 2010). The Black Sea system is very dynamic and is strongly affected by non-climate stressors in addition to climate change, making attribution of detected trends to climate change difficult.

30.5.3.1.4. Baltic Sea

Temperatures in the highly dynamic Baltic Sea increased substantially since the early 1980s (Aleksandrov et al., 2009; Belkin, 2009), with increases of 1.35°C (1982–2006) being among the highest rate of change seen in any SES (Belkin, 2009). Increases of this magnitude are not seen in longer records throughout the Baltic Sea (1861–2001: MacKenzie et al., 2007; MacKenzie and Schiedek, 2007a,b; 1900–1998: Madsen and Højerslev, 2009). The salinity of the surface and near bottom waters of the Baltic Sea, for example, Gdansk Basin (Aleksandrov et al., 2009) and central Baltic (Fonselius and Valderrama, 2003; Möllmann et al., 2003), decreased from 1975 to 2000, due to changing rainfall and river runoff, and a reduction in the pulses of seawater (vital for oxygenation and related chemical changes) from the North Sea through its opening via the Kattegat (*high confidence*; Samuelsson, 1996; Conley et al., 2009; Hänninen and Vuorinen, 2011). There is a strong vertical zonation within the Baltic Sea in terms of the availability of O_2. The shallow sub-regions of the Baltic are relatively well oxygenated. However, O_2 levels are low in the deeper basins, producing conditions in which organisms and ecosystems are exposed to prolonged hypoxia.

The annual biomass of phytoplankton has declined almost threefold in the Baltic Transition Zone (Kattegat, Belt Sea) and Western Baltic Sea since 1979 (Henriksen, 2009), reputedly due to changing nitrogen loads in the Danish Straits (*medium confidence*) in addition to increasing sea temperature (*very likely*; Madsen and Højerslev, 2009). Reduced phytoplankton production may have decreased the productivity of fisheries in the western Baltic Sea and the Transition Zone (*low* to *medium confidence*; Chassot et al., 2007). Decreasing salinity in the Baltic deep basins may also affect zooplankton reproduction, especially that of the copepod *Pseudocalanus acuspes*, contributing to density-dependent decrease in growth of the commercially important herring and sprat stocks (*high confidence*; Möllmann et al., 2003, 2005; Casini et al., 2011). The strong relationship between phytoplankton and fish production, and increasing sea temperature, decreasing salinity, and other environmental factors, suggests that major changes in fisheries production will occur as sea temperatures increase and the hydrological cycle in the Baltic region changes (*high confidence*; MacKenzie et al., 2012). A combination of climate change-induced oceanographic changes (i.e., decreased salinity and increased temperatures), eutrophication, and overfishing have resulted in major changes in trophic structure in the deep basins of the Baltic Sea (Möllmann et al., 2009). This had important implications for cod, a commercially important top predator (*medium confidence*; Lindegren et al., 2010).

30.5.3.1.5. Mediterranean Sea

The Mediterranean Sea is strongly linked to the climates of North Africa and Central Europe. SST within the Mediterranean increased by 0.43°C from 1957 to 2008 (supplementary material in Belkin, 2009), although analysis of data from 1950 to 2009 detected only a significant trend in summer temperature (0.11°C per decade, p-value ≤ 0.05; Table 30-1) due to large fluctuations in SST prior to the 1980s. Surface temperatures increased in the Mediterranean Sea consistent with significant increases in SST at a number of monitoring sites (*robust evidence*, *high agreement*; e.g., Coma et al., 2009; Conversi et al., 2010; Calvo et al., 2011). It is *likely* that temperatures, along with salinity, have also increased at depth (400 m or more) in the western Mediterranean Sea over the past 30 to 40 years which, when analyzed in the context of heat budget and water flux of the Mediterranean, is consistent with anthropogenic greenhouse warming (Bethoux et al., 1990; Rixen et al., 2005; Vargas-Yáñez et al., 2010). Large-scale variability such as the AMO and NAO can obscure or accentuate the overall warming trend (Marullo et al., 2011; WGI AR5 Sections 14.5.1, 14.7.6). Relatively warm episodes in the 1870s, 1930–1970s, and since the mid-1990s, for example, exhibit an influence of the AMO (Kerr, 2000; Moron, 2003). Reported temperature anomalies in the Mediterranean, often locally manifesting themselves as periods of low wind, increased water column stratification, and a deepening thermocline, are associated with positive phases of the NAO index (Molinero et al., 2005; Lejeusne et al., 2010).

Sea levels have increased rapidly in some areas over recent decades and are also strongly influenced by NAO phases. The rate has been approximately 3.4 mm yr^{-1} (1990–2009) in the northwest Mediterranean (*high confidence*; Calvo et al., 2011). These influences are reduced when measurements are pooled over longer time scales, resulting in a lower rate of SLR (Massuti et al., 2008). If the positive phase of the NAO is more frequent in the future (Terray et al., 2004; Kuzmina et al., 2005; WGI AR5 Section 14.4.2), then future SLR may be slightly suppressed as a result of atmospheric influences (*medium confidence*; Jordà et al., 2012). As temperatures have increased, the Mediterranean has become more saline (+0.035 to 0.040 psu from 1950 to 2000; Rixen et al., 2005) with the length of the thermal stratification period persisting twice as long in 2006 as it did in 1974 (Coma et al., 2009).

Conditions within the Mediterranean Sea changed abruptly and synchronously with similar changes across the North, Baltic, and Black Seas in the late 1980s (Conversi et al., 2010), which possibly explains the lack of trend in SES SST when examined from 1950 to 2009 (Table 30-1). These changes in physical conditions (increased temperature, higher sea level pressure, positive NAO index) also coincided with step changes in the diversity and abundance of zooplankton, decreases in stock abundance of anchovies and the frequency of "red tides," and increases in mucilage outbreaks (Conversi et al., 2010). Mucilage outbreaks are strongly associated with warmer and more stratified water columns (*high confidence*), and lead to a greater abundance and diversity of marine microbes and potentially disease-causing organisms (*likely*; Danovaro et al., 2009). Increasing temperatures are also driving the northward spread of warm-water species (*medium confidence*) such as the sardine *Sardinella aurita* (Sabatés et al., 2006; Tsikliras, 2008), and have contributed to the spread of the invading Atlantic coral *Oculina patagonia* (Serrano et al., 2013). The recent spread of warm-water species that have invaded through the Straits of Gibraltar and the Suez Canal into cooler northern areas is leading to the "tropicalization" of Mediterranean fauna (*high confidence*; Bianchi, 2007; Ben Rais Lasram and Mouillot, 2008; CIESM, 2008; Galil, 2008, 2011). Warming since the end of the 1990s has accelerated the spread of tropical invasive species from the eastern Mediterranean basin (Raitsos et al., 2010; Section 23.6.5).

In addition to general warming patterns, periods of extreme temperatures have had large-scale and negative consequences for Mediterranean marine ecosystems. Unprecedented mass mortality events, which affected at least 25 prominent invertebrate species, occurred during the summers of 1999, 2003, and 2006 across hundreds of kilometers of coastline in the northwest Mediterranean Sea (*very high confidence*; Cerrano et al., 2000; Garrabou et al., 2009; Calvo et al., 2011; Crisci et al., 2011). Events coincided with either short periods (2 to 5 days: 2003, 2006) of high sea temperatures (27°C) or longer periods (30 to 40 days) of modestly high temperatures (24°C: 1999; Bensoussan et al., 2010; Crisci et al., 2011). Impacts on marine organisms have been reported in response to the extreme conditions during these events (e.g., gorgonian coral mortality; Coma et al., 2009), shoot mortality, and anomalous flowering of seagrasses (*high confidence*; Diaz-Almela et al., 2007; Marbà and Duarte, 2010). The frequency and intensity of these types of heat stress events are expected to increase as sea temperatures increase (*high confidence*).

Longer-term data series (over several decades) of changes in relative acidity of the Mediterranean Sea are scarce (Calvo et al., 2011; The MerMex Group, 2011). Recent re-analysis, however, has concluded that the pH of Mediterranean waters has decreased by 0.05 to 0.14 pH units since the preindustrial period (*medium confidence*; Luchetta et al., 2010; Touratier and Goyet, 2011). Anthropogenic CO_2 has penetrated the entire Mediterranean water column, with the western basin being more contaminated than the eastern basin (Touratier and Goyet, 2011). Studies that have explored the consequences of ocean acidification for the biology and ecology of the Mediterranean Sea are rare (Martin and Gattuso, 2009; Rodolfo-Metalpa et al., 2010; Movilla et al., 2012), although insights have been gained by studying natural CO_2 seeps at Mediterranean sites such as Ischia in Italy, where biodiversity decreases with decreasing pH toward the vents, with a notable decline in calcifiers (Hall-Spencer et al., 2008). Transplants of corals, molluscs, and bryozoans along the acidification gradients around seeps reveal a low level of vulnerability to CO_2 levels expected over the next 100 years (*low confidence*; Rodolfo-Metalpa et al., 2010, 2011). However, periods of high temperature can increase vulnerability to ocean acidification, thereby increasing the long-term risk posed to Mediterranean organisms and ecosystems as temperatures warm. Significantly, some organisms such as seagrasses and some macroalgae appeared to benefit from local ocean acidification (Hall-Spencer et al., 2008).

30.5.3.2. Key Risks and Vulnerabilities

SES are highly vulnerable to changes in global temperature on account of their small volume and landlocked nature. Consequently, SES will respond faster than most other parts of the Ocean (*high confidence*). Risks to ecosystems within SES are *likely* to increase as water columns become further stratified under increased warming, promoting hypoxia at depth and reducing nutrient supply to the upper water column (*medium evidence*, *high agreement*). The impact of rising temperatures on SES is exacerbated by their vulnerability to other human influences such as over-exploitation, pollution, and enhanced runoff from modified coastlines. Due to a mixture of global and local human stressors, key fisheries have undergone fundamental changes in their abundance and distribution over the past 50 years (*medium confidence*). A major risk exists for SES from projected increases in the frequency of temperature extremes that drive mass mortality events, increasing water column stabilization leading to reduced mixing, and changes to the distribution and abundance of marine organisms. The vulnerability of marine ecosystems, fisheries, and human communities associated with the SES will continue to increase as global temperatures increase.

Sea temperatures are *very likely* to increase in the five SES under moderate (RCP6.0) to high (RCP8.5) future scenarios. Under BAU (RCP8.5; Table SM30-3), sea temperatures in the SES are projected to increase by 0.93°C to 1.24°C over 2010–2039 (Table SM30-4). Increases of 3.45°C to 4.37°C are projected over 2010–2099, with the greatest increases projected for the surface waters of the Baltic Sea (4.37°C) and Arabian Gulf (4.26°C), and lower yet substantial amounts of warming in the Red Sea (3.45°C) (Table SM30-4). The heat content added to these small ocean regions is *very likely* to increase stratification, which will reduce the nutrient supply to the upper layers of the water column, reducing primary productivity and driving major changes to the structure and productivity of fisheries. Reduced mixing and ventilation, along with increased microbial metabolism, will *very likely* increase hypoxia and expand the number and extent of "dead zones." Changing rainfall intensity (Section 23.3; WGI AR5 Section 12.4.5) can exert a strong influence on the physical and chemical conditions within SES, and in some cases will combine with other climatic changes to transform these areas. These changes are likely to increase the risk of reduced bottom-water O_2 levels to Baltic and Black Sea ecosystems (due to reduced solubility, increased stratification, and microbial respiration), which is *very likely* to affect fisheries. These changes will increase the frequency and intensity of impacts arising from heat stress, based on responses to temperature extremes seen over the past 30 years, such as the mass mortality of benthic organisms that occurred in the Mediterranean Sea during the summers of 1999, 2003, and 2006, and the Arabian Gulf in 1996 and 1998. Extreme temperature events such as heat waves are projected to increase (*high confidence*; Section 23.2; IPCC, 2012). Projections similar to those outlined in Section 30.5.4.2 can be applied to the coral reefs of the Arabian Gulf and the Red Sea, where temperatures are *very likely* to increase above established thresholds for mass coral bleaching and mortality (*very high confidence*; Figure 30-10).

30.5.4. Coastal Boundary Systems

The Coastal Boundary Systems (CBS) are highly productive regions, comprising 10.6% of primary production and 28.0% of global fisheries production (Table SM30-1; Figure 30-1b). The CBS include the marginal seas of the northwest Pacific, Indian, and Atlantic Oceans, encompassing the Bohai/Yellow Sea, East China Sea, South China Sea, and Southeast Asian Seas (e.g., the Timor, Arafura, and Sulu Seas, and the northern coast of Australia) in the Pacific; the Arabian Sea, Somali Current system, East Africa coast, Mozambique Channel, and Madagascar in the Indian Ocean; and the Caribbean Sea and Gulf of Mexico in the Atlantic Ocean). Some CBS are dominated by powerful currents such as the Kuroshio (Pacific), or are strongly influenced by monsoons (e.g., Asian-Australian and African monsoons).

30.5.4.1. Observed Changes and Potential Impacts

Many ecosystems within the CBS are strongly affected by the local activities of often-dense coastal human populations. Activities such as the overexploitation of fisheries, unsustainable coastal development, and pollution have resulted in the widespread degradation of CBS ecosystems (Burke et al., 2002, 2011). These influences have combined with steadily increasing ocean temperature and acidification to drive major changes to a range of important ecosystems over the past 50 years. Understanding the interactions between climate change and non-climate change drivers is a central part of the detection and attribution process within the CBS.

Overall, the CBS warmed by 0.14°C to 0.80°C from 1950 to 2009 (Table 30-1), although changes within the Gulf of Mexico/Caribbean Sea sub-region were not significant (p-value > 0.05) over this period. Key sub-regions within the CBS such as the Coral Triangle and Western Indian Ocean warmed by 0.79°C and 0.60°C, respectively, from 1950 to 2009 (Table 30-1). Rates of SLR vary from decreasing sea levels (–5 to –10 mm yr^{-1}) to low (2 to 3 mm yr^{-1}, Caribbean) to very high (10 mm yr^{-1}, Southeast Asia; Figure 30-5) rates of increase. Ocean acidification also varies from region to region (Figure SM30-2), and is influenced by oceanographic and coastal processes, which often have a large human component.

30.5.4.1.1. Bohai/Yellow Sea/East China Sea

The Bohai Sea, Yellow Sea, and the East China Sea (ECS) are shallow marginal seas along the edge of the northwest Pacific that are strongly influenced by the Kuroshio Current (Matsuno et al., 2009), the East Asian Monsoon (EAM), and major rivers such as the Yellow (Huang He) and Yangtze (Changjiang) Rivers. Upwelling of the Kuroshio sub-surface waters provides abundant nutrients that support high levels of primary productivity (Wong et al., 2000, 2001). The ecosystems of the ECS are heavily affected by human activities (e.g., overfishing and pollution), which tend to compound the influence and consequences of climate change.

SST within the ECS has increased rapidly since the early 1980s (*high confidence*; Lin et al., 2005; Jung, 2008; Cai et al., 2011; Tian et al., 2012). The largest increases in SST have occurred in the ECS in winter (1.96°C, 1955–2005) and in the Yellow Sea in summer (1.10°C, 1971–2006; Cai et al., 2011). These changes in SST are closely linked to a weakening of the EAM (e.g., Cai et al., 2006, 2011; Tang et al., 2009) and increasing warmth of the Kuroshio Current (Qi et al., 2010; Zhang et al., 2011; Wu et al., 2012). At the same time, dissolved O_2 has decreased (Lin et al., 2005; Jung, 2008; Qi et al., 2010), with an associated increase in the extent of the hypoxic areas in coastal areas of the Yellow Sea/ECS (Jung, 2008; Tang, 2009; Ning et al., 2011).

Primary productivity, biomass yields, and fish capture rates have experienced large changes within the ECS over the past decades (*limited evidence*, *medium agreement*; *low confidence*; Tang et al., 2003; Lin et al., 2005; Tang, 2009). Fluctuations in herring abundance appear to closely track SST shifts within the Yellow Sea (Tang, 2009). For plankton and fish species, the proportions of warm-water species relative to warm-temperate species in the Changjiang River Estuary (extending to the southern Taiwan Strait) have changed over past decades (Zhang et al., 2005; Ma et al., 2009; Lin and Yang, 2011). Northward shifts in catch distribution for some pelagic fish species in Korean waters were driven, in part, by warming SST (*medium confidence*; Jung et al., 2014). The frequency of harmful algal blooms and blooms of the giant jellyfish *Nemopilema nomurai* in the offshore area of the ECS have increased and have been associated with ocean warming and other factors such as eutrophication (Ye and Huang, 2003; Tang, 2009; Cai and Tan, 2010). Although attribution of these changes to anthropogenic climate change is complicated by the increasing influence of non-climate-related human activities, many of these changes are consistent with those expected as SST increases.

30.5.4.1.2. South China Sea

The South China Sea (SCS) is surrounded by continental areas and includes large numbers of islands, and is connected to the Pacific, ECS, and Sulu Sea by straits such as the Luzon and Taiwan Strait. The region is greatly influenced by cyclones/typhoons, and by the Pearl, Red, and Mekong Rivers. The region has a distinct seasonal circulation and is greatly influenced by the southwest monsoon (in summer), the Kuroshio Current, and northeast monsoon (in winter). The SCS includes significant commercial fisheries areas and includes coral reefs, mangroves, and seagrass beds.

The surface waters of the SCS have been warming steadily from 1945 to 1999 with the annual mean SST in the central SCS increasing by 0.92°C (1950–2006; Cai et al., 2009), a rate similar to that observed for the entire Indo-Pacific/Southeast Asian CBS from 1950 to 2009 (0.80°C; Table 30-1). Significant freshening in the SCS intermediate layer since the 1960s has been observed (Liu et al., 2007). The temperature change of the upper layers of the SCS has made a significant contribution to sea level variation, which is heterogeneous in space and time (Li et al., 2002; Cheng and Qi, 2007; Liu et al., 2007).

Identifying the extent to which climate change is influencing the SCS is difficult due to confounding non-climate change factors and their interactions (e.g., local human pollution, over-exploitation together with "natural" climate variability such as EAM, ENSO, and PDO). Changing sea temperatures have influenced the abundance of phytoplankton, benthic biomass, cephalopod fisheries, and the size of demersal trawl catches in the northern SCS observed over the period 1976–2004 (*limited evidence*, *medium agreement*; Ning et al., 2009). Coral reefs and mangroves are degrading rapidly as a result of both climate change and non-climate change-related factors (*very likely*; Box CC-CR; Chen et al., 2009; China-SNAP, 2011; Zhao et al., 2012). Mass coral bleaching and mortality of coral reefs within the SCS were triggered by elevated temperatures in 1998 and 2007 (Yu et al., 2006; Li et al., 2011). Conversely, warming enabled the establishment of a high-latitude, non-carbonate, coral community in Daya Bay in northern SCS, although this community has recently degraded as a result of increasing anthropogenic stresses (Chen et al., 2009; Qiu et al., 2010).

30.5.4.1.3. Southeast Asian Seas

The Southeast Asian Seas (SAS) include an archipelago of diverse islands that interact with the westward flow of the North Equatorial Current and the Indonesian Throughflow (Figure 30-1a). A large part of this region is referred to as the "Coral Triangle" (Veron et al., 2009). The world's most biologically diverse marine area, it includes parts of Malaysia, Indonesia, the Philippines, Timor Leste, the Solomon Islands, and Papua New Guinea. SST increased significantly from 1985 to 2006 (Peñaflor et al., 2009; McLeod et al., 2010), although with considerable spatial variation. Trends examined over longer periods (1950–2009) show significant warming (+0.80°C, *p*-value ≤ 0.05; Table 30-1). The sea level is rising by up to 10 mm yr^{-1} in much of this region (Church et al., 2004, 2006; Green et al., 2010). Like other tropical areas in the world, coral reefs within SAS have experienced periods of elevated temperature, which has driven several mass coral bleaching and mortality events since the early 1980s (*high confidence*; Hoegh-Guldberg et al., 2009; McLeod et al., 2010; Figure 30-10a). The most recent occurred during warm conditions in 2010 (Krishnan et al., 2011). These changes are the result of increasing ocean temperatures and are *very likely* to be a consequence of anthropogenic climate change (*high confidence*; Box CC-CR; WGI AR5 Section 10.4.1). Although calcification rates of some key organisms (e.g., reef-building corals; Tanzil et al., 2009) have slowed over the past 2 decades, it is not possible to conclude that the changes are due to ocean acidification. While a large part of the decline in coral reefs has been due to increasing local stresses (principally destructive fishing, declining water quality, and over-exploitation of key reef species), projected increases in SST represent a major challenge for these valuable ecosystems (*high agreement*; Burke et al., 2002; Burke and Maidens, 2004).

30.5.4.1.4. Arabian Sea and Somali Current

The Arabian Sea and Somali Current are relatively productive ocean areas, being strongly influenced by upwelling and the monsoonal system. Wind-generated upwelling enhances primary production in the western Arabian Sea (Prakash and Ramesh, 2007). Several key fisheries within this region are under escalating pressure from both fishing and climate change. SST increased by 0.18°C and 0.26°C in the Arabian Sea and Somali Current, respectively, from 1982 to 2006 (HadSST2; Rayner et al., 2003; Belkin, 2009), which is consistent with the overall warming of the Western Indian Ocean portion of the CBS from 1950 to 2009 (0.60°C; Table 30-1). Salinity of surface waters in the Arabian Sea increased by 0.5 to 1.0% over the past 60 years (Figure 30-6c), due to increased evaporation from warming seas and contributions from the outflows of the saline Red Sea and Arabian Gulf. As in other tropical sub-regions, increasing sea temperatures have increased the frequency of mass coral bleaching and mortality within this region (Wilkinson and Hodgson, 1999; Goreau et al., 2000; Wilkinson, 2004).

The aragonite saturation horizon in both the Arabian Sea and Bay of Bengal is now 100 to 200 m shallower than in preindustrial times as a result of ocean acidification (*medium confidence*; Feely et al., 2004). Shoaling of the aragonite saturation horizon is *likely* to affect a range of organisms and processes, such as the depth distribution of pteropods (zooplankton) in the western Arabian Sea (*medium confidence*; Hitchcock et al., 2002; Mohan et al., 2006). More than 50% of the area of OMZs in the world's oceans occur in the Arabian Sea and Bay of Bengal and long-term measurements reveal that O_2 concentrations are declining in this region (*high confidence*; Helly and Levin, 2004; Karstensen et al., 2008; Stramma et al., 2010; Section 30.3.2.3). The information regarding the consequences of climate change within this region is undeveloped and suggests that important physical, chemical, and biological responses to climate change need to be the focus of further investigation.

30.5.4.1.5. East Africa coast and Madagascar

The Western Indian Ocean strongly influences the coastal conditions associated with Kenya, Mozambique, Tanzania, Madagascar, La Réunion, Mayotte, and three archipelagos (Comoros, Mauritius, and the Seychelles). Sea temperatures in the Western Indian Ocean have increased by 0.60°C over 1950–2009 (*high confidence*; p-value ≤ 0.05; Table 30-1), increasing the frequency of positive thermal anomalies that have triggered mass coral bleaching and mortality events across the region over the past 2 decades (*high confidence*; Baker et al., 2008; Nakamura et al., 2011; Box CC-HS). Trends in changes in SST and surface salinity vary with location along the East African coastline, with faster rates at higher latitudes (Figure 30-2). Periods of heat stress over the past 20 years have triggered mass coral bleaching and mortality on coral reef ecosystems within this region (McClanahan et al., 2007, 2009a,b,c; Ateweberhan and McClanahan, 2010; Ateweberhan et al., 2011). Steadily increasing sea temperatures have also produced anomalous growth rates in long-lived corals such as *Porites* (*high confidence*; McClanahan et al., 2009b). Differences in the susceptibility of reef-building corals to stress from rising sea temperatures has also resulted in changes to the composition of coral (*high confidence*; p-value ≤ 0.05; McClanahan et al., 2007) and benthic fish communities (*high confidence*; p-value ≤ 0.05; Graham et al., 2008; Pratchett et al., 2011a). These changes are *very likely* to alter species composition and potentially the productivity of coastal fisheries (*robust evidence*, *high agreement*; *high confidence*; Jury et al., 2010), although there may be a significant lag between the loss of coral communities and the subsequent changes in the abundance and community structure of fish populations (p-value ≤ 0.05; Graham et al., 2007). Some of these potential changes can be adverted or reduced by interventions such as the establishment of marine protected areas and changes to fishing management (McClanahan et al., 2008; Cinner et al., 2009; Jury et al., 2010; MacNeil et al., 2010).

30.5.4.1.6. Gulf of Mexico and Caribbean Sea

The Gulf of Mexico and Caribbean Sea form a semi-contained maritime province within the Western Atlantic. These areas are dominated by a range of activities including mineral extraction, fishing, and tourism, which provide employment and opportunity for almost 75 million people who live in coastal areas of the USA, Mexico, and a range of other Caribbean nations (Adams et al., 2004). The Gulf of Mexico and Caribbean Sea have warmed by 0.31°C and 0.50°C, respectively, from 1982 to 2006 (*very likely*; Belkin, 2009). Warming trends are not significant from 1950 to 2009 (Table 30-1), which may be partly due to spatial variability in warming patterns (Section 30.5.3.1). The Caribbean region has experienced a sustained decrease in aragonite saturation state from 1996 to 2006 (*very likely*; Gledhill et al., 2008). Sea levels within the Gulf of Mexico and Caribbean Sea have increased at the rate of 2 to 3 mm yr^{-1} from 1950 to 2000 (Church et al., 2004; Zervas, 2009).

Understanding influences of climate change on ocean ecosystems in this region is complicated by the confounding influence of growing human populations and activities. The recent expansion of the seasonal hypoxic zone, and the associated "dead zone," in the Gulf of Mexico has been attributed to nitrogen inputs driven by land management (Turner and Rabalais, 1994; Donner et al., 2004) and changes to river flows, wind patterns, and thermal stratification of Gulf waters (*high confidence*; Justić et al., 1996, 2007; Levin et al., 2009; Rabalais et al., 2009). The increases in coastal pollution and fishing have potentially interacted with climate change to exacerbate impacts on marine ecosystems within this region (Sections 5.3.4, 29.3). These changes have often been abrupt and non-linear (Taylor et al., 2012).

A combination of local and global disturbances has driven a large-scale loss of reef-building corals across the Caribbean Sea since the late 1970s (*high confidence*; Hughes, 1994; Gardner et al., 2003). Record thermal stress in 2005 triggered the largest mass coral bleaching and mortality event on record for the region, damaging coral reefs across hundreds of square kilometers in the eastern Caribbean Sea (*high confidence*; Donner et al., 2007; Eakin et al., 2010). Although conditions in 2010 were milder than in 2005, elevated temperatures still occurred in some parts of the Caribbean (Smith et al., 2013). Increasing temperatures in the Caribbean have also been implicated in the spread of marine diseases (Harvell et al., 1999, 2002, 2004) and some introduced species (*likely*; Firth et al., 2011). As in other sub-regions, pelagic fish species are sensitive to changes in sea temperature and modify their distribution and abundance accordingly (Muhling et al., 2011). Fish and invertebrate assemblages in the Gulf of Mexico have shifted deeper in response to SST warming over 1970s–2011 (*medium confidence*; Pinsky et al., 2013).

Coral ecosystems in the Caribbean Sea are at risk from ocean acidification (*very likely*; Albright et al., 2010; Albright and Langdon, 2011), although impacts have yet to be observed under field conditions. Ocean acidification may also be altering patterns of fish recruitment to coral reefs, although direct evidence for how this has affected Caribbean species is lacking (*low confidence*; Dixson et al., 2008, 2010; Munday et al., 2009).

30.5.4.2. Key Risks and Vulnerabilities

Worldwide, 850 million people live within 100 km of tropical coastal ecosystems such as coral reefs and mangroves deriving multiple benefits including food, coastal protection, cultural services, and income from industries such as fishing and tourism (Burke et al., 2011). Marine ecosystems within the CBS are sensitive to increasing sea temperatures (Figure 30-10), although detection and attribution are complicated by the significant influence and interaction with non-climate change stressors (water quality, over-exploitation of fisheries, coastal degradation; Box CC-CR). Warming is likely to have changed the primary productivity of ocean waters, placing valuable ecosystems and fisheries within the ECS at risk (*low* to *medium confidence*). Other risks include the expansion of hypoxic conditions and associated dead zones in many parts of the CBS. Given the consequences for coastal ecosystems and fisheries, these changes are *very likely* to increase the vulnerability of coastal communities throughout the CBS.

Sea temperatures are increasing within many parts of CBS ecosystems (1950–2009; Table 30-1), and will continue to do so over the next few decades and century. Sea temperatures are projected to change by

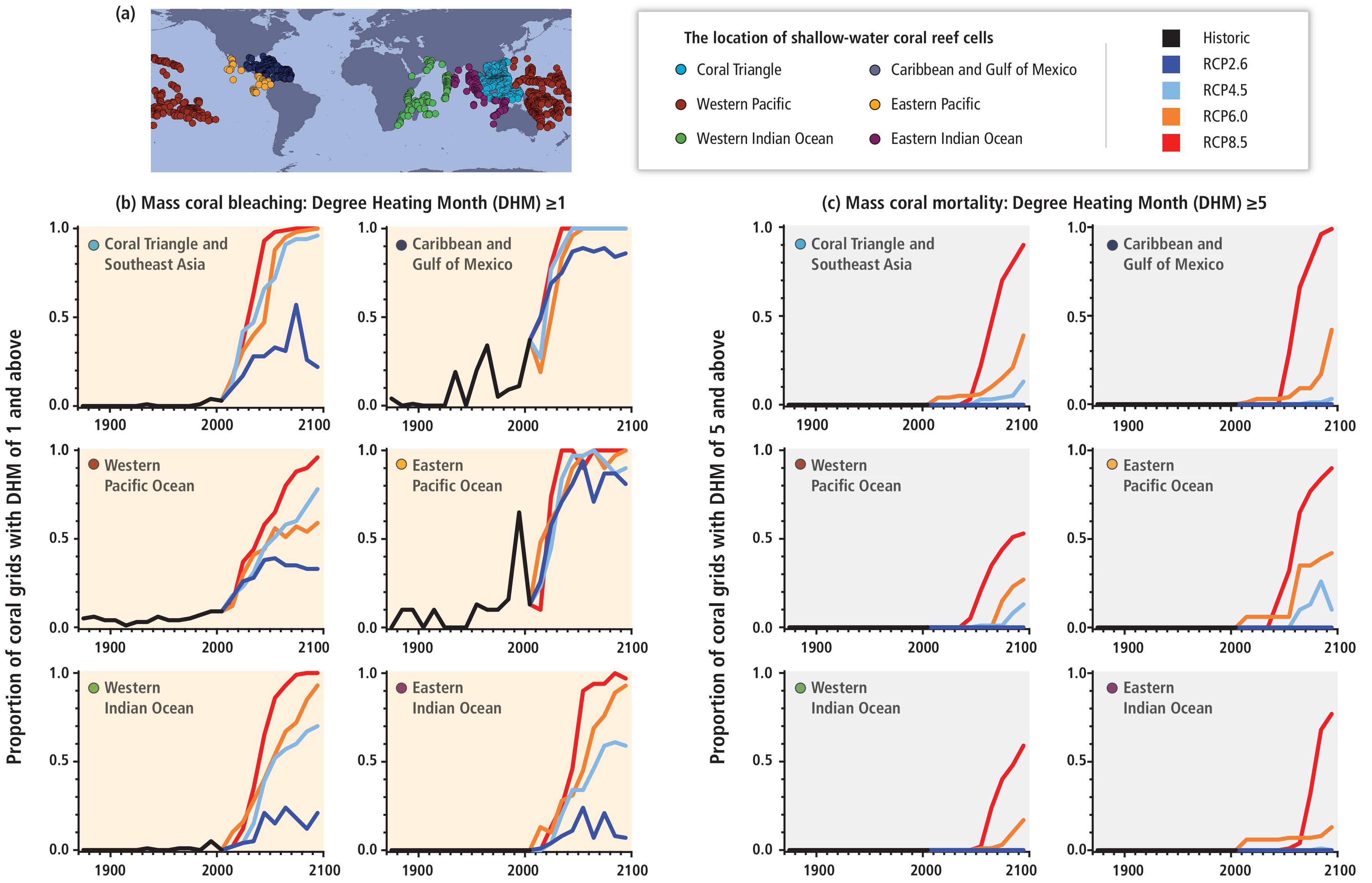


Figure 30-10 | Annual maximum proportions of reef pixels with Degree Heating Months (DHM, Donner et al., 2007) for each of the six coral regions (a, Figure 30-4b)—(b) DHM ≥1 (used for projecting the incidence of coral bleaching; Strong et al., 1997, 2011) and (c) DHM ≥5 (associated with bleaching followed by significant mortality; Eakin et al., 2010)—for the period 1870–2009 using the Hadley Centre Interpolated sea surface temperature 1.1 (HadISST1.1) data set. The black line on each graph is the maximum annual area value for each decade over the period 1870–2009. This value is continued through 2010–2099 using Coupled Model Intercomparison Project Phase 5 (CMIP5) data and splits into the four Representative Concentration Pathways (RCP2.6, 4.5, 6.0, and 8.5). DHM were produced for each of the four RCPs using the ensembles of CMIP models. From these global maps of DHM, the annual percentage of grid cells with DHM ≥1 and DHM ≥5 were calculated for each coral region. These data were then grouped into decades from which the maximum annual proportions were derived. The plotted lines for 2010–2099 are the average of these maximum proportion values for each RCP. Monthly sea surface temperature anomalies were derived using a 1985–2000 maximum monthly mean climatology derived in the calculations for Figure 30-4. This was done separately for HadISST1.1, the CMIP5 models, and each of the four RCPs, at each grid cell for every region. DHMs were then derived by adding up the monthly anomalies using a 4-month rolling sum. Figure SM30-3 presents past and future sea temperatures for the six major coral reef provinces under historic, un-forced, RCP4.5 and RCP8.5 scenarios.

0.34°C to 0.50°C over the near term (2010–2039) and by 0.23°C to 0.74°C over the long term (2010–2099) under the lowest RCP scenario (RCP2.6). Under BAU (RCP8.5), CBS sea temperatures are projected to increase by 0.62°C to 0.85°C over the near term and 2.44°C to 3.32°C over the long term (Table SM30-4). Given the large-scale impacts (e.g., mass coral bleaching and mortality events) that have occurred in response to much smaller changes in the past over CBS regions (0.14°C to 0.80°C from 1950–2009; Table 30-1), the projected changes of 2.44°C to 3.32°C over 2010–2099 are *very likely* to have large-scale and negative consequences for the structure and function of many CBS ecosystems (*virtually certain*), especially given the observed sensitivity of coral reefs to relatively small increases in temperature over the past 3 decades (Hoegh-Guldberg, 1999; Eakin et al., 2010; Lough, 2012).

It is *very likely* that coral-dominated reef ecosystems within the CBS (and elsewhere) will continue to decline and will consequently provide significantly less ecosystem goods and services for coastal communities if sea temperatures increase by more than 1°C above current temperatures (Box CC-CR; Figure 30-10). Combining the known sensitivity of coral reefs within the Caribbean and Coral Triangle sub-regions (Strong et al., 1997, 2011; Hoegh-Guldberg, 1999), with the exposure to higher temperatures that are projected under medium (RCP4.5) to high (RCP8.5) scenarios, reveals that both coral reef-rich regions are *virtually certain* to experience levels of thermal stress (DHM ≥ 1) that cause coral bleaching every 1 to 2 years by the mid- to late part of this century (*robust evidence*, *high agreement*; *very high confidence*; Figures 30-4b,c, 30-10, 30-12, SM30-3; van Hooidonk et al., 2013). The frequency of mass mortality events (DHM ≥ 5; Figure 30-10a,b,c) also increases toward a situation whereevents that occur every 1 to 2 years by the mid- to late part of this century under low to high climate change scenarios (*robust evidence*, *high agreement*; *very high confidence*; Hoegh-Guldberg, 1999; Donner et al., 2005; Frieler et al., 2012). Mass mortality events that affect coral reefs will result in changes to community composition in the near term (2010–2039; Berumen and Pratchett, 2006; Adjeroud et al., 2009) and a continuing downward trend in coral cover in the longer term (Gardner et al., 2003; Bruno and Selig, 2007; Baker et al., 2008).

It is *virtually certain* that composition of coral reef fish populations (Graham et al., 2007; Pratchett et al., 2008, 2011a,b) will change. The productivity of many fisheries will decrease (*limited evidence*, *medium agreement*) as waters warm, acidify, and stratify, and as crucial habitat, such as coral reefs, degrade (*low confidence*). These changes are *very likely* to increase the vulnerability of millions of people who live in coastal communities and depend directly on fisheries and other goods and services provided by ecosystems such as coral reefs (Hoegh-Guldberg et al., 2009; McLeod et al., 2010).

30.5.5. Eastern Boundary Upwelling Ecosystems

The Eastern Boundary Upwelling Ecosystems (EBUE) include the California, Peru/Humboldt, Canary/northwest Africa, and Benguela Currents. They are highly productive sub-regions with rates of primary productivity that may exceed 1000 g C m^{-2} yr^{-1}. Although these provinces comprise less than 2% of the Ocean area, they contribute nearly 7% of marine primary production (Figure 30-1b) and more than 20% of the world's marine capture fisheries (Pauly and Christensen, 1995). Catches in the EBUE are dominated by planktivorous sardine, anchovy, and horse/jack mackerel, and piscivorous benthic fish such as hake. Nutrient input from upwelling of cooler waters stimulates primary production that is transferred to mid and upper trophic levels, resulting in substantial fish, seabird, and marine mammal populations. As a result, the EBUE are considered "hotspots" of productivity and biodiversity (Block et al., 2011). The high level of productivity is a result of large-scale atmospheric pressure gradients and wind systems that advect surface waters offshore, leading to the upwelling of cold, nutrient-rich waters from depth (Box CC-UP; Chavez and Messie, 2009; Chavez et al., 2011). Upwelling waters are typically low in pH and high in CO$_2$, and are likely to continue to enhance changes in pH and CO$_2$ resulting from rising atmospheric CO$_2$ (Feely et al., 2008; Gruber, 2011).

30.5.5.1. Observed Changes and Potential Impacts

There are extensive studies of the coupled climate-ecosystem dynamics of individual EBUE (e.g., California Current). Decadal variability poses challenges to the detection and attribution of changes within the EBUE to anthropogenic climate change, although there are a number of long-term studies that have been able to provide insight into the patterns of change and their causes. Like other ocean sub-regions, EBUE are projected to warm under climate change, with increased stratification and intensified winds as westerly winds shift poleward (*likely*). However, cooling has also been predicted for some EBUE, resulting from the intensification of wind-driven upwelling (Bakun, 1990). The California and Canary Currents have warmed by 0.73°C and 0.53°C (*very likely*; p-value ≤ 0.05, 1950–2009; Table 30-1), respectively, while no significant trend was detected in the sea surface temperatures of the Benguela (p-value = 0.44) and Humboldt Currents (p-value = 0.21) from 1950 to 2009 (Table 30-1). These trends match shorter-term trends for various EBUE using Pathfinder version 5 data (Demarcq, 2009). These differences are *likely* to be the result of differences in the influence of long-term variability and the specific responses of coastal wind systems to warming, although an analysis of wind data over the same period did not pick up clear trends (*low confidence*, with respect to long-term wind trends; Demarcq, 2009; Barton et al., 2013).

How climate change will influence ocean upwelling is central to resolving ecosystem and fishery responses within each EBUE. There is considerable debate, however, as to whether or not climate change will drive an intensification of upwelling (e.g., Bakun et al., 2010; Narayan et al., 2010; Barton et al., 2013) in all regions. This debate is outlined in Box CC-UP. EBUE are also areas of naturally low pH and high CO$_2$ concentrations due to upwelling, and consequently may be vulnerable to ocean acidification and its synergistic impacts (Barton et al., 2012). A full understanding of the consequences of ocean acidification for marine organisms and ecosystems is discussed elsewhere (Boxes CC-OA, CC-UP; Sections 6.2, 6.3.2; Kroeker et al., 2013; WGI AR5 Section 6.4).

30.5.5.1.1. Canary Current

Part of the North Atlantic STG, the Canary Current extends from northern Morocco southwestward to the North Atlantic Equatorial Current. It is linked with the Portugal Current (which is sometimes considered part of the Canary Current) upstream. The coastal upwelling system, however, is limited to a narrow belt along the Saharan west coast to the coast of Guinea, with the most intense upwelling occurring centrally, along the coasts of Mauritania (15°N to 20°N) and Morocco (21°N to 26°N). Total fish catches, comprising mainly coastal pelagic sardines, sardinellas, anchovies, and mackerel, have fluctuated around 2 million tonnes yr^{-1} since the 1970s (www.seaaroundus.org/lme/27.aspx). Contrasting with the other EBUE, fishing productivity is modest, probably partly due to the legacy of uncontrolled fishing in the 1960s (Arístegui et al., 2009).

Most observations suggest that the Canary Current has warmed since the early 1980s (Arístegui et al., 2009; Belkin, 2009; Demarcq, 2009; Barton et al., 2013), with analysis of HadISST1.1 data from 1950 to 2009 indicating warming of 0.53°C from 1950–2009 (p-value ≤ 0.05; Table 30-1). Gómez-Gesteira et al. (2008) suggest a 20 and 45% decrease in the strength of upwelling in winter and summer, respectively, from 1967 to 2006, consistent with a decrease in wind strength and direction over the past 60 years. More recently, Barton et al. (2013) show no clear increasing or decreasing trend in wind strength over the past 60 years, and a lack of agreement among wind trends and variability from different wind products (e.g., Pacific Fisheries Environmental Laboratory (PFEL), International Comprehensive Ocean-Atmosphere Data Set (ICOADS), Wave- and Anemometer-based Sea Surface Wind (WASWind)). Barton et al. (2013) present no evidence for changes in upwelling intensity, with the exception of upwelling off northwest Spain, where winds are becoming slightly less favorable. Alteration of wind direction and strength influences upwelling and hence nutrient concentrations; however, nutrient levels can also change in response to other variables such as the supply of iron-laden dust from the Sahara (Alonso-Pérez et al., 2011). There is *medium evidence* and *medium agreement* that primary production in the Canary Current has decreased over the past 2 decades (Arístegui et al., 2009; Demarcq, 2009), in contrast to the nearby upwelling region off northwest Spain where no significant trend was observed (Bode et al., 2011). Satellite chlorophyll records (Sea-viewing Wide Field-of-view Sensor (SeaWiFS), Moderate Resolution Imaging Spectrometer (MODIS)) are relatively short, making it difficult to distinguish the influence of warming oceans from longer term patterns of variability (Arístegui et al., 2009; Henson et al., 2010). Changing temperature has resulted in changes to important fisheries species. For example, Mauritanian waters have become more suitable as feeding and spawning areas for some fisheries species (e.g., *Sardinella aurita*) as temperatures increased (Zeeberg et al., 2008). Clear attribution of these changes depends on the linkage between the Azores High and global temperature, and on longer records for both physical and biological systems, as pointed out for data sets in general (Arístegui et al., 2009; Henson et al., 2010).

30.5.5.1.2. Benguela Current

The Benguela Current originates from the eastward-flowing, cold South Atlantic Current, flows northward along the southwest coast of Africa, and is bounded north and south by the warm-water Angola and Agulhas Currents, respectively. Upwelling is strongest and most persistent toward the center of the system in the Lüderitz-Orange River upwelling cell (Hutchings et al., 2009). Fish catch reached a peak in the late 1970s of 2.8 million tonnes yr^{-1} (www.seaaroundus.org/lme/29/1.aspx), before declines in the northern Benguela, due to overfishing and inter-decadal environmental variability, resulted in a reduced catch of around 1 million tonnes yr^{-1} (present) (Cury and Shannon, 2004; Heymans et al., 2004; Hutchings et al., 2009). Offshore commercial fisheries currently comprise sardine, anchovy, horse mackerel, and hake, while the inshore artisanal and recreational fisheries comprise a variety of fish species mostly caught by hook and line.

Most research on the Benguela Current has focused on fisheries and oceanography, with little emphasis on climate change. As with the other EBUE, strong interannual and inter-decadal variability in physical oceanography make the detection and attribution of biophysical trends to climate change difficult. Nevertheless, the physical conditions of the Benguela Current are highly sensitive to climate variability over a range of scales, especially to atmospheric teleconnections that alter local wind stress (Hutchings et al., 2009; Leduc et al., 2010; Richter et al., 2010; Rouault et al., 2010). Consequently, there is *medium agreement*, despite *limited evidence* (Demarcq, 2009), that upwelling intensity and associated variables (e.g., temperature, nutrient, and O_2 concentrations) from the Benguela system will change as a result of climate change (Box CC-UP).

The temperature of the surface waters of the Benguela Current did not increase from 1950 to 2009 (p-value > +0.05; Table 30-1), although shorter records show an decrease in the south-central Benguela Current (0.35°C to 0.55°C per decade; Rouault et al., 2010) or an increase for the whole Benguela region (0.24°C; Belkin, 2009). These differences between short versus long records indicate the substantial influence of long-term variability on the Benguela system (Belkin, 2009). Information on other potential consequences of climate change within the Benguela system is sparse. SLR is similar to the global mean, although it has not been measured rigorously within the Benguela (Brundrit, 1995; Veitch, 2007). Although upwelling water in the northern and southern portions of the Benguela Current exhibits elevated and suppressed partial pressure of CO_2, respectively (Santana-Casiano et al., 2009), the consequences of changing upwelling intensity remain poorly explored with respect to ocean acidification. Finally, although periodic hypoxic events in the Benguela system are largely driven by natural advective processes, these may be exacerbated by future climate change (Monteiro et al., 2008; Bakun et al., 2010).

Despite its apparent sensitivity to environmental variability, there is *limited evidence* of ecological changes in the Benguela Current EBUE due to climate change (Poloczanska et al., 2013). For example, pelagic fish (Roy et al., 2007), benthic crustaceans (Cockcroft et al., 2008), and seabirds (Crawford et al., 2008) have demonstrated general eastward range shifts around the Cape of Good Hope. Although these may be associated with increased upwelling along the South African south coast, specific studies that attribute these changes to anthropogenic climate change are lacking. Trawl surveys of demersal fish and cephalopod species showed consistently predictable "hotspots" of species richness over a 20- to 30-year study period (the earliest surveys since 1984 off South Africa) that were associated with greater depths and cooler bottom waters (Kirkman et al., 2013). However, major changes in the structure and function of the demersal community have been shown in some parts of the Benguela Current EBUE in response to environmental change, for example, due predominantly to fishing pressure in the 1960s and environmental forcing in the early 2000s in the southern Benguela (Howard et al., 2007); therefore, changes driven by climate change may eventually affect the persistence of these biodiversity hotspots (Kirkman et al., 2013).

30.5.5.1.3. California Current

The California Current spans approximately 23° of latitude from central Baja California, Mexico, to central British Columbia, Canada, linking the North Pacific Current (West Wind Drift) with the North Equatorial and Kuroshio Currents to form the North Pacific Gyre. High productivity driven by advective transport and upwelling (Hickey, 1979; Chelton et al., 1982; Checkley and Barth, 2009; Auad et al., 2011) supports well-studied ecosystems and fisheries. Fish catches have been approximately 0.6 million tonnes yr^{-1} since 1950 (www.seaaroundus.org/lme/3.aspx), which makes it the lowest catch of the four EBUE. The ecosystem supports the foraging and reproductive activities of 2 to 6 million seabirds from around 100 species (Tyler et al., 1993). Marine mammals are diverse and relatively abundant, including recovering populations of humpback whales, among other species (Barlow et al., 2008).

The average temperature of the California Current warmed by 0.73°C from 1950 to 2009 (p-value ≤ 0.05; Table 30-1) and by 0.14°C to 0.80°C from 1985 to 2007 (Demarcq, 2009). Like other EBUE, the California Current is characterized by large-scale interannual and inter-decadal climate-ecosystem variability (McGowan et al., 1998; Hare and Mantua, 2000; Chavez et al., 2003; Checkley and Barth, 2009). During an El Niño, coastally trapped Kelvin waves from the tropics deepen the thermocline, thereby severely reducing upwelling and increasing ocean temperatures from California to Washington (Peterson and Schwing, 2003; King et al., 2011). Atmospheric teleconnections to the tropical Pacific alter wind stress and coastal upwelling. Therefore, the ENSO is intimately linked with Bakun's (1990) upwelling intensification hypothesis (Box CC-UP). Inter-decadal variability in the California Current stems from variability in the Pacific-North America pattern (Overland et al., 2010), which is influenced by the PDO (Mantua et al., 1997; Peterson and Schwing, 2003) and the NPGO (Di Lorenzo et al., 2008). The major effects of the PDO and NPGO appear north of 39°N (Di Lorenzo et al., 2008; Menge et al., 2009).

There is *robust evidence* and *medium agreement* that the California Current has experienced a decrease in the number of upwelling events (23 to 40%), but an increase in duration of individual events, resulting in an increase of the overall magnitude of upwelling events from 1967 to 2010 (*high confidence*; Demarcq, 2009; Iles et al., 2012). This is consistent with changes expected under climate change yet remains complicated by the influence of decadal-scale variability (*low confidence*; Iles et al., 2012). Oxygen concentrations have also undergone large and consistent decreases from 1984 to 2006 throughout the California Current, with the largest relative decreases occurring below the thermocline (21% at 300 m). The hypoxic boundary layer (<60 µmol kg^{-1}) has also shoaled by up to 90 m in some regions (Bograd et al., 2008). These changes are consistent with the increased input of organic carbon into deeper layers from enhanced upwelling and productivity, which stimulates microbial activity and results in the drawdown of O_2 (*likely*, Bakun et al., 2010; but see also McClatchie et al., 2010; Koslow et al., 2011; WGI AR5 Section 3.8.3). These changes are *likely* to have reduced the available habitat for key benthic communities as well as fish and other mobile species (Stramma et al., 2010). Increasing microbial activity will also increase the partial pressure of CO_2, decreasing the pH and carbonate concentration of seawater. Together with the shoaling of the saturation horizon, these changes have increased the incidence of low O_2 and low pH water flowing onto the continental shelf (*high confidence*; 40 to 120 m; Feely et al., 2008), causing problems for industries such as the shellfish aquaculture industry (Barton et al., 2012).

30.5.5.1.4. Humboldt Current

The Humboldt Current is the largest of the four EBUE, covering an area larger than the other three combined. It comprises the eastern edge of the South Pacific Gyre, linking the northern part of the Antarctic Circumpolar Current with the Pacific South Equatorial Current. Although the primary productivity per unit area is modest compared to that of the other EBUE, the total Humboldt Current system has very high levels of fish production. Current catches are in line with a long-term average (since the 1960s) of 8 million tonnes yr^{-1} (www.seaaroundus.org/lme/13/1.aspx), although decadal-scale variations range from 2.5 to 13 million tonnes yr^{-1}. While anchovies currently contribute 80% of the total catch, they alternate with sardines on a multi-decadal scale, with their dynamics mediated by the approach and retreat of subtropical waters to and from the coast (Alheit and Bakun, 2010). This variability does not appear to be changing due to anthropogenic climate change. Thus, from the late 1970s to the early 1990s, sardines were more important (Chavez et al., 2003). The other major commercial fish species are jack mackerel among the pelagic fish and hake among the demersal fish.

The Humboldt Current EBUE did not show an overall warming trend in SST over the last 60 years (p-value > 0.05; Table 30-1), which is consistent with other data sets (1982–2006, HadISST1.1: Belkin, 2009; 1985–2007, Pathfinder: Demarcq, 2009). Wind speed has increased in the central portions of the Humboldt Current, although wind has decreased in its southern and northern sections (Demarcq, 2009). The lack of a consistent warming signal may be due to the strong influence of adjacent ENSO activity exerting opposing drivers on upwelling and which, if they intensify, would decrease temperatures (*limited evidence*, *medium agreement*). Similar to the Canary Current EBUE, however, there was a significant increase in the temperatures of the warmest month of the year over the period 1950–2009 (p-value ≤ 0.05; Table 30-1).

Primary production is suppressed during warm El Niño events and amplified during cooler La Niña phases, these changes then propagate through to higher trophic levels (Chavez et al., 2003; Tam et al., 2008; Taylor et al., 2008). However, in addition to trophic changes, there is also a direct thermal impact on organisms, which varies depending on the thermal adaptation window for each species (*high confidence*). A 37-year zooplankton time series for the coast of Peru showed no persistent trend in abundance and diversity (Ayón et al., 2004), although observed shifts coincided with the shifts in the regional SST. As for other EBUE, there is lack of studies that have rigorously attempted to detect and attribute changes to anthropogenic climate change, although at least two studies (Mendelssohn and Schwing, 2002; Gutiérrez et al.,

2011) provide additional evidence that the northern Humboldt Current has cooled (due to upwelling intensification) since the 1950s, a trend matched by increasing primary production. This is not entirely consistent with the lack of significant change over the period 1950–2009 (p-value > 0.05; Table 30-1). Nevertheless, these relationships are *likely* to be complex in their origin, especially in their sensitivity to the long-term changes associated with ENSO and PDO, and the fact that areas within the Humboldt Current EBUE may be showing different behaviors.

30.5.5.2. Key Risks and Vulnerabilities

EBUE are vulnerable to changes that influence the intensity of currents, upwelling, and mixing (and hence changes in SST, wind strength and direction), as well as O_2 content, carbonate chemistry, nutrient content, and the supply of organic carbon to deep offshore locations (*robust evidence*, *high agreement*; *high confidence*). The extent to which any particular EBUE is vulnerable to these factors depends on location (Figure 3 from Gruber, 2011) and other factors such as alternative sources of nutrient input and fishing pressure (Bakun et al., 2010). This complex interplay between regional and global drivers means that our understanding of how factors such as upwelling within the EBUE will respond to further climate change is uncertain (Box CC-UP; Rykaczewski and Dunne, 2010).

In the GCM ensembles examined (Table SM30-3), modest rates of warming (0.22°C to 0.93°C) occur within the four EBUEs in the near term. Over 2010–2099, however, EBUE SSTs warm by 0.07°C to 1.02°C under RCP2.6, and 2.52°C to 3.51°C under RCP8.5 (Table SM30-4). These high temperatures have the potential to increase stratification of the water column and substantially reduce overall mixing in some areas. In contrast, the potential strengthening of coastal wind systems would intensify upwelling and stimulate primary productivity through the increased injection of nutrients into the photic zone of the EBUE (Box CC-UP). Garreaud and Falvey (2009) explored how wind stress along the South American coast would change by 2100 under SRES B2 and A2 scenarios. Using an ensemble of 15 GCMs, southerly wind systems upwelling increased along the subtropical coast of South America, extending and strengthening conditions for upwelling.

Changes in the intensity of upwelling within the EBUE will drive fundamental changes to the abundance, distribution, and viability of resident organisms, although an understanding of their nature and direction is limited. In some cases, large-scale decreases in primary productivity and dependent fisheries are projected to occur for EBUE ecosystems (Blanchard et al., 2012), while other projections question the strong connection between primary productivity and fisheries production (Arístegui et al., 2009). Increased upwelling intensity also has potential disadvantages. Elevated primary productivity may lead to decreasing trophic transfer efficiency, thus increasing the amount of organic carbon exported to the seabed, where it is *virtually certain* to increase microbial respiration and hence increase low O_2 stress (Weeks et al., 2002; Bakun et al., 2010). Increased wind stress may also increase turbulence, breaking up food concentrations (affecting trophic transfer), or causing excessive offshore advection, which could remove plankton from shelf habitats. The central issue for the EBUE is therefore whether or not upwelling will intensify and, if so, whether the negative consequences (e.g., reduced O_2 and elevated CO_2) associated with upwelling intensification will outweigh potential benefits from increased primary production and fisheries catch.

30.5.6. Subtropical Gyres

Subtropical gyres (STG) dominate the Pacific, Atlantic, and Indian Oceans (Figure 30-1a), and consist of large stable water masses that circulate clockwise (Northern Hemisphere) and anticlockwise (Southern Hemisphere) due to the Coriolis Effect. The oligotrophic areas at the core of the STG represent one of the largest habitats on Earth, contributing 21.2% of ocean primary productivity and 8.3% of the

Frequently Asked Questions

FAQ 30.4 | Will climate change increase the number of "dead zones" in the oceans?

Dissolved oxygen is a major determinant of the distribution and abundance of marine organisms. Dead zones are persistent hypoxic conditions where the water doesn't have enough dissolved oxygen to support oxygen-dependent marine species. These areas exist all over the world and are expanding, with impacts on coastal ecosystems and fisheries (*high confidence*). Dead zones are caused by several factors, particularly eutrophication where too many nutrients run off coastal cities and agricultural areas into rivers that carry these materials out to sea. This stimulates primary production, leading to a greater supply of organic carbon, which can sink into the deeper layers of the ocean. As microbial activity is stimulated, there is a sharp reduction in dissolved oxygen levels and an increased risk of dead zones (*high confidence*). Climate change can influence the distribution of dead zones by increasing water temperature and hence microbial activity, as well as reducing mixing (i.e., increasing layering or stratification) of the Ocean, thereby reducing mixing of oxygen-rich surface layers into the deeper parts of the Ocean. In other areas, increased upwelling can lead to stimulated productivity, which can also lead to more organic carbon entering the deep ocean, where it is consumed, decreasing oxygen levels (*medium confidence*). Managing local factors such as the input of nutrients into coastal regions can play an important role in reducing the rate at which dead zones are spreading across the world's oceans (*high agreement*).

global fish catch (Figure 30-1b; Table SM30-1). A number of small island nations are found within this region. While many of the observed changes within these nations have been described in previous chapters (e.g., Sections 5.3-4, 29.3-5), region-wide issues and consequences are discussed here due to the strong linkages between ocean and coastal issues.

30.5.6.1. Observed Changes and Potential Impacts

The central portions of the STG are oligotrophic (Figure SM30-1). Temperatures within the STG of the North Pacific (NPAC), South Pacific (SPAC), Indian Ocean (IOCE), North Atlantic (NATL), and South Atlantic (SATL) have increased at rates of 0.020°C, 0.024°C, 0.032°C, 0.025°C, and 0.027°C yr^{-1} from 1998 to 2010, respectively (Signorini and McClain, 2012). This is consistent with increases observed from 1950 to 2009 (0.25°C to 0.67°C; Table 30-1). However, differences among studies done over differing time periods emphasize the importance of long-term patterns of variability. Salinity has decreased across the North and South Pacific STG (Figure 30-6c; WGI AR5 Section 3.3.3.1), consistent with warmer sea temperatures and an intensification of the hydrological cycle (Boyer, 2005).

The North and South Pacific STG have expanded since 1993 (*high confidence*), with these changes *likely* being the consequence of a combination of wind forcing and long-term variability (Parrish et al., 2000; WGI AR5 Section 3.6.3). Chlorophyll levels, as determined by remote-sensing of ocean color (Box CC-UP), have decreased in the NPAC, IOCE, and NATL by 9, 12, and 11%, respectively (p-value ≤ 0.5; Signorini and McClain, 2012) over and above the inherent seasonal and interannual variability from 1998 to 2010 (Vantrepotte and Mélin, 2011). Chlorophyll levels did not change in the remaining two gyres (SPAC and SATL, and confirmed for SPAC by Lee and McPhaden (2010) and Lee et al. (2010)). Furthermore, over the period 1998–2007, median cell diameter of key phytoplankton species exhibited statistically significant linear declines of about 2% in the North and South Pacific, and 4% in the North Atlantic Ocean (Polovina and Woodworth, 2012). Changes in chlorophyll and primary productivity in these sub-regions have been noted before (McClain et al., 2004; Gregg et al., 2005; Polovina et al., 2008) and are influenced by seasonal and longer-term sources of variability (e.g., ENSO, PDO; Section 6.3.4; Figure 6-9). These changes represent a significant expansion of the world's most unproductive waters, although caution must be exercised given the limitations of satellite detection methods (Box CC-PP) and the shortness of records relative to longer-term patterns of climate variability. There is *high confidence* that changes that reduce the vertical transport of nutrients into the euphotic zone (e.g., decreased wind speed, increasing surface temperatures, and stratification) will reduce the rate of primary productivity and hence fisheries.

30.5.6.1.1. Pacific Ocean Subtropical Gyres

Pacific climate is heavily influenced by the position of the Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), which are part of the ascending branch of the Hadley circulation (WGI AR5 Section 14.3.1). These features are also strongly influenced by interannual to inter-decadal climate patterns of variability including ENSO and PDO. The current understanding of how ENSO and PDO will change as average global temperatures increase is not clear (*low confidence*; Collins et al., 2010; WGI AR5 Section 12.4.4.2). The position of both the ITCZ and SPCZ vary seasonally and with ENSO (Lough et al., 2011), with a northward migration during the Northern Hemisphere summer and a southward migration during the Southern Hemisphere summer. These changes, along with the West Pacific Monsoon, determine the timing and extent of the wet and dry seasons in SPAC and NPAC sub-regions (Ganachaud et al., 2011). Tropical cyclones are prominent in the Pacific (particularly the western Pacific), and CBS sub-regions between 10° and 30° north and south of the equator, although the associated storm systems may occasionally reach higher latitudes. Spatial patterns of cyclones vary with ENSO, spreading out from the Coral Sea to the Marquesas Islands during El Niño and contracting back to the Coral Sea, New Caledonia, and Vanuatu during La Niña (Lough et al., 2011). Historically, there have been almost twice as many land-falling tropical cyclones in La Niña as opposed to El Niño years off the east coast of Australia, with a declining trend in the number of severe tropical cyclones from 0.45 per year in the early 1870s to 0.17 per year in recent times (Callaghan and Power, 2011).

The Pacific Ocean underwent an abrupt shift to warmer sea temperatures in the mid-1970s as a result of both natural (e.g., IPO) and climate forcing (*high confidence*; Meehl et al., 2009). This change coincided with changes to total rainfall, rain days, and dry spells across the Pacific, with the direction of change depending on the location relative to the SPCZ. Countries such as the Cook Islands, Tonga, Samoa and American Samoa, and Fiji tend to experience drought conditions as the SPCZ (with cooler sea temperatures) moves toward the northeast during El Niño (*high confidence*). The opposite is true during La Niña conditions. The consequences of changing rainfall on the countries of the Pacific STG are discussed in greater detail elsewhere (Sections 5.4, 29.3; Table 29-1). Although these changes are due to different phases of long-term variability in the Pacific, they illustrate the ramifications and sensitivity of the Pacific to changes in climate change.

Elevated sea temperatures within the Pacific Ocean have increased the frequency of widespread mass coral bleaching and mortality since the early 1980s (*very high confidence*; Hoegh-Guldberg and Salvat, 1995; Hoegh-Guldberg, 1999; Mumby et al., 2001; Baker et al., 2008; Donner et al., 2010). There are few, if any, scientific records of mass coral bleaching and mortality prior to this period (*high confidence*; Hoegh-Guldberg, 1999). Rates of decline in coral cover on coastal coral reef ecosystems range between 0.5 and 2.0% per year depending on the location within the Indo-Pacific region (*high confidence*; Bruno and Selig, 2007; Hughes et al., 2011; Sweatman et al., 2011; De'ath et al., 2012). The reasons for this decline are complex and involve non-climate change-related factors (e.g., coastal pollution and overfishing) as well as global warming and possibly acidification. A recent comprehensive analysis of the ecological consequences of coral bleaching and mortality concluded that "bleaching episodes have resulted in catastrophic loss of coral reefs in some locations, and have changed coral community structure in many others, with a potentially critical influence on the maintenance of biodiversity in the marine tropics" (*high confidence*; Baker et al., 2008, p. 435). Increasing sea levels have also caused changes in seagrass and mangrove systems. Gilman et al. (2007) found a reduction in mangrove area with SLR, with the observed mean landward recession of three mangrove areas over 4 decades being 25, 64, and 72 mm yr^{-1}, 12 to 37 times faster than the observed rate of SLR. Significant interactions exist between climate change and coastal development, where migration shoreward depends on the extent to which coastlines have been modified or barriers to successful migration have been established.

Changes in sea temperature also lead to changes in the distribution of key pelagic fisheries such as skipjack tuna (*Katsuwonus pelamis*), yellowfin tuna (*Thunnus albacares*), big-eye tuna (*T. obesus*), and South Pacific albacore tuna (*T. alalunga*), which make up the majority of key fisheries in the Pacific Ocean. Changes in distribution and recruitment in response to changes in sea temperature as result of ENSO demonstrate the close association of pelagic fish stocks and water temperature. The shift in habitat for top predators in the northeast Pacific was examined by Hazen et al. (2012), who used tracking data from 23 marine species and associated environmental variables to predict changes of up to 35% in core habitat for these species within the North Pacific. Potential habitats are predicted to contract for the blue whale, salmon shark, loggerhead turtle, and blue and mako sharks, while potential habitats for the sooty shearwater; black -footed albatross; leatherback turtle; white shark; elephant seal; and albacore, bluefin and yellowfin tuna are predicted to expand (Hazen et al., 2012). However, expansion of OMZs in the Pacific STG is predicted to compress habitat (depth) for hypoxia-intolerant species such as tuna (Stramma et al., 2010, 2012).

Reduction of ocean productivity of the STG (Sarmiento et al., 2004; Signorini and McClain, 2012) reduces the flow of energy to higher trophic levels such as those of pelagic fish (Le Borgne et al., 2011). The distribution and abundance of fisheries stocks such as tuna are also sensitive to changes in sea temperature, and hence long-term variability such as ENSO and PDO. The redistribution of tuna in the western central equatorial region has been related to the position of the oceanic convergence zones, where the warm pool meets the cooler tongue of the Pacific. These changes have been reliably reproduced by population models that use temperature as a driver of the distribution and abundance of tuna (Lehodey et al., 1997, 2006). Projections of big-eye tuna (*T. obesus*) distributions under SRES A2 show an improvement in spawning and feeding habitats by 2100 in the eastern tropical Pacific and declines in the western tropical Pacific, leading to an eastern displacement of tuna stocks (Lehodey et al., 2008, 2010b).

30.5.6.1.2. Indian Ocean Subtropical Gyre

Like the Pacific Ocean, the Indian Ocean plays a crucial role in global weather patterns, with teleconnections throughout Africa, Australasia, Asia, and the Americas (e.g., Clark et al., 2000; Manhique et al., 2011; Meehl and Arblaster, 2011; Nakamura et al., 2011). Increasing sea level, temperature, storm distribution and intensity, and changing seawater chemistry all influence the broad range of physical, chemical, and biological aspects of the Indian Ocean. Coral reef ecosystems in the Indian Ocean gyre system were heavily affected by record positive sea temperature anomalies seen in the Southern Hemisphere between February to April 1998 (*robust evidence*, *high agreement*; *high confidence*; Ateweberhan et al., 2011). Coral cover across the Western Indian Ocean declined by an average of 37.7% after the 1998 heat stress event (Ateweberhan et al., 2011). Responses to the anomalously warm conditions in 1998 varied between sub-regions, with the central Indian Ocean islands (Maldives, Seychelles, Chagos, and Lakshadweep) experiencing major decreases in coral cover directly after the 1998 event (from 40 to 53% coral cover in 1977–1997 to 7% in 1999–2000; *high confidence*; Ateweberhan et al., 2011). Coral reefs lining the islands of southern India and Sri Lanka experienced similar decreases in coral cover (45%, 1977–1997 to 12%, 1999–2000). Corals in the southwestern Indian Ocean (Comoros, Madagascar, Mauritius, Mayotte, Réunion, and Rodrigues) showed less impact (44%, 1977–1997 to 40%, 1999–2000). Recovery from these increases in mortality has been variable, with sites such as those around the central Indian Ocean islands exhibiting fairly slow recovery (13% by 2001–2005) while those around southern India and Sri Lanka are showing much higher rates (achieving a mean coral cover of 37% by 2001–2005; Ateweberhan et al., 2011). These changes to the population size of key reef-building species will drive major changes in the abundance and composition of fish populations in coastal areas, and affect other ecosystem services that are important for underpinning tourism and coastal protection (*medium confidence*; Box CC-CR).

Fisheries that exploit tuna and other large pelagic species are very valuable to many small island states within the Indian Ocean. As with Pacific fisheries, the distribution and abundance of large pelagic fish in the Indian Ocean is greatly influenced by sea temperature. The anomalously high sea temperatures of 1997–1998 (leading to a deepening of the mixed layer in the west and a shoaling in the east) coincided with anomalously low primary production in the Western Indian Ocean and a major shift in tuna stocks (*high confidence*; Menard et al., 2007; Robinson et al., 2010). Fishing grounds in the Western Indian Ocean were deserted and fishing fleets underwent a massive shift toward the eastern basin, which was unprecedented for the tuna fishery (*high confidence*). As a result of these changes, many countries throughout the Indian Ocean lost significant tuna-related revenue (Robinson et al., 2010). In 2007, tuna fishing revenue was again reduced by strong surface warming and deepening of the mixed layer, and associated with a modest reduction in primary productivity in the west. These trends highlight the overall vulnerability of tuna fishing countries in the Indian Ocean to climate variability, a situation similar to that in the other major oceans of the world.

30.5.6.1.3. Atlantic Ocean Subtropical Gyres

SST has increased within the two STG of the Atlantic Ocean over the last 2 decades (Belkin, 2009; Signorini and McClain, 2012). Over longer periods of time (1950–2009), trends in average temperature are not significant for the North Atlantic STG (p-value > 0.05) while they remain so for the South Atlantic STG (*very likely*; 0.08°C per decade, p-value $\leq$ 0.05; Table 30-1). In both cases, however, temperatures in the coolest and warmest months increased significantly (Table 30-1). The difference between these studies (i.e., over 10 to 30 years vs. 60 years) emphasizes the importance of long-term patterns of variability in the North Atlantic region. Variability in SST at a period of about 60 to 80 years is associated with the Atlantic Multi-decadal Oscillation (AMO; Trenberth and Shea, 2006). Sea surface temperatures influence hurricane activity (*very likely*) with recent record SST associated with record hurricane activity in 2005 in the Atlantic (Trenberth and Shea, 2006) and mass coral bleaching and mortality in the eastern Caribbean (*high confidence*; Eakin et al., 2010). In the former case, analysis concluded that 0.1°C of the SST anomaly was attributable to the state of the AMO while 0.45°C was due to ocean warming as a result of anthropogenic influences (Trenberth and Shea, 2006).

These changes have influenced the distribution of key fishery species as well the ecology of coral reefs in Bermuda (Wilkinson and Hodgson, 1999; Baker et al., 2008) and in the eastern Caribbean (Eakin et al., 2010). Small island nations such as Bermuda depend on coral reefs for fisheries and tourism and are vulnerable to further increases in sea temperature that cause mass coral bleaching and mortality (*high confidence*; Box CC-CR; Figure 30-10). As with the other STG, phytoplankton communities and pelagic fish stocks are sensitive to temperature changes that have occurred over the past several decades. Observation of these changes has enabled development of models that have a high degree of accuracy in projecting the distribution and abundance of these elements within the Atlantic region in general (Cheung et al., 2011).

30.5.6.2. Key Risks and Vulnerabilities

SSTs of the vast STGs of the Atlantic, Pacific, and Indian Oceans are increasing, which is *very likely* to increase stratification of the water column. In turn, this is *likely* to reduce surface concentrations of nutrients and, consequently, primary productivity (*medium confidence*; Box CC-PP). Warming is projected to continue (Table SM30-4), with substantial increases in the vulnerability and risk associated with systems that have been observed to change so far (*high confidence*; Figure 30-12). Under RCP2.6, the temperatures of the STG are projected to increase by 0.17°C to 0.56°C in the near term (over 2010–2039) and between –0.03°C to 0.90°C in the long term (over 2010–2099) (Table SM30-4). Under RCP8.5, however, surface temperatures of the world's STG are projected to be 0.45°C to 0.91°C warmer in the near term and 1.90°C to 3.44°C warmer in the long term (Table SM30-4). These changes in temperature are *very likely* to increase water column stability, reduce the depth of the mixed layer, and influence key parameters such as nutrient availability and O_2 concentrations. It is not clear as to how longer-term sources of variability such as ENSO and PDO will change (WGI AR5 Sections 14.4, 14.7.6) and ultimately influence these trends.

The world's most oligotrophic ocean sub-regions are *likely* to continue to expand over coming decades, with consequences for ecosystem services such as gas exchange, fisheries, and carbon sequestration. Polovina et al. (2011) explored this question for the North Pacific using a climate model that included a coupled ocean biogeochemical component to investigate potential changes under an SRES A2 scenario (~RCP6.0 to RCP8.5; see also Figure 1.5 from Rogelj et al., 2012). Model projections indicated the STG expanding by approximately 30% by 2100, driven by the northward drift of the mid-latitude westerlies and enhanced stratification of the water column. The expansion of the STG occurred at the expense of the equatorial upwelling and other regions within the North Pacific. In the North Pacific STG, the total primary production is projected to decrease by 10 to 20% and large fish catch by 19 to 29% by 2100 under SRES A2 (Howell et al., 2013; Woodworth-Jefcoats et al., 2013). However, our understanding of how large-scale eddy systems will change in a warming world is incomplete, as are the implications for primary productivity of these large and important systems (Boxes CC-PP, CC-UP).

Understanding how storm frequency and intensity will change represents a key question for many countries and territories within the various STG. Projections of increasing sea temperature are *likely* to change the behavior of tropical cyclones. At the same time, the maximum wind speed and rainfall associated with cyclones is *likely* to increase, although future trends in cyclones and severe storms are *very likely* to vary from region to region (WGI AR5 Section 14.6). Patterns such as "temporal clustering" can have a strong influence on the impact of tropical cyclones on ecosystems such as coral reefs (Mumby et al., 2011), although how these patterns will change within all STG is uncertain at this point. However, an intensifying hydrological cycle is expected to increase precipitation in many areas (*high confidence*; WGI AR5 Sections 2.5, 14.2), although longer droughts are also expected in other STG (*medium confidence*). Changes in the hydrological cycle impact coastal ecosystems, increasing damage through coastal flooding and physical damage from storm waves (Mumby et al., 2011). Improving our understanding of how weather systems associated with features such as the SPCZ (WGI AR5 Section 14.3.1) will vary is critical to climate change adaptation of a large number of nations associated with the STG. Developing an understanding of how ocean temperature, climate systems such as the SPCZ and ITCZ, and climate change and variability (e.g., ENSO, PDO) interact will be essential in this regard. For example, variability in the latitude of the SPCZ is projected to increase, possibly leading to more extreme events in Pacific Island countries (Cai et al., 2012).

The consequences of projected sea temperatures on the frequency of coral bleaching and mortality within key sub-regions of the STG are outlined in Box CC-CR and Figures 30-10 and SM30-3. As with other sub-regions (particularly CBS, STG, and SES) dominated by coral reefs, mass coral bleaching and mortality becomes an annual risk under all scenarios, with mass mortality events beginning to occur every 1 to 2 years by 2100 (*virtually certain*; Box CC-CR; Figures 30-10, SM30-3). Coral-dominated reef ecosystems (areas with more than 30% coral cover) are *very likely* to disappear under these circumstances by the mid part of this century (van Hooidonk et al., 2013). The loss of substantial coral communities has implications for the three-dimensional structure of coral reefs (Box CC-CR) and the role of the latter as habitat for organisms such as fish (Hoegh-Guldberg, 2011; Hoegh-Guldberg et al., 2011a; Pratchett et al., 2011a; Bell et al., 2013b).

The consequences of increasing sea temperature can be exacerbated by increasing ocean acidification, with potential implications for reef calcification (*medium confidence*; Kleypas et al., 1999; Hoegh-Guldberg et al., 2007; Doney et al., 2009), reef metabolism and community calcification (Dove et al., 2013), and other key ecological processes (Pörtner et al., 2001, 2007; Munday et al., 2009). Ocean pH within the STG will continue to decrease as atmospheric CO_2 increases, bringing pH within the STG to 7.9 and 7.7 at atmospheric concentrations of 450 ppm and 800 ppm, respectively (Figure SM30-2a; Box CC-OA). Aragonite saturation states will decrease to around 1.6 (800 ppm) and 3.3 (450 ppm; Figure SM30-2b). Decreasing carbonate ion concentrations and saturation states pose serious risks to other marine calcifiers such as encrusting coralline algae, coccolithophores (phytoplankton), and a range of benthic invertebrates (Doney et al., 2009; Feely et al., 2009).

Increasing sea temperatures and sea level are also *likely* to influence other coastal ecosystems (e.g., mangroves, seagrass meadows) in the Pacific, although significant gaps and uncertainties exist (Section 29.3.1.2; Waycott et al., 2007, 2011). Many of the negative consequences for coral reefs, mangroves, and seagrass meadows are *likely* to have negative consequences for dependent coastal fisheries (through habitat destruction) and tourism industries (*medium confidence*; Bell et al., 2011a, 2013a; Pratchett et al., 2011a,b).

Populations of key large pelagic fish are projected to move many hundreds of kilometers east of where they are today in the Pacific STG (*high confidence*; Lehodey et al., 2008, 2010a, 2011, 2013), with implications for income, industry, and food security across multiple Pacific Island nations (*high confidence*; Cheung et al., 2010; McIlgorm et al., 2010; Bell et al., 2011b, 2013a; Section 7.4.2; Tables 29-2, 29-4). These predictions of species range displacements, contractions, and expansions in response to anticipated changes in the Ocean (Box CC-MB) present both a challenge and an opportunity for the development of large-scale management strategies to preserve these valuable species. Our understanding of the consequences of reduced O_2 for pelagic fish populations is not clear, although there is *high agreement* on the potential physiological outcomes (Section 6.3.3). Those species that are intolerant to hypoxia, such as skipjack and yellowfin tuna (Lehodey et al., 2011), will have their depth range compressed in the Pacific STG, which will increase their vulnerability to fisheries and reduce overall fisheries habitat and productivity (*medium confidence*; Stramma et al., 2010, 2011). Despite the importance of these potential changes, our understanding of the full range of consequences is *limited* at this point.

30.5.7. Deep Sea (>1000 m)

Assessments of the influence of climate change on the Deep Sea (DS) are challenging because of difficulty of access and scarcity of long-term, comprehensive observations (Smith, Jr. et al., 2009). The size of this habitat is also vast, covering well over 54% of the Earth's surface and stretching from the top of the mid-oceanic ridges to the bottom of deep ocean trenches (Smith, Jr. et al., 2009). The fossil record in marine sediments reveals that the DS has undergone large changes in response to climate change in the past (Knoll and Fischer, 2011). The paleo-skeletal record shows that it is the rate, not just the magnitude, of climate change (temperature, O_2, and CO_2) that is critical to marine life in DS. The current rate of change in key parameters *very likely* exceeds that of other major events in Earth history. Two primary time scales are of interest. The first is the slow rate (century-scale) of ocean circulation and mixing, and consequently the slow rate at which DS ecosystems experience physical climate change. The second is the rapid rate at which organic matter enters the deep ocean from primary productivity generated at the surface of the Ocean, which represents a critical food supply to DS animals (Smith, Jr. and Kaufmann, 1999; Smith, Jr. et al., 2009). It can also represent a potential risk in some circumstances where the flux of organic carbon into the deep ocean, coupled with increased sea temperatures, can lead to anoxic areas (dead zones) as metabolism is increased and O_2 decreased (Chan et al., 2008; Stramma et al., 2010).

30.5.7.1. Observed Changes and Potential Impacts

The greatest rate of change of temperature is occurring in the upper 700 m of the Ocean (*very high confidence*; WGI AR5 Section 3.2), although smaller yet significant changes are occurring at depth. The DS environment is typically cold (~–0.5°C to 3°C; Smith et al., 2008), although abyssal temperatures in the SES can be higher (e.g., Mediterranean DS ~12°C; Danovaro et al., 2010). In the latter case, DS organisms can thrive in these environments as well, illustrating the variety of temperature conditions that differing species of abyssal life have adapted to. Individual species, however, are typically constrained within a narrow thermal and O_2-demand window of tolerance (Pörtner, 2010) and therefore it is *likely* that shifts in the distribution of DS species and regional extinctions will occur. Warming over multiple decades has been observed below 700 m (Levitus et al., 2005, 2009), with warming being minimal at mid-range depths (2000 to 3000 m), and increasing toward the sea floor in some sub-regions (e.g., Southern Ocean; WGI AR5 Chapter 3). For the deep Atlantic Ocean, the mean age of deep waters (mean time since last exposure to the atmosphere) is approximately 250 years; the oldest deep waters of the Pacific Ocean are >1000 years old. The patterns of ocean circulation are clearly revealed by the penetration of tracers and the signal of CO_2 released from burning fossil fuel penetrating into the abyss (Sabine et al., 2004). It will take many centuries for full equilibration of deep ocean waters and their ecosystems with recent planetary warming and CO_2 levels (Wunsch and Heimbach, 2008).

Temperature accounts for approximately 86% of the variance in the export of organic matter to the DS (*medium confidence*; Laws et al., 2000). Consequently, upper ocean warming will reduce the export of organic matter to the DS (*medium confidence*), potentially changing the distribution and abundance of DS organisms and associated food webs, and ecosystem processes (Smith, Jr. and Kaufmann, 1999). Most organic matter entering the DS is recycled by microbial systems at relatively shallow depths (Buesseler et al., 2007), and at rates that are temperature dependent. Upper ocean warming will increase the rate of sub-surface decomposition of organic matter (*high confidence*), thus intensifying the intermediate depth OMZs (Stramma et al., 2008, 2010) and reducing food supply to the abyssal ocean.

Particulate organic carbon is exported from the surface to deeper layers of the Ocean (>500 m) with an efficiency of between 20 and 50% (Buesseler et al., 2007), much of it being recycled by microbes before it reaches 1000 m (Smith, Jr. et al., 2009). The export of organic carbon is dependent on surface net primary productivity, which is *likely* to vary (Box CC-PP), influencing the supply of food to DS (Laws et al., 2000; Smith et al., 2008). Warming of intermediate waters will also increase respiration at mid-water depths, reducing the flux of organic carbon. Our understanding of other components of DS ecosystems is also relatively poor. For example, there is *limited evidence* and *limited agreement* as to how ocean warming and acidification are *likely* to affect ecosystems such as those associated with hydrothermal vents (Van Dover, 2012).

Oxygen concentrations are decreasing in the DS (Stramma et al., 2008; Helm et al., 2011a). Although the largest signals occur at intermediate water depths < 1000 m (Nakanowatari et al., 2007; Whitney et al., 2007; Falkowski et al., 2011), some waters >1000 m depth are also experiencing a decline (Jenkins, 2008). The quantity of dissolved O_2

throughout the Ocean will be reduced with warming due to direct effects on solubility (*high confidence*), with these effects being widely distributed (Shaffer et al., 2009). It is also *virtually certain* that metabolic rates of all animals and microbial respiration rates will increase with temperature (Brown et al., 2004). Thus, increased microbial activity and reduced O_2 solubility at higher temperatures will have additive consequences for the decline of O_2 (*high confidence*) even in the DS. The DS waters are relatively well oxygenated owing to the higher solubility of O_2 in colder waters and the low supply rate of organic matter to great depths. The availability of oxygen to marine animals is governed by a combination of concentration, temperature, pressure, and related properties such as diffusivity. Analysis by Hofmann et al. (2013) reveals that the supply potential of oxygen to marine animals in cold deep waters is similar to that at much shallower depths (*very high confidence*).

Anthropogenic CO_2 has penetrated to at least 1000 m in all three ocean basins (particularly the Atlantic; Doney et al., 2009). Further declines of calcite and aragonite in already under-saturated DS water will presumably decrease biological carbonate structure formation and increase dissolution, as has happened many times in Earth's past (*high confidence*; Zeebe and Ridgwell, 2011). Some cold-water corals (reported down to 3500 m) already exist in waters under-saturated with respect to aragonite (Lundsten et al., 2009). Although initial investigations suggested that ocean acidification (reduced by 0.15 and 0.30 pH units) would result in a reduction in the calcification rate of deep water corals (30 and 56%, respectively), accumulating evidence shows that ocean acidification may have far less impact than previously anticipated on the calcification of some deep water corals (*limited evidence*, *medium agreement*; *low confidence*) although it may reduce important habitats given that dead unprotected coral mounds are *likely* to dissolve in under-saturated waters (Thresher et al., 2011; Form and Riebesell, 2012; Maier et al., 2013).

30.5.7.2. Key Risks and Vulnerabilities

Rising atmospheric CO_2 poses a risk to DS communities through increasing temperature, decreasing O_2 and pH, and changing carbonate chemistry (*high confidence*; Keeling et al., 2010). Risks associated with the DS have implications for the Ocean and planet given the high degree of inherent dependency and connectivity. The resulting changes to the flow of organic carbon to some parts of the DS (e.g., STG) are *very likely* to affect DS ecosystems (*medium confidence*; Smith et al., 2008). As with the Ocean generally, there is a need to fill in the substantial gaps that exist in our knowledge and understanding of the world's largest habitat and its responses to rapid anthropogenic climate change.

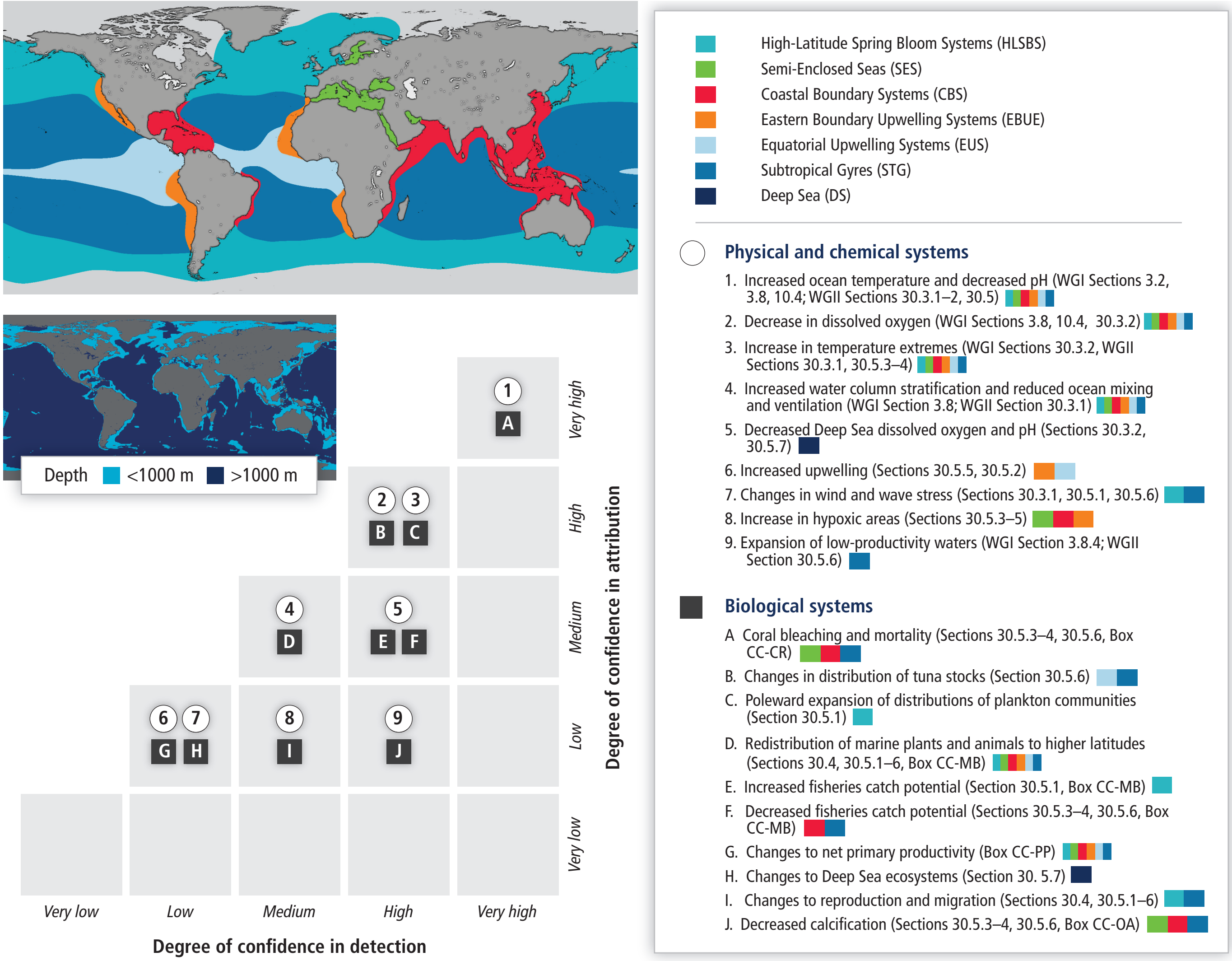
30.5.8. Detection and Attribution of Climate Change Impacts with Confidence Levels

The analysis in this chapter and elsewhere in AR5 has identified a wide range of physical, chemical, and ecological components that have changed over the last century (Box CC-MB). Figure 30-11 summarizes a number of examples from the Ocean as a region together with the degree of confidence in both the detection and attribution steps. For ocean warming and acidification, confidence is *very high* that changes are being detected and that they are due to changes to the atmospheric GHG content. There is considerable confidence in both the detection (*very high confidence*) and attribution (*high confidence*) of mass coral bleaching and mortality, given the well-developed understanding of environmental processes and physiological responses driving these events (Box CC-CR; Section 6.3.1). For other changes, confidence is lower, either because detection of changes has been difficult, or monitoring programs are not long established (e.g., field evidence of declining calcification), or because detection has been possible but models are in conflict (e.g., wind-driven upwelling). The detection and attribution of recent changes is discussed in further detail in Sections 18.3.3-4.

30.6. Sectoral Impacts, Adaptation, and Mitigation Responses

Human welfare is highly dependent on ecosystem services provided by the Ocean. Many of these services are provided by coastal and shelf areas, and are consequently addressed in other chapters (e.g., Sections 5.4.3, 7.3.2.4, 22.3.2.3). Oceans contribute provisioning (e.g., food, raw materials; see Section 30.6.2.1), regulating (e.g., gas exchange, nutrient recycling, carbon storage, climate regulation, water flux), supporting (e.g., habitat, genetic diversity), and cultural (e.g., recreational, religious) services (MEA, 2005; Tallis et al., 2013). The accumulating evidence indicating that fundamental ecosystem services within the Ocean are shifting rapidly should be of major concern, especially with respect to the ability of regulating and supporting ecosystem services to underpin current and future human population demands (Rockström et al., 2009; Ruckelshaus et al., 2013). Discussion here is restricted to environmental, economic, and social sectors that have direct relevance to the Ocean—namely natural ecosystems, fisheries and aquaculture, tourism, shipping, oil and gas, human health, maritime security, and renewable energy. The influences of climate change on Ocean sectors will be mediated through simultaneous changes in multiple environmental and ecological variables (see Figure 30-12), and the extent to which changes can be adapted to and/or risks mitigated (Table 30-3). Both short- and longer-term adaptation is necessary to address impacts arising from warming, even under the lowest stabilization scenarios assessed.

Sectoral approaches dominate resource use and management in the Ocean (e.g., shipping tends to be treated in isolation from fishing within an area), yet cumulative and interactive effects of individual stressors are known to be ubiquitous and substantial (Crain et al., 2008). Climate change consistently emerges as a dominant stressor in regional- to global-scale assessments, although land-based pollution, commercial fishing, invasive species, coastal habitat modification, and commercial activities such as shipping all rank high in many places around the world (e.g., Sections 5.3.4, 30.5.3-4; Halpern et al., 2009, 2010). Such cumulative effects pose challenges to managing for the full suite of stressors to marine systems, but also present opportunities where mitigating a few key stressors can potentially improve overall ecosystem condition (e.g., Halpern et al., 2010; Kelly et al., 2011). The latter has often been seen as a potential strategy for reducing negative consequences of climate impacts on marine ecosystems by boosting ecosystem resilience, thus buying time while the core issue of reducing GHG emissions is tackled (West et al., 2009).



High-Latitude Spring Bloom Systems (HLSBS)
Semi-Enclosed Seas (SES)
Coastal Boundary Systems (CBS)
Eastern Boundary Upwelling Systems (EBUE)
Equatorial Upwelling Systems (EUS)
Subtropical Gyres (STG)
Deep Sea (DS)

Physical and chemical systems

1. Increased ocean temperature and decreased pH (WGI Sections 3.2, 3.8, 10.4; WGII Sections 30.3.1–2, 30.5)
2. Decrease in dissolved oxygen (WGI Sections 3.8, 10.4, 30.3.2)
3. Increase in temperature extremes (WGI Sections 30.3.2, WGII Sections 30.3.1, 30.5.3–4)
4. Increased water column stratification and reduced ocean mixing and ventilation (WGI Section 3.8; WGII Section 30.3.1)
5. Decreased Deep Sea dissolved oxygen and pH (Sections 30.3.2, 30.5.7)
6. Increased upwelling (Sections 30.5.5, 30.5.2)
7. Changes in wind and wave stress (Sections 30.3.1, 30.5.1, 30.5.6)
8. Increase in hypoxic areas (Sections 30.5.3–5)
9. Expansion of low-productivity waters (WGI Section 3.8.4; WGII Section 30.5.6)

Biological systems

A Coral bleaching and mortality (Sections 30.5.3–4, 30.5.6, Box CC-CR)
B. Changes in distribution of tuna stocks (Section 30.5.6)
C. Poleward expansion of distributions of plankton communities (Section 30.5.1)
D. Redistribution of marine plants and animals to higher latitudes (Sections 30.4, 30.5.1–6, Box CC-MB)
E. Increased fisheries catch potential (Section 30.5.1, Box CC-MB)
F. Decreased fisheries catch potential (Sections 30.5.3–4, 30.5.6, Box CC-MB)
G. Changes to net primary productivity (Box CC-PP)
H. Changes to Deep Sea ecosystems (Section 30. 5.7)
I. Changes to reproduction and migration (Sections 30.4, 30.5.1–6)
J. Decreased calcification (Sections 30.5.3–4, 30.5.6, Box CC-OA)

Figure 30-11 | Expert assessment of degree of confidence in detection and attribution of physical and chemical changes (white circles) and ecological changes (dark gray squares) across sub-regions, as designated in Figure 30-1a, and processes in the Ocean (based on evidence explored throughout Chapter 30 and elsewhere in AR5). Further explanation of this figure is given in Sections 18.3.3-4 and 18.6.

30.6.1. Natural Ecosystems

Adaptation in natural ecosystems may occur autonomously, such as tracking shifts in species' composition and distributions (Poloczanska et al., 2013), or engineered by human intervention, such as assisted dispersal (Section 4.4.2.4; Hoegh-Guldberg et al., 2008). Currently, adaptation strategies for marine ecosystems include reducing additional stressors (e.g., maintaining water quality, adapting fisheries management) and maintaining resilience ecosystems (e.g., Marine Protected Areas), and are moving toward whole-of-ecosystem management approaches. Coral reefs, for example, will recover faster from mass coral bleaching and mortality if healthy populations of herbivorous fish are maintained (*medium confidence*; Hughes et al., 2003), indicating that reducing overfishing will help maintain coral-dominated reef systems while the international community reduces the emissions of GHGs to stabilize global temperature and ocean chemistry.

Approaches such as providing a formal valuation of ecological services from the Ocean have potential to facilitate adaptation by underpinning more effective governance, regulation, and ocean policy while at the same time potentially improving management of these often vulnerable services through the development of market mechanisms and incentives (Beaudoin and Pendleton, 2012). Supporting, regulating, and cultural ecosystem services tend to transcend the immediate demands placed on provisioning services and are difficult to value in formal economic terms owing to their complexity, problems such as double counting, and the value of non-market goods and services arising from marine ecosystems generally (Fu et al., 2011; Beaudoin and Pendleton, 2012).

"Blue Carbon" is defined as the organic carbon sequestered by marine ecosystems such as phytoplankton, mangrove, seagrass, and salt marsh ecosystems (Laffoley and Grimsditch, 2009; Nellemann et al., 2009). In this respect, Blue Carbon will provide opportunities for both adaptation to, and mitigation of, climate change if key uncertainties in inventories, methodologies, and policies for measuring, valuing, and implementing Blue Carbon strategies are resolved (McLeod et al., 2011). Sediment surface levels in vegetated coastal habitats can rise several meters over thousands of years, building carbon-rich deposits (Brevik and Homburg,

○ Examples of projected impacts and vulnerabilities associated with climate change in Ocean regions

● (1) Expansion of low-productivity areas as a consequence of thermal stratification and changes in wind stress. (*Low*)

● (2) Northward expansion of plankton, invertebrate, and fish communities with warming. Increase in fish biomass at high latitude fringes. (*High*)

● (6) Decline in dissolved oxygen through changes in oxygen solubility (temperature), and ocean ventilation and circulation. (*Medium*)

● (7) Expansion of seasonally hypoxic waters due to thermal stratification and eutrophication. Mass coral bleaching events leading to degradation of coral reefs and loss of associated biodiversity. (*High*)

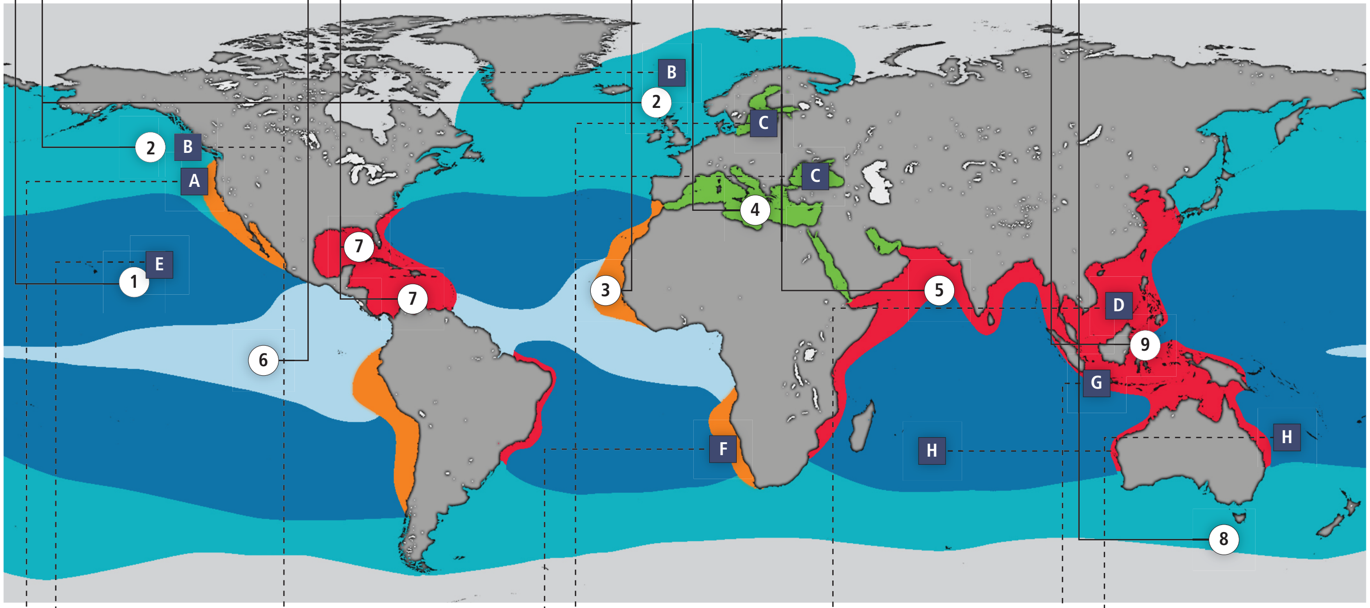
● (3) Upwelling, hence productivity, changes as a result of climate variability, particularly wind stress variability. (*Low*)

● (4) Spread of tropical species originating from Indian and Atlantic Oceans. Increased frequency of mass mortality events of benthic plants and animals linked to extreme temperature events. (*High*)

● (5) Declining dissolved oxygen concentrations and expansion of Oxygen Minimum Zones (OMZ). (*High*)

● (9) Mass coral bleaching and mortality in response to warming as well as temperature-driven decline in growth rates of some corals leading to degradation of coral reefs and loss of associated biodiversity. (*Very high*)

● (8) Shoaling of aragonite saturation horizon reduces biological calcification and alters plankton communities. (*High*)



■ (E) Warming leads to decline in primary production and reductions in fish catch. (*Low*)

■ (A) Acidification of seasonally upwelling waters impacts shellfish aquaculture. (*Medium*)

■ (B) Increased fish catches at high-latitude fringes with economic disruptions and jurisdictional tensions as some fish stocks shift distributions. (*Medium*)

■ (C) Thermal stratification and eutrophication reduces dissolved oxygen with impacts on fish stocks. (*Medium*)

■ (F) Increase in variability of upwelling in some EBUEs adds uncertainty to fisheries management. (*Medium*)

■ (D) Sea level rise modifies coastlines and increases flooding, challenging aquaculture such as shrimp farms. (*Medium*)

■ (H) Temperature-driven shifts in stocks of large pelagic fish create winners and losers among country and island economies. (*High*)

■ (G) Degradation of coral reefs and associated fish stocks as the extent and intensity of mass coral bleaching and mortality increases, increasing risks to regional food security. (*High*)

■ Examples of risks to fisheries from observed and projected impacts

🟦 High-Latitude Spring Bloom Systems
🟩 Semi-Enclosed Seas
🟥 Coastal Boundary Systems
🟧 Eastern Boundary Upwelling Systems
🟦 Equatorial Upwelling Systems
🟦 Subtropical Gyres

Figure 30-12 | Top: Examples of projected impacts and vulnerabilities associated with climate change in Ocean sub-regions. Bottom: Examples of risks to fisheries from observed and projected impacts across Ocean sub-regions. Words in parentheses indicate level of confidence. Details of sub-regions are given in Table 30-1a and Section 30.1.1.

2004; Lo Iacono et al., 2008). The degradation of coastal habitats not only liberates much of the carbon associated with vegetation loss, but can also release and oxidize buried organic carbon through erosion of cleared coastlines (*high confidence*; Duarte et al., 2005). Combining data on global area, land use conversion rates, and near-surface carbon stocks for marshes, mangroves, and seagrass meadows, Pendleton et al. (2012) revealed that the CO_2 emissions arising from destruction of these three ecosystems was equivalent to 3 to 19% of the emissions generated by deforestation globally, with economic damages estimated to be US$6 to US$42 billion annually. Similarly, Luisetti et al. (2013) estimate the carbon stock of seagrass and salt marshes in Europe, representing less than 4% of global carbon stocks in coastal vegetation, was valued at US$180 million, at EU Allowance price of €8/tCO_2 in June 2012. A reversal of EU Environmental Protection Directives could result in economic losses of US$1 billion by 2060. Blue Carbon strategies can also be justified in light of the numerous ecosystem services these ecosystems provide, such as protection against coastal erosion and storm damage, and provision of habitats for fisheries species (Section 5.5.7).

30.6.2. Economic Sectors

30.6.2.1. Fisheries and Aquaculture

The Ocean provided 64% of the production supplied by world fisheries (capture and aquaculture) in 2010, amounting to 148.5 million tonnes of fish and shellfish (FAO, 2012). This production, valued at US$217.5 billion, supplied, on average, 18.6 kg of protein-rich food per person to an estimated population of 6.9 billion (FAO, 2012). Marine capture fisheries supplied 77.4 million tonnes with highest production from the northwest Pacific (27%), west-central Pacific (15%), northeast Atlantic (11%), and southeast Pacific (10%) (FAO, 2012). World aquaculture production (59.9 million tonnes in 2010) is dominated by freshwater fishes; nevertheless, marine aquaculture supplied 18.1 million tonnes (30%) (FAO, 2012).

Marine capture fisheries production increased from 16.8 million tonnes in 1950 to a peak of 86.4 million tonnes in 1996, then declined before stabilizing around 80 million tonnes (FAO, 2012). The stagnation of marine capture fisheries production is attributed to full exploitation of around 60% of the world's marine fisheries and overexploitation of 30% (estimates for 2009) (FAO, 2012). Major issues for industrial fisheries include illegal, unreported, and unregulated fishing; ineffective implementation of monitoring, control, and surveillance; and overcapacity in fishing fleets (World Bank and FAO, 2008; FAO, 2012). Such problems are being progressively addressed in several developed and developing countries (Hilborn, 2007; Pitcher et al., 2009; Worm et al., 2009), where investments have been made in stock assessment, strong management, and application of the FAO Code of Conduct for Responsible Fisheries and the FAO Ecosystem Approach to Fisheries Management.

The significance of marine capture fisheries is illustrated powerfully by the number of people engaged in marine small-scale fisheries (SSF) in developing countries. SSF account for around half of the fish harvested from the Ocean, and provide jobs for more than 47 million people—about 12.5 million fishers and another 34.5 million people engaged in post-harvest activities (Mills et al., 2011). SSF are often characterized by large numbers of politically weak fishers operating from decentralized localities, with poor governance and insufficient data to monitor catches effectively (Kurien and Willmann, 2009; Cochrane et al., 2011; Pomeroy and Andrew, 2011). For these SSF, management that aims to avoid further depletion of overfished stocks may be more appropriate in the short-term than management aimed at maximizing sustainable production. These aims are achieved through adaptive management by (1) introduction of harvest controls (e.g., size limits, closed seasons and areas, gear restrictions, and protection of spawning aggregations) to avoid irreversible damage to stocks in the face of uncertainty (Cochrane et al., 2011); (2) flexible modification of these controls through monitoring (Plagányi et al., 2013); and (3) investing in the social capital and institutions needed for communities and governments to manage SSF (Makino et al., 2009; Pomeroy and Andrew, 2011).

Changes to ocean temperature, chemistry, and other factors are generating new challenges for fisheries resulting in loss of coastal and oceanic habitat (Hazen et al., 2012; Stramma et al., 2012), the movement of species (Cheung et al., 2011), the spread and increase of disease and invading species (Ling, 2008; Raitsos et al., 2010; Chan et al., 2011), and changes in primary production (Chassot et al., 2010). There is *medium evidence* and *medium agreement* that these changes will change both the nature of fisheries and their ability to provide food and protein for hundreds of millions of people (Section 7.2.1.2). The risks to ecosystems and fisheries vary from region to region (Section 7.3.2.4). Dynamic bioclimatic envelope models under SRES A1B project potential increases in fisheries production at high latitudes, and potential decreases at lower latitudes by the mid-21st century (Cheung et al., 2010; Section 6.5). Overall, warming temperatures are projected to shift optimal environments for individual species polewards and redistribute production; however, changes will be region specific (Cheung et al., 2010; Merino et al., 2012).

Fisheries, in particular shellfish, are also vulnerable to declining pH and carbonate ion concentrations. As a result, the global production of shellfish fisheries is *likely* to decrease (Cooley and Doney, 2009; Pickering et al., 2011) with further ocean acidification (*medium confidence*; Sections 6.3.2, 6.3.5, 6.4.1.1; Box CC-OA). Impacts may be first observed in EBUE where upwelled water is already relatively low in O_2 and under-saturated with aragonite (Section 30.5.5). Seasonal upwelling of acidified waters onto the continental shelf in the California Current region has recently affected oyster hatcheries along the coast of Washington and Oregon (Barton et al., 2012; Section 30.5.5.1.1). Whether declining pH and aragonite saturation due to climate change played a role is unclear; however, future declines will increase the risk of such events occurring.

Most marine aquaculture species are sensitive to changing ocean temperature (Section 6.3.1.4; exposed through pens, cages, and racks placed directly in the sea, utilization of seawater in land-based tanks, collection of wild spat) and, for molluscs particularly, changes in carbonate chemistry (Turley and Boot, 2011; Barton et al., 2012; Section 6.3.2.4). Environmental changes can therefore impact farm profitability, depending on target species and farm location. For example, a 1°C rise in SST is projected to shift production of Norwegian salmonids further north but may increase production overall (Hermansen and Heen, 2012). Industries for non-food products, which can be important for regional livelihoods such as Black Pearl in Polynesia, are also affected by rising SST. Higher temperatures are known to affect the quality of pearl nacre, and can increase levels of disease in adult oysters (Pickering et al., 2011; Bell et al., 2013b). Aquaculture production is also vulnerable to extreme events such as storms and floods (e.g., Chang et al., 2013). Flooding and inundation by seawater may be a problem to shore facilities on low-lying coasts. For example, shrimp farming operations in the tropics will be challenged by rising sea levels, which will be exacerbated by mangrove encroachment and a reduced ability for thorough-drying of ponds between crops (Della Patrona et al., 2011).

The impacts of climate change on marine fish stocks are expected to affect the economics of fisheries and livelihoods in fishing nations through changes in the price and value of catches, fishing costs, income to fishers and fishing companies, national labor markets, and industry re-organization (Sumaila et al., 2011; Section 6.4.1). A study of the potential vulnerabilities of national economies to the effects of climate change on fisheries, in terms of exposure to warming, relative importance of fisheries to national economies and diets, and limited societal capacity to adapt, concluded that a number countries including Malawi, Guinea, Senegal, Uganda, Sierra Leone, Mozambique, Tanzania, Peru, Colombia, Venezuela, Mauritania, Morocco, Bangladesh, Cambodia, Pakistan, Yemen, and Ukraine are most vulnerable (Allison et al., 2009).

Aquaculture production is expanding rapidly (Bostock et al., 2010) and will play an important role in food production and livelihoods as the human demand for protein grows. This may also add pressure on capture fisheries (FAO, 2012; Merino et al., 2012). Two-thirds of farmed food fish production (marine and freshwater) is achieved with the use of feed derived from wild-harvested, small, pelagic fish and shellfish. Fluctuations in the availability and price of fishmeal and fish oil for feeds, as well as their availability, pose challenges for the growth of sustainable aquaculture production, particularly given uncertainties in changes in EBUE upwelling dynamics to climate change (Section 30.5.5). Technological advances and changes in management such as increasing feed efficiencies, using alternatives to fishmeal and fish oil, and farming of herbivorous finfish, coupled with economic and regulatory incentives, will reduce the vulnerability of aquaculture to the impacts of climate change on small, pelagic fish abundance (Naylor et al., 2009; Merino et al., 2010; FAO, 2012).

The challenges of optimizing the economic and social benefits of both industrial fisheries and SSF and aquaculture operations, which often already include strategies to adapt to climatic variability (Salinger et al., 2013), are now made more complex by climate change (Cochrane et al., 2009; Brander, 2010, 2013). Nevertheless, adaptation options include establishment of early warning systems to aid decision making, diversification of enterprises, and development of adaptable management systems (Chang et al., 2013). Vulnerability assessments that link oceanographic, biological, and socioeconomic systems can be applied to identify practical adaptations to assist enterprises, communities, and households to reduce the risks from climate change and capitalize on the opportunities (Pecl et al., 2009; Bell et al., 2013b; Norman-López et al., 2013). The diversity of these adaptation options, and the policies needed to support them, are illustrated by the examples in the following subsections.

30.6.2.1.1. Tropical fisheries based on large pelagic fish

Fisheries for skipjack, yellowfin, big-eye, and albacore tuna provide substantial economic and social benefits to the people of Small Island Developing States (SIDS). For example, tuna fishing license fees contribute substantially (up to 40%) to the government revenue of several Pacific Island nations (Gillett, 2009; Bell et al., 2013b). Tuna fishing and processing operations also contribute up to 25% of gross domestic product in some of these nations and employ more than 12,000 people (Gillett, 2009; Bell et al., 2013b). Considerable economic benefits are also derived from fisheries for top pelagic predators in the Indian and Atlantic Oceans (FAO, 2012; Bell et al., 2013a). Increasing sea temperatures and changing patterns of upwelling are projected to cause shifts in the distribution and abundance of pelagic top predator fish stocks (Sections 30.5.2, 30.5.5-6), with potential to create "winners" and "losers" among island economies as catches of the transboundary tuna stocks change among and within their exclusive economic zones (EEZs; Bell et al., 2013a,b).

A number of practical adaptation options and supporting policies have been identified to minimize the risks and maximize the opportunities associated with the projected changes in distribution of the abundant skipjack tuna in the tropical Pacific (Bell et al., 2011b, 2013a; Lehodey et al., 2011; Table 30-2). These adaptation and policy options include (1) full implementation of the regional "vessel day scheme," designed to distribute the economic benefits from the resource in the face of climatic variability, and other schemes to control fishing effort in subtropical areas; (2) strategies for diversifying the supply of fish for canneries in the west of the region as tuna move progressively east; (3) continued effective fisheries management of all tuna species; (4) energy efficiency programs to assist domestic fleets to cope with increasing fuel costs and the possible need to fish further from port; and (5) the eventual restructuring of regional fisheries management organizations to help coordinate management measures across the entire tropical Pacific. Efforts to ensure provision of operational-level catch and effort data from all industrial fishing operations will improve models for projecting redistribution of tuna stocks and quotas under climate change (Nicol et al., 2013; Salinger et al., 2013). Similar adaptation options and policy responses are expected to be relevant to the challenges faced by tuna fisheries in the tropical and subtropical Indian and Atlantic Oceans.

30.6.2.1.2. Small-scale fisheries

Small-scale fisheries (SSF) account for 56% of catch and 91% of people working in fisheries in developing countries (Mills et al., 2011). SSF are fisheries that tend to operate at family or community level, have low levels of capitalization, and make an important contribution to food security and livelihoods. They are often dependent on coastal ecosystems, such as coral reefs, that provide habitats for a wide range of harvested fish and invertebrate species. Despite their importance to many developing countries, such ecosystems are under serious pressure from human activities including deteriorating coastal water quality, sedimentation, ocean warming, overfishing, and acidification (Sections 7.2.1.2, 30.3, 30.5; Box CC-CR). These pressures are translating into a steady decline in live coral cover, which is *very likely* to continue over the coming decades, even where integrated coastal zone management is in place

Table 30-2 | Examples of priority adaptation options and supporting policies to assist Pacific Island countries and territories to minimize the threats of climate change to the socioeconomic benefits derived from pelagic and coastal fisheries and aquaculture, and to maximize the opportunities. These measures are classified as win–win (W–W) adaptations, which address other drivers of the sector in the short term and climate change in the long term, or lose–win (L–W) adaptations, where benefits do not exceed costs in the short term but accrue under longer term climate change (modified from Bell et al., 2013b). WCPFC = Western and Central Pacific Fisheries Commission.

	Adaptation options	Supporting policies
Economic development	• Full implementation of the vessel day scheme to control fishing effort by the Parties to the Nauru Agreement[a] (W–W) • Diversifying sources of fish for canneries in the region and maintaining trade agreements, e.g., an economic partnership agreement with the European Union (W–W) • Continued conservation and management measures for all species of tuna to maintain stocks at healthy levels and make these valuable species more resilient to climate change (W–W) • Energy efficiency programs to assist fleets to cope with oil price rises and minimize CO_2 emissions and reduce costs of fishing further afield as tuna distributions shift east (W–W) • Pan-Pacific tuna management through merger of the WCPFC and Inter-American Tropical Tuna Commission to coordinate management measures across the tropical Pacific (L–W)	• Strengthen national capacity to administer the vessel day scheme. • Adjust national tuna management plans and marketing strategies to provide flexible arrangements to buy and sell tuna. • Include implications of climate change in management objectives of the WCPFC. • Apply national management measures to address climate change effects for subregional concentrations of tuna in archipelagic waters beyond the mandate of WCPFC. • Require all industrial tuna vessels to provide operational-level catch and effort data to improve the models for redistribution of tuna stocks during climate change.
Food security	• Manage catchment vegetation to reduce transfer of sediments and nutrients to coasts to reduce damage to adjacent coastal coral reefs, mangroves, and seagrasses that support coastal fisheries (W–W). • Foster the care of coral reefs, mangroves, and seagrasses by preventing pollution, managing waste, and eliminating direct damage to these coastal fish habitats (W–W). • Provide for migration of fish habitats by prohibiting construction adjacent to mangroves and seagrasses and installing culverts beneath roads to help the plants colonize landward areas as sea level rises (L–W). • Sustain and diversify catches of demersal coastal fish to maintain the replenishment potential of all stocks (L–W). • Increase access to tuna caught by industrial fleets through storing and selling tuna and by-catch landed at major ports to provide inexpensive fish for rapidly growing urban populations (W–W). • Install fish aggregating devices close to the coast to improve access to fish for rural communities as human populations increase and demersal fish decline (W–W). • Develop coastal fisheries for small pelagic fish species, e.g., mackerel, anchovies, pilchards, sardines, and scads (W–W?). • Promote simple post-harvest methods, such as traditional smoking, salting, and drying, to extend the shelf life of fish when abundant catches are landed (W–W).	• Strengthen governance for sustainable use of coastal fish habitats by (1) building national capacity to understand the threats of climate change; (2) empowering communities to manage fish habitats; and (3) changing agriculture, forestry, and mining practices to prevent sedimentation and pollution. • Minimize barriers to landward migration of coastal habitats during development of strategies to assist other sectors to respond to climate change. • Apply "primary fisheries management" to stocks of coastal fish and shellfish to maintain their potential for replenishment. • Allocate the necessary quantities of tuna from total national catches to food security to increase access to fish for both urban and coastal populations. • Dedicate a proportion of the revenue from fishing licences to improve access to tuna for food security. • Include anchored inshore fish aggregating devices as part of national infrastructure for food security.
Livelihoods	• Relocate pearl farming operations to deeper water and to sites closer to coral reefs and seagrass/algal areas where water temperatures and aragonite saturation levels are likely to be more suitable for good growth and survival of pearl oysters and formation of high-quality pearls (L–W). • Raise the walls and floor of shrimp ponds so that they drain adequately as sea level rises (L–W). • Identify which shrimp ponds may need to be rededicated to producing other commodities (L–W).	• Provide incentives for aquaculture enterprises to assess risks to infrastructure so that farming operations and facilities can be "climate-proofed" and relocated if necessary. • Strengthen environmental impact assessments for coastal aquaculture activities to include the additional risks posed by climate change. • Develop partnerships with regional technical agencies to provide support for development of sustainable aquaculture.

[a] The Parties to the Nauru Agreement are Federated States of Micronesia, Kiribati, Marshall Islands, Nauru, Palau, Papua New Guinea, Solomon Islands, and Tuvalu.

(Sections 30.5.4, 30.5.6). For example, coral losses around Pacific Islands are projected to be as high as 75% by 2050 (Hoegh-Guldberg et al., 2011a). Even under the most optimistic projections (a 50% loss of coral by 2050), changes to state of coral reefs (Box CC-CR; Figures 30-10, 30-12) are *very likely* to reduce the availability of associated fish and invertebrates that support many of the SSF in the tropics (*high confidence*). In the Pacific, the productivity of SSF on coral reefs has been projected to decrease by at least 20% by 2050 (Pratchett et al., 2011b), which is also *likely* to occur in other coral reef areas globally given the similar and growing stresses in these other regions (Table SM30-1; Section 30.5.4).

Adaptation options and policies for building the resilience of coral reef fisheries to climate change suggested for the tropical Pacific include (1) strengthening the management of catchment vegetation to improve water quality along coastlines; (2) reducing direct damage to coral reefs; (3) maintaining connectivity of coral reefs with mangrove and seagrass habitats; (4) sustaining and diversifying the catch of coral reef fish to maintain their replenishment potential; and (5) transferring fishing effort from coral reefs to skipjack and yellowfin tuna resources by installing anchored fish-aggregating devices (FAD) close to shore (Bell et al., 2011b; 2013a,b; Table 30-2). These adaptation options and policies represent a "no regrets" strategy in that they provide benefits for coral reef fisheries and fishers irrespective of climate change and ocean acidification.

30.6.2.1.3. Northern Hemisphere HLSBS fisheries

The high-latitude fisheries in the Northern Hemisphere span from around 30/35°N to 60°N in the North Pacific and 80°N in the North Atlantic, covering a wide range of thermal habitats supporting subtropical/temperate species to boreal/arctic species. The characteristics of these

HLSBS environments, as well as warming trends, are outlined in Section 30.5.1 and Table 30-1. In part, as a result of 30 years of increase in temperature (Belkin, 2009; Sherman et al., 2009), there has been an increase in the size of fish stocks associated with high-latitude fisheries in the Northern Hemisphere. This is particularly the case for the Norwegian spring-spawning herring, which has recovered from near-extinction as a result of overfishing and a cooler climate during the 1960s (Toresen and Østvedt, 2000). The major components of both pelagic and demersal high-latitude fish stocks are boreal species located north of 50°N. Climate change is projected to increase high-latitude plankton production and displace zooplankton and fish species poleward. As a combined result of these future changes, the abundance of fish (particularly boreal species) may increase in the northernmost part of the high-latitude region (Cheung et al., 2011), although increases will only be moderate in some areas.

The changes in distribution and migration of pelagic fish shows considerable spatial and temporal variability, which can increase tensions among fishing nations. In this regard, tension over the Atlantic mackerel fisheries has led to what many consider the first climate change-related conflict between fishing nations (Cheung et al., 2012; Section 30.6.5), and which has emphasized the importance of developing international collaboration and frameworks for decision making (Miller et al., 2013; Sections 15.4.3.3, 30.6.7). The Atlantic mackerel has over the recent decades been a shared stock between the EU and Norway. However, the recent advancement of the Atlantic mackerel into the Icelandic EEZ during summer has resulted in Icelandic fishers operating outside the agreement between the EU and Norway. Earlier records of mackerel from the first half of the 20th and second half of the 19th century show, however, that mackerel was present in Icelandic waters during the earlier warm periods (Astthorsson et al., 2012). In the Barents Sea, the northeast Arctic cod, *Gadus morhua*, reached record-high abundance in 2012 and also reached its northernmost-recorded distribution (82°N) (ICES, 2012). A further northward migration is impossible as this would be into the Deep Sea Polar Basin, beyond the habitat of shelf species. A further advancement eastwards to the Siberian shelf is, however, possible. The northeast Arctic cod stock is shared exclusively by Norway and Russia, and to date there has been a good agreement between those two nations on the management of the stock. These examples highlight the importance of international agreements and cooperation (Table 30-4).

The HLSBS fisheries constitute a large-scale high-tech industry, with large investments in highly mobile fishing vessels, equipment, and land-based industries with capacity for adapting fisheries management and industries for climate change (Frontiers Economics, Ltd., 2013). Knowledge of how climate fluctuations and change affect the growth, recruitment, and distribution of fish stocks is presently not incorporated into fisheries management strategies (Perry et al., 2010). These strategies are vital for fisheries that hope to cope with the challenges of a changing ocean environment, and are centrally important to any attempt to develop ecosystem-based management and sustainable fisheries under climate change. The large pelagic stocks, with their climate-dependent migration pattern, are shared among several nations. Developing equitable sharing of fish quotas through international treaties (Table 30-4) is a necessary adaptation for a sustainable fishery. Factors presently taken into account in determining the shares of quotas are the historical fishery, bilateral exchanges of quotas for various species, and the time that stocks are in the various EEZs.

30.6.2.2. Tourism

Tourism recreation represents one of the world's largest industries, accounting for 9% (>US$6 trillion) of global GDP and employing more than 255 million people. It is expected to grow by an average of 4% annually and reach 10% of global GDP within the next 10 years (WTTC, 2012). As with all tourism, that which is associated with the Ocean is heavily influenced by climate change, global economic and socio-political conditions, and their interactions (Scott et al., 2012b; Section 10.6.1). Climate change, through impacts on ecosystems (e.g., coral reef bleaching), can reduce the appeal of destinations, increase operating costs, and/or increase uncertainty in a highly sensitive business environment (Scott et al., 2012b).

Several facets of the influence of climate change on the Ocean directly impact tourism (Section 10.6). Tourism is susceptible to extreme events such as violent storms, long periods of drought, and/or extreme precipitation events (Sections 5.4.3.4, 10.6.1; IPCC, 2012). SLR, through its influence on coastal erosion and submergence, salinization of water supplies, and changes to storm surge, increases the vulnerability of coastal tourism infrastructure, tourist safety, and iconic ecosystems (*high confidence*; Sections 5.3.3.2, 5.4.3.4, 10.6; Table SPM.1; IPCC, 2012). For example, approximately 29% of resorts in the Caribbean are within 1 m of the high tide mark and 60% are at risk of beach erosion from rapid SLR (Scott et al., 2012a).

Increasing sea temperatures (Section 30.3.1.1) can change attractiveness of locations and the opportunities for tourism through their influence on the movement of organisms and the state of ecosystems such as coral reefs (Section 10.6.2; Box CC-CR; UNWTO and UNEP, 2008). Mass coral bleaching and mortality (triggered by elevated sea temperatures; *high confidence*) can decrease the appeal of destinations for diving-related tourism, although the level of awareness of tourists of impacts (e.g., <50% of tourists were concerned about coral bleaching during a major bleaching year, 1998) and expected economic impacts have been found to be uncertain (Scott et al., 2012b). Some studies, however, have noted reduced tourist satisfaction and identified "dead coral" as one of the reasons for disappointment at the end of the holiday (Westmacott et al., 2000). Tourists respond to changes in factors such as weather and opportunity by expressing different preferences. For example, preferred conditions and hence tourism are projected to shift toward higher latitudes with climate change, or from summer to cooler seasons (Amelung et al., 2007; Section 10.6.1).

Options for adaptation by the marine tourism sector include (1) identifying and responding to inundation risks with current infrastructure, and planning for projected SLR when building new tourism infrastructure (Section 5.5; Scott et al., 2012a); (2) promoting shoreline stability and natural barriers by preserving ecosystems such as mangroves, salt marshes, and coral reefs (Section 5.5; Scott et al., 2012b); (3) deploying forecasting and early-warning systems in order to anticipate challenges to tourism and natural ecosystems (Strong et al., 2011; IPCC, 2012); (4) preparation of risk management and disaster preparation plans in order to respond to extreme events; (5) reducing the effect of other stressors on ecosystems and building resilience in iconic tourism features such as coral reefs and mangroves; and (6) educating tourists to improve understanding of the negative consequences of climate change over those stemming from local stresses (Scott et al., 2012a,b). Adaptation plans for tourism industries need to address specific operators and regions. For example, some operators may have costly infrastructure at risk while others may have few assets but are dependent on the integrity of natural environments and ecosystems (Turton et al., 2010).

30.6.2.3. Shipping

International shipping accounts for more than 80% of world trade by volume (UNCTAD, 2009a,b) and approximately 3% of global CO_2 emissions from fuel combustion although CO_2 emissions are expected to increase two- to threefold by 2050 (Heitmann and Khalilian, 2010; WGIII AR5 Section 8.1). Changes in shipping routes (Borgerson, 2008) and variation in the transport network due to shifts in grain production and global markets, as well as new fuel and weather-monitoring technology, may alter these emission patterns (WGIII AR5 Sections 8.3, 8.5). Extreme weather events, intensified by climate change, may interrupt ports and transport routes more frequently, damaging infrastructure and introducing additional dangers to ships, crews, and the environment (UNCTAD, 2009a,b; Pinnegar et al., 2012; Section 10.4.4). These issues have been assessed by some countries which have raised concerns over the potential for costly delays and cancellation of services, and the implications for insurance premiums as storminess and other factors increase risks (Thornes et al., 2012).

Climate change may benefit maritime transport by reducing Arctic sea ice and consequently shorten travel distances between key ports (Borgerson, 2008), thus also decreasing total GHG emissions from ships (WGIII AR5 Section 8.5.1). Currently, the low level of reliability of this route limits its use (Schøyen and Bråthen, 2011), and the potential full operation of the Northwest Passage and Northern Sea Route would require a transit management regime, regulation (e.g., navigation, environmental, safety, and security issues), and a clear legal framework to address potential territorial claims that may arise, with a number of countries having direct interest in the Arctic. Further discussion of issues around melting Arctic sea ice and the Northern Sea Route are given in Chapter 28 (Sections 28.2.6, 28.3.4).

30.6.2.4. Offshore Energy and Mineral Resource Extraction and Supply

The marine oil and gas industry face potential impacts from climate change on its ocean-based activities. More than 100 oil and gas platforms were destroyed in the Gulf of Mexico by the unusually strong Hurricanes Katrina and Rita in 2005. Other consequences for oil pipelines and production facilities ultimately reduced US refining capacity by 20% (IPCC, 2012). The increasing demand for oil and gas has pushed operations to waters 2000 m deep or more, far beyond continental shelves. The very large-scale moored developments required are exposed to greater hazards and higher risks, most of which are not well understood by existing climate/weather projections. Although there is a strong trend toward seafloor well completions with a complex of wells, manifolds, and pipes that are not exposed to surface forcing, these systems face different hazards from instability and scouring of the unconsolidated sediments by DS currents (Randolph et al., 2010). The influence of warming oceans on sea floor stability is widely debated due largely to uncertainties about the effects of methane and methane hydrates (Sultan et al., 2004; Archer et al., 2009; Geresi et al., 2009). Declining sea ice is also opening up the Arctic to further oil and gas extraction. Discussion of potential expansion of oil and mineral production in the Arctic is made in Chapter 28 (Sections 28.2.5-6, 28.3.4).

The principal threat to oil and gas extraction and infrastructure in maritime settings is the impact of extreme weather (Kessler et al., 2011), which is *likely* to increase given that future storm systems are expected to have greater energy (Emanuel, 2005; Trenberth and Shea, 2006; Knutson et al., 2010). Events such as Hurricane Katrina have illustrated challenges which will arise for this industry with projected increases in storm intensity (Cruz and Krausmann, 2008). In this regard, early warning systems and integrated planning offer some potential to reduce the effect of extreme events (IPCC, 2012).

30.6.3. Human Health

Major threats to public health due to climate change include diminished security of water and food supplies, extreme weather events, and changes in the distribution and severity of diseases, including those due to marine biotoxins (Costello et al., 2009; Sections 5.4.3.5, 6.4.2.3, 11.2). The predominantly negative impacts of disease for human communities are expected to be more serious in low-income areas such as Southeast Asia, southern and east Africa, and various sub-regions of South America (Patz et al., 2005), which also have under-resourced health systems (Costello et al., 2009). Many of the influences are directly or indirectly related to basin-scale changes in the Ocean (e.g., temperature, rainfall, plankton populations, SLR, and ocean circulation; McMichael et al., 2006). Climate change in the Ocean may influence the distribution of diseases such as cholera (Section 11.5.2.1), and the distribution and occurrence of HABs. The frequency of cholera outbreaks induced by *Vibrio cholerae* and other enteric pathogens are correlated with sea surface temperatures, multi-decadal fluctuations of ENSO, and plankton blooms, which may provide insight into how this disease may change with projected rates of ocean warming (Colwell, 1996; Pascual et al., 2000; Rodó et al., 2002; Patz et al., 2005; Myers and Patz, 2009; Baker-Austin et al., 2012). The incidence of diseases such as ciguatera also shows links to ENSO, with ciguatera becoming more prominent after periods of elevated sea temperature. This indicates that ciguatera may become more frequent in a warmer climate (Llewellyn, 2010), particularly given the higher prevalence of ciguatera in areas with degraded coral reefs (*low confidence*; Pratchett et al., 2011a).

30.6.4. Ocean-Based Mitigation

30.6.4.1. Deep Sea Carbon Sequestration

Carbon dioxide capture and storage into the deep sea and geologic structures are also discussed in WGIII AR5 Chapter 7 (Sections 7.5.5, 7.8.2, 7.12). The economic impact of deliberate CO_2 sequestration beneath the sea floor has previously been reviewed (IPCC, 2005). Active CO_2 sequestration from co-produced CO_2 into sub-sea geologic formations is being instigated in the North Sea and in the Santos Basin offshore from Brazil. These activities will increase as offshore oil and gas production increasingly exploits fields with high CO_2 in the source gas and oil. Significant risks from the injection of high levels of CO_2 into deep ocean waters have been identified for DS organisms and ecosystems although chronic effects have not yet been studied. These risks are similar to those discussed previously with respect to ocean acidification and could further exacerbate declining O_2 levels and changing trophic networks in deep water areas (Seibel and Walsh, 2001; Section 6.4.2.2).

There are significant issues within the decision frameworks regulating these activities. Dumping of any waste or other matter in the sea, including the seabed and its subsoil, is strictly prohibited under the 1996 London Protocol (LP) except for those few materials listed in Annex I. Annex 1 was amended in 2006 to permit storage of CO_2 under the seabed. "Specific Guidelines for Assessment of Carbon Dioxide Streams for Disposal into Sub-Seabed Geological Formations" were adopted by the parties to the LP in 2007. The Guidelines take a precautionary approach to the process, requiring Contracting Parties under whose jurisdiction or control such activities are conducted to issue a permit for the disposal subject to stringent conditions being fulfilled (Rayfuse and Warner, 2012).

30.6.4.2. Offshore Renewable Energy

Renewable energy supply from the Ocean includes ocean energy and offshore wind turbines. The global technical potential for ocean and wind energy is not as high as solar energy although considerable potential still remains. Detailed discussion of the potential of renewable energy sources are given in WGIII AR5 Chapter 7 (Sections 7.4.2, 7.5.3, 7.8.2). There is an increasing trend in the renewable energy sector to offshore wind turbines (Section 10.2.2). At present, there is *high uncertainty* about how changes in wind intensity and patterns, and extreme events (from climate change), will impact the offshore wind energy sector. Given the design and engineering solutions available to combat climate change impacts (Tables 10-1, 10-7), it is *unlikely* that this sector will face insurmountable challenges from climate change.

30.6.5. Maritime Security and Related Operations

Climate change and its influence on the Ocean has become an area of increasing concern in terms of the maintenance of national security and the protection of citizens. These concerns have arisen as nation-states increasingly engage in operations ranging from humanitarian assistance in climate-related disasters to territorial issues exacerbated by changing coastlines, human communities, resource access, and new seaways (Kaye, 2012; Rahman, 2012; Section 12.6). In this regard, increasing sea levels along gently sloping coastlines can have the seemingly perverse outcome that the territorial limits to the maritime jurisdiction of the State might be open to question as the distance from national baselines to the outer limits of the EEZ increases beyond 200 nm over time (Schofield and Arsana, 2012).

Changes in coastal resources may also be coupled with decreasing food security to compound coastal poverty and lead, in some cases, to increased criminal activities such as piracy; IUU fishing; and human, arms, and drug trafficking (Kaye, 2012). While the linkages have not been clearly defined in all cases, it is possible that changes in the Ocean as result of climate change will increase pressure on resources aimed at maintaining maritime security and countering criminal activity, disaster relief operations, and freedom of navigation (Section 12.6.2). National maritime security capacity and infrastructure may also require rethinking as new challenges present themselves as a result of climate change and ocean acidification (Allen and Bergin, 2009; Rahman, 2012; Sections 12.6.1-2).

Opportunities may also arise from changes to international geography such as formation of new ice-free seaways through the Arctic, which may benefit some countries in terms of maintaining maritime security and access (Section 28.2.6). Conversely, such new features may also lead to increasing international tensions as States perceive new vulnerabilities from these changes to geography.

Like commercial shipping (Section 30.6.2.3), naval operations in many countries result in significant GHG emissions (e.g., the US Navy emits around 2% of the national GHG emissions; Mabus, 2010). As a result, there are a number of programs being implemented by navies around the world to try and reduce their carbon footprint and air pollution such as improving engine efficiency, reducing fouling of vessels, increasing the use of biofuels, and using nuclear technology for power generation, among other initiatives.

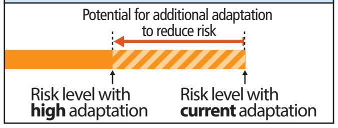
30.7. Synthesis and Conclusions

Evidence that human activities are fundamentally changing the Ocean is *virtually certain*. Sea temperatures have increased rapidly over the past 60 years at the same time as pH has declined, consistent with the expected influence of rising atmospheric concentrations of CO_2 and other GHGs (*very high confidence*). The rapid rate at which these fundamental physical and chemical parameters of the Ocean are changing is unprecedented within the last 65 Ma (*high confidence*) and possibly 300 Ma (*medium confidence*). As the heat content of the Ocean has increased, the Ocean has become more stratified (*very likely*), although there is considerable regional variability. In some cases, changing surface wind has influenced the extent of mixing and upwelling, although our understanding of where and why these differences occur regionally is uncertain. The changing structure and function of the Ocean has led to changes in parameters such as O_2, carbonate ion, and inorganic nutrient concentrations (*high confidence*). Not surprisingly, these fundamental changes have resulted in responses by key marine organisms, ecosystems, and ecological processes, with negative implications for hundreds of millions of people that depend on the ecosystem goods and services provided (*very likely*). Marine organisms are migrating at rapid rates toward higher latitudes, fisheries are transforming, and many organisms are shifting their reproductive and migratory activity in time and in concert with changes in temperature and other parameters. Ecosystems such as coral reefs are declining rapidly (*high confidence*). An extensive discussion of these changes is provided in previous sections and in other chapters of AR5.

Table 30-3 | Key risks to ocean and coastal issues from climate change and the potential for risk reduction through mitigation and adaptation. Key risks are identified based on assessment of the literature and expert judgments made by authors of the various WGII AR5 chapters, with supporting evaluation of evidence and agreement in the referenced chapter sections. Each key risk is characterized as very low, low, medium, high, or very high. Risk levels are presented for the near-term era of committed climate change (here, for 2030–2040), in which projected levels of global mean temperature increase do not diverge substantially across emissions scenarios. Risk levels are also presented for the longer term era of climate options (here, for 2080–2100), for global mean temperature increases of 2°C and 4°C above pre-industrial levels. For each time frame, risk levels are estimated for the current state of adaptation and for a hypothetical highly adapted state. As the assessment considers potential impacts on different physical, biological, and human systems, risk levels should not necessarily be used to evaluate relative risk across key risks. Relevant climate variables are indicated by symbols.

Climate-related drivers of impacts							
Warming trend	Extreme temperature	Extreme precipitation	Precipitation	Damaging cyclone	Sea level	Ocean acidification	Hypoxia

Level of risk & potential for adaptation



Risks to ecosystems and adaptation options

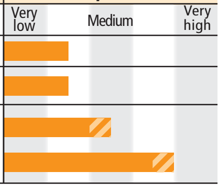
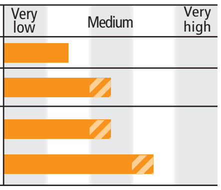
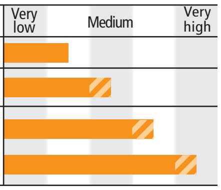
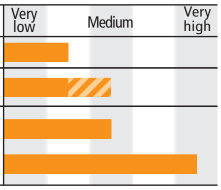
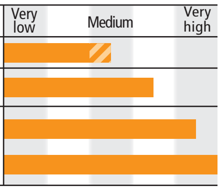
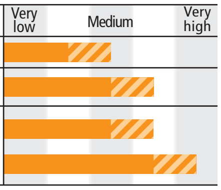
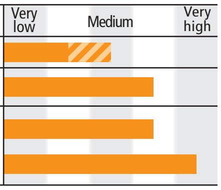
Key risk	Adaptation issues & prospects	Climatic drivers	Timeframe	Risk & potential for adaptation
Changes in ecosystem productivity associated with the redistribution and loss of net primary productivity in open oceans. (*medium confidence*) [6.5.1, 6.3.4, Box CC-PP]	Adaptation options are limited to the translocation of industrial fishing activities due to regional decreases (low latitude) versus increases (high latitude) in productivity, or to the expansion of aquaculture.	Warming trend Extreme temperature	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Distributional shift in fish and invertebrate species, fall in fisheries catch potential at low latitudes, e.g., in EUS, CBS, and STG regions. (*high confidence*) [6.3.1, Box CC-MB]	Evolutionary adaptation potential of fish and invertebrate species to warming is limited as indicated by their changes in distribution to maintain temperatures. Human adaptation options involve the large-scale translocation of industrial fishing activities following the regional decreases (low latitude) versus (possibly transient) increases (high latitude) in catch potential as well as deploying flexible management that can react to variability and change. Further options include improving fish resilience to thermal stress by reducing other stressors such as pollution and eutrophication, the expansion of sustainable aquaculture and development of alternative livelihoods in some regions.	Warming trend Extreme temperature	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
High mortalities and loss of habitat to larger fauna including commercial species due to hypoxia expansion and effects. (*high confidence*) [6.3.3, 30.5.3.2, 30.5.4.1-2]	Human adaptation options involve the large-scale translocation of industrial fishing activities as a consequence of the hypoxia-induced decreases in biodiversity and fisheries catch of pelagic fish and squid. Special fisheries may benefit (Humboldt squid). Reducing the amount of organic carbon running off of coastlines by controlling nutrients and pollution running off agricultural areas can reduce microbial activity and consequently limit the extent of the oxygen drawdown and the formation of coastal dead zones.	Warming trend Hypoxia	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Ocean acidification: Reduced growth and survival of commercially valuable shellfish and other calcifiers, e.g., reef building corals, calcareous red algae. (*high confidence*) [5.3.3.5, 6.1.1, 6.3.2, 6.4.1.1, 30.3.2.2, Box CC-OA]	Evidence for differential resistance and evolutionary adaptation of some species exists but is likely limited by the CO_2 concentrations and high temperatures reached; adaptation options shifting to exploit more resilient species or the protection of habitats with low natural CO_2 levels, as well as the reduction of other stresses, mainly pollution and limiting pressures from tourism and fishing.	Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Reduced biodiversity, fisheries abundance and coastal protection by coral reefs due to heat-induced mass coral bleaching and mortality increases, exacerbated by ocean acidification, e.g., in CBS, SES, and STG regions. (*high confidence*) [5.4.2.4, 6.4.2, 30.3.1.1, 30.3.2.2, 30.5.2, 30.5.3, 30.5.4, 30.5.6, Box CC-CR]	Evidence of rapid evolution by corals is very limited or nonexistent. Some corals may migrate to higher latitudes. However, the movement of entire reef systems is unlikely given estimates that they need to move at the speed of 10 – 20 km yr^{-1} to keep up with the pace of climate change. Human adaptation options are limited to reducing other stresses, mainly enhancing water quality and limiting pressures from tourism and fishing. This option will delay the impacts of climate change by a few decades but is likely to disappear as thermal stress increases.	Warming trend Damaging cyclone Extreme temperature Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Coastal inundation and habitat loss due to sea level rise, extreme events, changes in precipitation, and reduced ecological resilience, e.g., in CBS and STG subregions. (*medium to high confidence*) [5.5.2, 5.5.4, 30.5.6.1.3, 30.6.2.2, Box CC-CR]	Options to maintain ecosystem integrity are limited to the reduction of other stresses, mainly pollution and limiting pressures from tourism, fishing, physical destruction, and unsustainable aquaculture. Reducing deforestation and increasing reforestation of river catchments and coastal areas to retain sediments and nutrients. Increased mangrove, coral reef, and seagrass protection and restoration to protect numerous ecosystem goods and services such as coastal protection, tourist value, and fish habitat.	Warming trend Ocean acidification Sea level Extreme precipitation Damaging cyclone	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Marine biodiversity loss with high rate of climate change. (*medium confidence*) [6.3.1-3, 6.4.1.2-3, Table 30.4, Box CC-MB]	Adaptation options are limited to the reduction of other stresses, mainly to reducing pollution and to limiting pressures from tourism and fishing.	Warming trend Extreme temperature Ocean acidification	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

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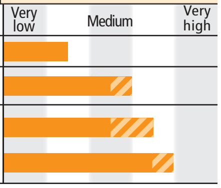
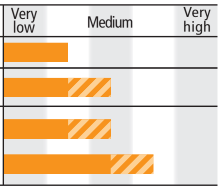
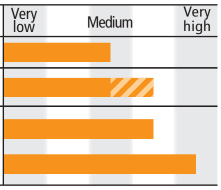
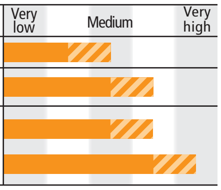
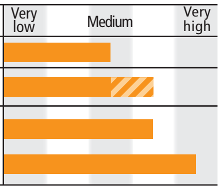
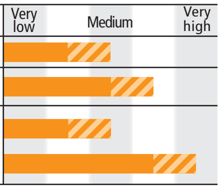
Table 30-3 (continued)

Risks to fisheries				
Key risk	**Adaptation issues & prospects**	**Climatic drivers**	**Timeframe**	**Risk & potential for adaptation**
Decreased production of global shellfish fisheries. (*high confidence*) [6.3.2, 6.3.5, 6.4.1.1, 30.5.5, 30.6.2.1, Box CC-OA]	Effective shift to alternative livelihoods, changes in food consumption patterns, and adjustment of (global) markets.	🌡 🌡! C	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Global redistribution and decrease of low-latitude fisheries yields are paralleled by a global trend to catches having smaller fishes. (*medium confidence*) [6.3.1, 6.4.1, 6.5.3, 30.5.4, 30.5.6, 30.6.2]	Increasing coastal poverty at low latitudes as fisheries becomes smaller – partially compensated by the growth of aquaculture and marine spatial planning, as well as enhanced industrialized fishing efforts.	🌡	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Redistribution of catch potential of large pelagic-highly migratory fish resources, such as tropical Pacific tuna fisheries. (*high confidence*) [6.3.1, 6.4.3, Table 30.4]	International fisheries agreements and instruments, such as the tuna commissions, may have limited success in establishing sustainable fisheries yields.	🌡	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Variability of small pelagic fishes in EBUEs is becoming more extreme at interannual to multidecadal scales, making industry and management decisions more uncertain. (*medium confidence*) [6.3.2, 6.3.3, 30.5.2, 30.5.5, Box CC-UP]	Development of new and specific management tools and models may have limited success to sustain yields. Reduction in fishing intensity increases resilience of the fisheries.	O_2↓ 🌡 C	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Decrease in catch and species diversity of fisheries in tropical coral reefs, exacerbated by interactions with other human drivers such as eutrophication and habitat destruction. (*high confidence*) [6.4.1, 30.5.3-4, 30.5.6, Box CC-CR]	Restoration of overexploited fisheries and reduction of other stressors on coral reefs delay ecosystem changes. Human adaptation includes the usage of alternative livelihoods and food sources (e.g., coastal aquaculture).	🌡 🌡! C	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Current spatial management units, especially the marine protected areas (MPAs), may fail in the future due to shifts in species distributions and community structure. (*high confidence*) [6.3.1, 6.4.2.1, 30.5.1, Box CC-MB]	Continuous revision and shifts of MPA borders, and of MPA goals and performance.	🌡 O_2↓ 🌡! C	Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

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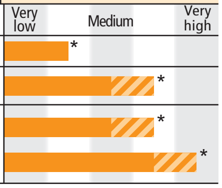
30.7.1. Key Risks and Vulnerabilities

The rapid changes in the physical, chemical, and biological state of the Ocean pose a number of key risks and vulnerabilities for ecosystems, communities, and nations worldwide. Table 30-3 and Figure 30-12 summarize risks and vulnerabilities from climate change and ocean acidification, along with adaptation issues and prospects, and a summary of expert opinion on how these risks will change under further changes in environmental conditions.

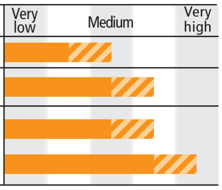
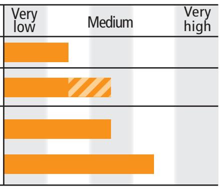
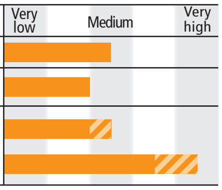
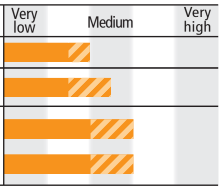
Rising ocean temperatures are changing the distribution, abundance, and phenology of many marine species and ecosystems, and consequently represent a key risk to food resources, coastal livelihoods, and industries such as tourism and fishing, especially for HLSBS, CBS, STG, and EBUE (Sections 6.3.1, 6.3.4, 7.3.2.4, 30.5; Figure 30-12; Table 30-3; Box CC-MB). Key risks involve changes in the distribution and abundance of key fishery species (*high confidence*; Section 30.6.2.1; Figure 30-12 A,B,G,H) as well as the spread of disease and invading organisms, each of which has the potential to impact ecosystems as well as aquaculture and fishing (Sections 6.3.5, 6.4.1.1, 6.5.3, 7.3.2.4, 7.4.2, 29.5.3-4; Table 30-3). Adaptation to these changes may be possible in the short-term through dynamic fisheries policy and management (i.e., relocation of fishing effort; Table 30-3), as well as monitoring and responding to potential invading species in coastal settings. The increasing frequency of thermal extremes (Box CC-HS) will also increase the risk that the thermal threshold of corals and other organisms is exceeded on a more frequent

Table 30-3 (continued)

Risks to humans and infrastructure (continued)

Key risk	Adaptation issues & prospects	Climatic drivers	Timeframe	Risk & potential for adaptation
Reduced coastal socioeconomic security. (*high confidence*) [5.5.2, 5.5.4, 30.6.5, 30.7.1]	Human adaptation options involve (1) protection using coastal defences (e.g. seawalls where appropriate and economic) and soft measures (e.g., mangrove replanting and enhancing coral growth); (2) accommodation to allow continued occupation of coastal areas by making changes to human activities and infrastructure; and (3) managed retreat as a last viable option. Vary from large-scale engineering works to smaller scale community projects. Options are available under the more traditional CZM (coastal zone management) framework but increasingly under DRR (disaster risk reduction) and CCA (climate change adaptation) frameworks.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

**High confidence* in existence of adaptation measures, *Low confidence* in magnitude of risk reduction

Key risk	Adaptation issues & prospects	Climatic drivers	Timeframe	Risk & potential for adaptation
Reduced livelihoods and increased poverty. (*medium confidence*) [6.4.1-2, 30.6.2, 30.6.5]	Human adaptation options involve the large-scale translocation of industrial fishing activities following the regional decreases (low latitude) versus increases (high latitude) in catch potential and shifts in biodiversity. Artisanal fisheries are extremely limited in their adaptation options by available financial resources and technical capacities, except for their potential shift to other species of interest.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts due to increased frequency of harmful algal blooms (*medium confidence*) [6.4.2.3]	Adaptation options include improved monitoring and early warning system, reduction of stresses favoring harmful algal blooms, mainly pollution and eutrophication, as well as the avoidance of contaminated areas and fisheries products.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts on marine resources threatening regional security as territorial disputes and food security challenges increase (*limited evidence, medium agreement*) [IPCC 2012, 30.6.5, 12.4-12.6, 29.3]	Decrease in marine resources, movements of fish stocks and opening of new seaways , and impacts of extreme events coupled with increasing populations will increase the potential for conflict in some regions, drive potential migration of people, and increase humanitarian crises.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	
Impacts on shipping and infrastructure for energy and mineral extraction increases as storm intensity and wave height increase in some regions (e.g., high latitudes) (*high confidence*) [IPCC 2012, 30.6.5, 12.4-12.6, 29.3]	Adaptation options are to limit activities to particular times of the year and/or develop strategies to decrease the vulnerability of structures and operations.		Present Near term (2030 – 2040) Long term (2080 – 2100) 2°C 4°C	

CBS = Coastal Boundary Systems; EBUE = Eastern Boundary Upwehlling Ecosystems; EUS = Equatorial Upwelling Systems; HLSBS = High-Latitude Spring Bloom Systems; SES = Semi-Enclosed Seas; STG = Subtropical Gyres.

basis (especially in CBS, STG, SES, HLSBS, and EUS regions; Sections 6.2, 30.5; Box CC-CR). These changes pose a key risk to vulnerable ecosystems such as mangroves and coral reefs, with potential to have a series of serious impacts on fisheries, tourism, and coastal ecosystem services such as coastal protection (Sections 5.4.2.4, 6.3.2, 6.3.5, 6.4.1.3, 7.2.1.2, 29.3.1.2, 30.5; Table 30-3; Box CC-CR). Genetic adaptation of species to increasing levels of stress may not occur fast enough given fairly long generation times of organisms such as reef-building corals and many other invertebrates and fish (Table 30-3). In this case, risks may be reduced by addressing stresses not related to climate change (e.g., pollution, overfishing), although this strategy could have minimal impact if further increases in sea temperature occur (*high confidence*).

Loss of these important coastal ecosystems is associated with emerging risks associated with the collapse of some coastal fisheries along with livelihoods, food, and regional security (*medium confidence*). These changes are *likely* to be exacerbated by other key risks such as coastal inundation and habitat loss due to SLR, as well as intensified precipitation events (*high confidence*; Section 5.4; Box CC-CR). Adaptation options in this case include engineered coastal defenses, reestablishing coastal vegetation such as mangroves, protecting water supplies from salination, and developing strategies for coastal communities to withdraw to less vulnerable locations over time (Section 5.5).

The recent decline in O_2 concentrations has been ascribed to warming through the effect on ocean mixing and ventilation, as well as the solubility of O_2 and its consumption by marine microbes (Sections 6.1.1.3, 6.3.3, 30.3.2.3, 30.5.7). This represents a key risk to ocean ecosystems (*medium confidence*; Figure 30-12 5,6,C). These changes increase the vulnerability of marine communities, especially those below the euphotic zone, to hypoxia and ultimately lead to a restriction of suitable habitat (*high confidence*; Figure 30-12 5). In the more extreme case, often exacerbated by the contribution of organic carbon from land-based sources, "dead zones" may form. Decreasing oxygen, consequently, is *very likely* to increase the vulnerability of fisheries and aquaculture (*medium confidence*; Figure 30-12 C), and consequently puts livelihoods at risk, particularly in EBUE (e.g., California and Humboldt Current ecosystems; Section 30.5.5), SES (e.g., Baltic and Black Seas; Section 30.5.3), and CBS (e.g., Gulf of Mexico, northeast Indian Ocean; Sections 30.3.2.3, 30.5.4). It is *very likely* that the warming of surface waters has also increased the stratification of the upper ocean by about 4% between 0 and 200 m from 1971 to 2010 in all oceans north of about 40°S. In many cases, there is significant adaptation opportunity to reduce hypoxia locally by reducing the flow of organic carbon, hence microbial activity, within these coastal systems (Section 30.5.4). Relocating fishing effort, and modifying procedures associated with industries such as aquaculture, may offer some opportunity to adapt to these changes (*likely*). Declining O_2 concentrations are *likely* to have significant impacts on DS habitats, where organisms are relatively sensitive to environmental changes of this nature owing to the very constant conditions under which they have evolved (Section 30.5.7).

Ocean acidification has increased the vulnerability of ocean ecosystems by affecting key aspects of the physiology and ecology of marine organisms (particularly in CBS, STG, and SES; Section 6.3.2; Table 30-3; Box CC-OA). Decreasing pH and carbonate ion concentrations reduce the ability of marine organisms to produce shells and skeletons, and may interfere with a range of biological processes such as reproduction, gas exchange, metabolism, navigation ability, and neural function in a broad range of marine organisms that show minor to major influences of ocean acidification on their biology (Sections 6.3.2, 30.3.2.2; Box CC-OA). Natural variability in ocean pH can interact with ocean acidification to create damaging periods of extremes (i.e., high CO_2, low O_2 and pH), which can have a strong effect on coastal activities such as aquaculture (*medium confidence*; Section 6.2; Figure 30-12 A; Box CC-UP). There may be opportunity to adapt aquaculture to increasingly acidic conditions by monitoring natural variability and restricting water intake to periods of optimal conditions. Reducing other non-climate change or ocean acidification associated stresses also represents an opportunity to build greater ecological resilience against the impacts of changing ocean carbonate chemistry. Ocean acidification is also an emerging risk for DS habitats as CO_2 continues to penetrate the Ocean, although the impacts and adaptation options are poorly understood and explored. Ocean acidification has heightened importance for some groups of organisms and ecosystems (Box CC-OA). In ecosystems that are heavily dependent on the accumulation of calcium carbonate over time (e.g., coral reefs, *Halimeda* beds), increasing ocean acidification puts at risk ecosystems services that are critical for hundreds of thousands of marine species, plus people and industries, particularly within CBS, STG, and SES (*high confidence*). Further risks may emerge from the non-linear interaction of different factors (e.g., increasing ocean temperature may amplify effects of ocean acidification, and vice versa) and via the interaction of local stressors with climate change (e.g., interacting changes may lead to greater ecosystems disturbances than each impact on its own). There is an urgent need to understand these types of interactions and impacts, especially given the long time it will take to return ocean ecosystems to preindustrial pH and carbonate chemistry (i.e., tens of thousands of years (FAQ 30.1) should CO_2 emissions continue at the current rate).

It is *very likely* that surface warming has increased stratification of the upper ocean, contributing to the decrease in O_2 along with the temperature-related decreases in oxygen solubility (WGI AR5 Section 3.8.3). Changes to wind speed, wave height, and storm intensity influence the location and rate of mixing within the upper layers of the Ocean and hence the concentration of inorganic nutrients (e.g., in EBUE, EUS; Figure 30-12 1,3). These changes to ocean structure increase the risks and vulnerability of food webs within the Ocean. However, our understanding of how primary productivity is going to change in a warming and more acidified ocean is limited, as is our understanding of how upwelling will respond to changing surface wind as the world continues to warm (Boxes CC-PP, CC-UP). As already discussed, these types of changes can have implications for the supply of O_2 into the Ocean and the upward transport of inorganic nutrients to the euphotic zone. Although our understanding is limited, there is significant potential for regional increases in wind speed to result in greater rates of upwelling and the supply of inorganic nutrients to the photic zone. Although this may increase productivity of phytoplankton communities and associated fisheries, greater rates of upwelling can increase the risk of hypoxic conditions developing at depth as excess primary production sinks into the Ocean and stimulates microbial activity at depth (Sections 6.1.1.3, 30.3.2.3, 30.5.5; Table 30-3). Changes in storm intensity may increase the risk of damage to shipping and industrial infrastructure, which increases the risk of accidents and delays to the transport of products between countries, security operations, and the extraction of minerals from coastal and oceanic areas (Section 30.6.2; IPCC, 2012).

Frequently Asked Questions

FAQ 30.5 | How can we use non-climate factors to manage climate change impacts on the oceans?

The Ocean is exposed to a range of stresses that may or may not be related to climate change. Human activities can result in pollution, eutrophication (too many nutrients), habitat destruction, invasive species, destructive fishing, and over-exploitation of marine resources. Sometimes, these activities can increase the impacts of climate change, although they can, in a few circumstances, dampen the effects as well. Understanding how these factors interact with climate change and ocean acidification is important in its own right. However, reducing the impact of these non-climate factors may reduce the overall rate of change within ocean ecosystems. Building ecological resilience through ecosystem-based approaches to the management of the marine environment, for example, may pay dividends in terms of reducing and delaying the effects of climate change (*high confidence*).

The proliferation of key risks and vulnerabilities to the goods and services provided by ocean ecosystems as a result of ocean warming and acidification generate a number of key risks for the citizens of almost every nation. Risks to food security and livelihoods are expected to increase over time, aggravating poverty and inequity (Table 30-3). As these problems increase, regional security is likely to deteriorate as disputes over resources increase, along with increasing insecurity of food and nutrition (Sections 12.4-6, 29.3.3, 30.6.5; Table 30-3; IPCC, 2012).

30.7.2. Global Frameworks for Decision Making

Global frameworks for decision making are central to management of vulnerability and risk at the scale and complexity of the world's oceans. General frameworks and conventions for policy development and decision making within oceanic and coastal regions are important in terms of the management of stressors not directly due to ocean warming or acidification, but that may influence the outcome of these two factors. Tables 30-3 and 30-4 outline a further set of challenges arising from multiple interacting stressors, as well as potential risks and vulnerabilities, ramifications, and adaptation options. In the latter case, examples of potential global frameworks and initiatives for beginning and managing these adaptation options are described. These frameworks represent opportunities for global cooperation and the development of international, regional, and national policy responses to the challenges posed by the changing ocean (Kenchington and Warner, 2012; Tsamenyi and Hanich, 2012; Warner and Schofield, 2012).

The United Nations Convention on the Law of the Sea (UNCLOS) was a major outcome of the third UN Conference on the Law of the Sea (UNCLOS III). The European Union and 164 countries have joined in the

Table 30-4 | Ramifications, adaptation options, and frameworks for decision making for ocean regions. Symbols for primary drivers: IC = ice cover; NU = nutrient concentration; OA = ocean acidification; SLR = sea level rise; SS = storm strength; T = sea temperature (↑ = increased; ↓ = decreased; * = uncertain).

Primary driver(s)	Biophysical change projected	Key risks and vulnerabilities	Ramifications	Adaptation options	Policy frameworks and initiatives (examples)	Key references and chapter sections
↑T, ↑OA	Spatial and temporal variation in primary productivity (*medium confidence* at global scales; Box CC-PP)	Reduced fisheries production impacts important sources of income to some countries while others may see increased productivity (e.g., as tuna stocks shift eastwards in the Pacific) (*medium confidence*).	Reduced national income, increased unemployment, plus increase in poverty. Potential increase in disputes over national ownership of key fishery resources (*likely*)	Increased international cooperation over key fisheries. Improved understanding of linkages between ocean productivity, recruitment, and fisheries stock levels. Implementation of the regional "vessel day scheme" provides social and economic incentives to fisheries and fishers for adaptation.	UNCLOS, PEMSEA, CTI, RFMO agreements, UNSFSA	Bell et al. (2011, 2013a); Tsamenyi and Hanich (2012); Sections 6.4.1, 6.5.3, 30.6.2.1, 30.7.2; Box CC-PP
↑T, ↑OA	Ecosystem regime shifts (e.g., coral to algal reefs; structural shifts in phytoplankton communities) (*medium confidence*)	Reduced fisheries production of coastal habitats and ecosystems such as coral reefs (*medium confidence*).	Decreased food and employment security and human migration away from coastal zone (*likely*)	Strengthen coastal zone management to reduce contributing stressors (e.g., coastal pollution, over-harvesting, and physical damage to coastal resources). Promote Blue Carbon[a] initiatives.	PEMSEA, CTI, PACC, MARPOL, UNHCR, CBD, International Organization for Migration, Global Environment Facility, International Labor Organization	Bell et al. (2013a); Sections 5.4.3, 6.3.1–2, 12.4, 29.3.1, 29.3.3, 30.5.2–4, 30.5.6, 30.6.1, 30.6.2.1; Box CC-CR
		Tourist appeal of coastal assets decreases as ecosystems change to less "desirable" state, reducing income to some countries (*low confidence*).	Increased levels of coastal poverty in some countries as tourist income decreases (*likely*)	As above, strengthen coastal zone management and reduce additional stressors on tourist sites; implement education programs and awareness among visitors. Diversify tourism activities.	CBD, PEMSEA, CTI, PACC, UNHCR, MARPOL	Kenchington and Warner (2012); Sections 5.5.4.1, 6.4.1–2, 10.6, 30.6.2.2
		Increased risk of some diseases (e.g., ciguatera, harmful algal blooms) as temperatures increase shift and ecosystems shift away from coral dominance (*low confidence*).	Increased disease and mortality; decreases in coastal food resources and fisheries income (*likely*)	Increase monitoring and education surrounding key risks (e.g., ciguatera); develop alternate fisheries and income for periods when disease incidence increases, and develop or update health response plans.	National policy strategies and regional cooperation needed	Llewellyn (2010); Sections 6.4.2.3, 10.6, 29.3.3.2, 29.5.3, 30.6.3
		Increased poverty and dislocation of coastal people (particularly in the tropics) as coastal resources such as fisheries degrade (*medium confidence*)	Increased population pressure on migration destinations (e.g., large regional cities), and reduced freedom to navigate in some areas (as criminal activity increases) (*likely*)	Develop alternative industries and income for affected coastal people. Strengthen coastal security both nationally and across regions. Increase cooperation over handling of criminal activities.	UNCLOS, PEMSEA, CTI, International Ship and Port Facility Security, IMO, Bali Process, Association of Southeast Asian Nations MLA Treaty and bilateral extradition and MLA agreements	Kaye (2012); Rahman (2012); Sections 12.4–6,29.3.3, 29.6.2, 30.6.5

Continued next page →

[a]Blue Carbon initiatives include conservation and restoration of mangroves, saltmarsh, and seagrass beds as carbon sinks (Section 30.6.1).

Notes: CBD = Convention on Biological Diversity; CTI = Coral Triangle Initiative; IHO = International Hydrographic Organization; IOM = International Organization of Migration; ISPS = International Ship and Port Facility Security; MARPOL = International Convention for the Prevention of Pollution From Ships; MLA = mutual legal assistance; PACC = Pacific Adaptation to Climate Change Project; PEMSEA = Partnerships in Environmental Management for the Seas of East Asia; RFMO = Regional Fisheries Management Organizations; UNCLOS = United Nations Convention on the Law of the Sea; UNHCR = United Nations High Commissioner for Refugees; UNSFSA = United Nations Straddling Fish Stocks Agreement.

Table 30-4 (continued)

Primary driver(s)	Biophysical change projected	Key risks and vulnerabilities	Ramifications	Adaptation options	Policy frameworks and initiatives (examples)	Key references and chapter sections
↑T	Migration of organisms and ecosystems to higher latitudes (*high confidence*)	Reorganization of commercial fish stocks and ecological regime shifts (*medium to high confidence*)	Social and economic disruption (*very likely*)	Increase international cooperation and improve understanding of regime changes; implement early-detection monitoring of physical and biological variables and regional seasonal forecasting; include related uncertainties into fisheries management; provide social and economic incentives for industry.	UNCLOS, CBD, RFMO agreements, UNSFSA	Sections 7.4.2, 6.5, 30.5, 30.6.2.1; Box CC-MB
		Increase in abundance, growing season, and distributional extent of pests and fouling species (*medium confidence*)	Increased disease risk to aquaculture and fisheries. Income loss and increased operating and maintenance costs (*very likely*)	Increase environmental monitoring; promote technological advances to deal with pest and fouling organisms; increase vigilance and control related to biosecurity.	IMO, ballast water management, Anti- Fouling Convention	Sections 6.4.1.5, 7.3.2.4, 29.5.3–4, 30.6.2.1; Box CC-MB
		Threats to human health increase due to expansion of pathogen distribution to higher latitudes (*low confidence*)	Increased disease and mortality in some coastal communities (*likely*)	Reduce exposure through increased monitoring and education, adoption, or update of health response plans to outbreaks.	UNICEF, World Health Ogranization, IHOs, and national governments	Myers and Patz (2009); Sections 6.4.3, 10.8.2, 11.7, 29.3.3, 30.6.3; Box CC-MB
↑T, ↑NU, ↑OA*	Increased incidence of harmful algal blooms (*low confidence*)	Increased threats to ecosystems, fisheries, and human health (*medium confidence*)	Reduced supply of marine fish and shellfish and greater incidence of disease among some coastal communities (*likely*)	Provide early-detection monitoring and improve predictive models; provide education and adoption or update of health response plans.	CTI, PEMSEA, World Health Organization, MARPOL	Llewellyn (2010);, Sections 30.6.3, 11.7, 6.4.2.3
↑T	Increased precipitation as a result of intensified hydrological cycle in some coastal areas (*medium confidence*)	Increased freshwater, sediment, and nutrient flow into coastal areas; increase in number and severity of flood events (*medium to high confidence*)	Increasing damage to coastal reef systems with ecological regime shifts in many cases (*very likely*)	Improve management of catchment and coastal processes; expand riparian vegetation along creeks and rivers; improve agricultural retention of soils and nutrients.	CTI, PEMSEA, Secretariat of the Pacific Regional Environment Programme	Sections 3.4, 29.3.1, 30.5.4, 30.6.1
↑T	Changing weather patterns, storm frequency (*medium confidence*)	Increased risk of damage to infrastructure such as that involved in shipping and oil and gas exploration and extraction (*medium to low confidence*)	Increased damage and associated costs (*likely*)	Adjust infrastructure specifications, develop early-warning systems, and update emergency response plans to extreme events.	IMO	IPCC (2012); Sections 10.4.4, 29.3, 30.6.2.3–4
↑SLR, ↑SS	Increased wave exposure of coastal areas and increased sea level (*high confidence*)	Exposure of coastal infrastructure and communities to damage and inundation, increased coastal erosion (*high confidence*)	Increased costs to human towns and settlements, numbers of displaced people, and human migration (*very likely*)	Develop integrated coastal management that considers SLR in planning and decision making; increase understanding of the issues through education.	UNICEF, IHOs, and national governments	Warner (2012); Sections 5.5, 12.4.1, 29.5.1, 30.3.1.2, 30.6.5
		Inundation of coastal aquifers reduces water supplies and decreases coastal agricultural productivity (*high confidence*).	Reduced food and water security leads to increased coastal poverty, reduced food security, and migration (*very likely*).	Assist communities in finding alternatives for food and water, or assist in relocation of populations and agriculture from vulnerable areas.	UNICEF, IHOs, and national governments.	Warner (2012); Sections 5.4.3, 12.4.1, 29.3.2, 30.3.1.2
↑SLR	Risk of inundation and coastal erosion, especially in low-lying countries (*high confidence*)	UNCLOS-defined limits of maritime jurisdiction will contract as national baselines shift inland. Potential uncertainty increases in some areas with respect to the international boundaries to maritime jurisdiction (*high confidence*).	Lack of clarity increases, as do disputes over maritime limits and maritime jurisdiction. Some nations at risk of major losses to their territorial waters (*very likely*)	Seek resolution of "shifting national baselines" issue (retreat and redefinition, stabilization, or fixation of exclusive economic zones and other currently defined maritime jurisdiction limits).	UNCLOS	IPCC (2012); Schofield and Arsana (2012); Warner and Schofield (2012); Sections 5.5, 30.6.5
↑T, ↓IC	Loss of summer sea ice (*high confidence*)	Access to northern coasts of Canada, USA, and Russia increases security concerns (*high confidence*).	Potential for increased tension on different interpretations of access rights and boundaries (*likely to very likely*)	Seek early resolution of areas in dispute currently and in the future.	UNCLOS	Chapter 28
		New resources become available as ice retreats, increasing vulnerability of international borders in some cases (*medium confidence*).	Tensions over maritime claims and ownership of resources (*likely*)	Sort out international agreements.		

Convention. UNCLOS replaced earlier frameworks that were built around the "freedom of the seas" concept and that limited territorial rights to 3 nm off a coastline. UNCLOS provides a comprehensive framework for the legitimate use of the Ocean and its resources, including maritime zones, navigational rights, protection and preservation of the marine environment, fishing activities, marine scientific research, and mineral resource extraction from the seabed beyond national jurisdiction. The relationship between climate change and UNCLOS is not clear and depends on interpretation of key elements within the UNFCCC (United Nations Framework Convention for Climate Change) and Kyoto Protocol (Boyle, 2012). However, UNCLOS provides mechanisms to help structural adaptation in response to challenges posed by climate change. In a similar way, there is a wide range of other policy and legal frameworks that structure and enable responses to the outcomes of rapid anthropogenic climate change in the Ocean.

There are many existing international conventions and agreements that explicitly recognize climate change (Table 30-4). The UN Straddling Fish Stocks Agreement (UNSFSA) aims at enhancing international cooperation of fisheries resources, with an explicit understanding under Article 6 that management needs to take account "existing and predicted oceanic, environmental and socio-economic conditions" and to undertake "relevant research, including surveys of abundance, biomass surveys, hydro-acoustic surveys, research on environmental factors affecting stock abundance, and oceanographic and ecological studies" (UNSFSA, Annex 1, Article 3). International conventions such as these will become increasingly important as changes to the distribution and abundance of fisheries are modified by climate change and ocean acidification.

Global frameworks for decision making are increasingly important in the case of the Ocean, most of which falls outside national boundaries (Oude Elferink, 2012; Warner, 2012). Approximately 64% of the Ocean (40% of the Earth's surface) is outside EEZs and continental shelves of the world's nations (high seas and seabed beyond national jurisdiction). With rapidly increasing levels of exploitation, there are increasing calls for more effective decision frameworks aimed at regulating fishing and other activities (e.g., bio-prospecting) within these ocean "commons." These international frameworks will become increasingly valuable as nations respond to impacts on fisheries resources that stretch across national boundaries. One such example is the multilateral cooperation that was driven by President Yudhoyono of Indonesia in August 2007 and led to the Coral Triangle Initiative on Coral Reefs, Fisheries, and Food Security (CTI), which involves region-wide (involving 6.8 million km^2 including 132,800 km of coastline) cooperation between the governments of Indonesia, Philippines, Malaysia, Papua New Guinea, the Solomon Islands, and Timor Leste on reversing the decline in coastal ecosystems such as coral reefs (Clifton, 2009; Hoegh-Guldberg et al., 2009; Veron et al., 2009). Partnerships, such as CTI, have the potential to provide key frameworks to address issues such as interaction between the over-exploitation of coastal fishing resources and the recovery of reefs from mass coral bleaching and mortality, and the implications of the movement of valuable fishery stocks beyond waters under national jurisdiction.

An initiative called the Global Partnership for Oceans set out to establish a global framework with which to share experience, resources, and expertise, as well as to engage governments, industry, civil, and public sector interests in both understanding and finding solutions to key issues such as overfishing, pollution, and habitat destruction (Hoegh-Guldberg et al., 2013). Similarly, the Areas Beyond National Jurisdiction (ABNJ, Global Environment Facility) Initiative has been established to promote the efficient, collaborative, and sustainable management of fisheries resources and biodiversity conservation across the Ocean.

Global partnerships are also essential for providing support to the many nations that often do not have the scientific or financial resources to solve the challenges that lie ahead (Busby, 2009; Mertz et al., 2009). In this regard, international networks and partnerships are particularly significant in terms of assisting nations in developing local adaptation solutions to their ocean resources. By sharing common experiences and strategies through global networks, nations have the chance to tap into a vast array of options with respect to responding to the negative consequences of climate change and ocean acidification on the world's ocean and coastal resources.

30.7.3. Emerging Issues, Data Gaps, and Research Needs

Although there has been an increase in the number of studies being undertaken to understand the physical, chemical, and biological changes within the Ocean in response to climate change and ocean acidification, the number of marine studies of ecological impacts and risks still lag behind terrestrial studies (Hoegh-Guldberg and Bruno, 2010; Poloczanska et al., 2013). Rectifying this gap should be a major international objective given the importance of the Ocean in terms of understanding and responding to future changes and consequences of ocean warming and acidification.

30.7.3.1. Changing Variability and Marine Impacts

Understanding the long-term variability of the Ocean is critically important in terms of the detection and attribution of changes to climate change (Sections 30.3, 30.5.8), but also in terms of the interaction between variability and anthropogenic climate change. Developing instrument systems that expand the spatial and temporal coverage of the Ocean and key processes will be critical to documenting and understanding its behavior under further increases in average global temperature and changes in the atmospheric concentration of CO_2. International collaborations such as the Argo network of oceanographic floats illustrate how international cooperation can rapidly improve our understanding of the physical behavior of the Ocean and will provide important insight into its long-term subsurface variability (Schofield et al., 2013).

30.7.3.2. Surface Wind, Storms, and Upwelling

Improving our understanding of the potential behavior of surface wind in a warming world is centrally important to our understanding of how upwelling will change in key regions (e.g., EUS, EBUE; Box CC-UP). Understanding these changes will provide important information for future fisheries management but will also illuminate the potential risks of intensified upwelling leading to hypoxia at depth and the potential expansion of "dead zones" (Sections 30.3.2, 30.5.2-4). Understanding surface wind in a warming climate will also yield important information on surface mixing as well as how surface wave height might also vary, improving our understanding of potential interactions in coastal areas between wind, waves, and SLR (Section 30.3.1). Given the importance of mixing and upwelling to the supply of inorganic nutrients to the surface layers of the ocean, understanding these important phenomena at the ocean-atmosphere interface will provide important insight into how ocean warming and acidification are likely to impact ecosystems, food webs, and ultimately important fisheries such as those found along the west coasts of Africa and the Americas.

30.7.3.3. Declining Oxygen Concentrations

The declining level of O_2 in the Ocean is an emerging issue of major importance (Section 30.3.2). Developing a better understanding of the role and temperature sensitivity of microbial systems in determining O_2 concentrations will enable a more coherent understanding of the changes and potential risks to marine ecosystems. Given the importance of microbial systems to the physical, chemical, and biological characteristics of the Ocean, it is extremely important that these systems receive greater focus, especially with regard to their response to ocean warming and acidification. This is particularly important for the DS (>1000 m), which is the most extensive habitat on the planet. In this respect, increasing our understanding of DS habitats and how they may be changing under the influence of climate change and ocean acidification is of great importance. Linkages between changes occurring in the surface layers and those associated with the DS are particularly important in light of our need to understand how rapidly changes are occurring and what the implications are for the metabolic activity and O_2 content of DS habitats.

30.7.3.4. Ocean Acidification

The rapid and largely unprecedented changes to ocean acidification represent an emerging issue given the central importance of pH and the concentration of ions such as carbonate in the biology of marine organisms (Box CC-OA). Despite the relatively short history of research on this issue, there are already a large number of laboratory and field studies that demonstrate a large range of effects across organisms, processes, and ecosystems. Key gaps (Gattuso et al., 2011) remain in our understanding of how ocean acidification will interact with other changes in the Ocean, and whether or not biological responses to ocean acidification are necessarily linear. The vulnerability of fishery species (e.g., molluscs) to ocean acidification represents an emerging issue, with a need for research to understand and develop strategies for fishery and aquaculture industries to minimize the impacts. Understanding of how carbonate structures such as coral reefs and *Halimeda* beds will respond to a rapidly acidifying ocean represents a key gap and research need, especially in understanding the rate at which consolidated carbonate structures and related habitats are likely to erode and dissolve. Interactions between ocean acidification, upwelling, and decreasing O_2 represent additional areas of concern and research. There is also a need to improve our understanding of the socioeconomic ramifications of ocean acidification (Turley and Boot, 2011; Hilmi et al., 2013).

30.7.3.5. Net Primary Productivity

Oceanic phytoplankton are responsible for approximately 50% of global net primary productivity. However, our understanding of how oceanic primary production is likely to change in a warmer and more acidified ocean is uncertain (Boxes CC-PP, CC-UP). Changes in net primary productivity will resonate through food webs and ultimately affect fisheries production. Given the central role that primary producers and their associated ecological processes play in ocean ecosystem functioning, the understanding of how net primary productivity is likely to vary at global and regional levels is improved (Sections 30.5.2, 30.5.5). At the same time, understanding how plankton communities will vary spatially and temporarily will be important in any attempt to understand how fish populations will fare in a warmer and more acidified ocean. The research challenge is to determine when and where net primary production is expected to change, coupled with research on adaptation strategies for coping with the changes to the global distribution of seafood procurement, management, and food security.

30.7.3.6. Movement of Marine Organisms and Ecosystems

Marine organisms are moving generally toward higher latitudes or deeper waters consistent with the expectation of a warming ocean. Our current understanding of which organisms and ecosystems are moving, ramifications for reorganization of ecosystems and communities, and the implications for nations is uncertain at best. Given the implications for fisheries, invasive species, and the spread of disease, it is imperative that our understanding of the movement of ecosystems is improved. Documentation of species' responses and a deeper understanding of the processes that lead to persistent range shifts, and a focus on the ecosystem, social, and economic implications of range shifts is an important research need.

30.7.3.7. Understanding Cumulative and Synergistic Impacts

Understanding cumulative and synergistic impacts is poorly developed for ocean systems. Much of our understanding has been built on experimental approaches that are focused on single stressors that respond gradually without interaction or impacts that accumulate over time (Table 30-3). Multifactorial experiments exploring the impact of combined variables (e.g., elevated temperature and acidification at the same time) will enable more realistic projections of the future to be established. Equally, developing a better understanding of how biological and ecological responses change in relation to key environmental variables should also be a goal of future research. In this regard, assumptions that responses are likely to be gradual and linear over time ultimately have little basis, yet are widespread within the scientific literature.

30.7.3.8. Reorganization of Ecosystems and Food Webs

The pervasive influence of ocean warming and acidification on the distribution, abundance, and function of organisms and processes has and will continue to drive the reorganization of ecosystems and food webs (*virtually certain*; Hoegh-Guldberg and Bruno, 2010; Poloczanska et al., 2013; Box CC-MB). One of the inevitable outcomes of differing tolerances and responses to climate change and ocean acidification is the development of novel assemblages of organisms in the near future. Such communities are likely to have no past or contemporary counterparts, and will consequently require new strategies for managing coastal areas and fisheries. Changes to a wide array of factors related or not related to climate change have the potential to drive extremely complex changes in community structure and, consequently, food web dynamics. Developing a greater capability for detecting and understanding these changes will be critical for future management of ocean and coastal resources.

30.7.3.9. Socio-ecological Resilience

Many communities depend on marine ecosystems for food and income yet our understanding of the consequences of environmental degradation is poor. For example, although there is *high confidence* that coral reefs will continue to deteriorate at current rates of climate change and ocean acidification (Gardner et al., 2003; Bruno and Selig, 2007; De'ath et al., 2012), there is relatively poor understanding of the implications for the hundreds of millions of people who depend on these important coastal ecosystems for food and livelihoods. Improving our understanding of how to reinforce socio-ecological resilience in communities affected by the deterioration of key coastal and oceanic ecosystems is central to developing effective adaptation responses to these growing challenges (Section 30.6, Tables 30-3, 30-4).

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The Ocean
Supplementary Material

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Additional Information on Past, Present, and Future Ocean Conditions

These supplementary materials include further information on primary productivity and fisheries as well as past, present, and future (over the next 100 years) pH, Aragonite Saturation State, and Sea Surface temperatures (SST).

SM30.1. Primary Productivity and Long-Term Fisheries Catch

Different ocean sub-regions have substantially different primary and fishery productivities. Notably, over 80% of fisheries production is associated with three Ocean sub-regions: Northern hemisphere High Latitude Spring Bloom Systems (HLSBS-North), Coastal Boundary Systems (CBS), and Eastern Boundary Upwelling Ecosystems (EBUE; Table SM30-1, Figure 30-1b).

SM30.2. Definition as well as Coolest and Warmest Months for Key Ocean Sub-Regions Examined in Chapter 30

The HadISST1.1 data set (Rayner et al., 2003) was used to explore SST trends over the past 60 years (1950–2009; main text, Table 30-1), particularly in terms of long-term trends in average temperature as well as long-term trends in the coolest and warmest months of the year (Table SM30-2). The regions are outlined in Figure SM30-1 and Table SM30-2 (column 1). These data are discussed in the main text of Chapter 30 (Table 30-1).

Table SM30-1 | Percentage surface area of the ocean, average primary production, and fisheries productivity of key ocean sub-regions (Figure 30-1). Also shown are the primary IPCC assessments (by chapter number and sections of Chapter 30) that are relevant to each of the sub-regions. Details of calculations are as follows: (1) Calculation of the surface areas of the ocean sub-regions was made by transferring the boundary lines of the sub-regions to Google Maps and then using a graphical planimeter freeware provided in Google Maps. The planimeter program was made by Europa Technologies, MapLink, Tele Atlas, INEGI. (2) Calculation of primary production for each sub-region was carried out using a similar approach by transferring the original map of world primary production in Field et al. (1998) to the planimeter tool (freeware provided in Google Maps). Areas were weighted for each color scale value (g C m^{-2}) to get numbers in g C. These were summed for each area within a sub-region to get total values of g C. (3) Calculation of fish catch for each sub-region was based on the FAO Statistics on world fish catch in their standard regional areas (1–88). However, as the FAO standard catch areas do not completely resolve the spatial areas of the sub-regions and partly cross different ocean sub-regions of Chapter 30, the division of fish catches in Large Marine Ecosystems (LMEs, as displayed in the project The Sea Around Us: http://www.seaaroundus.org/) was used to correct the numbers. The data from this source are, however, also based on the same FAO Fish Statistics. *Based on Field et al. (1998); **Average fish catch 1970–2006 Based on FAO; ***Not calculated.

Ocean sub-region		Area (%)	Primary productivity (%)*	Long-term fish catch (%)**	Relevant IPCC regions (chapters)	Chapter 30 sections
1. High Latitude Spring Bloom System (HLSBS)	Northern Hemisphere	10.60	22.74	29.20	23–24, 26, 28	30.5.6, 30.6.2.1, Box CC-MB
	Southern Hemisphere	14.40	20.55	6.82	22, 25, 28	30.5.6
2. Equatorial Upwelling Systems (EUS)		8.20	9.01	4.68	22, 27, 29	30.5.3, 30.6.2.1, Box CC-CR
3. Semi-Enclosed Seas (SES)		1.12	2.35	3.28	22, 23	30.5.5
4. Coastal Boundary Systems (CBS)		6.29	10.64	28.02	22, 24–26, 29	30.5.4, 30.6.2.1, Box CC-CR
5. Eastern Boundary Upwelling Systems (EBUE)		1.80	6.97	19.21	22, 26, 27	30.5.5, Box CC-UP
6. Sub-Tropical Gyres (STG)		40.55	21.20	8.26	22, 24–26, 29	30.5.6, 30.6.2.1, Box CC-PP, Box CC-CR
7. Deep Sea (DS)***		N/A	N/A	N/A	22–29	30.5.7
Arctic and Antarctic System		17.04	6.54	0.53	23, 24, 25, 26	

Table SM30-2 | The coolest and warmest months for ocean sub-regions identified in Figure SM30-1. Entire regions (e.g., Indian Ocean) or parts of sub-regions (e.g., eastern portion of the North Pacific Ocean) are indicated by letters in the first column that relate to those inscribed on Figure SM30-1. Coolest and hottest months were identified from an analysis of the last 60 years of sea surface temperature using the HadISST1.1 dataset (Rayner et al., 2003).

Sub-region	Map component (see Figure SM30-1)	Area	Coolest month	Warmest month
1. High Latitude Spring Bloom Systems (HLSBS)	A	Indian Ocean	September	February
	B	North Atlantic Ocean	March	August
	C	South Atlantic Ocean	August	February
	D	North Pacific Ocean (west)	March	August
	E	North Pacific Ocean (east)	March	August
	D+E	North Pacific Ocean	March	August
	F	South Pacific Ocean (west)	September	February
	G	South Pacific Ocean (east)	September	February
	F+G	South Pacific Ocean	September	February
2. Equatorial Upwelling Systems (EUS)	H	Atlantic Equatorial Upwelling	August	April
	I	Pacific Equatorial Upwelling	September	April
3. Semi-Enclosed Seas (SES)	J	Arabian Gulf	February	August
	K	Baltic Sea	March	August
	L	Black Sea	March	August
	M	Mediterranean Sea	February	August
	N	Red Sea	February	August
4. Coastal Boundary Systems (CBS)	O	Western Atlantic	August	March
	P	Caribbean Sea/Gulf of Mexico	February	September
	Q	Indian Ocean (west)	August	May
	R	Indian Ocean (east)	August	April
	S	Indian Ocean (east)/Southeast Asia/Pacific Ocean (west)	February	August
5. Eastern Boundary Upwelling Ecosystems (EBUE)	T	Benguela Current	August	March
	U	California Current	March	September
	V	Canary Current	February	September
	W	Humboldt Current	September	February
6. Sub-Tropical Gyres (STG)	X	Indian Ocean	August	March
	Y	North Atlantic Ocean	March	August
	Z	South Atlantic Ocean	September	March
	AA	North Pacific Ocean (west)	February	August
	AB	North Pacific Ocean (east)	February	September
	AA+AB	North Pacific Ocean	February	September
	AC	South Pacific Ocean (west)	August	February
	AD	South Pacific Ocean (east)	September	February
	AC+AD	South Pacific Ocean	September	February
Coral Reef Provinces	See Figure 30-4(b)	Caribbean Sea/Gulf of Mexico	February	September
		Coral Triangle and Southeast Asia	February	May
		Indian Ocean (east)	August	April
		Indian Ocean (west)	August	April
		Pacific Ocean (east)	December	August
		Pacific Ocean (west)	August	February
Basin Scale	B+Y	North Atlantic Ocean	March	August
	C+Z	South Atlantic Ocean	September	March
	B+Y+H+Z+C	Atlantic Ocean	December	August
	E+AB+D+AA	North Pacific Ocean	March	August
	AD+G+AC+F	South Pacific Ocean	August	February
	E+AB+I+AD+G+D+AA+AC+F	Pacific Ocean	December	August
	Q+X+A	Indian Ocean	August	April

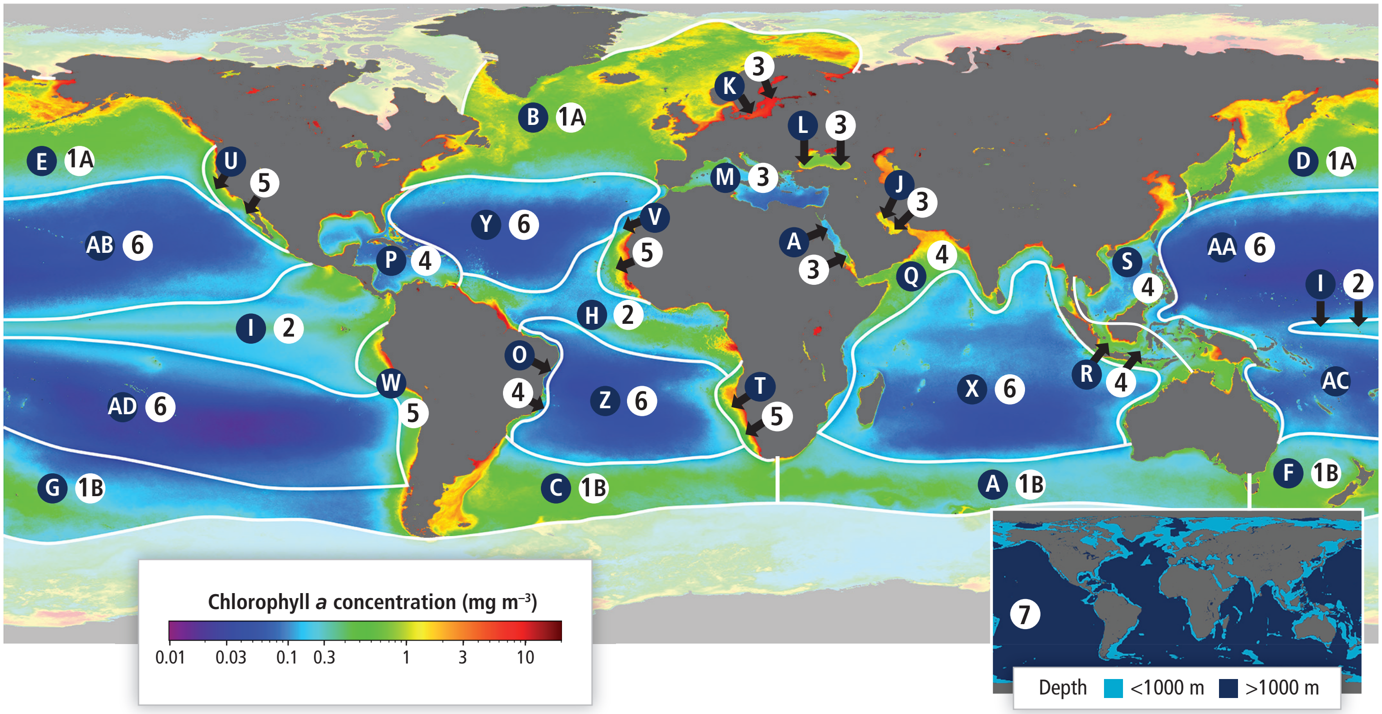


Figure SM30-1 | The seven major sub-regions of the Ocean used in Chapter 30 (numbered) and areas identified in Table SM30-2 (marked with letters). The chlorophyll-*a* concentration averaged over the period from SeaWiFS (Sep 1997–30 Nov 2010; NASA) is provides a proxy for differences in marine productivity (with the caveats provided in Box CC-PP). Key oceanographic features and primary production was the basis for separating the ocean into the sub-regions shown (Section 30.1.1, Table SM30-1). The map insert shows the distribution of Deep Sea habitat (1000 m; Bathypelagic and Abyssopelagic habitats combined).

SM30.3. Sea Surface Temperatures under RCP2.6, RCP4.5, RCP6.0, and RCP8.5

Projections of future SST changes were examined for sub-regions and areas within sub-regions (Figure SM30-1) using ensemble averages from AOGCM simulations available in the CMIP5 archive (Table SM30-3) for the four representative concentration pathways (RCP2.6, RCP4.5, RCP6.0, and RCP8.5; van Vuuren et al., 2011). Ensemble averages for each RCP are based on simulations from 10 to 16 individual models (Table SM30-3). Model hind-casts matched those observed for ocean sub-regions for the period 1980–2009 (HadISST1.1; Table SM30-2, Figure SM30-1), but with the AOGCM ensemble slightly overestimating the extent of change across the different ocean sub-regions (slope of observed/model = 0.81, $r^2 = 0.76$, $p<0.001$). In this way, the absolute change projected to occur in the ocean sub-regions was calculated for near-term (over 2010–2039) and long-term (over 2070–2099) periods (Table SM30-4). In the near term, projected changes in SST are largely indistinguishable between the different RCP pathways due to similarity in forcing until 2030s. Over the long-term, however, SSTs across the Ocean sub-regions were 1–3°C higher under RCP8.5 than those projected to occur under RCP2.6 (Table SM30-4). The implications of these projected changes on the structure and function of oceanic systems are extensively discussed in Chapters 6 and 30.

Table SM30-3 | CMIP5 model simulations used to create the Representative Concentration Pathway (RCP) 2.6, 4.5, 6.0, and 8.5 SST ensembles used in Chapter 30. The subset of CMIP5 models was chosen because each has historic runs enabling the derivation of the Maximum Monthly Mean (MMM) climatology over 1985 to 2000, ensuring that all anomalies were comparable across time periods and across RCPs. All models indicated were used to ensemble SSTs as well as to produce Degree Heating Month (DHM) measurements, with the exception of 2 model outputs, denoted by a * (which could not generate reliable MMM climatologies).

CMIP5 Model	RCP2.6	RCP4.5	RCP6.0	RCP8.5
ACCESS1-0		1		1
ACCESS1-3		1		1
BCC-CSM1-1	1	1	1	1
CanESM2	1	1		1
CMCC-CM		1		1
CMCC-CMS		1		1
CNRM-CM5	1	1		1
CSIRO-Mk3-6-0	1	1	1	1
EC-EARTH	*	1		1
FIO-ESM	1	1	1	1
GFDL-CM3	1	1	1	1
GFDL-ESM2G	1	1	1	1
GFDL-ESM2M	1	1		1
GISS-E2-R-p1	1	1	1	1
GISS-E2-R-p2	1	1	1	1
GISS-E2-R-p3	1	1	1	1
HadGEM2-AO	1	1	1	1
HadGEM2-CC		1		1
HadGEM2-ES	1	1	1	1
INMCM4		1		1
IPSL-CM5A-LR	1	1	*	1
IPSL-CM5A-MR	1	1		1
MPI-ESM-LR	1	1		1
MPI-ESM-MR	1	1		1
CESM1-BGC		1		1
CESM1-CAM5	1	1	1	1
NorESM1-M	1	1	1	1
NorESM1-ME	1	1	1	1
Number of models	20+1*	28	13+1*	28

Table SM30-4 | Projected changes in sea surface temperature (SST °C) over the next 90 years for ocean sub-regions (Figure SM30-1) from AOGCM model simulations from the Coupled Model Intercomparison Project Phase 5 (CMIP5, http://cmip-pcmdi.llnl.gov/cmip5/). Simulations were available for four Representative Concentration Pathways (RCPs): RCP2.6, RCP4.5, RCP6.0, and RCP8.5. The CMIP5 models used in this analysis are listed in Table SM30-3. For each ocean sub-region, a linear regression was fitted to all 1x1 degree monthly SST data extracted from the models for each of three periods; 2010–2039, 2040–2069, and 2070–2099. The average change in SST was calculated by multiplying the slope of each linear regression by 360 (months) to derive the average change over each successive 30-year period. The table is divided into two sections: "Near-term (2010–2039)" (the average change in SST over the next 30 years) and "Long-term (2010–2099)" (the total change over 2010–2099, which was calculated by adding the average change of the three 30-year periods from 2010 to 2099). This is a simplified method to account for slight non-linearity in SST change over the 90-year period.

Sub-region	Area	Near-term (2010–2039)				Long-term (2010–2099)				RCP8.5 minus RCP2.6
		RCP2.6	RCP4.5	RCP6.0	RCP8.5	RCP2.6	RCP4.5	RCP6.0	RCP8.5	
1. High Latitude Spring Bloom Systems (HLSBS)	Indian Ocean	0.13	0.29	0.18	0.41	–0.16	0.49	0.83	2.01	2.17
	North Atlantic Ocean	0.31	0.56	0.52	0.65	0.54	1.54	1.95	3.02	2.48
	South Atlantic Ocean	0.17	0.36	0.20	0.45	–0.09	0.67	0.88	2.26	2.36
	North Pacific Ocean (west)	0.79	0.96	0.91	1.17	1.46	2.47	3.07	4.84	3.38
	North Pacific Ocean (east)	0.79	0.81	0.93	1.06	1.31	2.17	2.96	4.39	3.08
	North Pacific Ocean	0.79	0.88	0.92	1.11	1.35	2.31	3.01	4.60	3.25
	South Pacific Ocean (west)	0.17	0.40	0.25	0.50	–0.16	0.63	0.85	2.37	2.53
	South Pacific Ocean (east)	0.12	0.23	0.13	0.35	–0.09	0.45	0.75	1.70	1.79
	South Pacific Ocean	0.14	0.28	0.17	0.40	–0.12	0.51	0.78	1.91	2.03
2. Equatorial Upwelling Systems (EUS)	Atlantic Equatorial Upwelling	0.43	0.58	0.49	0.81	0.46	1.19	1.61	3.03	2.56
	Pacific Equatorial Upwelling	0.35	0.55	0.54	0.77	0.43	1.22	1.75	3.01	2.57
3. Semi-Enclosed Seas (SES)	Arabian Gulf	0.82	0.97	0.89	1.20	1.30	2.39	2.96	4.26	2.96
	Baltic Sea	0.73	1.24	0.92	1.20	1.32	2.74	3.06	4.37	3.05
	Black Sea	0.74	1.01	0.86	1.24	1.37	2.61	3.16	4.19	2.82
	Mediterranean Sea	0.72	0.87	0.84	1.09	1.37	2.10	2.82	4.08	2.70
	Red Sea	0.56	0.72	0.71	0.93	0.88	1.65	2.39	3.45	2.57
4. Coastal Boundary Systems (CBS)	Atlantic Ocean (west)	0.34	0.40	0.45	0.62	0.23	0.81	1.33	2.44	2.21
	Caribbean Sea/Gulf of Mexico	0.50	0.67	0.64	0.85	0.74	1.53	1.97	3.23	2.49
	Indian Ocean (west)	0.46	0.59	0.56	0.85	0.63	1.39	1.95	3.32	2.69
	Indian Ocean (east)	0.34	0.57	0.46	0.69	0.38	1.22	1.59	2.80	2.42
	Indian Ocean (east), Southeast Asia, Pacific Ocean (west)	0.48	0.66	0.57	0.82	0.66	1.47	1.89	3.12	2.46
5. Eastern Boundary Upwelling Ecosystems (EBUE)	Benguela Current	0.30	0.43	0.45	0.71	0.07	0.70	1.41	2.52	2.45
	California Current	0.62	0.71	0.84	0.93	1.02	1.86	2.46	3.51	2.49
	Canary Current	0.55	0.62	0.58	0.82	0.97	1.30	1.83	3.18	2.21
	Humboldt Current	0.22	0.43	0.34	0.60	0.11	0.91	1.22	2.58	2.47
6. Sub-Tropical Gyres (STG)	Indian Ocean	0.30	0.44	0.37	0.63	0.19	0.89	1.35	2.62	2.43
	North Atlantic Ocean	0.49	0.66	0.60	0.85	0.87	1.62	1.98	3.30	2.43
	South Atlantic Ocean	0.25	0.33	0.33	0.55	0.03	0.58	1.03	2.20	2.18
	North Pacific Ocean (west)	0.54	0.70	0.64	0.90	0.84	1.62	2.08	3.39	2.55
	North Pacific Ocean (east)	0.56	0.66	0.71	0.91	0.90	1.56	1.50	3.44	2.54
	North Pacific Ocean	0.55	0.68	0.68	0.90	0.87	1.58	2.09	3.42	2.55
	South Pacific Ocean (west)	0.31	0.44	0.34	0.62	0.12	0.88	1.19	2.56	2.44
	South Pacific Ocean (east)	0.17	0.27	0.21	0.45	–0.03	0.52	0.89	1.90	1.93
	South Pacific Ocean	0.20	0.31	0.24	0.49	0.00	0.60	0.96	2.05	2.05
Coral Reef Provinces; see Figure 30-4b	Caribbean Sea/Gulf of Mexico	0.48	0.64	0.61	0.83	0.68	1.43	1.87	3.14	2.46
	Coral Triangle and Southeast Asia	0.42	0.61	0.52	0.76	0.58	1.35	1.75	2.95	2.37
	Indian Ocean (east)	0.32	0.56	0.46	0.67	0.37	1.18	1.59	2.76	2.40
	Indian Ocean (west)	0.39	0.51	0.50	0.77	0.43	1.18	1.71	2.97	2.54
	Pacific Ocean (east)	0.46	0.64	0.64	0.83	0.63	1.44	1.99	3.23	2.60
	Pacific Ocean (west)	0.35	0.48	0.40	0.68	0.30	1.02	1.39	2.66	2.35

Continued next page →

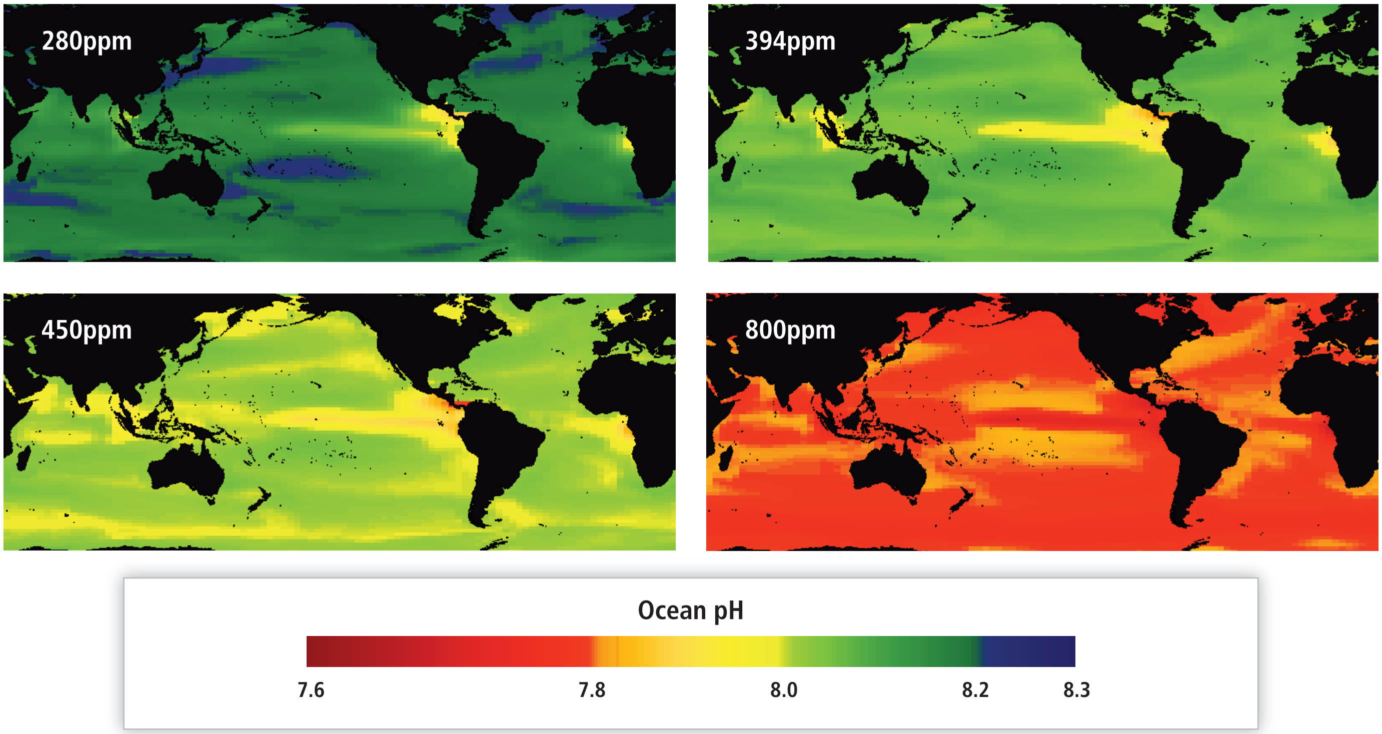
Table SM30-4 (continued)

Sub-region	Area	Near-term (2010–2039)				Long-term (2010–2099)				RCP8.5 minus RCP2.6
		RCP2.6	RCP4.5	RCP6.0	RCP8.5	RCP2.6	RCP4.5	RCP6.0	RCP8.5	
Basin Scale	North Atlantic Ocean	0.37	0.60	0.55	0.72	0.66	1.57	1.96	3.12	2.46
	South Atlantic Ocean	0.21	0.35	0.27	0.51	–0.03	0.62	0.76	2.23	2.26
	Atlantic Ocean	0.32	0.50	0.44	0.65	0.38	1.17	1.54	2.78	2.40
	North Pacific Ocean	0.64	0.75	0.77	0.98	1.06	1.85	2.43	3.86	2.80
	South Pacific Ocean	0.18	0.30	0.21	0.45	–0.04	0.56	0.89	2.00	2.04
	Pacific Ocean	0.41	0.54	0.51	0.73	0.52	1.23	1.70	2.97	2.45
	Indian Ocean	0.30	0.44	0.37	0.63	0.19	0.89	1.35	2.62	2.43

SM30.4. Changes to Surface pH and Aragonite Saturation State under Different Concentrations of Atmospheric CO_2

The relative changes in pH and the aragonite saturation state of seawater varies in concert with increases in the partial pressure of CO_2 above the ocean. Observations of ocean chemistry (Doney et al., 2009; Feely et al., 2009) are highly consistent with models of the carbonate chemistry of the upper ocean (Caldeira and Wickett, 2003). Notably, high latitude areas, as well as regions where upwelling is dominant, show naturally lower pH and aragonite saturation states. These regions are expected to reach critical levels in terms of pH and aragonite saturation sooner than lower latitudes and non-upwelling regions (Section 30.3.2.2).

(a) Ocean pH as a function of atmospheric CO_2 concentration



(b) Aragonite saturation state

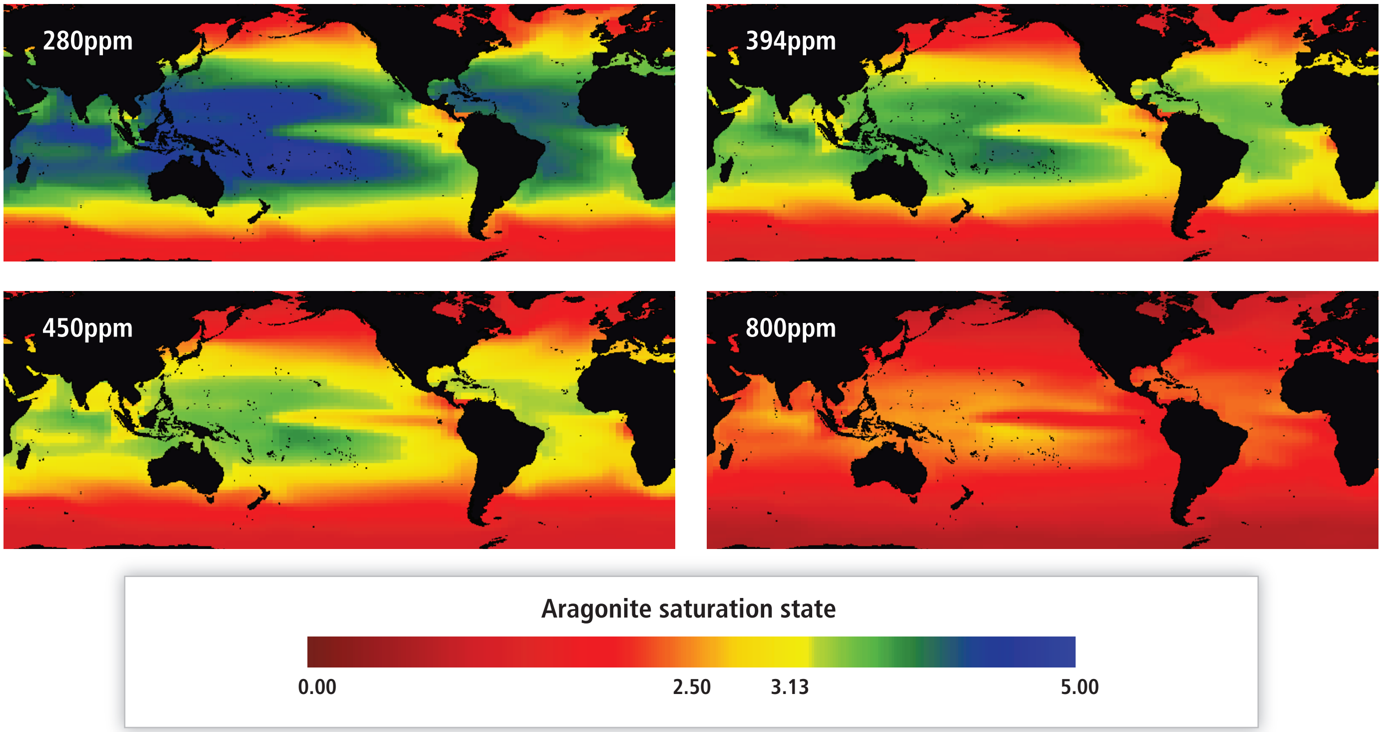


Figure SM30-2 | The carbonate chemistry of the Ocean under current different atmospheric concentration s of CO_2. 280ppm represents pre-industrial and 394ppm present-day levels (WGI Annex II). (a) Surface pH and (b) Aragonite saturation state of the Ocean simulated by the University of Victoria Earth System Model. The fields of pH and aragonite saturation state are calculated from the model output of dissolved inorganic carbon concentration, alkalinity concentration, temperature, and salinity, together with the chemistry routine from the OCMIP-3 project (http://www.ipsl.jussieu.fr/OCMIP/phase3).

SM30.5. Projections of Changes to Sea Surface Temperatures (RCP2.6 and RCP8.5) for Different Regions that have Coral Reefs

Warm-water coral reefs throughout the world (but particularly in CBS, SES, and STG; Figure SM30-1) are rapidly declining as result of local perturbations (i.e., coastal pollution, overexploitation) and climate change (*high confidence*; Sections 30.5.3-4, 30.5.6). Reef-building corals, which are responsible for building the carbonate framework of coral reefs, are sensitive to both elevated sea temperatures as well as reduced pH and carbonate concentrations (*high confidence*; Section 6.3.2; Boxes CC-CR, CC-OA). Continued increases in sea temperature will increase the incidence of impacts such as mass coral bleaching and mortality (*virtually certain*), with the CMIP5 ensemble projecting the irreversible degradation of coral reefs from most sites globally by 2050 (*very likely*; Section 30.5; Figure 30-10; Box CC-CR). Investigating past, present, and future sea temperatures in six major coral reef areas (Figure 30-4b) reveals that future sea temperatures will exceed established thresholds of coral bleaching and mortality around the middle to late part of this century (Figure SM30-3).

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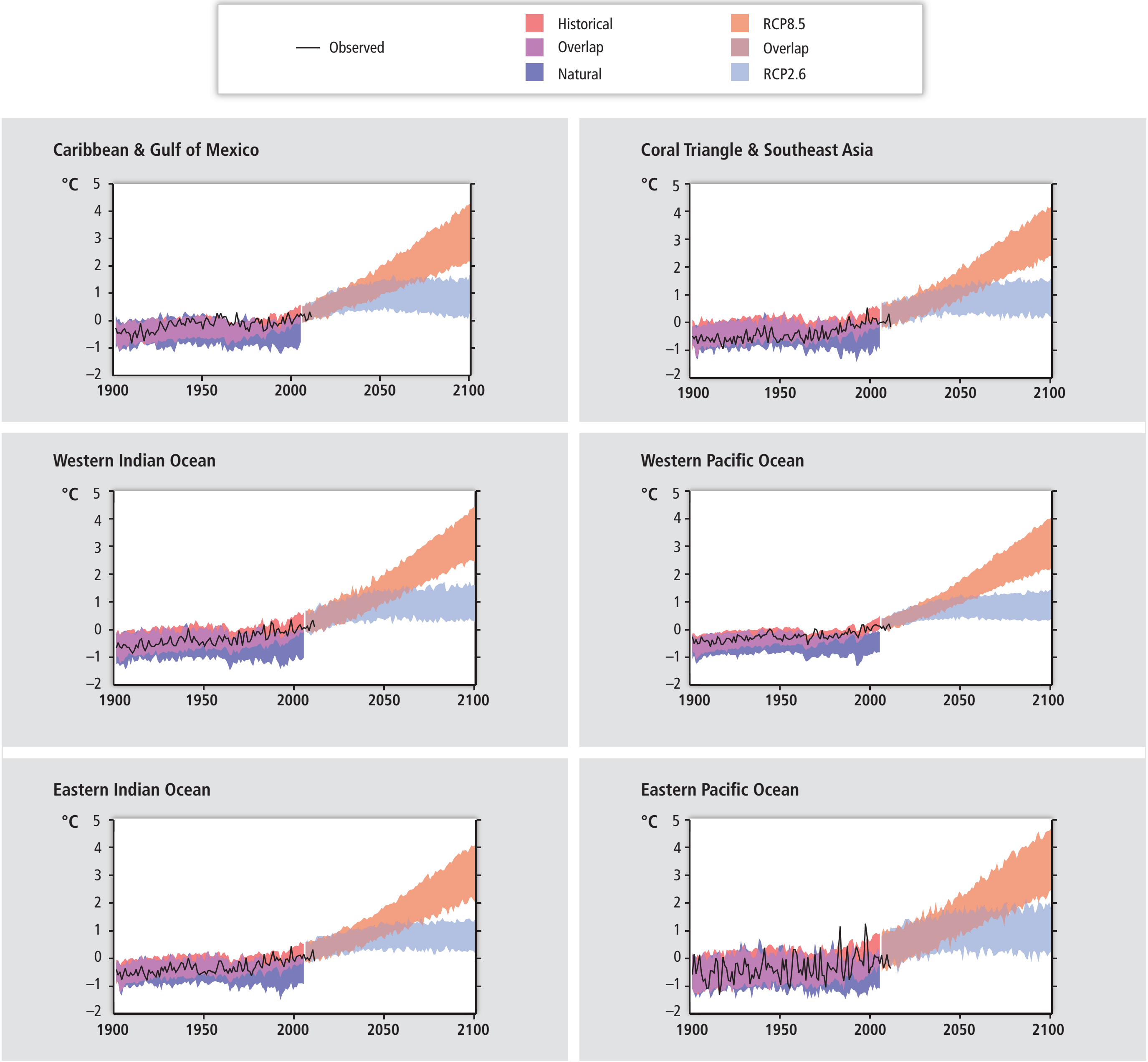

Figure SM30-3 | Past and future sea surface temperatures (SST) in six major coral reef provinces and locations (Figure 30-4b) under historic, un-forced (natural), RCP2.6 and RCP8.5 scenarios from CMIP5 ensembles (Table SM30-3). Observed and simulated variations in past and projected future annual average SST over various sites where coral reefs are prominent ecosystems (locations shown in Figure 30-4b). The black line shows estimates from HadISST1.1 [Rayner et al., 2003] reconstructed historical SST dataset. Shading denotes the 5–95 percentile range of climate model simulations driven with 'historical' changes in anthropogenic and natural drivers (62 simulations), historical changes in 'natural' drivers only (25), the RCP4.5 emissions scenario (62), and the RCP8.5 (62). Data are anomalies from the 1986–2006 average of the HadISST1.1 data (for the HadISST1.q time series) or of the corresponding historical all- forcing simulations. Further details are given in Box 21.1, 21.3.

CR

Coral Reefs

Jean-Pierre Gattuso (France), Ove Hoegh-Guldberg (Australia), Hans-Otto Pörtner (Germany)

Coral reefs are shallow-water ecosystems that consist of reefs made of calcium carbonate which is mostly secreted by reef-building corals and encrusting macroalgae. They occupy less than 0.1% of the ocean floor yet play multiple important roles throughout the tropics, housing high levels of biological diversity as well as providing key ecosystem goods and services such as habitat for fisheries, coastal protection, and appealing environments for tourism (Wild et al., 2011). About 275 million people live within 30 km of a coral reef (Burke et al., 2011) and derive some benefits from the ecosystem services that coral reefs provide (Hoegh-Guldberg, 2011), including provisioning (food, livelihoods, construction material, medicine), regulating (shoreline protection, water quality), supporting (primary production, nutrient cycling), and cultural (religion, tourism) services. This is especially true for the many coastal and small island nations in the world's tropical regions (Section 29.3.3.1).

Coral reefs are one of the most vulnerable marine ecosystems (*high confidence*; Sections 5.4.2.4, 6.3.1, 6.3.2, 6.3.5, 25.6.2, and 30.5), and more than half of the world's reefs are under medium or high risk of degradation (Burke et al., 2011). Most human-induced disturbances to coral reefs were local until the early 1980s (e.g., unsustainable coastal development, pollution, nutrient enrichment, and overfishing) when disturbances from ocean warming (principally mass coral bleaching and mortality) began to become widespread (Glynn, 1984). Concern about the impact of ocean acidification on coral reefs developed over the same period, primarily over the implications of ocean acidification for the building and maintenance of the calcium carbonate reef framework (Box CC-OA).

A wide range of climatic and non-climatic drivers affect corals and coral reefs and negative impacts have already been observed (Sections 5.4.2.4, 6.3.1, 6.3.2, 25.6.2.1, 30.5.3, 30.5.6). Bleaching involves the breakdown and loss of endosymbiotic algae, which live in the coral tissues and play a key role in supplying the coral host with energy (see Section 6.3.1. for physiological details and Section 30.5 for a regional analysis). Mass coral bleaching and mortality, triggered by positive temperature anomalies (*high confidence*), is the most widespread and conspicuous impact of climate change (Figure CR-1A and B, Figure 5-3; Sections 5.4.2.4, 6.3.1, 6.3.5, 25.6.2.1, 30.5, and 30.8.2). For example, the level of thermal stress at most of the 47 reef sites where bleaching occurred during 1997–1998 was unmatched in the period 1903–1999 (Lough, 2000). Ocean acidification reduces biodiversity (Figure CR-1C and D) and the calcification rate of corals (*high confidence*; Sections 5.4.2.4, 6.3.2, 6.3.5) while at the same time increasing the rate of dissolution of the reef framework (*medium confidence*; Section 5.2.2.4) through stimulation of biological erosion and chemical dissolution. Taken together, these changes will tip the calcium carbonate balance of coral reefs toward net dissolution (*medium confidence*; Section 5.4.2.4).

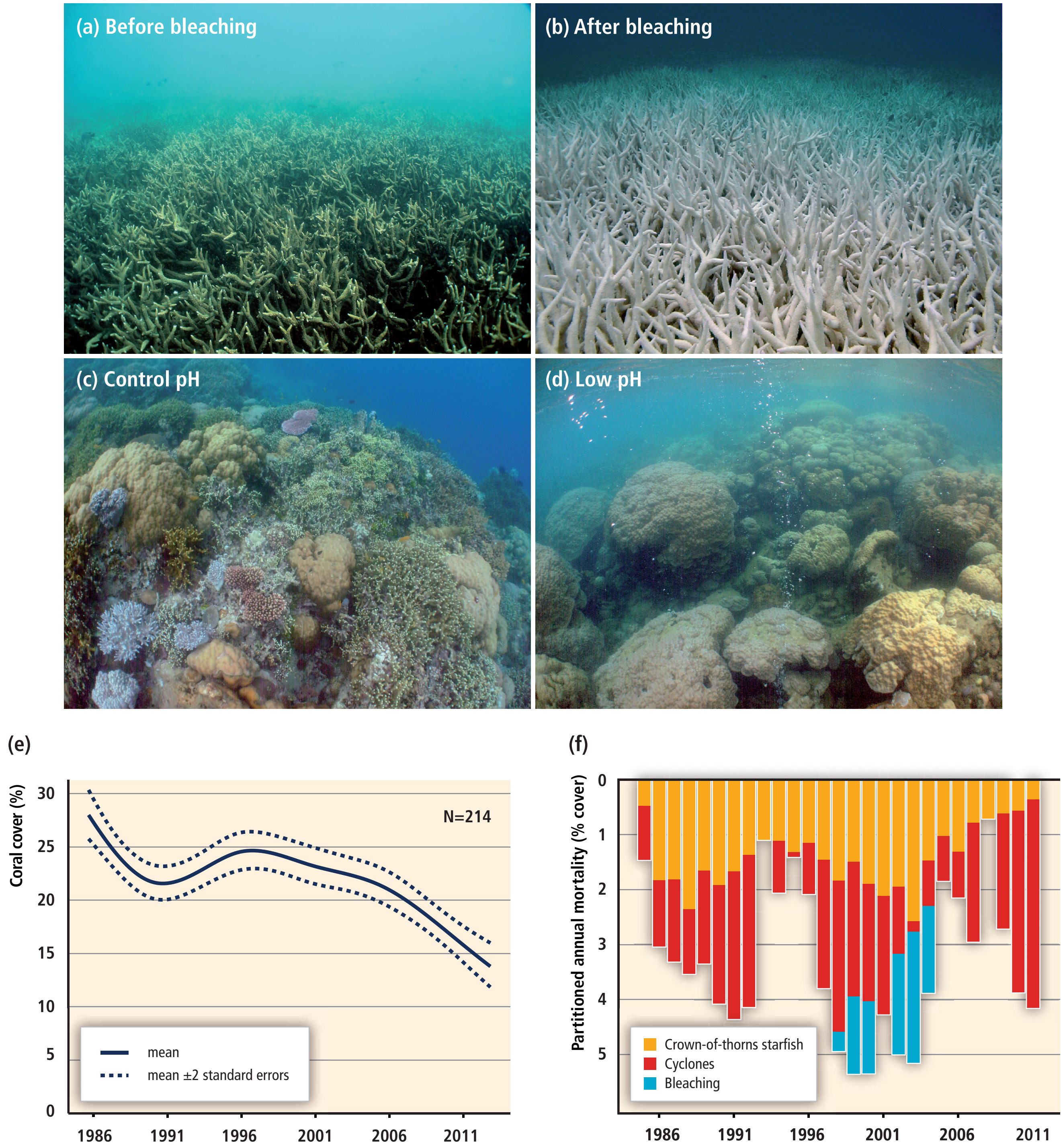

Figure CR-1 | (a, b) The same coral community before and after a bleaching event in February 2002 at 5 m depth, Halfway Island, Great Barrier Reef. Approximately 95% of the coral community was severely bleached in 2002 (Elvidge et al., 2004). Corals experience increasing mortality as the intensity of a heating event increases. A few coral species show the ability to shuffle symbiotic communities of dinoflagellates and appear to be more tolerant of warmer conditions (Berkelmans and van Oppen, 2006; Jones et al., 2008). (c, d) Three CO_2 seeps in Milne Bay Province, Papua New Guinea show that prolonged exposure to high CO_2 is related to fundamental changes in the ecology of coral reefs (Fabricius et al., 2011), including reduced coral diversity (–39%), severely reduced structural complexity (–67%), lower density of young corals (–66%), and fewer crustose coralline algae (–85%). At high CO_2 sites (d; median pH_T ~7.8, where pH_T is pH on the total scale), reefs are dominated by massive corals while corals with high morphological complexity are underrepresented compared with control sites (c; median pH_T ~8.0). Reef development ceases at pH_T values below 7.7. (e) Temporal trend in coral cover for the whole Great Barrier Reef over the period 1985–2012 (N=number of reefs, De'ath et al., 2012). (f) Composite bars indicate the estimated mean coral mortality for each year, and the sub-bars indicate the relative mortality due to crown-of-thorns starfish, cyclones, and bleaching for the whole Great Barrier Reef (De'ath et al., 2012). (Photo credit: R. Berkelmans (a and b) and K. Fabricius (c and d).)

Ocean warming and acidification have synergistic effects in several reef-builders (Section 5.2.4.2, 6.3.5). Taken together, these changes will erode habitats for reef-based fisheries, increase the exposure of coastlines to waves and storms, as well as degrading environmental features important to industries such as tourism (*high confidence*; Section 6.4.1.3, 25.6.2, 30.5).

A growing number of studies have reported regional scale changes in coral calcification and mortality that are consistent with the scale and impact of ocean warming and acidification when compared to local factors such as declining water quality and overfishing (Hoegh-Guldberg et al., 2007). The abundance of reef building corals is in rapid decline in many Pacific and Southeast Asian regions (*very high confidence*, 1 to 2% per year for 1968–2004; Bruno and Selig, 2007). Similarly, the abundance of reef-building corals has decreased by more than 80% on many Caribbean reefs (1977–2001; Gardner et al., 2003), with a dramatic phase shift from corals to seaweeds occurring on Jamaican reefs (Hughes, 1994). Tropical cyclones, coral predators, and thermal stress-related coral bleaching and mortality have led to a decline in coral cover on the Great Barrier Reef by about 51% between 1985 and 2012 (Figure CR-1E and F). Although less well documented, benthic invertebrates other than corals are also at risk (Przeslawski et al., 2008). Fish biodiversity is threatened by the permanent degradation of coral reefs, including in a marine reserve (Jones et al., 2004).

Future impacts of climate-related drivers (ocean warming, acidification, sea level rise as well as more intense tropical cyclones and rainfall events) will exacerbate the impacts of non-climate–related drivers (*high confidence*). Even under optimistic assumptions regarding corals being able to rapidly adapt to thermal stress, one-third (9 to 60%, 68% uncertainty range) of the world's coral reefs are projected to be subject to long-term degradation (next few decades) under the Representative Concentration Pathway (RCP)3-PD scenario (Frieler et al., 2013). Under the RCP4.5 scenario, this fraction increases to two-thirds (30 to 88%, 68% uncertainty range). If present-day corals have residual capacity to acclimate and/or adapt, half of the coral reefs may avoid high-frequency bleaching through 2100 (*limited evidence*, *limited agreement*; Logan et al., 2014). Evidence of corals adapting rapidly, however, to climate change is missing or equivocal (Hoegh-Guldberg, 2012).

Damage to coral reefs has implications for several key regional services:

- *Resources*: Coral reefs account for 10 to 12% of the fish caught in tropical countries, and 20 to 25% of the fish caught by developing nations (Garcia and de Leiva Moreno, 2003). More than half (55%) of the 49 island countries considered by Newton et al. (2007) are already exploiting their coral reef fisheries in an unsustainable way and the production of coral reef fish in the Pacific is projected to decrease 20% by 2050 under the Special Report on Emission Scenarios (SRES) A2 emissions scenario (Bell et al., 2013).
- *Coastal protection*: Coral reefs contribute to protecting the shoreline from the destructive action of storm surges and cyclones (Sheppard et al., 2005), sheltering the only habitable land for several island nations, habitats suitable for the establishment and maintenance of mangroves and wetlands, as well as areas for recreational activities. This role is threatened by future sea level rise, the decrease in coral cover, reduced rates of calcification, and higher rates of dissolution and bioerosion due to ocean warming and acidification (Sections 5.4.2.4, 6.4.1, 30.5).
- *Tourism*: More than 100 countries benefit from the recreational value provided by their coral reefs (Burke et al., 2011). For example, the Great Barrier Reef Marine Park attracts about 1.9 million visits each year and generates A$5.4 billion to the Australian economy and 54,000 jobs (90% in the tourism sector; Biggs, 2011).

Coral reefs make a modest contribution to the global gross domestic product (GDP) but their economic importance can be high at the country and regional scales (Pratchett et al., 2008). For example, tourism and fisheries represent 5% of the GDP of South Pacific islands (average for 2001–2011; Laurans et al., 2013). At the local scale, these two services provided in 2009–2011 at least 25% of the annual income of villages in Vanuatu and Fiji (Pascal, 2011; Laurans et al., 2013).

Isolated reefs can recover from major disturbance, and the benefits of their isolation from chronic anthropogenic pressures can outweigh the costs of limited connectivity (Gilmour et al., 2013). Marine protected areas (MPAs) and fisheries management have the potential to increase ecosystem resilience and increase the recovery of coral reefs after climate change impacts such as mass coral bleaching (McLeod et al., 2009). Although they are key conservation and management tools, they are unable to protect corals directly from thermal stress (Selig et al., 2012), suggesting that they need to be complemented with additional and alternative strategies (Rau et al., 2012; Billé et al., 2013). While MPA networks are a critical management tool, they should be established considering other forms of resource management (e.g., fishery catch limits and gear restrictions) and integrated ocean and coastal management to control land-based threats such as pollution and sedimentation. There is *medium confidence* that networks of highly protected areas nested within a broader management framework can contribute to preserving coral reefs under increasing human pressure at local and global scales (Salm et al., 2006). Locally, controlling the input of nutrients and sediment from land is an important complementary management strategy (Mcleod et al., 2009) because nutrient enrichment can increase the susceptibility of corals to bleaching (Wiedenmann et al., 2013) and coastal pollutants enriched with fertilizers can increase acidification (Kelly et al., 2011). In the long term, limiting the amount of ocean warming and acidification is central to ensuring the viability of coral reefs and dependent communities (*high confidence*; Section 5.2.4.4, 30.5).

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HS

Heat Stress and Heat Waves

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According to WGI, it is *very likely* that the number and intensity of hot days have increased markedly in the last three decades and *virtually certain* that this increase will continue into the late 21st century. In addition, it is *likely* (*medium confidence*) that the occurrence of heat waves (multiple days of hot weather in a row) has more than doubled in some locations, but *very likely* that there will be more frequent heat waves over most land areas after mid-century. Under a medium warming scenario, Coumou et al. (2013) predicted that the number of monthly heat records will be more than 12 times more common by the 2040s compared to a non-warming world. In a longer time perspective, if the global mean temperature increases to +7°C or more, the habitability of parts of the tropics and mid-latitudes will be at risk (Sherwood and Huber, 2010). Heat waves affect natural and human systems directly, often with severe losses of lives and assets as a result, and may act as triggers of tipping points (Hughes et al., 2013). Consequently, heat stress plays an important role in several key risks noted in Chapter 19 and CC-KR.

Economy and Society (Chapters 10, 11, 12, 13)

Environmental heat stress has already reduced the global labor capacity to 90% in peak months with a further predicted reduction to 80% in peak months by 2050. Under a high warming scenario (RCP8.5), labor capacity is expected to be less than 40% of present-day conditions in peak months by 2200 (Dunne et al., 2013). Adaptation costs for securing cooling capacities and emergency shelters during heat waves will be substantial.

Heat waves are associated with social predicaments such as increasing violence (Anderson, 2012) as well as overall health and psychological distress and low life satisfaction (Tawatsupa et al., 2012). Impacts are highly differential with disproportional burdens on poor people, elderly people, and those who are marginalized (Wilhelmi et al., 2012). Urban areas are expected to suffer more due to the combined effect of climate and the urban heat island effect (Fischer et al., 2012; see also Section 8.2.3.1). In low- and medium-income countries, adaptation to heat stress is severely restricted for most people in poverty and particularly those who are dependent on working outdoors in agriculture, fisheries, and construction. In small-scale agriculture, women and children are particularly at risk due to the gendered division of labor (Croppenstedt et al., 2013). The expected increase in wildfires as a result of heat waves (Pechony and Shindell, 2010) is a concern for human security, health, and ecosystems. Air pollution from wildfires already causes an estimated 339,000 premature deaths per year worldwide (Johnston et al., 2012).

Human Health (Chapter 11)

Morbidity and mortality due to heat stress is now common all over the world (Barriopedro et al., 2011; Nitschke et al., 2011; Rahmstorf and Coumou, 2011; Diboulo et al., 2012; Hansen et al., 2012). Elderly people and people with circulatory and respiratory diseases are also vulnerable even in developed countries; they can become victims even inside their own houses (Honda et al., 2011). People in physical work are at particular risk as such work produces substantial heat within the body, which cannot be released if the outside temperature and humidity is above certain limits (Kjellstrom et al., 2009). The risk of non-melanoma skin cancer from exposure to UV radiation during summer months increases with temperature (van der Leun, et al., 2008). High temperatures are also associated with an increase in air-borne allergens acting as triggers for respiratory illnesses such as asthma, allergic rhinitis, conjunctivitis, and dermatitis (Beggs, 2010).

Ecosystems (Chapters 4, 5, 6, 30)

Tree mortality is increasing globally (Williams et al., 2013) and can be linked to climate impacts, especially heat and drought (Reichstein et al., 2013), even though attribution to climate change is difficult owing to lack of time series and confounding factors. In the Mediterranean region, higher fire risk, longer fire season, and more frequent large, severe fires are expected as a result of increasing heat waves in combination with drought (Duguy et al., 2013; see also Box 4.2).

Marine ecosystem shifts attributed to climate change are often caused by temperature extremes rather than changes in the average (Pörtner and Knust, 2007). During heat exposure near biogeographical limits, even small (<0.5°C) shifts in temperature extremes can have large effects, often exacerbated by concomitant exposures to hypoxia and/or elevated CO_2 levels and associated acidification (*medium confidence*; Hoegh-Guldberg et al., 2007; see also Figure 6-5; Sections 6.3.1, 6.3.5, 30.4, 30.5; CC-MB).

Most coral reefs have experienced heat stress sufficient to cause frequent mass coral bleaching events in the last 30 years, sometimes followed by mass mortality (Baker et al., 2008). The interaction of acidification and warming exacerbates coral bleaching and mortality (*very high confidence*).Temperate seagrass and kelp ecosystems will decline with the increased frequency of heat waves and through the impact of invasive subtropical species (*high confidence*; Sections 5, 6, 30.4, 30.5, CC-CR, CC-MB).

Agriculture (Chapter 7)

Excessive heat interacts with key physiological processes in crops. Negative yield impacts for all crops past +3°C of local warming without adaptation, even with benefits of higher CO_2 and rainfall, are expected even in cool environments (Teixeira et al., 2013). For tropical systems where moisture availability or extreme heat limits the length of the growing season, there is a high potential for a decline in the length of the growing season and suitability for crops (*medium evidence*, *medium agreement*; Jones and Thornton, 2009). For example, half of the wheat-growing area of the Indo-Gangetic Plains could become significantly heat-stressed by the 2050s.

There is *high confidence* that high temperatures reduce animal feeding and growth rates (Thornton et al., 2009). Heat stress reduces reproductive rates of livestock (Hansen, 2009), weakens their overall performance (Henry et al., 2012), and may cause mass mortality of animals in feedlots during heat waves (Polley et al., 2013). In the USA, current economic losses due to heat stress of livestock are estimated at several billion US$ annually (St-Pierre et al., 2003).

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MB

Observed Global Responses of Marine Biogeography, Abundance, and Phenology to Climate Change

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IPCC WGII AR4 presented the detection of a global fingerprint on natural systems and its attribution to climate change (AR4, Chapter 1, SPM Figure 1), but studies from marine systems were mostly absent. Since AR4, there has been a rapid increase in studies that focus on climate change impacts on marine species, which represents an opportunity to move from more anecdotal evidence to examining and potentially attributing detected biological changes within the ocean to climate change (Section 6.3; Figure MB-1). Recent changes in populations of marine species and the associated shifts in diversity patterns are resulting, at least partly, from climate change–mediated biological responses across ocean regions (*robust evidence, high agreement, high confidence*; Sections 6.2, 30.5; Table 6-7).

Poloczanska et al. (2013) assess a potential pattern in responses of ocean life to recent climate change using a global database of 208 peer-reviewed papers. Observed responses (n = 1735) were recorded from 857 species or assemblages across regions and taxonomic groups, from phytoplankton to marine reptiles and mammals (Figure MB-1). Observations were defined as those where the authors of a particular paper assessed the change in a biological parameter (including distribution, phenology, abundance, demography, or community composition) and, if change occurred, the consistency of the change with that expected under climate change. Studies from the peer-reviewed literature were selected using three criteria: (1) authors inferred or directly tested for trends in biological and climatic variables; (2) authors included data after 1990; and (3) observations spanned at least 19 years, to reduce bias resulting from biological responses to short-term climate variability.

The results of this meta-analysis show that climate change has already had widespread impacts on species' distribution, abundance, phenology, and subsequently, species richness and community composition across a broad range of taxonomic groups (plankton to top predators). Of the observations that showed a response in either direction, changes in phenology, distribution and abundance were overwhelmingly (81%) in a direction that was consistent with theoretical responses to climate change (Section 6.2). Knowledge gaps exist, especially in equatorial sub-regions and the Southern Hemisphere (Figure MB-1).

The timing of many biological events (phenology) had an earlier onset. For example, over the last 50 years, spring events shifted earlier for many species with an average advancement of 4.4 ± 0.7 days per decade (mean ± SE) and summer events by 4.4 ± 1.1 days per decade (*robust evidence, high agreement, high confidence*) (Figure MB-2). Phenological observations included in the study range from shifts in peak abundance of phytoplankton and zooplankton, to reproduction and migration of invertebrates, fishes, and seabirds (Sections 6.3.2, 30.5).

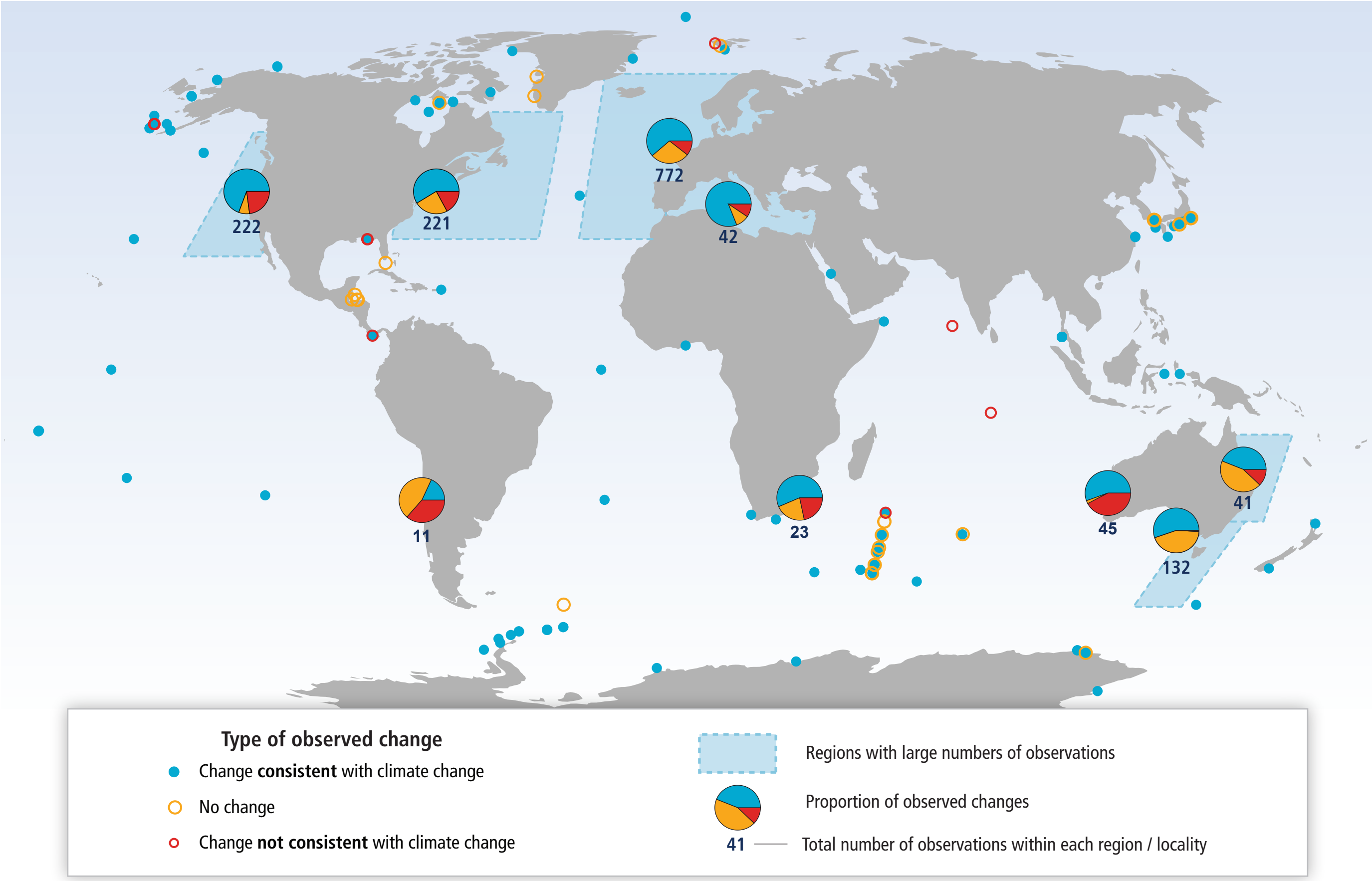

Figure MB-1 | 1735 observed responses to climate change from 208 single- and multi-species studies. Data shown include changes that are attributed (at least partly) to climate change (blue), changes that are inconsistent with climate change (red), and no change (orange). Each circle represents the center of a study area. Where points fall on land, it is because they are centroids of distributions that surround an island or peninsula. Studies encompass areas from single sites (e.g., seabird breeding colony) to large ocean regions (e.g., continuous plankton recorder surveys in north-east Atlantic). For regions (indicated by blue shading) and localities with large numbers of observations, pie charts summarize the relative proportions of the three types of observed changes (consistent with climate change, inconsistent with climate change, and no change) in those regions or localities. The numbers indicate the total observations within each region or locality. Note: 57% of the studies included were published since AR4. (From Poloczanska et al., 2013).

The distributions of benthic, pelagic, and demersal species and communities have shifted by up to a thousand kilometers, although the range shifts have not been uniform across taxonomic groups or ocean regions (Sections 6.3.2, 30.5) (*robust evidence, high agreement, high confidence*). Overall, leading range edges expanded in a poleward direction at 72.0 ± 13.5 km per decade and trailing edges contracted in a poleward direction at 15.8 ± 8.7 km per decade (Figure MB-2), revealing much higher current rates of migration than the potential maximum rates reported for terrestrial species (Figure 4-6) despite slower warming of the ocean than land surface (WGI Section 3.2).

Poleward distribution shifts have resulted in increased species richness in mid- to high-latitude regions (Hiddink and ter Hofstede, 2008) and changing community structure (Simpson et al., 2011; see also Section 28.2.2). Increases in warm-water components of communities concurrent with regional warming have been observed in mid- to high-latitude ocean regions including the Bering Sea, Barents Sea, Nordic Sea, North Sea, and Tasman Sea (Box 6.1; Section 30.5). Observed changes in species composition of catches from 1970–2006 that are partly attributed to long-term ocean warming suggest increasing dominance of warmer water species in subtropical and higher latitude regions, and reduction in abundance of subtropical species in equatorial waters (Cheung et al., 2013), with implications for fisheries (Sections 6.5, 7.4.2, 30.6.2.1).

The magnitude and direction of distribution shifts can be related to temperature velocities (i.e., the speed and direction at which isotherms propagate across the ocean's surface (Section 30.3.1.1; Burrows et al., 2011). Pinsky et al. (2013) showed that shifts in both latitude and depth of benthic fish and crustaceans could be explained by climate velocity with remarkable accuracy, using a database of 128 million individuals across 360 marine taxa from surveys of North American coastal waters conducted over 1968–2011. Poloczanska et al. (2013) found that faster distribution shifts generally occur in regions of highest surface temperature velocity, such as the North Sea and sub-Arctic Pacific Ocean. Observed marine species shifts, since approximately the 1950s, have generally been able to track observed velocities (Figure MB-3), with phyto- and zooplankton distribution shifts vastly exceeding climate velocities observed over most of the ocean surface, but with considerable variability within and among taxonomic groups (Poloczanska et al., 2013).

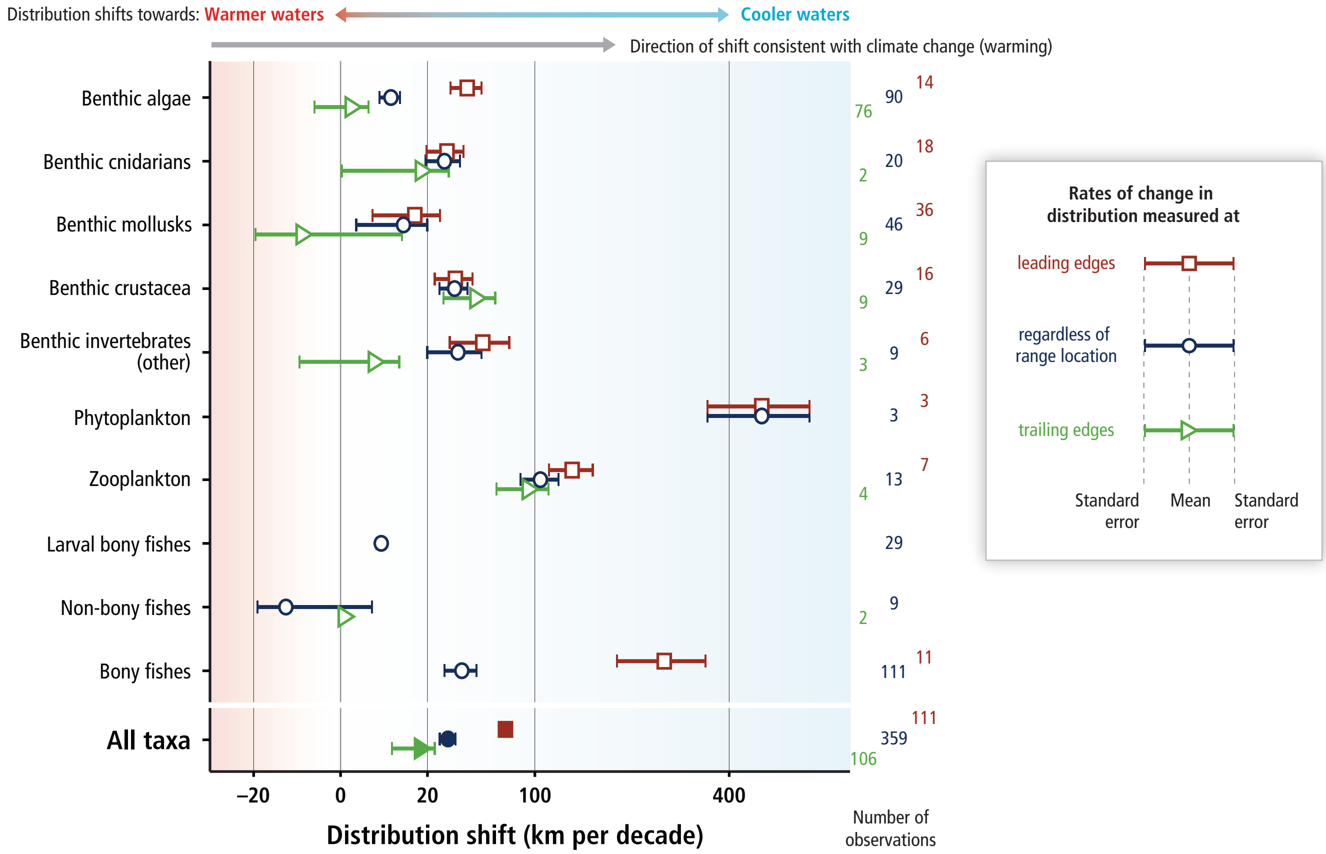

Figure MB-2 | Rates of change in distribution (kilometers per decade) for marine taxonomic groups, measured at the leading edges (red) and trailing edges (green). Average distribution shifts were calculated using all data, regardless of range location, and are in dark blue. Distribution shifts have been square-root transformed; standard errors may be asymmetric as a result. Positive distribution changes are consistent with warming (into previously cooler waters, generally poleward). Means ± standard error are shown, along with number of observations. Non-bony fishes include sharks, rays, lampreys, and hagfish. (From Poloczanska et al., 2013).

Biogeographic shifts are also influenced by other factors such as currents, nutrient and stratification changes, light levels, sea ice, species' interactions, habitat availability and fishing, some of which can be independently influenced by climate change (Section 6.3). Rate and pattern of biogeographic shifts in sedentary organisms and benthic macroalgae are complicated by the influence of local dynamics and topographic features (islands, channels, coastal lagoons, e.g., of the Mediterranean (Bianchi, 2007), coastal upwelling e.g., (Lima et al., 2007)). Geographical barriers constrain range shifts and may cause a loss of endemic species (Ben Rais Lasram et al., 2010), with associated niches filled by alien species, either naturally migrating or artificially introduced (Philippart et al., 2011).

Whether marine species can continue to keep pace as rates of warming, hence climate velocities, increase (Figure MB-3b) is a key uncertainty. Climate velocities on land are expected to outpace the ability of many terrestrial species to track climate velocities this century (Section 4.3.2.5; Figure 4-6). For marine species, the observed rates of shift are generally much faster than those for land species, particularly for primary producers and lower trophic levels (Poloczanska et al., 2013). Phyto- and zooplankton communities (excluding larval fish) have extended distributions at remarkable rates (Figure MB-3b), such as in the Northeast Atlantic (Section 30.5.1) with implications for marine food webs.

Geographical range shifts and depth distribution vary between coexisting marine species (Genner et al., 2004; Perry et al., 2005; Simpson et al., 2011) as a consequence of the width of species-specific thermal windows and associated vulnerabilities (Figure 6-5). Warming therefore causes differential changes in growth, reproductive success, larval output, early juvenile survival, and recruitment, implying shifts in the relative performance of animal species and, thus, their competitiveness (Pörtner and Farrell, 2008; Figure 6-7A). Such effects may underlie abundance losses or local extinctions, "regime shifts" between coexisting species, or critical mismatches between predator and prey organisms, resulting in changes in local and regional species richness, abundance, community composition, productivity, energy flows, and invasion resistance. Even among Antarctic stenotherms, differences in biological responses related to mode of life, phylogeny and associated metabolic capacities exist (Section 6.3.1.4). As a consequence, marine ecosystem functions may be substantially reorganized at the regional scale, potentially triggering a range of cascading effects (Hoegh-Guldberg and Bruno, 2010). A focus on understanding the mechanisms underpinning the nature and magnitude of responses of marine organisms to climate change can help forecast impacts and the associated costs to society as well as facilitate adaptive management strategies formitigating these impacts (Sections 6.3, 6.4).

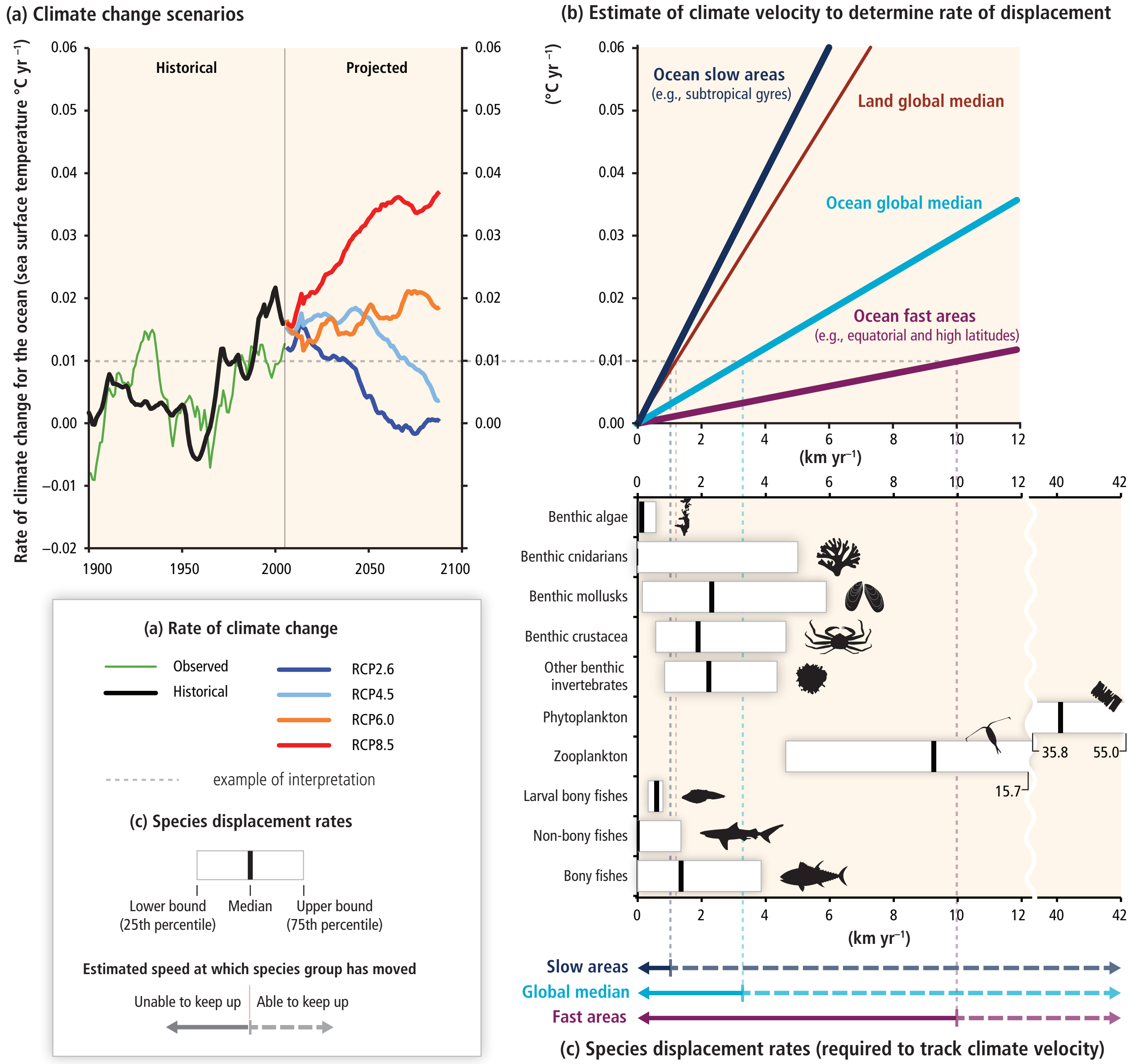

Figure MB-3 | (a) Rate of climate change for the ocean (sea surface temperature (SST) °C yr^{-1}). (b) Corresponding climate velocities for the ocean and median velocity from land (adapted from Burrows et al., 2011). (c) Observed rates of displacement of marine taxonomic groups based on observations over 1900–2010. The dotted bands give an example of interpretation. Rates of climate change of 0.01 °C yr^{-1} correspond to approximately 3.3 km yr^{-1} median climate velocity in the ocean. When compared to observed rates of displacement (c), many marine taxonomic groups have been able to track these velocities. For phytoplankton and zooplankton the rates of displacement greatly exceed median climate velocity for the ocean and, for phytoplankton exceed velocities in fast areas of the ocean approximately 10.0 km yr^{-1}. All values are calculated for ocean surface with the exclusion of polar seas (Figure 30-1a). (a) Observed rates of climate change for ocean SST (green line) are derived from the Hadley Centre Interpolated SST 1.1 (HadISST1.1) data set, and all other rates are calculated based on the average of the Coupled Model Intercomparison Project Phase 5 (CMIP5) climate model ensembles (Table SM30-3) for the historical period and for the future based on the four Representative Concentration Pathway (RCP) scenarios. Data were smoothed using a 20-year sliding window. (b) Median climate velocity over the global ocean surface (light blue line; excluding polar seas) calculated from HadSST1.1 data set over 1960–2009 using the methods of Burrows et al. (2011). Median velocities representative of ocean regions of slow velocities such as the Pacific subtropical gyre (dark blue line) and of high velocities such as the Coral Triangle or the North Sea (purple line) shown. Median rates over global land surface (red line) over 1960–2009 calculated using Climate Research Unit data set CRU TS3.1. Figure 30-3 shows climate velocities over the ocean surface calculated over 1960–2009. (c) Rates of displacement for marine taxonomic groups estimated by Poloczanska et al. (2013) using published studies. Note the displacement rates for phytoplankton exceed the axis, so values are given.

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OA

Ocean Acidification

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Anthropogenic ocean acidification and global warming share the same primary cause, which is the increase of atmospheric CO_2 (Figure OA-1A; WGI, Section 2.2.1). Eutrophication, loss of sea ice, upwelling and deposition of atmospheric nitrogen and sulfur all exacerbate ocean acidification locally (Sections 5.3.3.6, 6.1.1, 30.3.2.2).

Chemistry and Projections

The fundamental chemistry of ocean acidification is well understood (*robust evidence, high agreement*). Increasing atmospheric concentrations of CO_2 result in an increased flux of CO_2 into a mildly alkaline ocean, resulting in a reduction in pH, carbonate ion concentration, and the capacity of seawater to buffer changes in its chemistry (*very high confidence*). The changing chemistry of the surface layers of the open ocean can be projected at the global scale with high accuracy using projections of atmospheric CO_2 levels (Figure CC-OA-1B). Observations of changing upper ocean CO_2 chemistry over time support this linkage (WGI Table 3.2 and Figure 3.18; Figures 30-8, 30-9). Projected changes in open ocean, surface water chemistry for the year 2100 based on representative concentration pathways (WGI, Figure 6.28) compared to pre-industrial values range from a pH change of –0.14 units with Representative Concentration Pathway (RCP)2.6 (421 ppm CO_2, +1°C, 22% reduction of carbonate ion concentration) to a pH change of –0.43 units with RCP8.5 (936 ppm CO_2, +3.7°C, 56% reduction of carbonate ion concentration). Projections of regional changes, especially in the highly complex coastal systems (Sections 5.3.3.5, 30.3.2.2), in polar regions (WGI Section 6.4.4), and at depth are more difficult but generally follow similar trends.

Biological, Ecological, and Biogeochemical Impacts

Investigations of the effect of ocean acidification on marine organisms and ecosystems have a relatively short history, recently analyzed in several meta-analyses (Sections 6.3.2.1, 6.3.5.1). A wide range of sensitivities to projected rates of ocean acidification exists within and across diverse groups of organisms, with a trend for greater sensitivity in early life stages (*high confidence*; Sections 5.4.2.2, 5.4.2.4, 6.3.2). A pattern of positive and negative impacts emerges (*high confidence*; Figure OA-1C) but key uncertainties remain in our understanding of the impacts on organisms, life histories, and ecosystems. Responses can be influenced, often exacerbated by other drivers, such as warming, hypoxia, nutrient concentration, and light availability (*high confidence*; Sections 5.4.2.4, 6.3.5).

Growth and primary production are stimulated in seagrass and some phytoplankton (*high confidence*; Sections 5.4.2.3, 6.3.2.2, 6.3.2.3, 30.5.6). Harmful algal blooms could become more frequent (*limited evidence, medium agreement*). Ocean acidification may stimulate nitrogen fixation (*limited evidence, low agreement*; 6.3.2.2). It decreases the rate of calcification of most, but not all, sea floor calcifiers (*medium agreement, robust evidence*) such as reef-building corals (Box CC-CR), coralline algae, bivalves, and gastropods, reducing the competitiveness with non-calcifiers (Sections 5.4.2.2, 5.4.2.4, 6.3.2.5). Ocean warming and acidification promote higher rates of calcium carbonate dissolution resulting in the net dissolution of carbonate sediments and frameworks and loss of associated habitat (*medium confidence*; 5.4.2.4, 6.3.2.5, 6.3.5.4). Some corals and temperate fishes experience disturbances to behavior, navigation, and their ability to tell conspecifics from predators (Section 6.3.2.4). However, there is no evidence for these effects to persist on evolutionary time scales in the few groups analyzed (Section 6.3.2).

Some phytoplankton and molluscs displayed adaptation to ocean acidification in long-term experiments (*limited evidence, medium agreement*; Section 6.3.2.1), indicating that the long-term responses could be less than responses obtained in short-term experiments. However, mass extinctions in Earth history occurred during much slower rates of ocean acidification, combined with other drivers changing, suggesting that evolutionary rates are not fast enough for sensitive animals and plants to adapt to the projected rate of future change (*medium confidence*; Section 6.1.2).

Projections of ocean acidification effects at the ecosystem level are made difficult by the diversity of species-level responses. Differential sensitivities and associated shifts in performance and distribution will change predator–prey relationships and competitive interactions (Sections

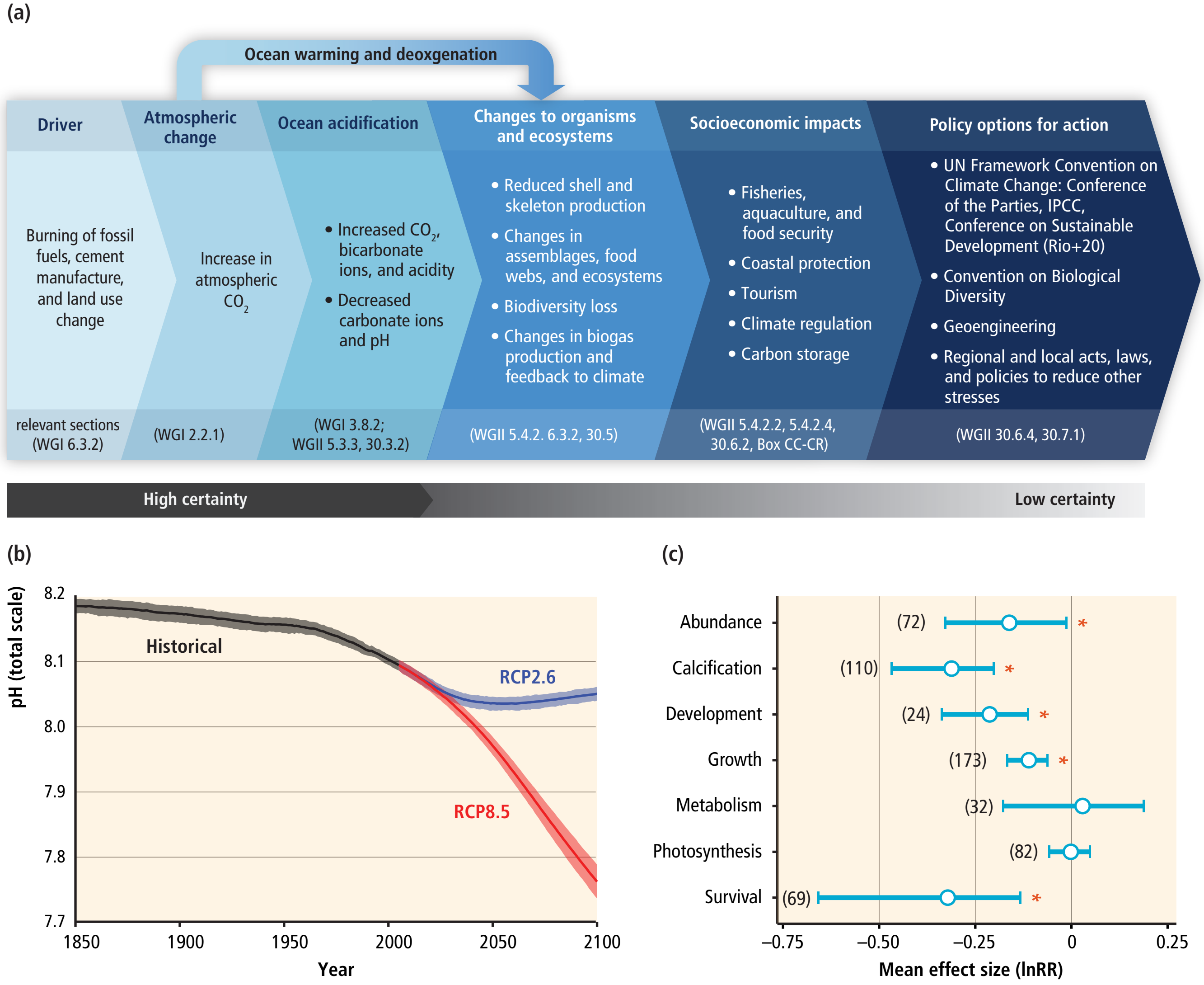

Figure OA-1 | (a) Overview of the chemical, biological, and socio-economic impacts of ocean acidification and of policy options (adapted from Turley and Gattuso, 2012). (b) Multi-model simulated time series of global mean ocean surface pH (on the total scale) from Coupled Model Intercomparison Project Phase 5 (CMIP5) climate model simulations from 1850 to 2100. Projections are shown for emission scenarios Representative Concentration Pathway (RCP)2.6 (blue) and RCP8.5 (red) for the multi-model mean (solid lines) and range across the distribution of individual model simulations (shading). Black (gray shading) is the modeled historical evolution using historical reconstructed forcings. The models that are included are those from CMIP5 that simulate the global carbon cycle while being driven by prescribed atmospheric CO_2 concentrations (WGI AR5 Figures SPM.7 and TS.20). (c) Effect of near-future acidification (seawater pH reduction of ≤0.5 units) on major response variables estimated using weighted random effects meta-analyses, with the exception of survival, which is not weighted (Kroeker et al., 2013). The log-transformed response ratio (lnRR) is the ratio of the mean effect in the acidification treatment to the mean effect in a control group. It indicates which process is most uniformly affected by ocean acidification, but large variability exists between species. Significance is determined when the 95% bootstrapped confidence interval does not cross zero. The number of experiments used in the analyses is shown in parentheses. The * denotes a statistically significant effect.

6.3.2.5, 6.3.5, 6.3.6), which could impact food webs and higher trophic levels (*limited evidence, high agreement*). Natural analogues at CO_2 vents indicate decreased species diversity, biomass, and trophic complexity of communities (Box CC-CR; Sections 5.4.2.3, 6.3.2.5, 30.3.2.2, 30.5). Shifts in community structure have also been documented in regions with rapidly declining pH (Section 5.4.2.2).

Owing to an incomplete understanding of species-specific responses and trophic interactions, the effect of ocean acidification on global biogeochemical cycles is not well understood (*limited evidence, low agreement*) and represents an important knowledge gap. The additive, synergistic, or antagonistic interactions of factors such as temperature, concentrations of oxygen and nutrients, and light are not sufficiently investigated yet.

Risks, Socioeconomic Impacts, and Costs

The risks of ocean acidification to marine organisms, ecosystems, and ultimately to human societies, include both the probability that ocean acidification will affect fundamental physiological and ecological processes of organisms (Section 6.3.2.1), and the magnitude of the resulting impacts on ecosystems and the ecosystem services they provide to society (Box 19-2). For example, ocean acidification under RCP4.5 to RCP8.5 will impact formation and maintenance of coral reefs (*high confidence*; Box CC-CR, Section 5.4.2.4) and the goods and services that they provide such as fisheries, tourism, and coastal protection (*limited evidence, high agreement*; Box CC-CR; Sections 6.4.1.1,19.5.2, 27.3.3, 30.5, 30.6). Ocean acidification poses many other potential risks, but these cannot yet be quantitatively assessed because of the small number of studies available, particularly on the magnitude of the ecological and socioeconomic impacts (Section 19.5.2).

Global estimates of observed or projected economic costs of ocean acidification do not exist. The largest uncertainty is how the impacts on lower trophic levels will propagate through the food webs and to top predators. However, there are a number of instructive examples that illustrate the magnitude of potential impacts of ocean acidification. A decrease of the production of commercially exploited shelled molluscs (Section 6.4.1.1) would result in a reduction of USA production of 3 to 13% according to the Special Report on Emission Scenarios (SRES) A1FI emission scenario (*low confidence*). The global cost of production loss of molluscs could be more than US$100 billion by 2100 (*limited evidence, medium agreement*). Models suggest that ocean acidification will generally reduce fish biomass and catch (*low confidence*) and that complex additive, antagonistic, and/or synergistic interactions will occur with other environmental (warming) and human (fisheries management) factors (Section 6.4.1.1). The annual economic damage of ocean-acidification–induced coral reef loss by 2100 has been estimated, in 2012, to be US$870 and 528 billion, respectively for the A1 and B2 SRES emission scenarios (*low confidence*; Section 6.4.1). Although this number is small compared to global gross domestic product (GDP), it can represent a very large GDP loss for the economies of many coastal regions or small islands that rely on the ecological goods and services of coral reefs (Sections 25.7.5, 29.3.1.2).

Mitigation and Adaptation

Successful management of the impacts of ocean acidification includes two approaches: mitigation of the source of the problem (i.e., reduce anthropogenic emissions of CO_2) and/or adaptation by reducing the consequences of past and future ocean acidification (Section 6.4.2.1). Mitigation of ocean acidification through reduction of atmospheric CO_2 is the most effective and the least risky method to limit ocean acidification and its impacts (Section 6.4.2.1). Climate geoengineering techniques based on solar radiation management will not abate ocean acidification and could increase it under some circumstances (Section 6.4.2.2). Geoengineering techniques to remove CO_2 from the atmosphere could directly address the problem but are very costly and may be limited by the lack of CO_2 storage capacity (Section 6.4.2.2). In addition, some ocean-based approaches, such as iron fertilization, would only relocate ocean acidification from the upper ocean to the ocean interior, with potential ramifications on deep water oxygen levels (Sections 6.4.2.2, 30.3.2.3, 30.5.7). A low-regret approach, with relatively limited effectiveness, is to limit the number and the magnitude of drivers other than CO_2, such as nutrient pollution (Section 6.4.2.1). Mitigation of ocean acidification at the local level could involve the reduction of anthropogenic inputs of nutrients and organic matter in the coastal ocean (Section 5.3.4.2). Some adaptation strategies include drawing water for aquaculture from local watersheds only when pH is in the right range, selecting for less sensitive species or strains, or relocating industries elsewhere (Section 6.4.2.1).

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PP

Net Primary Production in the Ocean

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Net Primary Production (NPP) is the rate of photosynthetic carbon fixation minus the fraction of fixed carbon used for cellular respiration and maintenance by autotrophic planktonic microbes and benthic plants (Sections 6.2.1, 6.3.1). Environmental drivers of NPP include light, nutrients, micronutrients, CO_2, and temperature (Figure PP-1a). These drivers, in turn, are influenced by oceanic and atmospheric processes, including cloud cover; sea ice extent; mixing by winds, waves, and currents; convection; density stratification; and various forms of upwelling induced by eddies, frontal activity, and boundary currents. Temperature has multiple roles as it influences rates of phytoplankton physiology and heterotrophic bacterial recycling of nutrients, in addition to stratification of the water column and sea ice extent (Figure PP-1a). Climate change is projected to strongly impact NPP through a multitude of ways that depend on the regional and local physical settings (WGI AR5, Chapter 3), and on ecosystem structure and functioning (*medium confidence*; Sections 6.3.4, 6.5.1). The influence of environmental drivers on NPP causes as much as a 10-fold variation in regional productivity with nutrient-poor subtropical waters and light-limited Arctic waters at the lower range and productive upwelling regions and highly eutrophic coastal regions at the upper range (Figure PP-1b).

The oceans currently provide ~50×10^{15} g C yr^{-1}, or about half of global NPP (Field et al., 1998). Global estimates of NPP are obtained mainly from satellite remote sensing (Section 6.1.2), which provides unprecedented spatial and temporal coverage, and may be validated regionally against oceanic measurements. Observations reveal significant changes in rates of NPP when environmental controls are altered by episodic natural perturbations, such as volcanic eruptions enhancing iron supply, as observed in high-nitrate low-chlorophyll waters of the Northeast Pacific (Hamme et al., 2010). Climate variability can drive pronounced changes in NPP (Chavez et al., 2011), such as from El Niño to La Niña transitions in Equatorial Pacific, when vertical nutrient and trace element supply are enhanced (Chavez et al., 1999).

Multi-year time series records of NPP have been used to assess spatial trends in NPP in recent decades. Behrenfeld et al. (2006), using satellite data, reported a prolonged and sustained global NPP decrease of 190×10^{12} g C yr^{-1}, for the period 1999–2005—an annual reduction of 0.57% of global NPP. In contrast, a time series of directly measured NPP between 1988 and 2007 by Saba et al. (2010) (i.e., *in situ* incubations using the radiotracer ^{14}C-bicarbonate) revealed an increase (2% yr^{-1}) in NPP for two low-latitude open ocean sites. This discrepancy between *in situ* and remotely sensed NPP trends points to uncertainties in either the methodology used and/or the extent to which discrete sites are representative of oceanic provinces (Saba et al., 2010, 2011). Modeling studies have subsequently revealed that the <15-year archive of satellite-

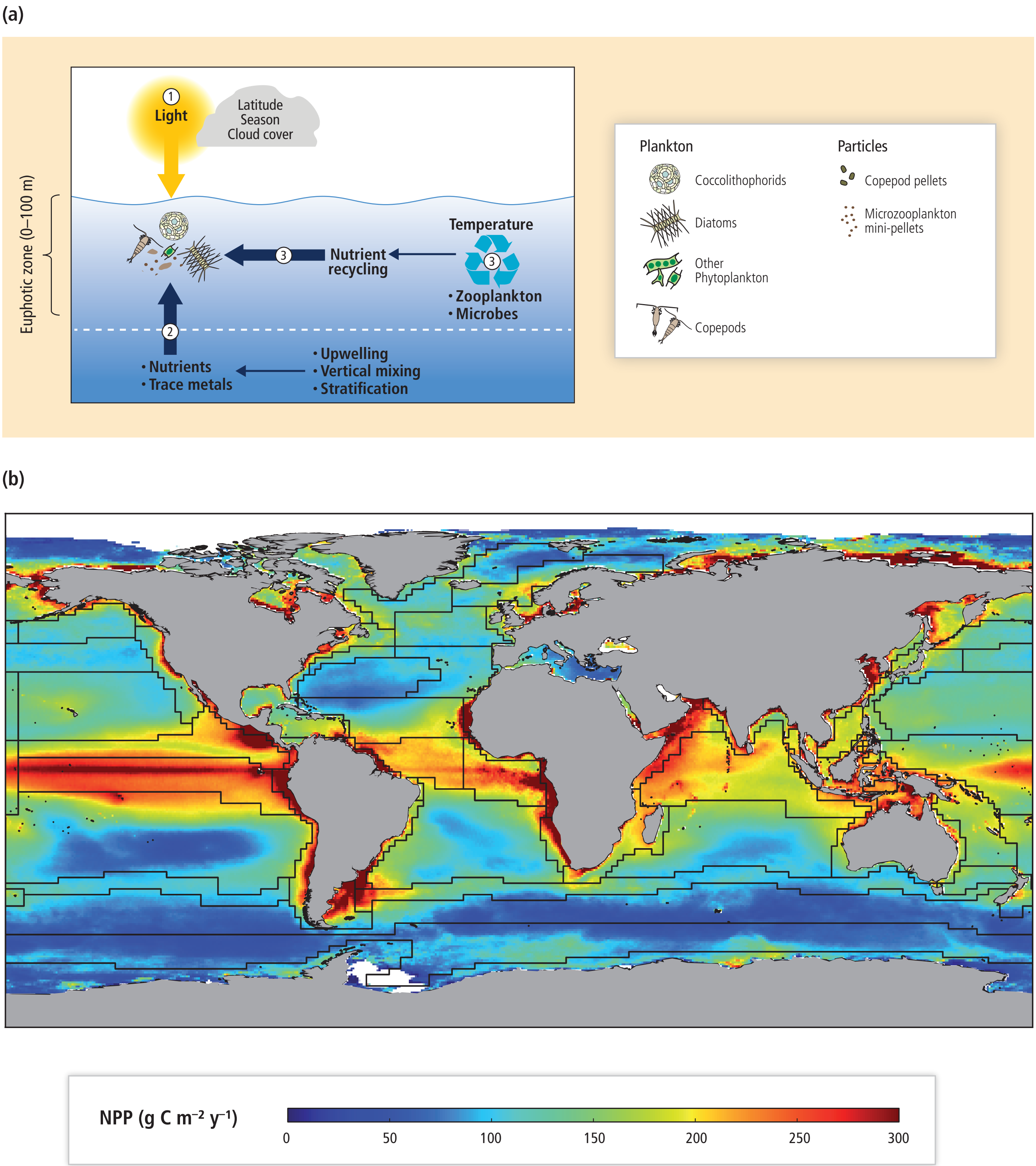

Figure PP-1 | (a) Environmental factors controlling Net Primary Production (NPP). NPP is controlled mainly by three basic processes: (1) light conditions in the surface ocean, that is, the photic zone where photosynthesis occurs; (2) upward flux of nutrients and micronutrients from underlying waters into the photic zone, and (3) regeneration of nutrients and micronutrients via the breakdown and recycling of organic material before it sinks out of the photic zone. All three processes are influenced by physical, chemical, and biological processes and vary across regional ecosystems. In addition, water temperature strongly influences the upper rate of photosynthesis for cells that are resource-replete. Predictions of alteration of primary productivity under climate change depend on correct parameterizations and simulations of each of these variables and processes for each region. (b) Annual composite map of global areal NPP rates (derived from Moderate Resolution Imaging Spectroradiometer (MODIS) Aqua satellite climatology from 2003–2012; NPP was calculated with the Carbon-based Productivity Model (CbPM; Westberry et al., 2008)). Overlaid is a grid of (thin black lines) that represent 51 distinct global ocean biogeographical provinces (after Longhurst, 1998 and based on Boyd and Doney, 2002). The characteristics and boundaries of each province are primarily set by the underlying regional ocean physics and chemistry. White areas = no data. (Figure courtesy of Toby Westberry (OSU) and Ivan Lima (WHOI), satellite data courtesy of NASA Ocean Biology Processing Group.)

derived NPP is insufficient to distinguish climate-change mediated shifts in NPP from those driven by natural climate variability (Henson et al., 2010; Beaulieu et al., 2013). Although multi-decadal, the available time series of oceanic NPP measurements are also not of sufficient duration relative to the time scales of longer-term climate variability modes as for example Atlantic Multi-decadal Oscillation (AMO), with periodicity of 60-70 years, Figure 6-1). Recent attempts to synthesize longer (i.e., centennial) records of chlorophyll as a proxy for phytoplankton stocks (e.g., Boyce et al., 2010) have been criticized for relying on questionable linkages between different proxies for chlorophyll over a century of records (e.g., Rykaczewski and Dunne, 2011).

Models in which projected climate change alters the environmental drivers of NPP provide estimates of spatial changes and of the rate of change of NPP. For example, four global coupled climate–ocean biogeochemical Earth System Models (WGI AR5 Chapter 6) projected an increase in NPP at high latitudes as a result of alleviation of light and temperature limitation of NPP, particularly in the high-latitude biomes (Steinacher et al., 2010). However, this regional increase in NPP was more than offset by decreases in NPP at lower latitudes and at mid-latitudes due to the reduced input of macronutrients into the photic zone. The reduced mixed-layer depth and reduced rate of circulation may cause a decrease in the flux of macronutrients to the euphotic zone (Figure 6-2). These changes to oceanic conditions result in a reduction in global mean NPP by 2 to 13% by 2100 relative to 2000 under a high emission scenario (Polovina et al., 2011; SRES (Special Report on Emission Scenarios) A2, between RCP6.0 and RCP8.5). This is consistent with a more recent analysis based on 10 Earth System Models (Bopp et al., 2013), which project decreases in global NPP by 8.6 (±7.9), 3.9 (±5.7), 3.6 (±5.7), and 2.0 (±4.1) % in the 2090s relative to the 1990s, under the scenarios RCP8.5, RCP6.0, RCP4.5, and RCP2.6, respectively. However, the magnitude of projected changes varies widely between models (e.g., from 0 to 20% decrease in NPP globally under RCP 8.5). The various models show very large differences in NPP at regional scales (i.e., provinces, see Figure PP-1b).

Model projections had predicted a range of changes in global NPP from an increase (relative to preindustrial rates) of up to 8.1% under an intermediate scenario (SRES A1B, similar to RCP6.0; Sarmiento et al., 2004; Schmittner et al., 2008) to a decrease of 2-20% under the SRES A2 emission scenario (Steinacher et al., 2010). These projections did not consider the potential contribution of primary production derived from atmospheric nitrogen fixation in tropical and subtropical regions, favoured by increasing stratification and reduced nutrient inputs from mixing. This mechanism is potentially important, although such episodic increases in nitrogen fixation are not sustainable without the presence of excess phosphate (e.g., Moore et al., 2009; Boyd et al., 2010). This may lead to an underestimation of NPP (Mohr et al., 2010; Mulholland et al., 2012; Wilson et al., 2012), however, the extent of such underestimation is unknown (Luo et al., 2012).

Care must be taken when comparing global, provincial (e.g., low-latitude waters, e.g., Behrenfeld et al., 2006) and regional trends in NPP derived from observations, as some regions have additional local environmental influences such as enhanced density stratification of the upper ocean from melting sea ice. For example, a longer phytoplankton growing season, due to more sea ice–free days, may have increased NPP (based on a regionally validated time-series of satellite NPP) in Arctic waters (Arrigo and van Dijken, 2011) by an average of 8.1×10^{12} g C yr^{-1} between 1998 and 2009. Other regional trends in NPP are reported in Sections 30.5.1 to 30.5.6. In addition, although future model projections of global NPP from different models (Steinacher et al., 2010; Bopp et al., 2013) are comparable, regional projections from each of the models differ substantially. This raises concerns as to which aspect(s) of the different model NPP parameterizations are responsible for driving regional differences in NPP, and moreover, how accurate model projections are of global NPP.

From a global perspective, open ocean NPP will decrease moderately by 2100 under both low- (SRES B1 or RCP4.5) and high-emission scenarios (*medium confidence*; SRES A2 or RCPs 6.0, 8.5, Sections 6.3.4, 6.5.1), paralleled by an increase in NPP at high latitudes and a decrease in the tropics (*medium confidence*). However, there is *limited evidence* and *low agreement* on the direction, magnitude and differences of a change of NPP in various ocean regions and coastal waters projected by 2100 (*low confidence*).

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UP

Uncertain Trends in Major Upwelling Ecosystems

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Upwelling is the vertical transport of cold, dense, nutrient-rich, relatively low-pH and often oxygen-poor waters to the euphotic zone where light is abundant. These conditions trigger high levels of primary production and a high biomass of benthic and pelagic organisms. The driving forces of upwelling include wind stress and the interaction of ocean currents with bottom topography. Upwelling intensity also depends on water column stratification. The major upwelling systems of the planet, the Equatorial Upwelling System (EUS; Section 30.5.2, Figure 30.1A) and the Eastern Boundary Upwelling Ecosystems (EBUE; Section 30.5.5, Figure 30.1A), represent only 10% of the ocean surface but contribute nearly 25% to global fish production (Figure 30.1B, Table SM30.1).

Marine ecosystems associated with upwelling systems can be influenced by a range of "bottom-up" trophic mechanisms, with upwelling, transport, and chlorophyll concentrations showing strong seasonal and interannual couplings and variability. These, in turn, influence trophic transfer up the food chain, affecting zooplankton, foraging fish, seabirds, and marine mammals.

There is considerable speculation as to how upwelling systems might change in a warming and acidifying ocean. Globally, the heat gain of the surface ocean has increased stratification by 4% (WGI Sections 3.2, 3.3, 3.8), which means that more wind energy is needed to bring deep waters to the surface. It is as yet unclear to what extent wind stress can offset the increased stratification, owing to the uncertainty in wind speed trends (WGI Section 3.4.4). In the tropics, observations of reductions in trade winds over several decades contrast more recent evidence indicating their strengthening since the late 1990s (WGI Section 3.4.4). Observations and modeling efforts in fact show diverging trends in coastal upwelling at the eastern boundaries of the Pacific and the Atlantic. Bakun (1990) proposed that the difference in rates of heat gain between land and ocean causes an increase in the pressure gradient, which results in increased alongshore winds and leads to intensified offshore transport of surface water through Ekman pumping and the upwelling of nutrient-rich, cold waters (Figure CC-UP). Some regional records support this hypothesis; others do not. There is considerable variability in warming and cooling trends over the past decades both within and among systems, making it difficult to predict changes in the intensity of all Eastern EBUEs (Section 30.5.5).

Understanding whether upwelling and climate change will impact resident biota in an additive, synergistic, or antagonistic manner is important for projections of how ecological goods and services provided for human society will change. Even though upwellings may prove more resilient to climate change than other ocean ecosystems because of their ability to function under extremely variable conditions (Capone and Hutchins, 2013), consequences of their shifts are highly relevant because these systems provide a significant portion of global primary productivity and fishery catch (Figure 30.1 A, B; Table SM30.1). Increased upwelling would enhance fisheries yields. However, the export of organic material from surface to deeper layers of the ocean may increase and stimulate its decomposition by microbial activity, thereby enhancing oxygen depletion and CO_2 enrichment in deeper water layers. Once this water returns to the surface through upwelling, benthic and pelagic coastal communities will be exposed to acidified and deoxygenated water which may combine with anthropogenic impact to negatively affect marine biota and ecosystem structure of the upper ocean (*high confidence*; Sections 6.3.2, 6.3.3, 30.3.2.2, 30.3.2.3). Extreme hypoxia may result in abnormal mortalities of fishes and invertebrates (Keller et al., 2010), reduce fisheries' catch potential, and impact aquaculture in coastal areas (Barton et al., 2012; see also Sections 5.4.3.3, 6.3.3, 6.4.1, 30.5.1.1.2, 30.5.5.1.3). Shifts in upwelling also coincide with an apparent increase in the frequency of submarine eruptions of methane and hydrogen sulfide gas, caused by enhanced formation and sinking of phytoplankton biomass to the hypoxic or anoxic sea floor. This combination of factors has been implicated in the extensive mortality of coastal fishes and invertebrates (Bakun and Weeks, 2004; Bakun et al., 2010), resulting in significant reductions in fishing productivity, such as Cape hake (*Merluccius capensis*), Namibia's most valuable fishery (Hamukuaya et al., 1998).

Reduced upwelling would also reduce the productivity of important pelagic fisheries, such as for sardines, anchovies and mackerel, with major consequences for the economies of several countries (Section 6.4.1, Chapter 7, Figure 30.1A, B, Table S30.1). However, under projected scenarios of reduced upward supply of nutrients due to stratification of the open ocean, upwelling of both nutrients and trace elements may become increasingly important to maintaining upper ocean nutrient and trace metal inventories. It has been suggested that upwelling areas may also increase nutrient content and productivity under enhanced stratification, and that upwelled and partially denitrified waters containing excess phosphate may select for N_2-fixing microorganisms (Deutsch et al., 2007; Deutsch and Weber, 2012), but field observations of N_2 fixation in these regions have not supported these predictions (Fernandez et al., 2011; Franz et al., 2012). The role of this process in global primary production thus needs to be validated (*low confidence*).

The central question therefore is whether or not upwelling will intensify, and if so, whether the effects of intensified upwelling on O_2 and CO_2 inventories will outweigh its benefits for primary production and associated fisheries and aquaculture (*low confidence*). In any case increasing atmospheric CO_2 concentrations will equilibrate with upwelling waters that may cause them to become more corrosive, depending on pCO_2 of the upwelled water, and potentially increasingly impact the biota of EBUEs.

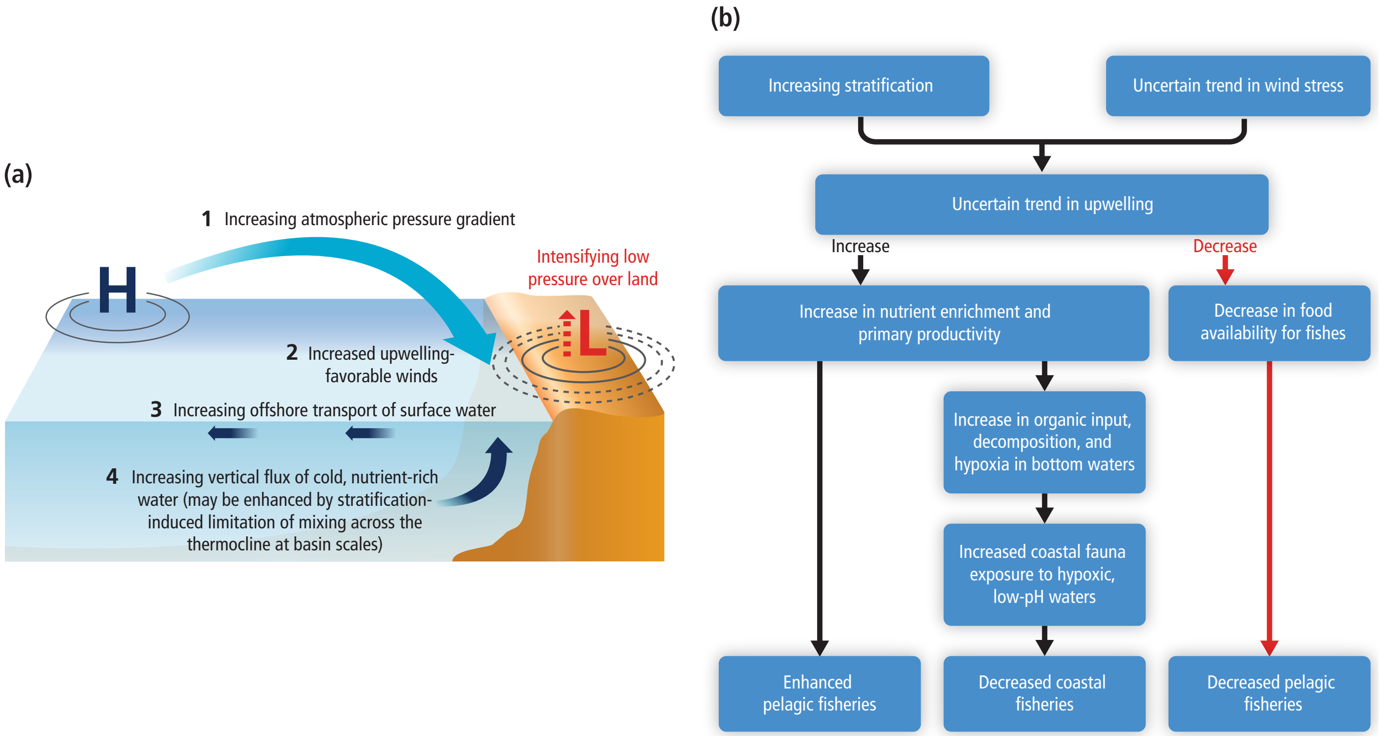

Figure UP-1 | (a) Hypothetic mechanism of increasing coastal wind–driven upwelling at Equatorial and Eastern Boundary upwelling systems (EUS, EBUE, Figure 30-1), where differential warming rates between land and ocean results in increased land–ocean (1) pressure gradients that produce (2) stronger alongshore winds and (3) offshore movement of surface water through Ekman transport, and (4) increased upwelling of deep cold nutrient rich waters to replace it. (b) Potential consequences of climate change in upwelling systems. Increasing stratification and uncertainty in wind stress trends result in uncertain trends in upwelling. Increasing upwelling may result in higher input of nutrients to the euphotic zone, and increased primary production, which in turn may enhance pelagic fisheries, but also decrease coastal fisheries due to an increased exposure of coastal fauna to hypoxic, low pH waters. Decreased upwelling may result in lower primary production in these systems with direct impacts on pelagic fisheries productivity.

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