Chapter 30. The Ocean

Coordinating Lead Authors
Ove Hoegh-Guldberg (Australia), Rongshuo Cai (China)

Lead Authors
Peter G. Brewer (USA), Victoria J. Fabry (USA), Karim Hilmi (Morocco), Sukgeun Jung (Republic of Korea), Elvira Poloczanska (Australia), Svein Sundby (Norway)

Contributing Authors
Johann Bell (New Caledonia), Christopher J. Brown (Australia), Michael T. Burrows (UK), Long Cao (USA), Simon Donner (Canada), C. Mark Eakin (USA), Arne Eide (Norway), Benjamin Halpern (USA), Charles R. McClain (USA), Skip McKinnell (Canada), Mary O'Connor (Canada), Camille Parmesan (USA), R. Ian Perry (Canada), Anthony J. Richardson (Australia), David Schoeman (Australia), Sergio Signorini (USA), William Skirving (USA / Australia), Dáithí Stone (Canada / South Africa / USA), William Sydeman (USA), Rui Zhang (China), Ruben van Hooidonk (USA)

Review Editors
Ly Omar (Senegal), Carol Turley (UK)

Volunteer Chapter Scientists
Jo Davy (New Zealand), Sarah Lee (USA)

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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 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), Sub-Tropical Gyres (STG), Semi-Enclosed Seas (SES), and the Deep Sea (DS; >1000 m). The eighth region, Polar Seas, is dealt with by Chapter 28 [Figure 30-1; WGI 3.2.5, Box 3.1, 3.8].

Global average sea surface temperatures have increased since both the beginning of the 20th Century and the 1950s (*virtually certain*). The average sea surface temperature (SST) of the Indian, Atlantic and Pacific Oceans has increased by 0.65, 0.41 and 0.31°C respectively over the period 1950–2009 (*very likely*, p-value$\leq$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 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 2.4.2-3, 3.2–3.8, 10.3.4, 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, 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*) [6.2, 6.3, 30.3.1, 30.3.2; 6.1.2; WGI 3.8.2, Box 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 greenhouse gas emissions (*low* to *medium confidences*). Marine organisms and ecosystems are *likely* to change in response to these regional changes, although evidence is limited and responses uncertain [30.3, 30.5; 6.2, 6.3; WGI 2.5, 2.7 3.3–3.8, 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 1–3°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, 30.3.2, 30.7.1 Figure 30-2e–g, Table 30-3].

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 [Box CC-MB, 30.4, 30.5, 6.3, 6.4].

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 [30.1.2, 30.5, 5.3.4, 18.3.3–4].

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 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 [Box CC-PP, 30.3.1, 30.3.2, 30.5.5–6, 6.3.3; WGI 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 SE 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, 30.6.1, 30.5.3–4, 30.5.6, Figure 30-4, Box CC-CR, IPCC 2012].

The highly productive HLSBS in the North-eastern Atlantic has changed in response to warming oceans (*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 [Box CC-MB, Box 6-1, 6.4.1.2, 30.5.1, 30.6.2.1].

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*, $p\text{-value} \leq 0.05$). While 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 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 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 [Table 30-1, 30.5.3, 5.3.4.3].

Coral reefs within CBS, SES, and STG are rapidly declining as result of local (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 CMIP5 ensemble projecting the loss of coral reefs from most sites globally by 2050 under mid to high rates of ocean warming (*very likely*) [Figure 30-10, 30.5.3–4, 30.5.6, 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 increases 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 livelihoods along the west coasts of South America, Africa, and North America (*low* to *medium confidence*) [30.3.2.2, 30.5.2, 30.5.5, Box CC-UP, Box 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 1998–2010 (*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 declines in nutrient availability and ecosystem productivity (*limited evidence*, *medium agreement*). The short timeframe of these studies against well-established patterns of long-term variability lead to the conclusion that these changes are *about as likely as not* due to climate change [30.5.6, Table 30-1, Box CC-UP, 6.3.4].

The world's most abundant yet difficult to access habitat, the DS, is changing (*limited evidence*, *medium agreement*), with warming between 700–2000 m from 1957–2010 *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 3.2.4, Figure 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 [30.3.1, 30.6.2, 28.2.2, 28.2.5, 28.3.4, 10.2.2, 10.4.4, 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 [30.6, 30.7, 7.4.2, 29.4].

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*) [30.6, 7.3.2.4].

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 provide 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 [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 fishing, IUU), 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, and indirectly through the regulation of atmospheric gas content and the distribution of heat and water across the planet. Chapter 30 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 (particularly Chapters 2, 3, 13, and 14). In Chapter 30, 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.

While 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., WGII 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), 30.5.7). There are considerable challenges in assessing the regional impacts of climate change on the Ocean. Devising an appropriate structure in order 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 over 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, we have used comparable principles but have divided the non-polar ocean into seven larger sub-regions similar to *Barber* [1988]. We recognize 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, over 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 (>1000m) is included as a separate category that overlaps with the six other ocean sub-regions dealt with in Chapter 30.

[INSERT FIGURE 30-1 HERE
Figure 30-1: (a) Separation of the world's non-polar oceans into seven major sub-regions (excluding the polar oceans, which are considered in Chapter 28). The chlorophyll-*a* signal measured by SeaWiFS and averaged over the period from Sep 4, 1997 to 30 Nov 2010 (NASA) is 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). Numbers refer to: 1 = High Latitude Spring Bloom Systems (HLSBS); 2 = Equatorial Upwelling Systems (EUS); 3 = Semi-Enclosed Seas (SES); 4 = Coastal Boundary Systems (CBS); 5 = Eastern Boundary Upwelling Ecosystems (EBUE); 6 = Sub-Tropical Gyres (STG); and 7 = DS (>1000 m). (b) Relationship between fish catch and area for each ocean sub-region is shown in (a). Red columns: average fish catch (as millions tons yr^{-1}) for the period 1970–2006. Blue columns: area (millions km^2). The four left-hand columns (sub-regions HLSBS-North, CBS, EBUE, and SES) cover 20 % of the world oceans' area and deliver 80% of the world's fish catches. The values for the percent area of the Ocean, primary productivity, and fishery catch for the major sub-regions are listed in Table SM30-1.].

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

The central goal of Chapter 30 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 Chapters 2, 3, 6, 10, 13, and 14 (AR5). The detection and attribution of climate change and ocean acidification on marine organisms and ecosystems is addressed in Chapter 6. Chapter 30 draws on these chapters to investigate regional changes in the physical, chemical, ecological, and socio-economic 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 [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. We attempt to do this 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 [18.2.1] along with challenges [18.2.2]. Different approaches and 'best practice' guidelines are discussed in WGI Chapters 10 and 18 as well as in several other places [*Hegerl et al.*, 2007; *Hegerl et al.*, 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, overexploitation 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; *Hegerl et al.*, 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 socio-economic 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 also concluded that sea levels had risen due to the thermal expansion of the Ocean but recognized that 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" (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" (AR4, WGI Chapter 5, Box 5.1), despite limited evidence of a slowing MOC.
According to AR4, "sea-level rise over the last 100–150 years is probably contributing to coastal erosion in many places", including the east coast of the United States and the United Kingdom (AR4, WGII Chapter 1). The AR4 assessment was *virtually certain* that rising atmospheric 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]; AR4, 5.4.2.3, WGI Table 7.3). Large-scale changes in ocean salinity were also observed from 1955–1998 and were "characterized by a global freshening in sub-polar latitudes and salinification of shallower parts of the tropical and subtropical oceans". In this case, freshening was observed in the Pacific, with increased salinity being observed in the Atlantic and Indian Oceans (AR4, WGI 5.3.2–5.3.5). 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".

Substantial evidence indicated that changing ocean conditions have extensively influenced marine ecosystems (AR4, WGII 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" (AR4, WGII 1.3.4.2). 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 and kelp forests, and the distribution of pathogens and invasive species. Changes were also observed in coral reefs (primarily increased mass coral bleaching and mortality), migratory patterns and trophic interactions of marine birds, reptiles, and mammals, as well as of a range of other marine organisms and ecosystems (AR4, WGII Table 1.5), although a separate exercise in detection and attribution of changes 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 (0–700 m) ocean (1971–2010; WGI Section 3.2.3, Figure 3.2, Box 3.1). It is *virtually certain* that global average sea surface temperatures (SST) 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 2.4.2). It is *virtually certain* that the upper ocean (0–700m depth) has warmed from 1971–2010 (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 decade^{-1} in the upper 75 m from 1971 to 2010) decreasing with depth (0.015°C decade^{-1} 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 0–200 m depth from 1971–2010 in all oceans north of 40°S. It is *likely* that the Ocean has warmed between 700–2000 m from 1957–2010, with the warming signal becoming less apparent or non-existent at deeper depths (WGI 3.2.1–3.2.3, Figures 3.1–3.2, Figure 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 all exceed those expected if there were no changes to greenhouse gas forcing over the past century (Figure 30-2e-g). In this respect, the observed record also falls within the range of historical model outputs that include observed increases in the concentration of greenhouse gases as opposed to models that do not (Figure 30-2 e-g).

Data archives such as HadISST1.1 contain sea surface temperatures (SST) 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]. We used the published HadISST1.1 data set (higher temporal and spatial resolution than HadSST3) to explore trends in historic SST within our sub-regions (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 date.

The surface layers of the three ocean basins have warmed (p-value≤0.05, *very likely*), with the Indian Ocean (0.11℃ decade^{-1}) warming faster than the Atlantic (0.07℃ decade^{-1}) and Pacific (0.05℃ decade^{-1}) Oceans *(high confidence)* (Table 30-1). This is consistent with the depth-averaged (0–700 m) temperature trend observed from 1971–2010 (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 1950–2009 (except in the Indian Ocean; Table 30-1) yet the temperatures of the warmest month (North and South Atlantic, and South-eastern 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 1950–2009 (Pacific EUS, 0.07℃ decade^{-1} and Atlantic EUS, 0.09℃ decade^{-1}; 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 decade^{-1} or 2.11°C from 1950–2009), as did the temperatures of the warmest months in the Black (0.14°C decade^{-1} or 0.83°C from 1950–2009), Mediterranean (0.11°C decade^{-1} or 0.66°C from 1950–2009) and Red (0.05°C decade^{-1} or 0.28°C from 1950–2009) 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 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–0.13°C.decade^{-1}, 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 decade^{-1} and 0.12°C decade^{-1}, respectively; p-value≤0.05), while the Benguela and Humboldt currents did not show significant temperature changes from 1950–2009 (p-value>0.05; Table 30-1). There was some variability between current systems in terms of the behavior of the coolest and warmest months. The temperature of the coolest month increased significantly from 1950–2009 in the case of the Benguela and California currents (0.06°C decade^{-1} and 0.12°C decade^{-1} 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 decade^{-1} and 0.10°C decade^{-1}, 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, 0.08, and 0.06°C decade^{-1}, 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, we found that coral reef waters (with the notable exception of the Gulf of Mexico and Caribbean) have shown strong increases in average temperature (0.07–0.13°C decade^{-1}) as well as the temperature of the coolest (0.07–0.14°C decade^{-1}) and warmest months (*very likely*) (0.07–0.12°C decade^{-1}; Table 30-1). These trends in temperature have resulted in an absolute increase in sea temperature of 0.44–0.79°C from 1950–2009.

[INSERT FIGURE 30-2 HERE
Figure 30-2: (a) Depth-averaged 0–700 m temperature trend for 1971–2010 (longitudinal versus latitude, colors and gray contours in °C per decade). (b) Zonally averaged temperature trends (latitude versus depth, colors and gray contours in °C per decade) for 1971–2010, with zonally averaged mean temperature over plotted (black contours in °C). (c) Globally-averaged temperature anomaly (Time versus depth, colors and grey contours in °C) relative to the 1971–2010 mean. (d) Globally-averaged temperature difference between the Ocean surface and 200 m depth (Black: annual values; red: five year running mean). Panels (a)–(d) from WGI Figure 3.1. (e)–(g) Observed and simulated variations in past and projected future annual average SST over three ocean basins (excluding regions within 300 km of the coast). The black line shows estimates from HadISST1.1 observational measurements. 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), and the Representative Concentration Pathways: Dark Blue: RCP2.6; Light Blue: RCP4.5; Green: RCP6.0, and Red: 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 Box 21-2.]

[INSERT FIGURE 30-3 HERE
Figure 30-3: Velocity at which sea surface temperature (SST) isotherms shifted (km decade^{-1}) over the period 1960–2009 calculated using HaDISST1.1, with arrows indicating the direction and magnitude of shifts. Velocity of climate change is obtained by dividing the temperature trend in °C decade^{-1} by the local spatial gradient °C km^{-1}. The direction of movement of SST is denoted by the direction of the spatial gradient and the sign of the temperature trend: towards locally cooler areas with a local warming trend or towards locally warmer areas where temperatures are cooling. Adapted from [*Burrows et al.*, 2011].]

[INSERT TABLE 30-1 HERE
Table 30-1: Regional changes in sea surface temperature (SST) over the period 1950–2009 using the Ocean regionalization specified in Figure 30-1a (for further detail of regions defined for analysis, see Figure SM30-1 and Table 30-2, column 1). A linear regression was fitted to the average of all 1×1 degree monthly SST data extracted from the HadISST1.1 data set [*Rayner et al.*, 2003] for each sub-region over the period 1950–2009. All SST values less than -1.8^{o}C, together with all SST pixels that were flagged as being sea ice, were reset to the freezing point of seawater (-1.8^{o}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 decade^{-1}), 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 6 decades) for each category. The p-values that exceed 0.05 plus the associated slope and change values have a gray background, denoting the lower statistical confidence in the slope being different from zero (no slope). Note, changes with higher p-values may still describe informative trends although the level of confidence is lower that the slope is different from zero.]

Given the essential role that temperature plays in the biology and ecology of marine organisms (Box CC-MB, 6.2, 6.3, [*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). While many of these temperature trajectories are towards 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 (up to 200 km decade^{-1}), 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.

Significant 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 has proven useful in understanding recent changes to 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 four 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 the 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.

[INSERT FIGURE 30-4 HERE
Figure 30-4: Recent changes in thermal stress calculated using HadISST1.1 data. A monthly climatology was created by averaging the HadISST monthly SST values over the period 1985–2000 to create twelve averages, one for each month of the year. The Maximum Monthly Mean (MMM) climatology was created by selecting the hottest month for each pixel. Anomalies were then created by subtracting this value from each SST value, but only allowing values to be recorded if they were greater than zero [*Donner et al.*, 2007]. Two measures of the change in thermal stress were calculated as a result: (a) The total thermal stress for the period 1981–2010, calculated by summing all monthly thermal anomalies for each grid cell. (b) The location of coral reef grid cells used in Table 30-1 and for comparison to regional heat stress here. 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 & Gulf of Mexico; Green – Western Indian Ocean; Pink – Eastern Indian Ocean; and Light Blue – Coral Triangle & SE Asia. (c) Proportion of years with thermal stress, which is defined as any year that has a thermal anomaly, for the periods 1951–1980 and (d) 1981–2010.]

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 30-2 e-g; WGI, AI.4–AI.8). Projected changes were also examined for specific ocean sub-regions using ensemble averages from AOGCM simulations available in the CMIP5 archive (Table SM30-3) for the four scenarios of the future (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 10–16 individual models (Table SM30-3). The subset of CMIP5 models were chosen because each has historic runs enabling the derivation of the MMM climatology from 1985–2000, 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$\leq$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 are largely indistinguishable between the different RCP pathways due to the similarity in forcing until 2040. By the end of the century, however, SST across the ocean sub-regions were 1.8–3.3°C higher under RCP8.5 than those projected to occur under RCP2.6 (Table SM30-4; Figure 30-2 e–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 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 (0.17–0.21) m (WGI Figure SPM, 3.7, 5.6, 13.2). It is *very likely* that the mean rate of global averaged sea level rise was 1.7 [1.5–1.9] mm yr^{-1} between 1901 and 2010, 2.0 [1.7–2.3] mm yr^{-1} between 1971 and 2010 and 3.2 [2.8–3.6] mm yr^{-1} between 1993 and 2010 (WGI SPM, 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 sea level rise 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 sea level rise are up to three times higher than the GMSL in the Western Pacific and South-east 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 13.6.5).

Sea level rise under increasing atmospheric greenhouse gas concentrations will continue for hundreds of years, with the extent and rate of the increase in GMSL being dependent on the emission scenario followed. Central to this analysis is the millennial-scale commitment to further sea level rise that is *likely* to arise from the loss of mass of the Greenland and Antarctic ice sheets (WGI 13.5.4, Figure 13.13). Sea level rise 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 sea level rise by 2081–2100 relative to 1986-2005 will be (5–95% 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 sea level rise 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 13.5.2–3; 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 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, although there is *medium agreement* that the global frequency of tropical cyclones is *likely* to decrease or remain constant (WGI 14.6, 14.8). While understanding of associated risks is relatively undeveloped, coastal and low-lying areas, particularly in the southern Asia, Pacific Ocean and North Atlantic regions, face increased flood risk (5.3.3.2, 8.2.3.4, 9.3.4.4. Future impacts of sea level rise include increasing penetration of storm surges into coastal areas and changing patterns of shoreline erosion (5.3), as well as the inundation of coastal aquifers by saltwater (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 (5.4.2, Box CC-CR), especially where rates of sea level rise are highest (e.g., South-east Asia and the Western Pacific).

[INSERT FIGURE 30-5 HERE
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 (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 2.7.5, 2.7.6, 2.7.8, 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 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 2.7.5, 2.7.8, Box 2.5). Wind stress has increased since the early 1980s over the Southern Ocean (*medium confidence*) (WGI 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 3.4.4; [*Vecchi et al.*, 2006]). For example, it is *very likely* that the sub-tropical 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 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 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 3.6, Figures 3.10, 3.11].

Winds speed may have increased within the regions of EBUE (*low confidence* in attribution to climate; e.g. California Current, WGI 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 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 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 NCEP/NCAR Reanalysis Project, *Kalnay et al.* [1996]) decreased by 4.3 W m^{-2} $decade^{-1}$ from the 1950s until 1991, after when it increased at 3.3 W m^{-2} $decade^{-1}$ until 1999 [*Ohmura*, 2009; *Wild*, 2009], matching a broad suite of evidence from many land-based sites (WGI, 2.3.3). While 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 2.5.7). 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 (ENSO, PDO), short observation time series and uneven spatial sampling, particularly in the early record (before 1950; WGI 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]. While 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 2.6.3). There is *medium agreement* that the frequency of the most intense cyclones in the Atlantic has increased since 1987 (WGI 2.6.3) and *robust evidence* of interdecadal changes in the storm track activity within the North Pacific and North Atlantic [*Lee et al.*, 2012]. It is also *very 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 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 extra-tropical cyclone systems since the 1950s in the North Pacific. Similar trends have been reported for the Asian region, although analyzes are limited in terms of the spatial and temporal coverage of reliable records (WGI 2.6.4). There is *low confidence,* however, in large-scale trends in storminess or storminess proxies over the last century due to the lack of long-term data and inconsistencies between studies (WGI 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 (0–200 m, 40-year record) north of 40°S (WGI 3.2.2). Increasing thermal stratification has reduced ocean ventilation and the depth of mixing in many ocean sub-regions (*medium confidence*) WGI 3.8.3). This in turn reduces the availability of inorganic nutrients and consequently primary productivity (*medium confidence*) (6.3.4). In the STG, which dominate the three major ocean basins (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 or 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, 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).

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 an outcome of the balance between evaporation and precipitation ([*Durack and Wijffels*, 2010]; WGI 3.3). Evaporation-dominated regions (Figure 30-6b) such as the STG, and Atlantic and Western Indian Oceans (WGI 3.3.3) have elevated salinity, while areas of high precipitation such as the North Pacific, North-eastern Indian Ocean, South-east Asia, and the eastern Pacific have relatively low salinities (WGI 3.3.3, Figure 30-6a). It is *very 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 3.3.2–3.3.4).

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 2.6.2.1, 3.3.4; [*IPCC*, 2012]). Spatial patterns in salinity and evaporation-precipitation are similar, providing indirect evidence that these processes have been enhanced since the 1950s [WGI 3.3.2–3.3.4, Figures 3.4, 3.5 and 3.20d, FAQ 3.3]. These trends in salinity are *very likely* to have a discernible contribution from anthropogenic climate change (WGI 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 Atlantic Meridional Overturning Circulation (AMOC).

[INSERT FIGURE 30-6 HERE
Figure 30-6: (a) The 1955–2005 climatological-mean sea surface salinity [*Antonov et al.*, 2010] color contoured at 0.5 PSS78 intervals (black lines). (b) Annual mean evaporation-precipitation averaged over the period 1950–2000 (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 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) 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 Figure 3.4 in WGI.]

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, WG1 Box 3.2; Figure SM30-2). The chemical response to increased CO_2 dissolving into the Ocean from the atmosphere is known with *very high confidence* (WGI 6.4.4). Factors such as temperature, biological processes, and sea ice (WGI 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., eastern Pacific upwelling, Californian Current'), where organisms and ecosystems may be relatively more exposed to ocean acidification as a result ([*Feely et al.*, 2012; *Gruber et al.*, 2012]; Figure 30-7a, b; Figure SM30-2). Aragonite and calcite concentrations vary with depth, with under-saturation occurring at deeper depths in the Atlantic (calcite: 3500–4500 m, aragonite: 400–3000 m) as opposed to the Pacific and Indian Oceans (calcite: 100–3000 m, aragonite: 100–1200 m; [*Feely et al.*, 2004; *Orr et al.*, 2005; *Feely et al.*, 2009]; Figure 30-8).

[INSERT FIGURE 30-7 HERE
Figure 30-7: Projected ocean acidification from 11 CMIP5 Earth System models under RCP8.5 (other 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 (filled), 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 1850–2100. 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 4, e.g., 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 Figure 6.28 in WGI.]

[INSERT FIGURE 30-8 HERE
Figure 30-8: Projected aragonite saturation state from 11 CMIP5 Earth System models under RCP8.5 scenario: (a) time series of surface carbonate ion concentration shown as the mean (solid line) and range of models (filled), 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 ΩA in (b) 2010, (d) 2050, and (f) 2100; and zonal mean sections (latitude versus depth) of ΩA 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. Reprinted from Figure 6.29 in WGI.]

Surface ocean pH has decreased by approximately 0.1 pH units since the beginning of the Industrial Revolution (*high confidence*) (Figure 30-7a; WGI 3.8.2, Box 3.2), with pH decreasing at the rate of -0.0013 and -0.0024 pH units yr^{-1} (WGI 3.8.2, Table 3.2). The presence of anthropogenic CO_2 diminishes with depth. The saturation horizons of both polymorphs of calcium carbonate, however, are shoaling rapidly (1–2 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 S30.2, 30.7 and 30.8). Doubling atmospheric CO_2 (~RCP4.5; [*Rogelj et al.*, 2012]) will decrease ocean pH by another 0.1 units 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, +1ºC, 22% reduction of carbonate ion concentration) to a pH change of -0.43 unit with RCP8.5 (936 ppm CO_2, +3.7ºC, 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 versus equator) in the Pacific Ocean ([*Sabine et al.*, 2004; *Orr et al.*, 2005]; WGI 6.4, Figure 6.28). Trends towards 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–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, 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 (100–2000 m) communities of scleractinian corals and other benthic organisms as atmospheric CO_2 increases ([*Orr et al.*, 2005; *Guinotte et al.*, 2006]; WGI 6.4), although studies from the Mediterranean and of seamounts off SW 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 functions (6.3.2). Due 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 impacts 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 10–100 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]; 6.1.2).

30.3.2.3. Oxygen Concentration

Dissolved O_2 is a major determinant of the distribution and abundance of marine organisms (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, 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 analyzes 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 Figure 3.20, 3.8.3). 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, consistent with greater mixing and ventilation due to strengthening wind systems (WGI 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 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 Box 6.5 Figure 1). 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 (generally defined as O_2 concentrations below ~60 μmol kg^{-1}) over recent decades has been documented across 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 mmol^{-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 (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 Table 6.14 [*Andrews et al.*, 2013]), with implications for nutrient and carbon cycling, ocean productivity, marine habitats, and ecosystem structure (6.3.5). The outcomes of these global changes are *very likely* to be influenced by regional differences such as wind stress, coastal processes, and the supply of organic matter.

[INSERT FIGURE 30-9 HERE
Figure 30-9: (a) Simulated changes in dissolved O_2 (mean and model range as shading) relative to 1990s for RCP2.6, RCP4.5, RCP6.0, and RCP8.5. (b) Multi-model mean dissolved O_2 (mmol m^{-3}) in the main thermocline (200–600 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 inter-comparison using the SRES-A2 scenario and have been applied to CMIP5 models here. Models used: CESM1-BGC, GFDL-ESM2G, GFDL-ESM2M, HadGEM2-ES, IPSL-CM5A-LR, IPSL-CM5A-MR, MPI-ESM-LR, MPI-ESM-MR, NorESM1. Figure originally presented in WGI Figure 6.30 in WGI.]

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 (6.3, Box CC-MB, Box 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 a rapid 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 have 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 decade^{-1} (mean ± SE) and the leading edges of species' distributions have extended (generally poleward) by 72.0 ± 0.35 km decade^{-1}. 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], 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. While 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 by 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 (HLSBS) stretch from 35ºN to the edge of the winter sea ice (and from 35ºS 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 both the magnitude and the spatial and temporal synchrony between successive trophic production peaks [*Hjort*, 1914; *Cushing*, 1990; *Beaugrand et al.*, 2003; *Beaugrand and Reid*, 2003].

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 decade^{-1}, 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 decade^{-1} and 0.12°C decade^{-1} respectively, $p\text{-value} \leq 0.05$). Since the 1970s, the Atlantic Ocean has warmed more than any other ocean basin (0.3ºC decade^{-1}; Figure 30-2a, WGI 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*, 2007b; a; *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 Multidecadal Oscillation (AMO) ([*Wang and Dong*, 2010]; WGI 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 35º–80ºN 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 1995–2060 (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 like 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 eastwards 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 (28.2.2.1, Nordic, and North Seas (Box 6-1; 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 North-east Atlantic have contracted towards the Arctic since 1958, and were 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], 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]. 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, 23.6.5).

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]; 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 tons.

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 decade^{-1} 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; *Beare et al.*, 2005].

In the North-east 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 15–50 km decade^{-1} 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 North-east 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 datasets (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 bought 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 30–50 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; *Tåning*, 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-yr instrumental record shows distinct warm phases of the Atlantic Multidecadal Oscillation (AMO) during approximately 1930–1965 and from 1995, and cool phases between approximately 1900–1930 and 1960–1995 (WGI 14.7.6). However, it is *virtually certain* that the enhanced warming in recent decades cannot be explained without external forcing (WGI 10.3.1.1.3). Understanding the changes in interdecadal 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 14.7.6, 11.3.2.4.1). 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 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 14.7.3), with periodicities in SST of both 15–25 y and 50–70 y [*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 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 three 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 1976–78, 1987–89 and 1998–99. These periods were associated with regime shifts in foraging fish that occurred in 1979–82, 1988–92 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; *Link et al.*, 2009b]. 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 towards more occurrences of the negative phase but credibility of projections remains uncertain (WGI 14.7.3). It is *about as likely as not* that the PDO will change its form or behavior in the future (WGI 14.7.3).

The Kuroshio-Oyashio Extension (KOE) in the North-west 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.*, 2013]. 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 39–71 km poleward from the 2000s to 2030s under SRES A1B [*Tian et al.*, 2012; *Jung et al.*, 2013]. 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 (SLP) in the North Pacific is *likely* influenced by anthropogenic forcing [*Gillett et al.*, 2003; *Gillett and Stott*, 2009] (WGI 10.3.3.4) and SLP 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]; 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 south-eastern Bering Sea shelf became earlier [*Wang et al.*, 2007a; *Wang et al.*, 2007b]. 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 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 m and ~100 m were attributed in approximately equal measure to anthropogenic and natural variations ([*Byrne et al.*, 2010]; WGI 3.8.2, 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 of smaller magnitude than those at northern hemisphere high latitudes [*Yoder et al.*, 1993]. The southern hemisphere HLSBS is broadly bounded by the sub-tropical front (STF) and the sub-Antarctic front. Associated with the STF 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 South-west Pacific (141–145°E) over the two 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 and Arctic 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]; 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 30–35% lower than those from sediment cores representing pre-industrial 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, show severe levels of shell dissolution consistent with the shoaling of the aragonite saturation horizon and indicate 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 decade^{-1} from 1950–2009, p-value $\leq$ 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]; WG1 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; *Ling et al.*, 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 0.35–1.17°C in the near-term (2010–2039) and by 1.70–4.84°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 (0.12-0.79°C) in the near-term, with slight cooling for some regions in the long-term (-0.16–1.46°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 (AR4 WGII 10.4.1, 12.4.7). A large number of publications since then has substantially extended such documentation and 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 (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].

HLSBS 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 to other upwelling systems (e.g., EBUE; 30.5.5). The associated upwelling is a consequence of the Earth’s rotation and easterly (westward) winds and currents, which drive water northwards and southwards 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 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 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 interdecadal variability (e.g., Interdecadal Pacific Oscillation (IPO); [*Power et al.*, 1999]; WGI 11.2.2, 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 1950–2009 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 El Niño-Southern Oscillation 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 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]. Sea level rise 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 induced 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 1997/8, 2002/3, 2004/5, and 2009/10 (WGI 14.4.2, 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 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≤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 affecting some 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 12.4.4.2). There is also *low agreement* between different CMIP5 GCMs on how ocean warming will affect ENSO, with no significant change to ENSO amplitude in half 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 due to the migration of fish stocks deriving from changing distribution of particular sea temperatures [*Lehodey et al.*, 2006; *Lehodey et al.*, 2008; *Cheung et al.*, 2010; *Lehodey et al.*, 2011; *Sumaila et al.*, 2011; *Bell et al.*, 2013b]. These projections suggest that fisheries within EUS will experience increased vulnerability due to elevated variability in space and time 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 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 km² 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 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 Red Sea, is the world's warmest sea, with both extreme negative and positive temperature excursions (annual temperature range of 12–35°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 AVHRR (NOAA) satellite data) of $0.6^{\circ}C\ decade^{-1}$ [*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 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 the Red Sea ecosystems to climate change (including ocean acidification). The Red Sea warmed by 0.74°C from 1982–2006 [*Belkin*, 2009], although trends in the average SST, however, are not significant from 1950–2009 ($p\text{-value}>0.05$, Table 30-1) due to a high degree of variability 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 decade^{-1}; 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*) ($p\text{-value}\leq 0.05$; [*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 1982-2006 [*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; *Silverman et al.*, 2009].

30.5.3.1.3. Black Sea

The temperature of the surface waters of the Black Sea increased by 0.96ºC from 1982–2006 [*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 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 for 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*, 2007b; a]; 1900–1998, [*Madsen and Højerslev*, 2009]). The salinity of the surface and near bottom waters of the Baltic Sea (e.g., Gdansk Basin, [*Aleksandrov et al.*, 2009]; central Baltic [*Fonselius and Valderrama*, 2003; *Möllmann et al.*, 2003] decreased from 1975–2000, due to changing rainfall and river run-off, and a reduction in the pulses of sea water (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 where 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 reduced 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; *Möllmann et al.*, 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℃ from 1957–2008 (supplementary material, *[Belkin, 2009]*), although analysis of data from 1950–2009 only detected a significant trend in summer temperature (0.11℃ decade^{-1}, p-value$\leq$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 (*high agreement, robust evidence*) (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 30-40 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 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 North-west 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 sea level rise [*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 14.4.2), then future sea level rise may be slightly suppressed due to atmospheric influences (*medium confidence*) [*Jordà et al.*, 2012]. As temperatures have increased, the Mediterranean has become more saline (+0.035–0.040 psu from 1950–2000, [*Rixen et al.*, 2005]) and the length of the thermal stratification period persisted 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 1950–2009 (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, decreases in 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 fast 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 ‘tropicalisation’ 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]; 23.6.5).

In addition to general patterns of warming, periods of extreme temperatures have had large-scale and negative consequences for Mediterranean marine ecosystems. Unprecedented mass mortality events, that affected at least 25 prominent invertebrate species, occurred during the summers of 1999, 2003, and 2006 across hundreds of kilometers of coastline in the North-west 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 (2–5 days: 2003, 2006) of high sea temperatures (27°C) or longer periods (30–40 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; *MerMex-Group*, 2011]. Recent re-analysis, however, has concluded that the pH of Mediterranean waters has decreased by 0.05–0.14 pH units since the pre-industrial 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 Gatusso*, 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 towards the vents, with a notable decline in calcifiers. [*Hall-Spencer et al.*, 2008]. Transplants of corals, mollusks, 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 ; *Rodolfo-Metalpa et al.*, 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 overexploitation, pollution, and enhanced run-off 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 0.93–1.24°C over 2010–2039 (Table SM30-4). Increases of 3.45–4.37°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 (23.3, WGI 12.4.) 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*) (23.2, [*IPCC*, 2012]). Similar projections to those outlined in 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 North-west Pacific, Indian, and Atlantic Oceans, encompassing: the Bohai/Yellow Sea, East China Sea, South China Sea, and South-east 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 wide-spread degradation of CBS ecosystems [*Burke et al.*, 2002; *Burke et al.*, 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 0.14–0.80°C from 1950–2009 (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 and 0.60°C, respectively, from 1950–2009 (Table 30-1). Rates of sea level rise vary from decreasing sea levels (-5 to -10 mm yr^{-1}) to low (2–3 mm yr^{-1}, Caribbean) to very high (10 mm yr^{-1}, South-east 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.

30.4.4.1.1. 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 North-west 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) River and Yangtze (Changjiang) River. Upwelling of the Kuroshio sub-surface waters provides abundant nutrients that support high levels of primary productivity [*Wong et al.*, 2000; *Wong et al.*, 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℃, 1955–2005) and in the Yellow Sea in summer (1.10℃, 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; *Tang et al.*, 2009; *Cai et al.*, 2011]) 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 in 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.*, 2013]. The frequency of harmful algal blooms (HAB) 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]. While 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 a large number 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 seagrasses.

The surface waters of the SCS have been warming steadily from 1945–1999 with the annual mean SST in the central SCS increasing by 0.92℃ (1950–2006, [*Cai et al.*, 2009]), a rate similar to that observed for the entire Indo-Pacific/SE Asian CBS from 1950–2009 (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 spatially non-homogeneous and varies in 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 the northern SCS, although this community has recently degraded due to increasing anthropogenic stresses [*Chen et al.*, 2009; *Qiu et al.*, 2010].

30.5.4.1.3. South-east Asian Seas

The South-east 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 1985–2006 [*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, $p\text{-value} \leq 0.05$, Table 30-1). The sea level is rising by up to 10 mm yr^{-1} in much of this region [*Church et al.*, 2004; *Church et al.*, 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 10.4.1). Although calcification rates of some key organisms (e.g., reef-building corals; [*Tanzil et al.*, 2009]) have slowed over the past two 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 overexploitation 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. Sea surface temperature increased by 0.18℃ and 0.26℃ in the Arabian Sea and Somali Current, respectively, from 1982–2006 (HadSST2, [*Rayner et al.*, 2003; *Belkin*, 2009]), which is consistent with the overall warming of the Western Indian Ocean portion of the CBS from 1950–2009 (0.60℃, Table 30-1). Salinity of surface waters in the Arabian Sea increased by 0.5–1.0% 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 100–200 m shallower than in pre-industrial 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 oxygen minimum zones (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] (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℃ over 1950–2009 (*high confidence*) (p-value$\leq$0.05; Table 30-1), increasing the frequency of positive thermal anomalies which have triggered mass coral bleaching and mortality events across the region over the past two decades (*high confidence*,[*Baker et al.*, 2008; *Nakamura et al.*, 2011](CC-HS). Trends in changes in SST and surface salinity vary with location along the East African coastline, with faster rates at higher latitudes (Figure 30-2). Periods of heat stress over the past 20 years have triggered mass coral bleaching and mortality on coral reef ecosystems within this region [*McClanahan et al.*, 2007; *McClanahan et al.*, 2009a; *McClanahan et al.*, 2009c; *McClanahan et al.*, 2009b; *Ateweberhan and McClanahan*, 2010; *Ateweberhan et al.*, 2011]. Steadily increasing sea temperatures have also produced anomalous growth rates in long-lived corals such as *Porites* (*high confidence*) [*McClanahan et al.*, 2009b]. Differences in the susceptibility of reef-building corals to stress from rising sea temperatures has also resulted in changes to the composition of coral (*high confidence*) (p-value$\leq$0.05; [*McClanahan et al.*, 2007]) and benthic fish communities (*high confidence*) (p-value$\leq$0.05; [*Graham et al.*, 2008; *Pratchett et al.*, 2011a]). These changes are *very likely* to alter species composition and potentially the productivity of coastal fisheries (*robust evidence, high agreement, high confidence*) [*Jury et al.*, 2010], although there may be a significant lag between the loss of coral communities and the subsequent changes in the abundance and community structure of fish (p-value$\leq$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 US, 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 and respectively from 1982–2006 (*very likely*) [*Belkin*, 2009]. Warming trends are not significant from 1950–2009 (Table 30-1), which may be partly due to spatial variability in warming patterns (30.5.3.1). The Caribbean region has experienced a sustained decrease in aragonite saturation state from 1996–2006 (*very likely*)

[*Gledhill et al.*, 2008]. Sea levels within the Gulf of Mexico and Caribbean Sea have increased at the rate of 2–3 mm y^{-1} from 1950–2000 [*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; *Justić et al.*, 2007; *Levin et al.*, 2009; *Rabalais et al.*, 2009; *Rabalais et al.*, 2010]. The increases in coastal pollution and fishing have potentially interacted with climate change to exacerbate impacts on marine ecosystems within this region (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 km^2 in the eastern Caribbean Sea (*high confidence*) [*Donner et al.*, 2007; *Eakin et al.*, 2010]. Although conditions in 2010 were milder than 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; *Harvell et al.*, 2002; *Harvell et al.*, 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 [*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 are 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; *Munday et al.*, 2009; *Dixson et al.*, 2010].

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 is 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 the 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 0.34–0.50°C over the near-term (2010–2039) and by 0.23–0.74°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 0.62–0.85°C over tne near-term and 2.44–3.32°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 the CBS regions (0.14–0.80°C from 1950–2009, Table 30-1), the projected changes of 2.44–3.32°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 sensitivity of coral reefs to relatively small increases in temperature over the past three 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; *Hoegh-Guldberg*, 1999; *Strong et al.*, 2011], 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 ($DHM\geq1$) that cause coral bleaching every 1–2 years by the mid to late part of this century (*robust evidence, high levels of agreement, very high confidence*) (Figure 30-4b, c; Figure 30-10, Figure 30-12, Figure SM30-3; [*van Hooidonk et al.*, 2013]). The frequency of mass mortality events ($DHM\geq5$, Figure 30-10a-c) also climbs towards events that occur every 1–2 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 reef-building coral stocks in the longer term [*Gardner et al.*, 2003; *Bruno and Selig*, 2007; *Baker et al.*, 2008].

It is *virtually certain* that composition of fisheries catches [*Graham et al.*, 2007; *Pratchett et al.*, 2011a] [*Pratchett et al.*, 2008; *Pratchett et al.*, 2011b] 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 degrades (*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 ecological goods and services [*Hoegh-Guldberg et al.*, 2009; *McLeod et al.*, 2010].

[INSERT FIGURE 30-10 HERE
Figure 30-10: Annual maximum proportions of reef pixels with Degree Heating Months [*Donner et al.*, 2007]; DHM)≥1 (used for projecting coral bleaching; [*Strong et al.*, 1997; *Strong et al.*, 2011]) and DHM≥5 (associated with bleaching across 100% of affected areas with significant mortality, [*Eakin et al.*, 2010] for the period 1870–2009 for each of the six coral regions (Figure 30-4d) using the 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 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 SST anomalies were derived using a 1985–2000 maximum monthly mean (MMM) 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.]

30.5.5. Eastern Boundary Upwelling Ecosystems

The Eastern Boundary Upwelling Ecosystems (EBUE) include the California, Peru/Humboldt, Canary/North-west 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 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 and 0.53°C (*very likely*) ($p\text{-value} \leq 0.05$, 1950-2009, Table 30-1), respectively, while no significant trend was detected in the sea surface temperatures of the Benguela ($p\text{-value} = 0.44$) and Humboldt Currents ($p\text{-value} = 0.21$) from 1950–2009 (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 (Box CC-OA, Box CC-UP, 6.2, 6.3.2, [*Kroeker et al.*, 2013], WGI 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 and extends downstream to the Atlantic Equatorial Current. 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 (15–20º N) and Morocco (21–26° N). Total fish catches, comprising mainly coastal pelagic sardines, sardinellas, anchovies, and mackerel, have fluctuated around 2 million tons yr^{-1} since the 1970s (http://www.seaaroundus.org/lme/27.aspx). Contrasting with the other EBUE, fishing productivity is modest, probably 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 1950–2009 indicating warming of 0.53°C from 1950–2009 ($p\text{-value} \leq 0.05$, 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 1967–2006, 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. PFEL, ICOADS, WASWind). This study presents no evidence for changes in upwelling intensity, with the exception of upwelling off North-west Spain, where winds are becoming slightly less favorable. Alteration of wind direction and strength influences upwelling and hence nutrient concentrations, howevernutrient 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 two decades [*Arístegui et al.*, 2009; *Demarcq*, 2009], in contrast to the nearby upwelling region off North-west Spain, where no significant trend was observed

[*Bode et al.*, 2011]. Satellite chlorophyll records (SeaWiFS, 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 tons yr^{-1} (http://www.seaaroundus.org/lme/29/1.aspx), before declines in the northern Benguela, due to overfishing and interdecadal environmental variability, resulted in a reduced catch of around 1 million tons 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 interdecadal 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 (0.35–0.55 ºC $decade^{-1}$ [*Rouault et al.*, 2010] or an increase for the whole Benguela region (0.24ºC, *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. Sea-level rise 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 $p\text{CO}_2$, respectively [*Santana-Casiano et al.*, 2009]), the consequences of changing upwelling intensity remain poorly explored with respect to ocean acidification. Finally, while 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 20–30 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 in response to environmental change e.g., due to predominately 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 ~23° of latitude from central Baja California, Mexico, to central British Columbia, Canada, linking the North Pacific Current (Westwind 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 from the California Current have been approximately 0.6 million tonnes yr^{-1} since 1950 (http://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 2–6 million seabirds from around 100 species [*Tyler et al.*, 1993]. Marine mammals are diverse and relatively abundant, including recovering populations of humpback whales, among others [*Barlow et al.*, 2008].

The average temperature of the California Current warmed by 0.73°C from 1950–2009 (p-value $\leq$0.05, Table 30-1) and by 0.14–0.80°C from 1985–2007 [*Demarcq*, 2009]. Like other EBUE, the California Current is characterized by large-scale interannual and interdecadal 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). Interdecadal 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 (23–40%), but an increase duration of individual events resulting in an increase of the overall magnitude of upwelling events from 1967–2010 (*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 1984–2006 throughout the California Current, with the largest relative decreases occurring below the thermocline (21% at 300 m). The hypoxic boundary layer (<60 $\mu mol\ kg^{-1}$) 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 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 pH and the carbonate chemistry 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*) (40–120 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 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 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 tons yr^{-1} (http://www.seaaroundus.org/lme/13/1.aspx), although decadal-scale variations range from 2.5–13 million tons yr^{-1}. While the 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 sub-tropical waters to and from the coast [*Alheit and Bakun*, 2010]. This variability does not appear to be altering 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; *Gutierrez 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 to 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 (0.22–0.93°C) occur within the four EBUEs in the near-term. Over 2010-2099, however, EBUE SSTs warm by 0.07–1.02ºC under RCP2.6, and 2.52–3.51ºC 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 B2 and A2 IPCC scenarios. Using an ensemble of 15 GCMs, southerly wind systems favoring upwelling increased along the sub-tropical 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 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. Sub-Tropical Gyres

Sub-Tropical 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 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., 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, 0.024, 0.032, 0.025, and 0.027°C yr^{-1} from 1998–2010, respectively ([*Signorini and McClain*, 2012]. This is consistent with increases observed from 1950–2009 (0.25–0.67ºC, 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 3.3.3.1), consistent with warmer sea temperatures and an intensification of the hydrological cycle [*Boyer*, 2005].

The North and South Pacific STGs 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 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$\leq$0.5; [*Signorini and McClain*, 2012]) over and above the inherent seasonal and interannual variability from 1998–2010 [*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; *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, 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 STG

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 14.3.1). These features are also strongly influenced by interannual to interdecadal 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 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°–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., Interdecadal Pacific Oscillation (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 (5.4, 29.3, Table 29.1). While 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 0.5–2.0% 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]. Increasing sea levels have also caused changes in seagrass and mangrove systems. *Gilman et al.* [2007] found a reduction in mangrove area with sea level rise, with the observed mean landward recession of three mangrove areas over four decades being 25, 64, and 72 mm yr^{-1}, 12–37 times faster than the observed rate of sea level rise. 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; *Stramma et al.*, 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 cold 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; *Lehodey et al.*, 2006]. Projections of big-eye tuna (*T. obesus*) distributions under SRES A2 show an improvement in spawning and feeding habitats to 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; *Lehodey et al.*, 2010b].

30.5.6.1.2. Indian Ocean STG

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 carbonate 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–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 hot 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 40–53% 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 1997–98 (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 the other major oceans of the world.

30.5.6.1.3. Atlantic Ocean STG

SST has increased within the two STG of the Atlantic Ocean over the last two 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 decade^{-1}, p-value≤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 10–30 years versus 60 years) emphasizes the importance of long-term patterns of variability in the North Atlantic region. Variability in SST at a period of about 60–80 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

Sea surface temperatures of the vast STG 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 0.17–0.56°C in the near-term (over 2010–2039) and between -0.03–0.90°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 0.45–0.91°C warmer in the near-term and 1.90–3.44°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 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 (~RCP6.0–8.5; Figure 1.5 [*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 10-20% and large fish catch by 19–29% 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 (Box CC-PP, Box 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 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 2.5, 14.2), although longer droughts are also expected in other STG (*medium* *confidence*). Changes in the hydrological cycle impact on 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 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, Figures 30.10 and Figure 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 1–2 years by 2100 (*virtually certain,* Box CC-CR, Figure 30-10, Figure 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 ([*Kleypas et al.*, 1999; *Hoegh-Guldberg et al.*, 2007; *Doney et al.*, 2009], *medium confidence*), reef metabolism and community calcification [*Dove et al.*, 2013], and other key ecological processes ([*Pörtner et al.*, 2001; *Pörtner et al.*, 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 (29.3.4 [*Waycott et al.*, 2007; *Waycott et al.*, 2011]. Many of the negative consequences for coral reefs, mangroves, and seagrass meadows are *likely* to have negative consequences for dependent coastal fisheries (through destruction of habitat) and tourism industries (*medium confidence*) ([*Bell et al.*, 2011b; *Pratchett et al.*, 2011a; *Pratchett et al.*, 2011b; *Bell et al.*, 2013a].

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; *Lehodey et al.*, 2010a; *Lehodey et al.*, 2011; *Lehodey et al.*, 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; *Bell et al.*, 2013a], 7.4.2, Table 29.2-29.3). 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 (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; *Stramma et al.*, 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 due to difficulty of access and scarcity of long-term, comprehensive observations [*Smith 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 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 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 surface of the Ocean, which represents a critical food supply to DS animals [*Smith and Kaufmann*, 1999; *Smith 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 (WGI 3.2, *very high confidence*), although smaller yet significant changes are occurring at depth. The DS environment is typically cold (~-0.5–3°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; *Levitus et al.*, 2009], with warming being minimal at mid-range depths (2000–3000 m), and increasing towards the sea floor in some sub-regions (e.g., Southern Ocean, WGI Chapter 3). For the deep Atlantic Ocean, the mean age of deep waters (mean time since last exposure to the atmosphere) is ~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 ~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 and Kaufmann*, 1999]. Most organic matter entering the DS is recycled by microbial systems at relatively shallow depths [*Buesseler et al.*, 2007], 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 oxygen minimum zones [*Stramma et al.*, 2008; *Stramma et al.*, 2010] and reducing food supply to the abyssal ocean.

Particulate organic carbon is exported from the surface to deeper layers of the Ocean (>500m) with an efficiency of between 20–50% [*Buesseler et al.*, 2007], much of it being recycled by microbes before it reaches 1000m [*Smith 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 due 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 3500m) already exist in waters under-saturated with respect to aragonite [*Lundsten et al.*, 2009]. While 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 6%, respectively [*Maier et al.*, 2009]), there is accumulating evidence 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, and decreasing O_2, carbonate chemistry and pH (*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 Chapter 30 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 greenhouse gas content. There is considerable confidence in both the detection (*very high confidence*) and attribution (*high confidence*) of mass coral bleaching and mortality (Figure 30-11b), given the well-developed understanding of environmental processes and physiological responses driving these events (Box CC-CR, 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 (18.3.3–4).

[INSERT FIGURE 30-11 HERE
Figure 30-11: Expert assessment of degree of confidence in detection and attribution of physical and chemical changes (a) and ecological changes (b) 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 18.3.3–4 and 18.6.]

30.6. Sectorial 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., 5.4.3, 7.3.2.4, 22.3.2.3) Oceans contribute provisioning (e.g., food, raw materials; see 30.6.4.1), regulating (e.g., gas exchange, nutrient recycling, carbon storage, climate regulation, water flux), supporting (e.g., habitat, genetic diversity) and cultural (e.g., recreational, cultural) services [*Millennium-Ecosystem-Assessment*, 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 (Figure 30-12), and the extent to which changes can be adapted to and/or risks mitigated (Table 30-3). Both short-term and longer-term adaptation is necessary to address impacts arising from warming, even under the lowest stabilization scenarios assessed.

[INSERT FIGURE 30-12 HERE
Figure 30-12: (a) Examples of projected impacts and vulnerabilities associated with climate change in Ocean sub-regions. (b) Examples of risks to fisheries from observed and projected impacts across Ocean sub-regions. Letters indicate level of confidence: (vL): Very low, (L): Low, (M): Medium, (H): High and (vH): Very high. Details of sub-regions are given in Table 30-1a and 30.1.1.

Sectorial approaches dominate resource 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., 30.5.3, 30.5.4, 5.3.4) [*Halpern et al.*, 2009; *Halpern et al.*, 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 greenhouse gas emissions is tackled [*West et al.*, 2009].

30.6.1. Natural Ecosystems

Adaptation in natural ecosystems may occur autonomously, such as shifts in species' composition and distributions [*Poloczanska et al.*, 2013] or engineered by human intervention, such as assisted dispersal (4.4, [*Hoegh-Guldberg et al.*, 2008]. Currently, adaptation strategies for marine ecosystems included reducing additional stressors (e.g. maintaining water quality, adapting fisheries management) and maintaining resilience ecosystems (e.g., Marine Protected Areas) and are moving towards 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 greenhouse gases to stabilize global temperature and ocean chemistry.

Approaches such as providing a formal valuation of ecological services from the Ocean have the potential to facilitate adaptation by underpinning more effective governance, regulation, and ocean policy while at the same time potentially improving the 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 due 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 and Corcoran*, 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*, 2004; *Lo Iacono et al.*, 2008]. The degradation of coastal habitats not only liberates much of the carbon associated with vegetation loss, but can 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 3–19% of the emissions generated by deforestation globally, with economic damages estimated to be US$6–42 billion annually. Similarly, *Luisetti et al.* [2013] estimate the carbon stock of seagrass and salt marshes in Europe, representing <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 (17.4).

30.6.2. Economic Sectors

30.6.2.1. Fisheries and Aquaculture

The Ocean provided 64% of the production (in tonnes) 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, and 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 stabilising 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 [*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 (7.2.1.2). The risks to ecosystems and fisheries vary from region to region (7.3.2.4). Dynamic bioclimatic envelope models under SRES A1B project potential increases in fisheries production high latitudes, and potential decreases at lower latitudes by the mid-21st century [*Cheung et al.*, 2010] (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*) (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 undersaturated with aragonite (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 (30.5.5.1.1[*Barton et al.*, 2012]). 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 (6.3.1.4, exposed through pens, cages and racks placed directly in the sea, utilization of seawater in land-based tanks or collection of wild spat) and, for mollusks particularly, changes in carbonate chemistry (6.3.2.4 [*Turley et al.*, 2011; *Barton et al.*, 2012]). 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 [*Bell et al.*, 2011a; *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 reduce the 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 fishing 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] (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, Columbia, 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 (7.3.2.6) [*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 (30.5.5). Technological advances and management change 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, 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 socio-economic 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 opportunists [*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 following examples.

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 over 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 (30.5.2, 30.5.5, 30.5.6), with potential to create ‘winners’ and ‘losers’ among island economies as catches of the trans-boundary tuna stocks change among and within their exclusive economic zones (EEZs;[*Bell et al.*, 2013b; *Bell et al.*, 2013a].

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.*, 2011; *Bell et al*., 2013a], 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. Provision of operational-level catch and effort data from all industrial fishing operations to improve models for projecting redistribution of tuna stocks and quotas under climate change [*Salinger et al*. 2013][*Nicol 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 sub-tropical Indian and Atlantic Oceans.

[INSERT TABLE 30-2 HERE
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 socio-economic 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 exceed costs in the short-term but accrue under longer-term climate change (modified from [*Bell et al.*, 2013b]).

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 (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 (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, Figure 30-10, Figure 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, 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 (FADs) close to shore [*Bell et al.*, 2011b; *Bell et al.*, 2013b; *Bell et al.*, 2013a], 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 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 the pelagic fishes 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], 30.6.5), and which has emphasized the importance of developing international collaboration and frameworks for decision making [*Miller et al.*, 2013](30.6.7, 15.4.3.3). 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 North-east Arctic cod, *Gadus morhua*, reached record-high abundance during 2012 and also reached its northernmost-recorded distribution (82°N)[*ICES*, 2012]. A further northward migration is impossible since 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 North-east 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 occupation time of the stocks 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 over 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]; 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 (10.6.1, 10.6.2). Tourism is susceptible to extreme events such as violent storms, long periods of drought, and/or extreme precipitation events (5.3.3, 10.6.1, [*IPCC*, 2012]). Sea level rise 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*) (5.3.3.2 10.6.1, [*IPCC*, 2012], Table SPM.1). 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 sea level rise [*Scott et al.*, 2012a].

Increasing sea temperatures (30.3.1.1) can change the 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 (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 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.*, 2001]. 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 towards higher latitudes with climate change, or from summer to cooler seasons [*Amelung et al.*, 2007] (10.6.2).

Options for adaptation by the marine tourism sector include: (1) identifying and responding to inundation risks with current infrastructure, and planning for projected sea level rise when building new tourism infrastructure (5.5, [*Scott et al.*, 2012a]); (2) promoting shoreline stability and natural barriers by preserving ecosystems such as mangroves, salt marshes and coral reefs (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 the negative consequences of climate change over those stemming from local stresses [*Scott et al.*, 2012b; *Scott et al.*, 2012a]. 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 [*Turton et al.*, 2010].

30.6.2.3. Shipping

International shipping accounts for >80% of world trade by volume [*UNCTAD*, 2009a; b] and ~3% of global CO_2 emissions from fuel combustion although CO_2 emissions are expected to increase 2-3 fold by 2050 [*Heitmann and Khalilian*, 2010], WGIII 8.1, 8.2). Changes in shipping routes [*Borgerson*, 2008], 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 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](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 change 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 8.5.2). Currently, 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), 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 (28.2.5, 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. Over 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 towards 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 the potential expansion of oil and mineral production in the Arctic is made in Chapter 28 (28.2.5, 28.2.6, 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

The 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], 5.4.3.5, 6.4.2.3, 11.2). The predominately negative impacts of disease for human communities are expected to be more serious in low-income areas such as South-east 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, sea level rise, and ocean circulation [*McMichael et al.*, 2006]). Climate change in the Ocean may influence the distribution of diseases like cholera (11.5.2.1), and the distribution and occurrence of harmful algal blooms (HAB). The frequency of cholera outbreaks induced by *Vibrio cholerae* and other enteric pathogens are correlated with sea surface temperatures, multidecadal 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 a link 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 Chapter 7 (7.5.5, 7.8.2, 7.10, 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] (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 Chapter 7 (7.4.2, 7.5.3, 7.8.2). There is an increasing trend in the renewable energy sector to offshore wind turbines (10.2.2). At present, there is *high uncertainty* about how changes in wind intensity and patterns, and extreme events, will impact the offshore wind energy sector. Given the design and engineering solutions available to combat climate change impacts (Table 10.1, Table 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], 12.6.1). 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 (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 (12.6.1-2) [*Allen and Bergin*, 2009; *Rahman*, 2012]. 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 (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 (30.6.2.3), naval operations in many countries result in significant greenhouse emissions (e.g., the US Navy emits around 2% of the national greenhouse gas 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, amongst 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 greenhouse gases (*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 ions, and inorganic nutrients 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 by the Ocean (*very likely*). Marine organisms are migrating at rapid rates towards higher latitudes, fisheries are transforming, and many organisms are shifting their reproductive and migratory activity in concert with the 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.

[INSERT TABLE 30-3 HERE
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 timeframe, 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.

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 (Table 30-3, Figure 30-12a-b, 30.5, Box CC-MB, 6.3.1 6.3.4, 7.3.2.4). Key risks involve changes in the distribution and abundance of key fishery species (Figure 30-12a 2, 4; Figure 30-12b *2*; *high confidence*) 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 (Table 30-3, 6.3.5, 6.4.1.1, 6.5.3, 7.3.2.4, 7.4.2, 29.5.3, 29.5.4). 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 basis (especially in CBS, STG, SES, HLSBS, and EUS ocean regions; 30.5, Box CC-CR, 6.2). 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 (Table 30-3, 30.5, Box CC-CR, 5.4.2.4, 6.3.2. 6.3.5, 6.4.1.3, 7.2.1.2, 29.3.1.2). 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 non-climate change related stresses (e.g., pollution, overfishing), although this strategy could have minimal impact if further increases in sea temperature occur (*high confidence*). The loss of these important coastal ecosystems is associated with the 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 sea level rise, as well is intensified precipitation events (*high confidence*) (5.4, Box CC-CR). Adaptation options in this case include engineered coastal defences, re-establishing coastal vegetation such as mangroves, protecting water supplies from salination, and developing strategies for coastal communities to withdraw to less vulnerable locations over time (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 (30.3.2.3, 30.5.7, 6.1.1.3, 6.3.3). This represents a key risk to ocean ecosystems (Figure 30-12a *6*, Figure 30-12b *3; medium confidence*). 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 (Figure 30-12a *7*, *high confidence*). 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 (Figure 30-12b *1, 3; medium confidence*), and consequently puts livelihoods at risk, particularly in EBUE (e.g., California and Humboldt Current ecosystems; 30.5.5), SES (e.g., Baltic and Black Seas; 30.5.3) and CBS (e.g., Gulf of Mexico, NE Indian Ocean; 30.5.4, 30.3.2.3). 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 1971–2010 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 and hence microbial activity within these coastal systems (30.5.4). Relocating fishing effort, and modifying procedures associated with industries like aquaculture, may offer some opportunity to adapt to these changes (*likely*). Declining O_2 concentrations is likely to have significant impacts on DS habitats, where organisms are relatively sensitive to environmental changes of this nature due to the very constant conditions under which they have evolved (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 regions; Table 30-3; 6.3.2, 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 broad range of important processes such as reproduction, navigation, and neural function in a broad range of marine organisms which show minor to major influences of ocean acidification on their biology (30.3.2.2, 6.3.2, 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* Figure 30-12b *1*; Box CC-UP, 6.2). 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 which 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 pre-industrial 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 is contributing to the decrease in O_2 along with the temperature related decreases in oxygen solubility (WGI 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 30-12a *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 (Box CC-PP, Box 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. While 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. While 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 (Table 30-3, 30.3.2.3, 30.5.5, 6.1.1.3). Changes in storm intensity can 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 (30.6.2, [*IPCC*, 2012].

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 (Table 30-3, [*IPCC*, 2012], 30.6.5, 12.4-12.6, 29.3).

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 which may influence the outcome of these two factors. Table 30-3 and Table 30-4 outlines 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 initiating 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].

[INSERT TABLE 30-4 HERE
Table 30-4: Ramifications, adaptation options and frameworks for decision-making for ocean regions. Symbols are as follows: T = sea temperature; UW = upwelling; OA = ocean acidification; NU = nutrient concentration; IC = ice cover; SS = storm strength; SLR = sea level rise (↑ = Increased; ↓ = decreased; italics = uncertain). Acronyms are: CBD (Convention on Biological Diversity); CTI (Coral Triangle Initiative); GEF (Global Environment Facility); IHO (International Hydrographic Organization); ILO (International Labor Organization); IOM (International Organization of Migration); ISPS (International Ship and Port Facility Security); MARPOL (International Convention for the Prevention of Pollution From Ships); PACC (Pacific Adaptation to Climate Change Project); PEMSEA (Partnerships in Environmental Management for the Seas of East Asia); RFMO (Regional Fisheries Management Organizations); SPREP (Secretariat of the Pacific Regional Environment Programme); UNCLOS (United Nations Convention on the Law of the Sea); UNHCR (United Nations High Commissioner for Refugees); UNSFSA (Straddling Fish Stocks Agreement); and WHO (World Health Organization).]

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 Convention. UNCLOS replaced earlier frameworks that were built around the 'freedom of the seas' concept and which 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 the 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 United Nations 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" (Annex one, 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 [*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, the 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 the 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 (www.globalpartnershipforoceans.org, March 28, 2012) 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 [*GPO*, 2012]. 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

While 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 still lag significantly behind terrestrial and atmospheric 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 from 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 (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 the atmospheric concentration of CO_2. International collaborations such as the Argo network of oceanographic floats are rapidly improving 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 needed for improving 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' (30.3.2; 30.5.2-30.5.4). 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 sea level rise (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 central important fisheries such as those found along the west coasts of Africa and the Americas.

30.7.3.3. Declining O_2 Concentrations

The declining level of O_2 in the Ocean is an emerging issue of major importance (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 regards 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 laboratories 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., mollusks) to ocean acidification represents an emerging issue, with a need for research to understand and develop strategies for industry to minimize the impacts. Understanding of how carbonate structures like 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 socio-economic ramifications of ocean acidification [*Turley et al.*, 2011; *Hilmi et al.*, 2013].

30.7.3.5. Net Primary Productivity

Oceanic phytoplankton are responsible for 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 (Box CC-PP; Box CC-UP). Changes in net primary productivity will resonate through food webs and ultimately affect fish production. Given the central role that primary producers and their associated ecological processes play in ocean ecosystem functioning, it is crucial that we improve our understanding of how net primary productivity is likely to vary at global and regional levels (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 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 towards higher latitudes or deeper waters, consistent with the expectation of a warming ocean. Our current understanding of which organisms and ecosystems are moving, and the ramifications for reorganisation 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 an 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 a 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 have 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, while 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 (30.6 Table 30-3, Table 30-4).

Frequently Asked Questions

FAQ 30.1: Can we reverse the climate change impacts on the ocean? *[to be inserted after Section 30.3.2]*
In less than 150 years, greenhouse gas 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 up to 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 greenhouse gas

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 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*).

FAQ30.2: Does slower warming mean less impact on plants and animals? *[to be inserted in Section 30.4]*
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 1–2°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 long-lived organisms such as corals and fish are concerned (*high confidence*).

FAQ30.3: How will marine primary productivity change? *[to be inserted after Section 30.5.2.2]*
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, Sub-tropical 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 to understand how these systems are changing as a result of climate change are a priority (*high agreement*).

FAQ30.4: Will climate change cause 'dead zones' in the oceans? *[to be inserted after Section 30.5.5.2]*
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 of the ocean (i.e., increasing layering or stratification) of the Ocean – which have different temperatures, densities, salinities – and 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).*

FAQ30.5: How can we use non-climate factors to manage climate change impacts on the oceans?
[to be inserted after Section 30.7.1]
Like most natural system, 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*).

Cross-Chapter Boxes

Box CC-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 (29.3.3.1).

Coral reefs are one of the most vulnerable marine ecosystems (*high confidence*; 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).

[INSERT FIGURE CR-1 HERE
Figure CR-1: A and B: the same coral community before and after a bleaching event in February 2002 at 5 m depth, Halfway Island, Great Barrier Reef. Coral cover at the time of bleaching was 95% bleached almost all of it severely bleached, resulting in mortality of 20.9% (Elvidge *et al*., 2004). Mortality was comparatively low due in part because these coral communities were able to shuffle their symbiont to more thermo-tolerant types (Berkelmans and van Oppen, 2006; Jones *et al*., 2008). C and 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 (panel D; median pH_T ~7.8), reefs are dominated by massive corals while corals with high morphological complexity are underrepresented compared with control sites (D; median pH ~8.0). Reef development ceases at pH_T values below 7.7. pH_T: pH on the total scale. E: temporal trend in coral cover for the whole Great Barrier Reef over the period 1985–2012 (N, number of reefs, mean ± 2 standard errors; 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).]

A wide range of climatic and non-climatic drivers affect corals and coral reefs and negative impacts have already been observed (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 6.3.1. for physiological details and 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; 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 1997-98 was unmatched in the period 1903 to 1999 (Lough, 2000). Ocean acidification reduces biodiversity (Figure CR-1C and D) and the calcification rate of corals (*high confidence*; 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*; 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 towards net dissolution (*medium confidence*; 5.4.2.4). Ocean warming and acidification have synergistic effects in several reef-builders (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*; 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 SE Asian regions (*very high confidence*, 1-2% per year for 1968-2004; Bruno and Selig, 2007). Similarly, the abundance of reef-building corals has decreased by over 80% on many Caribbean reefs (1977 to 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 RCP3-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*., 2013). 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 Moreno, 2003). Over 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 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 (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 Domestic Product 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*., 2012) 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*; 5.2.4.4, 30.5).

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Box CC-HS. Heat Stress and Heat Waves

[Lennart Olsson (Sweden), Dave Chadee (Trinidad and Tobago), Ove Hoegh-Guldberg (Australia), John Porter (Denmark), Hans-O. Pörtner (Germany), Kirk Smith (USA), Maria Isabel Travasso (Argentina), Petra Tschakert (USA)]

Heat waves are periods of abnormally and uncomfortably hot weather during which the risk of heat stress on people and ecosystems is high. The number and intensity of hot days have increased markedly in the last three decades (Coumou et al., 2013) (*high confidence*). According to WG I, it is *likely* that the occurrence of heat waves has more than doubled in some locations due to human influence and it is *virtually certain* that there will be more frequent hot extremes over most land areas in the latter half of the 21st century. Coumou et al. (2013) predicted that, under a medium warming scenario, the number of monthly heat records will be over 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 +10C or more, the habitability of large 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 they may act as triggers for tipping points (Hughes et al., 2013). Consequently, heat waves play an important role in several key risks noted in Chapter 19 and CC-KR.

Economy and Society [Ch 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). In LICs and MICs, 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 [Ch 11]

Morbidity and mortality due to heat stress is now common all over the world (Barriopedro *et al.*, 2011; Rahmstorf and Coumou, 2011; Nitschke et al., 2011; Diboulo et al., 2012; Hansen et al., 2012). 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, Jan C et al., 2008). Increase in ozone concentrations due to high temperatures affects health (Smith et al., 2010), leading to premature mortality, e.g. cardiopulmonary mortality (Smith et al., 2010). High temperatures are also associated with an increase in air-borne allergens acting as a trigger for respiratory illnesses such as asthma, allergic rhinitis, conjunctivitis, and dermatitis (Beggs, 2010).

Ecosystems [Ch 4, 5, 6, 30]

Tree mortality is increasing globally (Williams et al., 2012) and can be linked to climate impacts, especially heat and drought (Reichstein et al., 2013), even though attribution to climate change is difficult due 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), 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 (Hoegh-Guldberg et al., 2007), Figure 6-5, (*medium confidence*) [Ch 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*). [Ch 5, 6, 30.4-30.5, CC-CR, CC-MB]

Agriculture [Ch 7]

Excessive heat interacts with key physiological processes in crops. Negative yield impacts for all crops past +3C of local warming without adaptation, even with benefits of higher CO_2 and rainfall, are expected even in cool environments (Teixeira et al., 2011). 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 U.S., current economic losses due to heat stress of livestock are estimated at several billion USD annually (St-Pierre et al., 2003).

Box CC-HS References

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Box CC-MB. Observed Global Responses of Marine Biogeography, Abundance, and Phenology to Climate Change

[Elvira Poloczanska (Australia), Ove Hoegh-Guldberg (Australia), William Cheung (Canada), Hans O. Pörtner (Germany), Michael Burrows (UK)]

WGII AR4 presented detection and attribution of a global climate change fingerprint on natural systems (AR4, Ch 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 changes within the Ocean to climate change (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 (6.2, Table 6.7, 30.5) (*robust evidence, high agreement, high confidence*).

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 occurrence change in a biological parameter (including distribution, phenology, abundance, demography or community composition) and, if change occurs, 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) included data after 1990; and (3) observations spanned at least 19 years, to reduce bias resulting from biological responses to short-term-climate variability.

[INSERT FIGURE MB-1 HERE
Figure MB-1: 1735 observed responses to climate change from 208 single- and multi-species studies. Changes attributed to climate change (blue), inconsistent with climate change (red) and are equivocal (yellow). Each circle

represents the centre of a study area. Where points fall on land, it is because they are centroids of distribution that surround an island or peninsula. Pie charts show the proportions within regions bounded by red squares and in the Mediterranean; numbers indicate the total (consistent, opposite or equivocal) observations within each region. Note: 57% of the studies included were published since AR4 (from Poloczanska *et al.*, 2013).]

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 (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 decade^{-1} (mean ± SE) and summer events by 4.4 ± 1.1 days decade^{-1} (*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 (6.3.2, 30.5).

The distributions of benthic, pelagic and demersal species and communities have shifted by 10s to 1000s of km, although the range shifts have not been uniform across taxonomic groups or ocean regions (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 decade^{-1} and trailing edges contracted in a poleward direction at 15.8 ± 8.7 km decade^{-1} (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 (WG1 3.2).

[INSERT FIGURE MB-2 HERE
Figure MB-2. Rates of change in distribution (km decade^{-1}) for marine taxonomic groups, measured at the leading edges (red) and trailing edges (brown). Average distribution shifts calculated using all data, regardless of range location, are in black. Distribution rates 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 (from Poloczanska et al., 2013).]

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) (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, 30.5). Observed changes in species composition of catches from 1970–2006 that is partly attributed to long-term ocean warming suggest increasing dominance of warmer water species in sub-tropical and higher latitude regions, and reduction in abundance of sub-tropical species in equatorial waters (Cheung *et al*., 2013), with implications for fisheries (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 (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 to 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 1950s, have generally been able to track observed velocities (Fig MB-3), with phyto- and zooplankton distribution shifts vastly exceeding climate velocities, but with considerable variability within and among taxonomic groups (Poloczanska *et al.* 2013).

Biogeographic shifts are also be influenced by other factors such as nutrient and stratification changes, species' interactions, habitat availability and fishing (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 warming rates, hence climate velocities, increase (Fig 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 (4.3.2.5, Figure 4.6) For marine species, the observed rates of shift are generally much faster than those land 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 North-east Atlantic (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 species-specific thermal window widths 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. Changes in local and regional species richness, abundance, community composition, productivity, energy flows and invasion resistance result. Even among Antarctic stenotherms such differences related to mode of life, phylogeny and associated metabolic capacities exist (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 and facilitate adaptive management strategies effective in mitigating these impacts (6.3, 6.4).

[INSERT FIGURE MB-3 HERE
Figure MB-3. A. Rate of climate change for the Ocean (sea surface temperature (SST) °C); B. corresponding climate velocities for the Ocean and median velocity from land (adapted from Burrows et al., 2011); and C. observed rates of displacement of marine taxonomic groups over several decades until 2010. The thin dotted red arrows give an example of interpretation. Rates of climate change of 0.008 °C yr^{-1} correspond to ca. 2.4 km yr^{-1}median climate velocity in the Ocean. When compared to observed rates of displacement, many marine taxonomic groups have been able to track these velocities, except phyto- and zooplankton where rates of displacement greatly exceed climate velocity. 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 (Black dotted line) are derived from HadISST1.1 data set, all other rates are calculated based on the average of the CMIP5 climate model ensembles (Table S30-3) for the historical period and for the future based on the four RCP emissions scenarios. Data were smoothed using a 20-year sliding window. (B) Median climate velocity calculated from HadISST1.1 dataset over 1960–2010 using the methods of Burrows et al., 2011. The three axes represent estimated median climate velocities are representative of areas of slow velocities such as Pacific subtropical gyre (STG) system (Purple line), the global Ocean surface (excluding polar seas, Blue line), and areas of high velocities such as the Coral Triangle and North Sea (Orange line). Figure 30-3 shows climate velocities over the ocean surface calculated over 1960–2010. The Red line corresponds to the median rate over global land surface calculated using historical surface temperatures from the CMIP5 model ensemble (Table S30-3). (C) Rates of displacement for marine taxonomic groups estimated by Poloczanska et al. 2013 using published studies (Figure MB-2 Black data set). Note the displacement rates for phytoplankton exceed the axis, so values are given.]

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Box CC-OA. Ocean Acidification

[Jean-Pierre Gattuso (France), Peter Brewer (USA), Ove Hoegh-Guldberg (Australia), Joan A. Kleypas (USA), Hans-Otto Pörtner (Germany), Daniela Schmidt (UK)]

Anthropogenic ocean acidification and global warming share the same primary cause, which is the increase of atmospheric CO_2 (Figure OA-1A; WGI, 2.2.1). Eutrophication, loss of sea ice, upwelling and deposition of atmospheric nitrogen and sulphur all exacerbate ocean acidification locally (5.3.3.6, 6.1.1, 30.3.2.2).

[INSERT FIGURE OA-1 HERE
Figure OA-1: A: Overview of the chemical, biological, 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 CMIP5 climate model simulations from 1850 to 2100. Projections are shown for emission scenarios RCP2.6 (blue) and RCP8.5 (red) for the multi-model mean (solid lines) and range across the distribution of individual model simulations (shading). Black (grey shading) is the modelled 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. The number of CMIP5 models to calculate the multi-model mean is indicated for each time period/scenario (WGI AR5 Figure 6.28). C: Effect of near future acidification (seawater pH reduction of 0.5 unit or less) 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. * denotes a statistically significant effect.]

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 (Fig. 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 year 2100 based on representative concentration pathways (WGI, Figure 6.28) compared to preindustrial values range from a pH change of -0.14 unit with RCP 2.6 (421 ppm CO_2, +1 ºC, 22% reduction of carbonate ion concentration) to a pH change of -0.43 unit with RCP 8.5 (936 ppm CO_2, +3.7 ºC, 56% reduction of carbonate ion concentration). Projections of regional changes, especially in the highly complex coastal systems (5.3.3.6, 30.3.2.2), in polar regions (WGI 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 metaanalyses (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*; 5.4.2.2, 5.4.2.4, 6.3.2). A pattern of positive and negative impacts emerges (*high confidence*; Fig. 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*; 5.4.2.4, 6.3.5).

Growth and primary production are stimulated in seagrass and some phytoplankton (*high confidence*; 5.4.2.3, 6.3.2.2-3, 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 (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, 6.3.5.4-5). Some corals and temperate fishes experience disturbances to behavior, navigation and their ability to tell conspecifics from predators (6.3.2.4). However, there is no evidence for these effects to persist on evolutionary timescales in the few groups analyzed (6.3.2).

Some phytoplankton and mollusks displayed adaptation to ocean acidification in long-term experiments (*limited evidence, medium agreement*; 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*; 6.1.2).

Projections of ocean acidification effects at 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 (6.3.2.5, 6.3.5-6), 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; 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 (5.4.2.2).

Due 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 (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, 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, 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 due to the small number of studies available, particularly on the magnitude of the ecological and socioeconomic impacts (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 mollusks (6.4.1.1) would result in a reduction of US production of 3 to 13% according to the SRES A1FI emission scenario (*low confidence*). The global cost of production loss of mollusks could be over 100 billion USD 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 (6.4.1.1). The annual economic damage of ocean-acidification-induced coral reef loss by 2100 has been estimated, in 2009, to be 870 and 528 billion USD, respectively for the A1 and B2 SRES emission scenarios (*low confidence*; 6.4.1). Although this number is small compared to global 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 (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 (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 (6.4.2.1). Climate geoengineering techniques based on solar radiation management will not abate ocean acidification and could increase it under some circumstances (6.4.2.2). Geoengineering techniques to remove carbon dioxide from the atmosphere could directly address the problem but are very costly and may be limited by the lack of CO_2 storage capacity (6.4.2.2). Additionally, some ocean-based approaches, such as iron fertilization, would only re-locate ocean acidification from the upper ocean to the ocean interior, with potential ramifications on deep-water oxygen levels (6.4.2.2; 30.3.2.3 and 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 (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 (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 (6.4.2.1).

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Box CC-PP. Net Primary Production in the Ocean

[Philip W. Boyd (New Zealand), Svein Sundby (Norway), Hans-Otto Pörtner (Germany)]

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 (6.2.1, 6.3.1). Environmental drivers of NPP include light, nutrients, micronutrients, carbon dioxide, and temperature (Panel A). 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 (Panel A). Climate change is projected to strongly impact NPP through a multitude of ways that depend on the regional and local physical settings (WGI, Ch. 3), and on ecosystem structure and functioning (*medium confidence*, 6.3.4, 6.5.1). The influence of environmental drivers on NPP causes as much as a 10-fold variation in regional productivity: from <50 g C m^{-2} $year^{-1}$ in nutrient-poor subtropical waters and light-limited Arctic waters to >> 300 g C m^{-2} $year^{-1}$ in productive upwelling regions and highly eutrophic coastal regions (Panel B).

The oceans currently provide ~50 x 10^{15} g C $year^{-1}$, or about half of global NPP (Field *et al*. 1998). Global estimates of NPP are mainly obtained from satellite remote-sensing (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 NE Pacific (Hamme *et al*., 2010). Climate variability can drive pronounced changes in NPP (Chavez *et al*., 2011), such as during 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 x 10^{12} g C $year^{-1}$, for the period 1999 to 2005 - an annual reduction of ~0.4 % of global NPP. In contrast, a time-series of directly measured NPP between 1988 to 2007 by Saba *et al*. (2010) (i.e. *in situ* incubations using the radiotracer ^{14}C-bicarbonate) revealed an increase (2 % $year^{-1}$) 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-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 multidecadal, the available time-series of oceanic NPP measurements are also not of sufficient duration relative to the timescales of climate variability modes (up to 60-70 years for AMO, for example, 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 Ch. 6) projected an increase in NPP at high latitudes as a result of alleviation of light and temperature limitation of NPP particularly in Northern and Southern Hemisphere ‘subpolar gyre’ 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 macro-nutrients 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 photic 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 1860 under a high emission scenario (Polovina *et al*., 2011; SRES 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), 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 (i.e. provinces, see panel B) scales.

Earlier model projections had predicted changes in global NPP from a decrease of > 10% (Field *et al*., 1998; Boyd and Doney, 2002) to an increase of up to 8.1% under an intermediate scenario (SRES A1B, similar to RCP6.0) (Sarmiento *et al*., 2004; Schmittner *et al*., 2008). 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, for example 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 Tg C year^{-1} between 1998 and 2009. Other regional trends in NPP are reported in 30.5.1-6. 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 (A2 or RCP6.0 - 8.5, 6.3.4, 6.5.1, *medium confidence*), 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*).

[INSERT FIGURE PP-1 HERE
Figure PP-1: A) Environmental factors controlling Net Primary Production (NPP). NPP is mainly controlled by three basic processes: 1) Light conditions in the surface ocean, i.e. the photic zone where photosynthesis occurs, 2) upward flux of nutrients and micronutrients from underlying waters into the photic zone, 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 MODIS Aqua satellite climatology from 2003-2012; NPP was calculated with the Carbon-based Production 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. Figure courtesy of Toby Westberry (OSU) and Ivan Lima (WHOI), satellite data courtesy of NASA Ocean Biology Processing Group.]

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Box CC-UP. Uncertain Trends in Major Upwelling Ecosystems

[Salvador E. Lluch-Cota (Mexico), Ove Hoegh-Guldberg (Australia), David Karl (USA), Hans O. Pörtner (Germany), Svein Sundby (Norway), Jean-Pierre Gatusso (France)]

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 waters 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, 30.5.2, Figure 30.1A) and the Eastern Boundary Upwelling Ecosystems (EBUE, 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 S30.1).

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 3.2, 3.4.4, 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, due to the uncertainty in wind speed trends (WGI, 3.4.4). In the tropics, observations of reductions in trade winds over several decades contrast more recent evidence indicating their strengthening since the early 1990s (WGI, 9.4.1.3.4). Observations and modelling efforts in fact show diverging trends in coastal upwelling at the eastern boundaries of the Pacific and the Atlantic. Bakun (1990) proposed that the the difference in heat gaining rates 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 Boundary Upwelling Ecosystems (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 since these are the most biologically active systems in the ocean. 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, 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 the fisheries catch potential and impact aquaculture in coastal areas (5.4.3.3, 6.3.7, 30.5.1.1.2, 30.5.5.1.3, Barton *et al.*, 2012). Shifts in upwelling also coincide with an apparent increase in the frequency of submarine eruptions of methane and hydrogen sulphide 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), 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 (6.4.1, Chp 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 N2 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 upon pCO_2 of the upwelled water, and potentially increasingly impact the biota of Eastern Boundary Upwelling Ecosystems.

[INSERT FIGURE UP-1 HERE
Figure UP-1: Upper panel: Schematic hypothetic mechanism of increasing coastal wind-driven upwelling at eastern boundary systems, where differential warming rates between land and ocean results in increased land-ocean pressure gradients (1) that produce stronger alongshore winds (2) and offshore movement of surface water through Ekman transport (3), and increased upwelling of deep cold nutrient rich waters to replace it (4). Lower panel: 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 decreased coastal fisheries due to an augmented 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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Table 30-1: Regional changes in sea surface temperature (SST) over the period 1950–2009 using the Ocean regionalization specified in Figure 30-1a (for further detail of regions defined for analysis, see Figure SM30-1 and Table 30-2, column 1). A linear regression was fitted to the average of all 1×1 degree monthly SST data extracted from the 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 decade^{-1}), 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 6 decades) for each category. The p-values that exceed 0.05 plus the associated slope and change values have a gray background, denoting the lower statistical confidence in the slope being different from zero (no slope). Note, changes with higher p-values may still describe informative trends although the level of confidence is lower that the slope is different from zero.

	Regression slope			**Total change over 60 years**			**p value, slope different from zero**		
Sub-region	**°C/Decade Coolest Month**	**°C/Decade All months**	**°C/Decade Warmest Month**	**Change over 60 years (coolest month)**	**Change over 60 years (all months)**	**Change over 60 years (warmest month)**	**°C/Decade Coolest Month**	**°C/Decade All months**	**°C/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	**0.054**	0.073	**0.116**	**0.324**	0.438	**0.696**	**0.001**	0.15	**0.000**
South Atlantic	**0.087**	0.063	**0.097**	**0.522**	0.378	**0.582**	**0.000**	0.098	**0.000**
North Pacific (west)	0.052	0.071	0.013	0.312	0.426	0.078	0.52	0.403	0.462
North Pacific (east)	0.016	0.04	0.016	0.096	0.24	0.096	0.643	0.53	0.444
Total North Pacific	0.033	0.055	0.015	0.198	0.33	0.09	0.284	0.456	0.319
South Pacific (west)	**0.043**	0.017	0.044	**0.258**	0.102	0.264	**0.016**	0.652	0.147
South Pacific (east)	**0.047**	0.031	**0.052**	**0.282**	0.186	**0.312**	**0.000**	0.396	**0.003**
Total South Pacific	**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	**0.101**	**0.090**	**0.079**	**0.606**	**0.540**	**0.474**	**0.000**	**0.000**	**0.000**
Pacific Equatorial	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)											
	Western Atlantic	**0.137**	**0.123**	**0.127**	**0.822**	**0.738**	**0.762**	**0.000**	**0.000**	**0.000**	
	Caribbean/Gulf of Mexico	0.023	0.024	0.019	0.138	0.144	0.114	0.193	0.498	0.281	
	Western Indian Ocean	**0.097**	**0.100**	**0.096**	**0.582**	**0.600**	**0.576**	**0.000**	**0.000**	**0.000**	
	Eastern Indian Ocean	**0.099**	**0.092**	**0.080**	**0.594**	**0.552**	**0.480**	**0.000**	**0.000**	**0.000**	
	E Indian/SE Asia/W Pacific	**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. Sub-Tropical Gyres											
	Indian Ocean	**0.141**	**0.112**	**0.103**	**0.846**	**0.672**	**0.618**	**0.000**	**0.000**	**0.000**	
	North Atlantic	**0.042**	0.046	**0.029**	**0.252**	0.276	**0.174**	**0.048**	0.276	**0.038**	
	South Atlantic	**0.079**	**0.083**	**0.098**	**0.474**	**0.498**	**0.588**	**0.000**	**0.017**	**0.000**	
	North Pacific (west)	**0.065**	**0.071**	**0.059**	**0.390**	**0.426**	**0.354**	**0.000**	**0.018**	**0.000**	
	North Pacific (east)	0.008	0.042	**0.051**	0.048	0.252	**0.306**	0.617	0.133	**0.014**	
	Total North Pacific	**0.034**	0.055	**0.051**	**0.204**	0.33	**0.306**	**0.001**	0.053	**0.000**	
	South Pacific (west)	**0.060**	**0.076**	**0.092**	**0.360**	**0.456**	**0.552**	**0.002**	**0.000**	**0.000**	
	South Pacific (east)	**0.055**	0.056	**0.088**	**0.330**	0.336	**0.528**	**0.000**	0.058	**0.000**	
	Total South Pacific	**0.056**	**0.060**	**0.089**	**0.336**	**0.360**	**0.534**	**0.000**	**0.027**	**0.000**	

7. Coral Reef Provinces (Figure 30-3)											
	Caribbean/Gulf of Mexico	0.026	0.024	0.023	0.156	0.144	0.138	0.107	0.382	0.203	
	Coral Triangle & SE Asia	**0.137**	**0.131**	**0.098**	**0.822**	**0.786**	**0.588**	**0.000**	**0.000**	**0.000**	
	Eastern Indian Ocean	**0.081**	**0.097**	**0.116**	**0.486**	**0.582**	**0.696**	**0.000**	**0.000**	**0.000**	
	Western Indian Ocean	**0.091**	**0.100**	**0.102**	**0.546**	**0.600**	**0.612**	**0.000**	**0.000**	**0.000**	
	Eastern Pacific Ocean	**0.079**	**0.094**	**0.101**	**0.474**	**0.564**	**0.606**	**0.106**	**0.000**	**0.023**	
	Western Pacific Ocean	**0.072**	**0.073**	**0.073**	**0.432**	**0.438**	**0.438**	**0.000**	**0.000**	**0.000**	
8. Basin Scale											
	North Atlantic (combined)	**0.045**	0.061	**0.090**	**0.270**	0.366	**0.540**	**0.002**	0.198	**0.000**	
	South Atlantic (combined)	**0.076**	**0.074**	**0.101**	**0.456**	**0.444**	**0.606**	**0.000**	**0.041**	**0.000**	
	Atlantic Ocean Basin	**0.060**	**0.068**	**0.091**	**0.360**	**0.408**	**0.546**	**0.000**	**0.000**	**0.000**	
	Total North Pacific	**0.030**	0.052	**0.046**	**0.180**	0.312	**0.276**	**0.000**	0.248	**0.006**	
	Total South Pacific	**0.055**	0.048	**0.075**	**0.330**	0.288	**0.450**	**0.000**	0.115	**0.000**	
	Pacific Ocean Basin	**0.043**	**0.052**	**0.046**	**0.258**	**0.312**	**0.276**	**0.000**	**0.000**	**0.006**	
	Indian Ocean Basin	**0.130**	**0.108**	**0.106**	**0.780**	**0.648**	**0.636**	**0.000**	**0.000**	**0.000**	

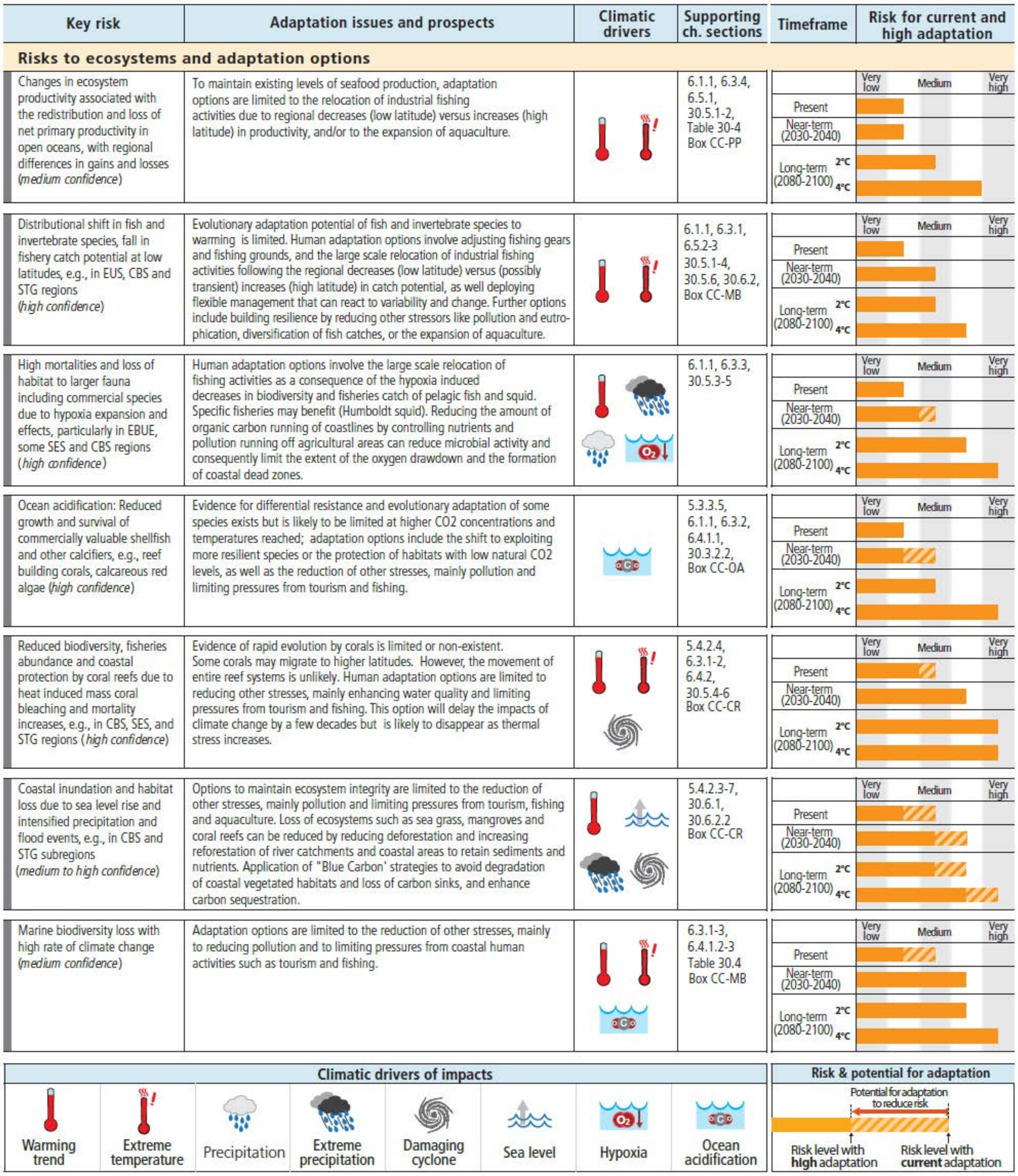
Table 30-2: Examples of priority adaptation options and supporting policies to assist Pacific Island countries and territories to minimize the threats of climate change to the socio-economic benefits derived from pelagic and coastal fisheries and aquaculture, and to maximize the opportunities. These 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 exceed costs in the short-term but accrue under longer-term climate change (modified from [*Bell et al.*, 2013b]).

Adaptation options	Supporting policies
Economic development	
• Full implementation of the vessel day scheme (VDS) to control fishing effort by the Parties to the Nauru Agreements (W-W). • Diversify sources of fish for canneries in the region and maintain 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 programmes to assist fleets to cope with oil price rises and minimise CO_2 emissions, and reduce costs of fishing further afield as tuna distributionsshift east (W-W). • Pan-Pacific tuna management through merger of the Western and Central Pacific Fisheries Commission (WCPFC) and Inter-American Tropical Tuna Commission to coordinate management measures across the tropical Pacific (L-W).	• Strengthen national capacity to administer the VDS. • 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 colonise 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 (FADs) close to the coast to improve access to fish for rural communities	• 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. • Minimise 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 FADs as part of national infrastructure for food security.

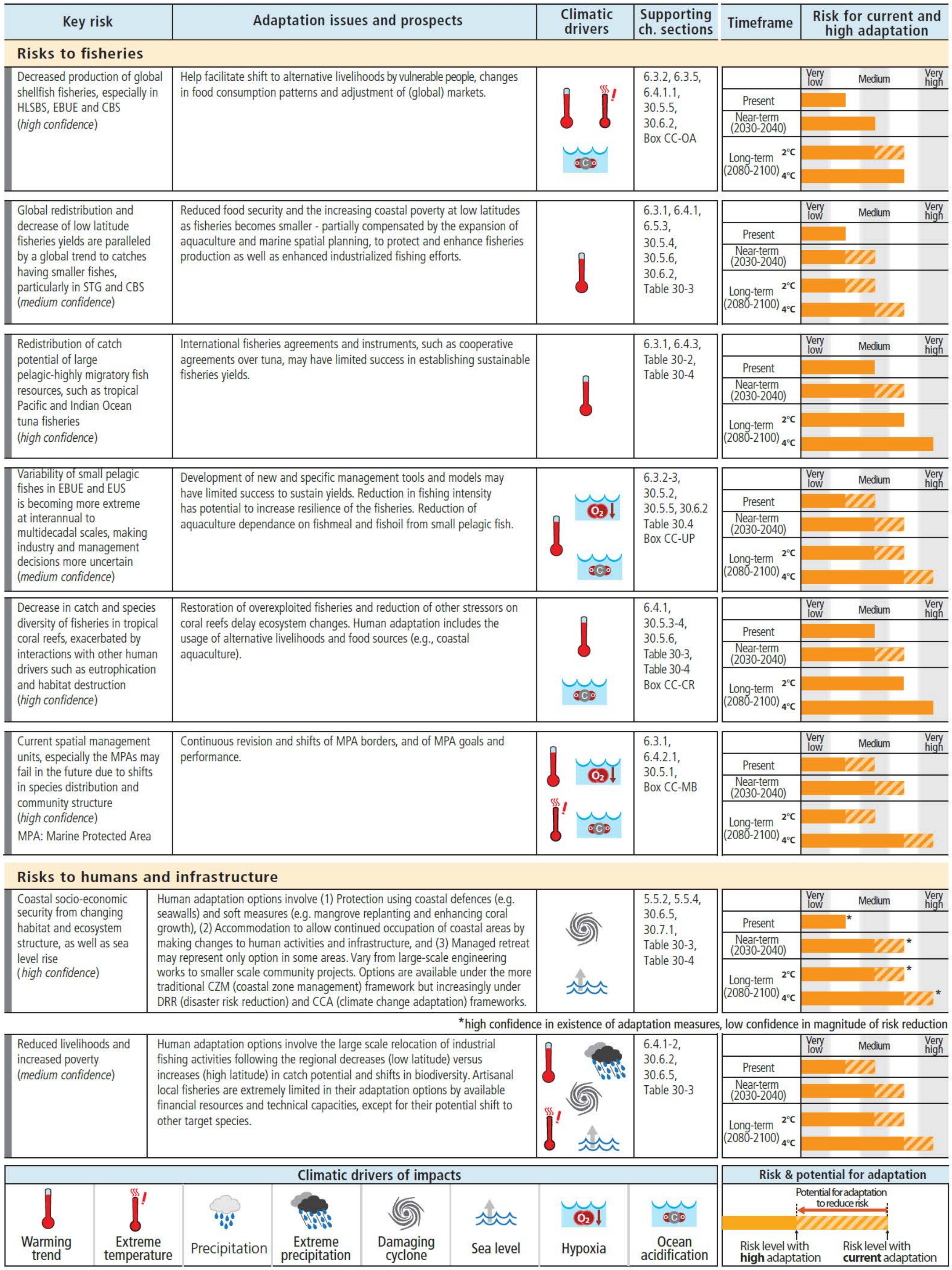
as human populations increase and demersal fish decline (W-W). • Develop coastal fisheries for small pelagic fish species, e.g. mackerel, anchovies, pilchards, sardines and scads (W-W?). • Promote simple post-harvest methods, such as traditional smoking, salting and drying, to extend the shelf life of fish when abundant catches are landed (W-W).	
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 (PNA) are Palau, Federated States of Micronesia, Papua New Guinea, Solomon Islands, Marshall Islands, Nauru, Kiribati, and Tuvalu.

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 timeframe, 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.

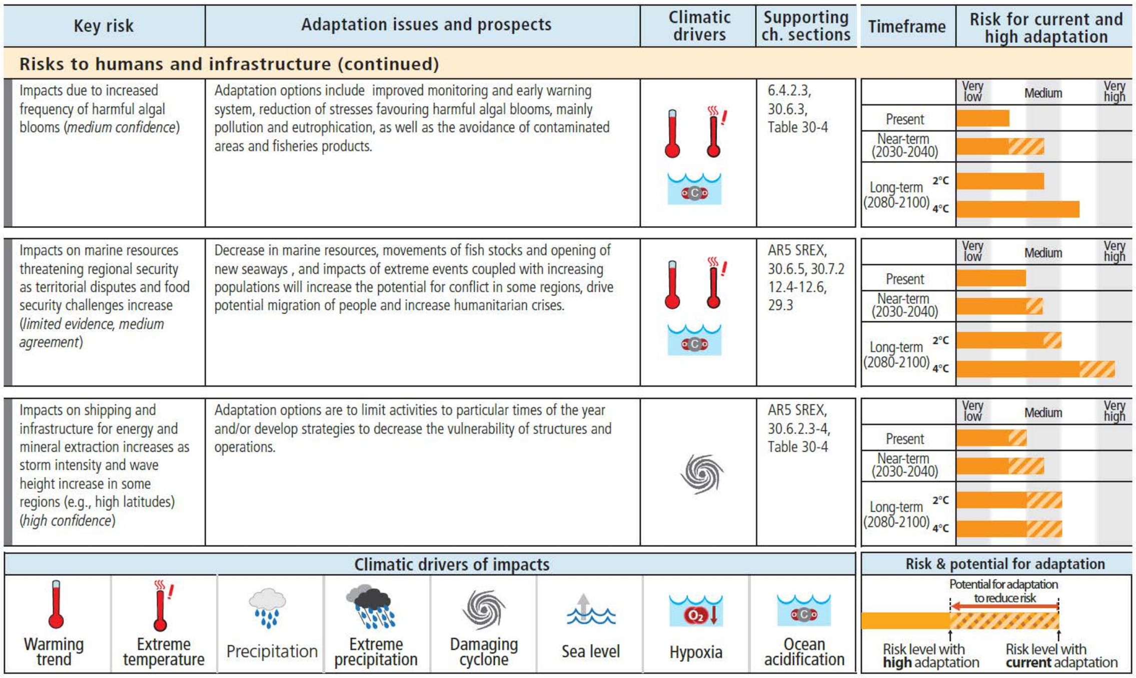


Key risk	Adaptation issues and prospects	Climatic drivers	Supporting ch. sections
Risks to ecosystems and adaptation options			
Changes in ecosystem productivity associated with the redistribution and loss of net primary productivity in open oceans, with regional differences in gains and losses (*medium confidence*)	To maintain existing levels of seafood production, adaptation options are limited to the relocation of industrial fishing activities due to regional decreases (low latitude) versus increases (high latitude) in productivity, and/or to the expansion of aquaculture.	🌡 🌡❗	6.1.1, 6.3.4, 6.5.1, 30.5.1-2, Table 30-4 Box CC-PP
Distributional shift in fish and invertebrate species, fall in fishery catch potential at low latitudes, e.g., in EUS, CBS and STG regions (*high confidence*)	Evolutionary adaptation potential of fish and invertebrate species to warming is limited. Human adaptation options involve adjusting fishing gears and fishing grounds, and the large scale relocation of industrial fishing activities following the regional decreases (low latitude) versus (possibly transient) increases (high latitude) in catch potential, as well deploying flexible management that can react to variability and change. Further options include building resilience by reducing other stressors like pollution and eutrophication, diversification of fish catches, or the expansion of aquaculture.	🌡 🌡❗	6.1.1, 6.3.1, 6.5.2-3 30.5.1-4, 30.5.6, 30.6.2, Box CC-MB
High mortalities and loss of habitat to larger fauna including commercial species due to hypoxia expansion and effects, particularly in EBUE, some SES and CBS regions (*high confidence*)	Human adaptation options involve the large scale relocation of fishing activities as a consequence of the hypoxia induced decreases in biodiversity and fisheries catch of pelagic fish and squid. Specific fisheries may benefit (Humboldt squid). Reducing the amount of organic carbon running 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.	🌡 ⛈ 🌧 O_2↓	6.1.1, 6.3.3, 30.5.3-5
Ocean acidification: Reduced growth and survival of commercially valuable shellfish and other calcifiers, e.g., reef building corals, calcareous red algae (*high confidence*)	Evidence for differential resistance and evolutionary adaptation of some species exists but is likely to be limited at higher CO2 concentrations and temperatures reached; adaptation options include the shift to exploiting more resilient species or the protection of habitats with low natural CO2 levels, as well as the reduction of other stresses, mainly pollution and limiting pressures from tourism and fishing.	OCO	5.3.3.5, 6.1.1, 6.3.2, 6.4.1.1, 30.3.2.2, Box CC-OA
Reduced biodiversity, fisheries abundance and coastal protection by coral reefs due to heat induced mass coral bleaching and mortality increases, e.g., in CBS, SES, and STG regions (*high confidence*)	Evidence of rapid evolution by corals is limited or non-existent. Some corals may migrate to higher latitudes. However, the movement of entire reef systems is unlikely. 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.	🌡 🌡❗ 🌀	5.4.2.4, 6.3.1-2, 6.4.2, 30.5.4-6 Box CC-CR
Coastal inundation and habitat loss due to sea level rise and intensified precipitation and flood events, e.g., in CBS and STG subregions (*medium to high confidence*)	Options to maintain ecosystem integrity are limited to the reduction of other stresses, mainly pollution and limiting pressures from tourism, fishing and aquaculture. Loss of ecosystems such as sea grass, mangroves and coral reefs can be reduced by reducing deforestation and increasing reforestation of river catchments and coastal areas to retain sediments and nutrients. Application of "Blue Carbon' strategies to avoid degradation of coastal vegetated habitats and loss of carbon sinks, and enhance carbon sequestration.	🌡 🌊↑ ⛈ 🌀	5.4.2.3-7, 30.6.1, 30.6.2.2 Box CC-CR
Marine biodiversity loss with high rate of climate change (*medium confidence*)	Adaptation options are limited to the reduction of other stresses, mainly to reducing pollution and to limiting pressures from coastal human activities such as tourism and fishing.	🌡 🌡❗ OCO	6.3.1-3, 6.4.1.2-3 Table 30.4 Box CC-MB



Key risk	Adaptation issues and prospects	Climatic drivers	Supporting ch. sections	Timeframe	Risk for current and high adaptation
Risks to fisheries					
Decreased production of global shellfish fisheries, especially in HLSBS, EBUE and CBS (*high confidence*)	Help facilitate shift to alternative livelihoods by vulnerable people, changes in food consumption patterns and adjustment of (global) markets.	Warming trend Extreme temperature Ocean acidification	6.3.2, 6.3.5, 6.4.1.1, 30.5.5, 30.6.2, Box CC-OA	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
Global redistribution and decrease of low latitude fisheries yields are paralleled by a global trend to catches having smaller fishes, particularly in STG and CBS (*medium confidence*)	Reduced food security and the increasing coastal poverty at low latitudes as fisheries becomes smaller - partially compensated by the expansion of aquaculture and marine spatial planning, to protect and enhance fisheries production as well as enhanced industrialized fishing efforts.	Warming trend	6.3.1, 6.4.1, 6.5.3, 30.5.4, 30.5.6, 30.6.2, Table 30-3	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
Redistribution of catch potential of large pelagic-highly migratory fish resources, such as tropical Pacific and Indian Ocean tuna fisheries (*high confidence*)	International fisheries agreements and instruments, such as cooperative agreements over tuna, may have limited success in establishing sustainable fisheries yields.	Warming trend	6.3.1, 6.4.3, Table 30-2, Table 30-4	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
Variability of small pelagic fishes in EBUE and EUS is becoming more extreme at interannual to multidecadal scales, making industry and management decisions more uncertain (*medium confidence*)	Development of new and specific management tools and models may have limited success to sustain yields. Reduction in fishing intensity has potential to increase resilience of the fisheries. Reduction of aquaculture dependance on fishmeal and fishoil from small pelagic fish.	Warming trend Hypoxia Ocean acidification	6.3.2-3, 30.5.2, 30.5.5, 30.6.2 Table 30.4 Box CC-UP	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
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*)	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).	Warming trend Ocean acidification	6.4.1, 30.5.3-4, 30.5.6, Table 30-3, Table 30-4 Box CC-CR	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
Current spatial management units, especially the MPAs may fail in the future due to shifts in species distribution and community structure (*high confidence*) MPA: Marine Protected Area	Continuous revision and shifts of MPA borders, and of MPA goals and performance.	Warming trend Hypoxia Extreme temperature Ocean acidification	6.3.1, 6.4.2.1, 30.5.1, Box CC-MB	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high
Risks to humans and infrastructure					
Coastal socio-economic security from changing habitat and ecosystem structure, as well as sea level rise (*high confidence*)	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 may represent only option in some areas. 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.	Damaging cyclone Sea level	5.5.2, 5.5.4, 30.6.5, 30.7.1, Table 30-3, Table 30-4	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high *
Reduced livelihoods and increased poverty (*medium confidence*)	Human adaptation options involve the large scale relocation 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 target species.	Warming trend Extreme precipitation Damaging cyclone Extreme temperature Sea level	6.4.1-2, 30.6.2, 30.6.5, Table 30-3	Present Near-term (2030-2040) Long-term (2080-2100) 2°C 4°C	Very low / Medium / Very high

*high confidence in existence of adaptation measures, low confidence in magnitude of risk reduction



Key risk	Adaptation issues and prospects	Climatic drivers	Supporting ch. sections	Timeframe	Risk for current and high adaptation
Risks to humans and infrastructure (continued)					
Impacts due to increased frequency of harmful algal blooms (*medium confidence*)	Adaptation options include improved monitoring and early warning system, reduction of stresses favouring harmful algal blooms, mainly pollution and eutrophication, as well as the avoidance of contaminated areas and fisheries products.	Warming trend Extreme temperature Ocean acidification	6.4.2.3, 30.6.3, Table 30-4	Present Near-term (2030-2040) Long-term (2080-2100) 2°C Long-term (2080-2100) 4°C	Very low – Medium – Very high
Impacts on marine resources threatening regional security as territorial disputes and food security challenges increase (*limited evidence, medium agreement*)	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.	Warming trend Extreme temperature Ocean acidification	AR5 SREX, 30.6.5, 30.7.2 12.4-12.6, 29.3	Present Near-term (2030-2040) Long-term (2080-2100) 2°C Long-term (2080-2100) 4°C	Very low – Medium – Very high
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*)	Adaptation options are to limit activities to particular times of the year and/or develop strategies to decrease the vulnerability of structures and operations.	Damaging cyclone	AR5 SREX, 30.6.2.3-4, Table 30-4	Present Near-term (2030-2040) Long-term (2080-2100) 2°C Long-term (2080-2100) 4°C	Very low – Medium – Very high

Climatic drivers of impacts

Warming trend	Extreme temperature	Precipitation	Extreme precipitation	Damaging cyclone	Sea level	Hypoxia	Ocean acidification

Risk & potential for adaptation

Table 30-4: Ramifications, adaptation options and frameworks for decision-making for ocean regions. Symbols are as follows: T = sea temperature; UW = upwelling; OA = ocean acidification; NU = nutrient concentration; IC = ice cover; SS = storm strength; SLR = sea level rise (↑ = Increased; ↓ = decreased; italics = uncertain). Acronyms are: CBD (Convention on Biological Diversity); CTI (Coral Triangle Initiative); GEF (Global Environment Facility); IHO (International Hydrographic Organization); ILO (International Labor Organization); IOM (International Organization of Migration); ISPS (International Ship and Port Facility Security); MARPOL (International Convention for the Prevention of Pollution From Ships); PACC (Pacific Adaptation to Climate Change Project); PEMSEA (Partnerships in Environmental Management for the Seas of East Asia); RFMO (Regional Fisheries Management Organizations); SPREP (Secretariat of the Pacific Regional Environment Programme); UNCLOS (United Nations Convention on the Law of the Sea); UNHCR (United Nations High Commissioner for Refugees); UNSFSA (Straddling Fish Stocks Agreement); and WHO (World Health Organization).

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 increase 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', provide social and economic incentives to fisheries and fishersfor adaptation,.	LOSC, PEMSEA, CTI, RFMO agreements, UNSFSA,	[*Tsamenyi and Hanich*, 2012] [*Bell et al.*, 2011; *Bell et al.*, 2013a] 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 as 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* initiatives.	PEMSEA, CTI, PACC, MARPOL, UNHCR, CBD, IOM, GEF, ILO	[*Bell et al.*, 2013a]; 5.4.3, 6.3.1-2, 12.4, 30.5.2-4, 29.3.1, 29.3.3, 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], 5.5.4.1, 6.4.1–2, 10.6, 30.6.2.2,

Primary driver(s)	**Biophysical change projected**	**Key risks and vulnerabilities**	**Ramifications**	**Adaptation options**	**Policy frameworks and initiatives (examples)**	**Key References and Chapter sections**
		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*).	Increased monitoring and education surrounding key risks (e.g., ciguatera); develop alternate fisheries and income for periods when disease incidence increases, develop or update health response plans.	National policy strategies as well as and regional cooperation needed	[*Llewellyn*, 2010] 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 criminal activities.	UNCLOS, PEMSEA, CTI, ISPS, IMO, Bali Process on Transnational Crime ASEAN Mutual legal Assistance treaty and bilateral extradition and MLA agreements	[*Kaye*, 2012; *Rahman*, 2012] 12.4-6,29.3.3, 29.6.2, 30.6.5
↑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*)	Increased international cooperation and improve understanding of regime changes; early-detection monitoring of physical and biological variables and regional seasonal forecasting; include related uncertainties into fisheries management; social and economic incentives for industry.	UNCLOS, CBD, RFMO agreements, UNSFSA	7.4.2, 6.5, 30.5, 30.6.2.1, Box CC-MB,
		Increased in abundance, growing season and distributional extent ofpests 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; ; technological advances to deal with pest and fouling organisms; increase vigilance and control related to biosecurity.	IMO, BWM, Anti Fouling Convention	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 pathogens 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, WHO, IHOs, and national governments.	[*Myers and Patz*, 2009]; 6.4.3, 10.8.2, 11.7, 29.3.3, 30.6.3, Box CC-MB

Primary driver(s)	**Biophysical change projected**	**Key risks and vulnerabilities**	**Ramifications**	**Adaptation options**	**Policy frameworks and initiatives (examples)**	**Key References and Chapter sections**
↑T, ↑NU, ↑*OA*	Increased incidence of harmful algal blooms (HABs, *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, WHO, MARPOL	[*Llewellyn*, 2010], 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 nutrients 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, SPREP	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, Developearly-warning systems and update emergency response plans to extreme events.	IMO	[*IPCC*, 2012], 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, increase 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 consider SLR in planning and decision-making; increase understanding of the issues through education.	UNICEF, IHOs, and national governments.	[*Warner*, 2012] 5.5, 12.4.1, 29.5.1, 30.3.1.2, 30.6.5
		Inundation of coastal aquifers reduces water supplies and decreased coastal agricultural productivity (*high confidence*).	Reduced food and water security leads to increased coastal poverty, reduced food security, and migration (*very likely*).	Assist communities to find alternatives for food and water, or assist in relocation of populations and agriculture from vulnerable areas.	UNICEF, IHOs, and national governments.	[*Warner*, 2012], 5.4.3, 12.4.1, 29.3.2, 30.3.1.2

Primary driver(s)	**Biophysical change projected**	**Key risks and vulnerabilities**	**Ramifications**	**Adaptation options**	**Policy frameworks and initiatives (examples)**	**Key References and Chapter sections**
↑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 EEZ and other currently defined maritime jurisdiction limits.	UNCLOS	[*IPCC*, 2012; *Schofield and Arsana*, 2012; *Warner and Schofield*, 2012] 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*).	International agreements need to be sorted.	UNCLOS	Chapter 28

*Blue Carbon initiatives include conservation and restoration of mangroves, saltmarsh and seagrass beds as carbon sinks (30.6.1).

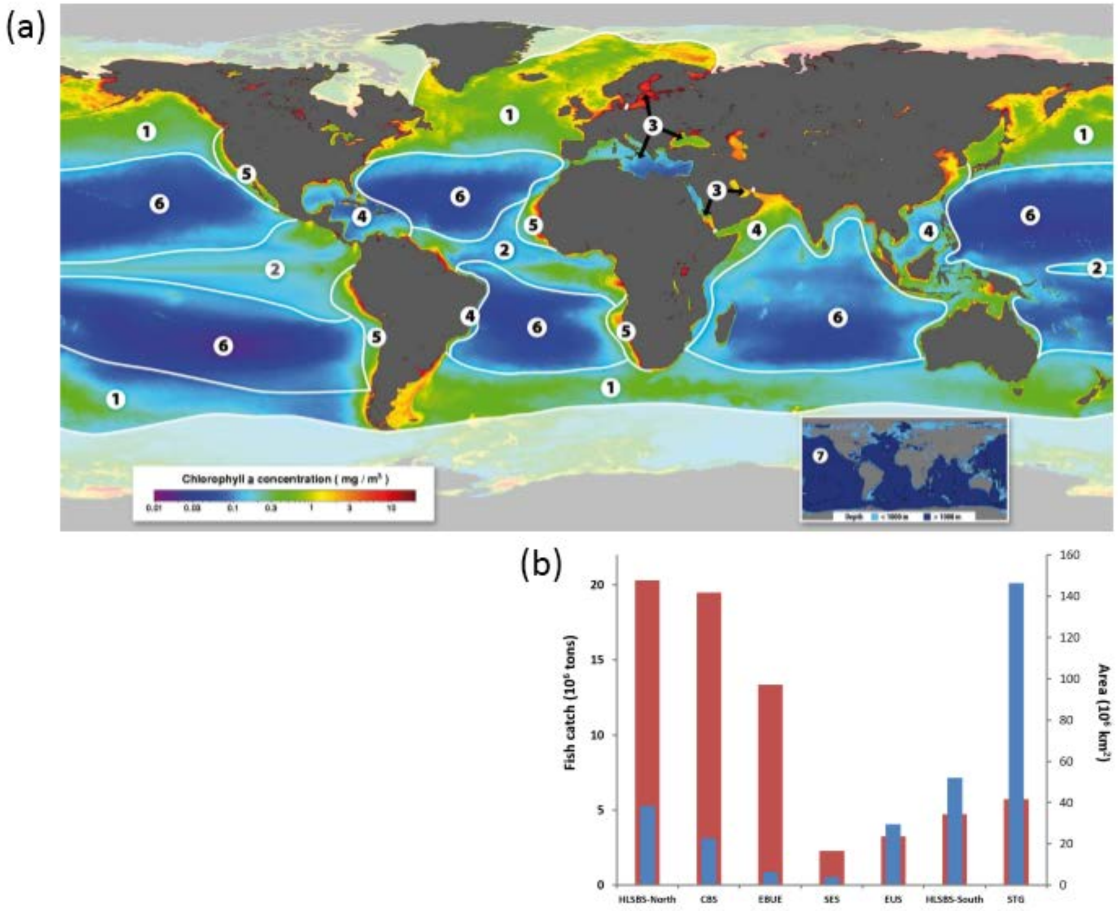


Figure 30-1: (a) Separation of the world's non-polar oceans into seven major sub-regions (excluding the polar oceans, which are considered in Chapter 28). The chlorophyll-*a* signal measured by SeaWiFS and averaged over the period from Sep 4, 1997 to 30 Nov 2010 (NASA) is provides a proxy for differences 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). Numbers refer to: 1 = High Latitude Spring Bloom Systems (HLSBS); 2 = Equatorial Upwelling Systems (EUS); 3 = Semi-Enclosed Seas (SES); 4 = Coastal Boundary Systems (CBS); 5 = Eastern Boundary Upwelling Ecosystems (EBUE); 6 = Sub-Tropical Gyres (STG); and 7 = DS (>1000 m). (b) Relationship between fish catch and area for each ocean sub-region is shown in (a). Red columns: average fish catch (as millions tons yr^{-1}) for the period 1970–2006. Blue columns: area (millions km^{2}). The four left-hand columns (sub-regions HLSBS-North, CBS, EBUE, and SES) cover 20 % of the world oceans' area and deliver 80% of the world's fish catches. The values for the percent area of the Ocean, primary productivity, and fishery catch for the major sub-regions are listed in Table SM30-1.
[Illustration to be redrawn to conform to IPCC publication specifications.]

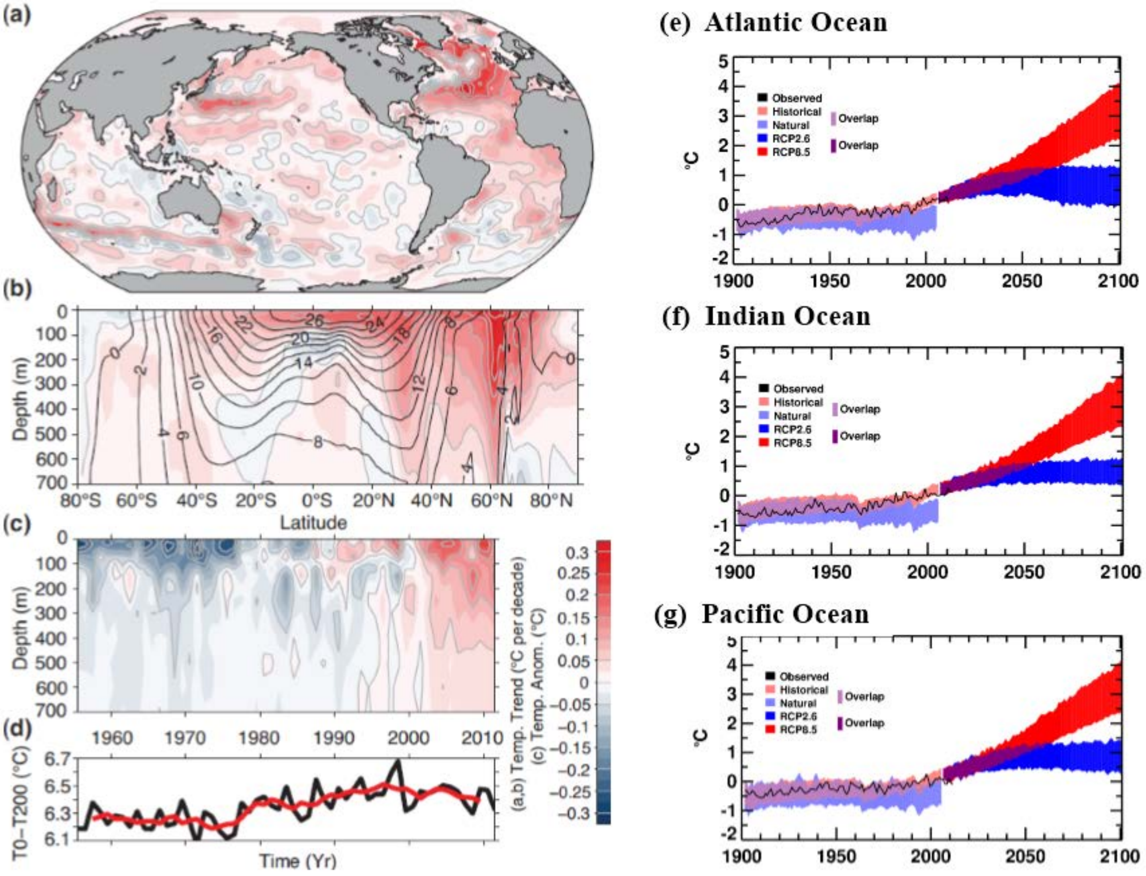


Figure 30-2: (a) Depth-averaged 0–700 m temperature trend for 1971–2010 (longitudinal versus latitude, colors and gray contours in °C per decade). (b) Zonally averaged temperature trends (latitude versus depth, colors and gray contours in °C per decade) for 1971–2010, with zonally averaged mean temperature over plotted (black contours in °C). (c) Globally-averaged temperature anomaly (Time versus depth, colors and grey contours in °C) relative to the 1971–2010 mean. (d) Globally-averaged temperature difference between the Ocean surface and 200 m depth (Black: annual values; red: five year running mean). Panels (a)–(d) from WGI Figure 3.1. (e)–(g) Observed and simulated variations in past and projected future annual average SST over three ocean basins (excluding regions within 300 km of the coast). The black line shows estimates from HadISST1.1 observational measurements. 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), and the Representative Concentration Pathways: Dark Blue: RCP2.6; Light Blue: RCP4.5; Green: RCP6.0, and Red: 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 Box 21-2.

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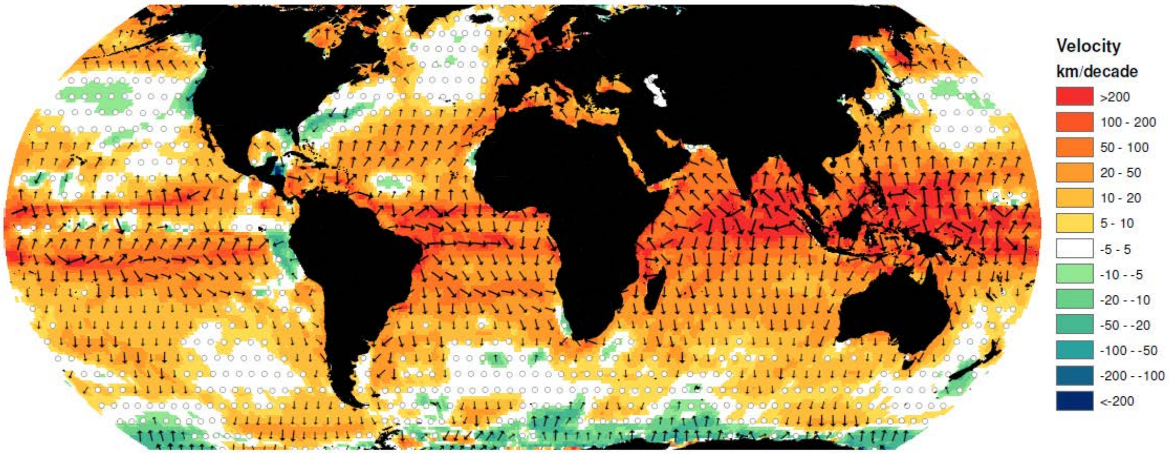

Figure 30-3: Velocity at which sea surface temperature (SST) isotherms shifted (km decade^{-1}) over the period 1960–2009 calculated using HaDISST1.1, with arrows indicating the direction and magnitude of shifts. Velocity of climate change is obtained by dividing the temperature trend in °C decade^{-1} by the local spatial gradient °C km^{-1}. The direction of movement of SST is denoted by the direction of the spatial gradient and the sign of the temperature trend: towards locally cooler areas with a local warming trend or towards locally warmer areas where temperatures are cooling. Adapted from [*Burrows et al.*, 2011].

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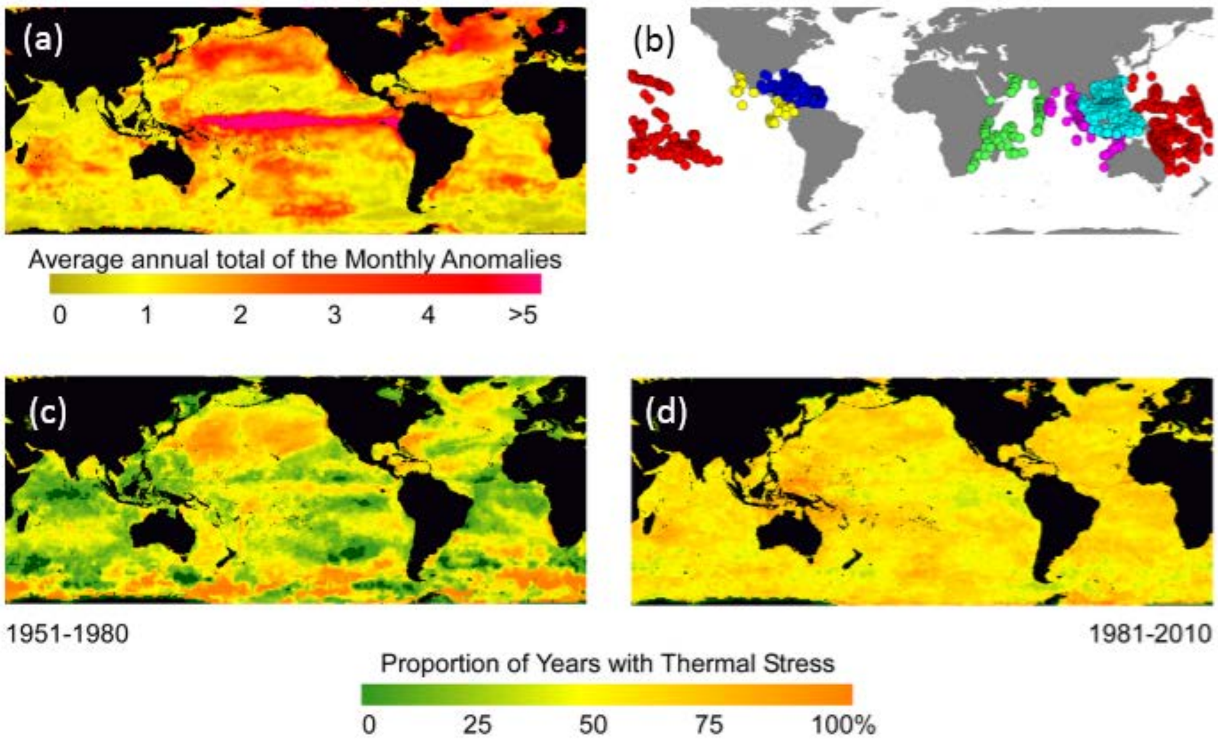

Figure 30-4: Recent changes in thermal stress calculated using HadISST1.1 data. A monthly climatology was created by averaging the HadISST monthly SST values over the period 1985–2000 to create twelve averages, one for each month of the year. The Maximum Monthly Mean (MMM) climatology was created by selecting the hottest month for each pixel. Anomalies were then created by subtracting this value from each SST value, but only allowing values to be recorded if they were greater than zero [*Donner et al.*, 2007]. Two measures of the change in thermal stress were calculated as a result: (a) The total thermal stress for the period 1981–2010, calculated by summing all monthly thermal anomalies for each grid cell. (b) The location of coral reef grid cells used in Table 30-1 and for comparison to regional heat stress here. 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 & Gulf of Mexico; Green – Western Indian Ocean; Pink – Eastern Indian Ocean; and Light Blue – Coral Triangle & SE Asia. (c) Proportion of years with thermal stress, which is defined as any year that has a thermal anomaly, for the periods 1951–1980 and (d) 1981–2010. **[Illustration to be redrawn to conform to IPCC publication specifications.]**

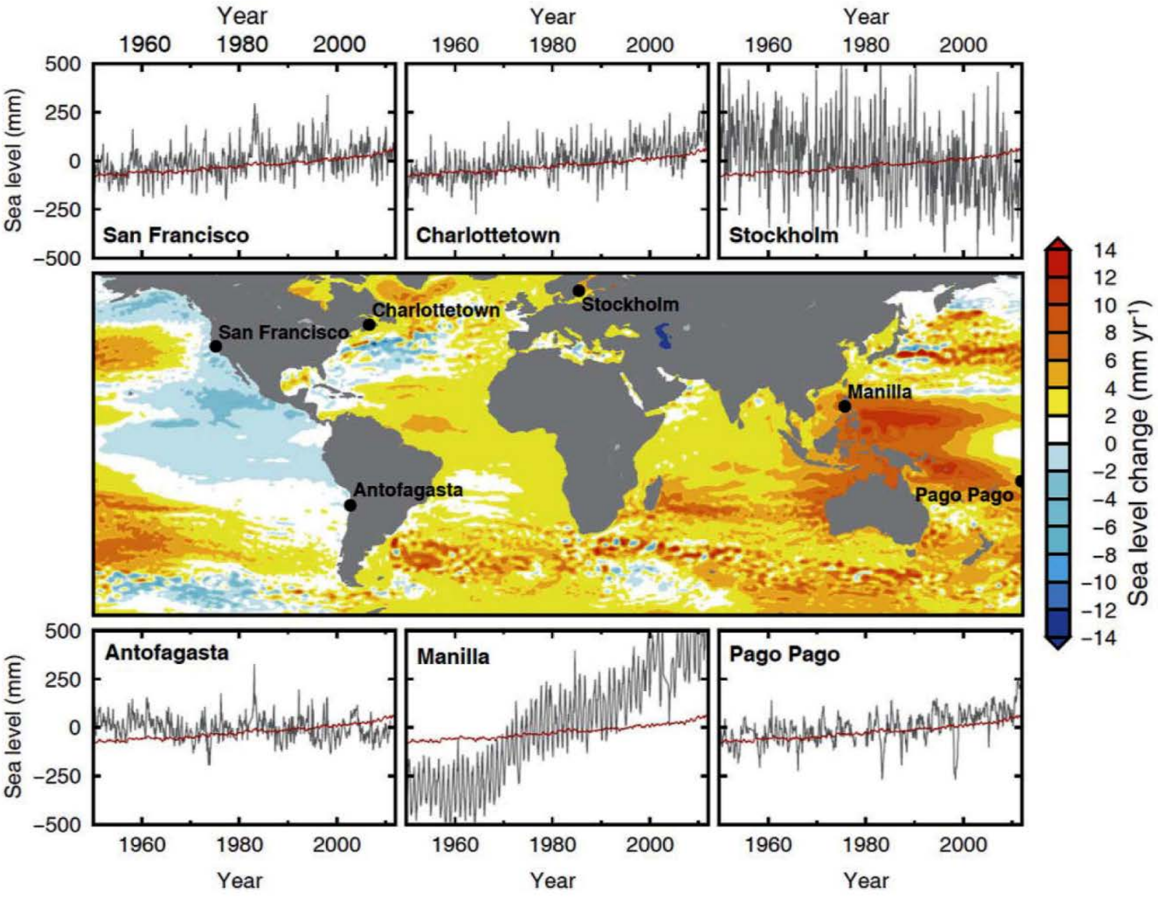

Figure 30-5. Map of the rate of change in sea surface height (geocentric sea level) for the period 1993–2012 derived from satellite altimetry. Also shown are relative sea level changes (gray lines) from selected tide gauge stations for the period 1950–2012. For comparison, an estimate of global mean 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 FAQ 13.1, Figure 1).

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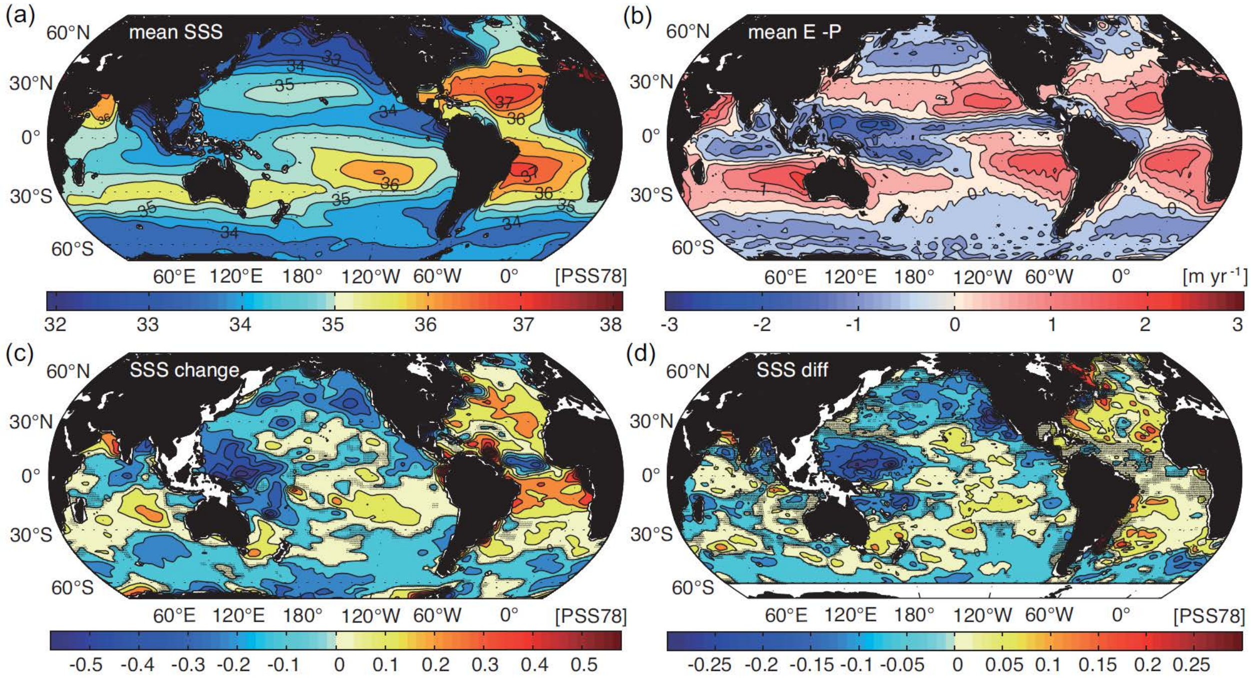

Figure 30-6: (a) The 1955–2005 climatological-mean sea surface salinity [*Antonov et al.*, 2010] color contoured at 0.5 PSS78 intervals (black lines). (b) Annual mean evaporation-precipitation averaged over the period 1950–2000 (NCEP) color contoured at 0.5 m yr–1 intervals (black lines). (c) The 58-year (2008 minus 1950) sea surface salinity change derived from the linear trend (PSS78), with seasonal and 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) 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 Figure 3.4 in WGI.
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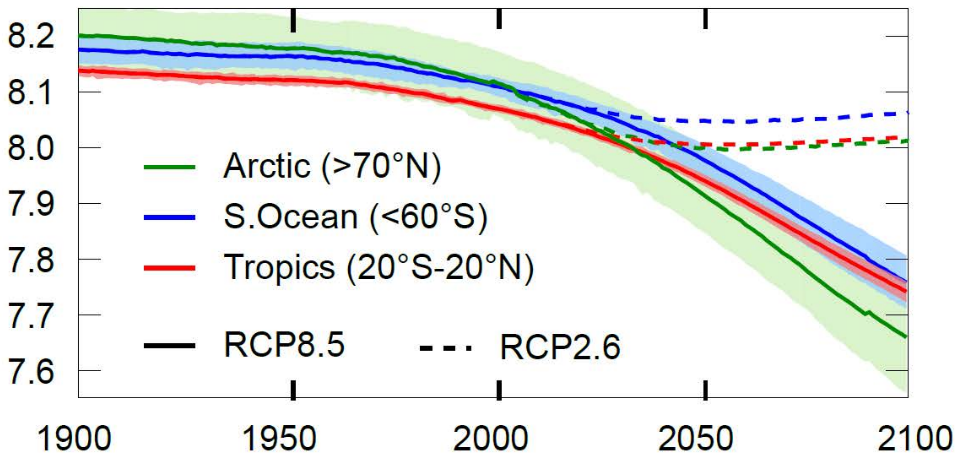
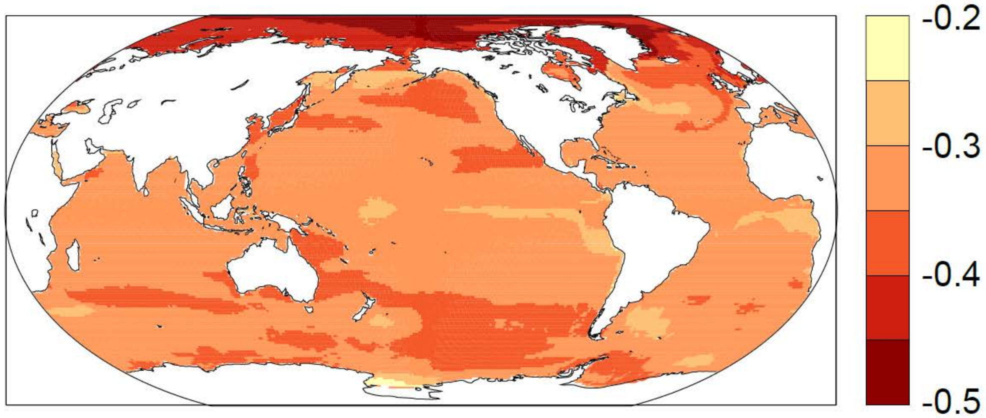


a. Surface pH



b. Surface pH in 2090s (RCP8.5, changed from 1990s)

Figure 30-7: Projected ocean acidification from 11 CMIP5 Earth System models under RCP8.5 (other 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 (filled), 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 1850–2100. 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 4, e.g., 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 Figure 6.28 in WGI.

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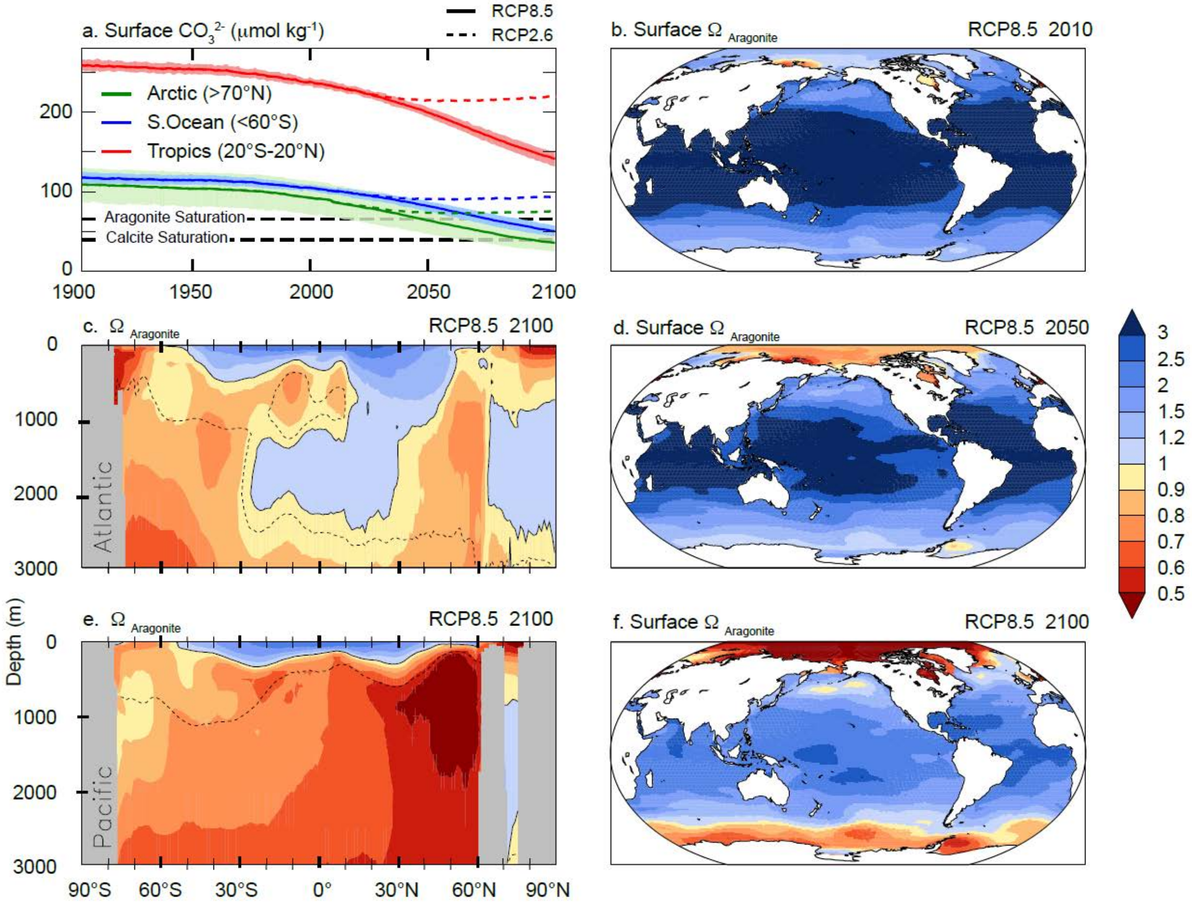

Figure 30-8: Projected aragonite saturation state from 11 CMIP5 Earth System models under RCP8.5 scenario: (a) time series of surface carbonate ion concentration shown as the mean (solid line) and range of models (filled), given as area weighted averages over the Arctic Ocean (green), the tropical oceans (red), and the Southern Ocean (blue); maps of the median model's surface ΩA in (b) 2010, (d) 2050, and (f) 2100; and zonal mean sections (latitude versus depth) of ΩA 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. Reprinted from Figure 6.29 in WGI.
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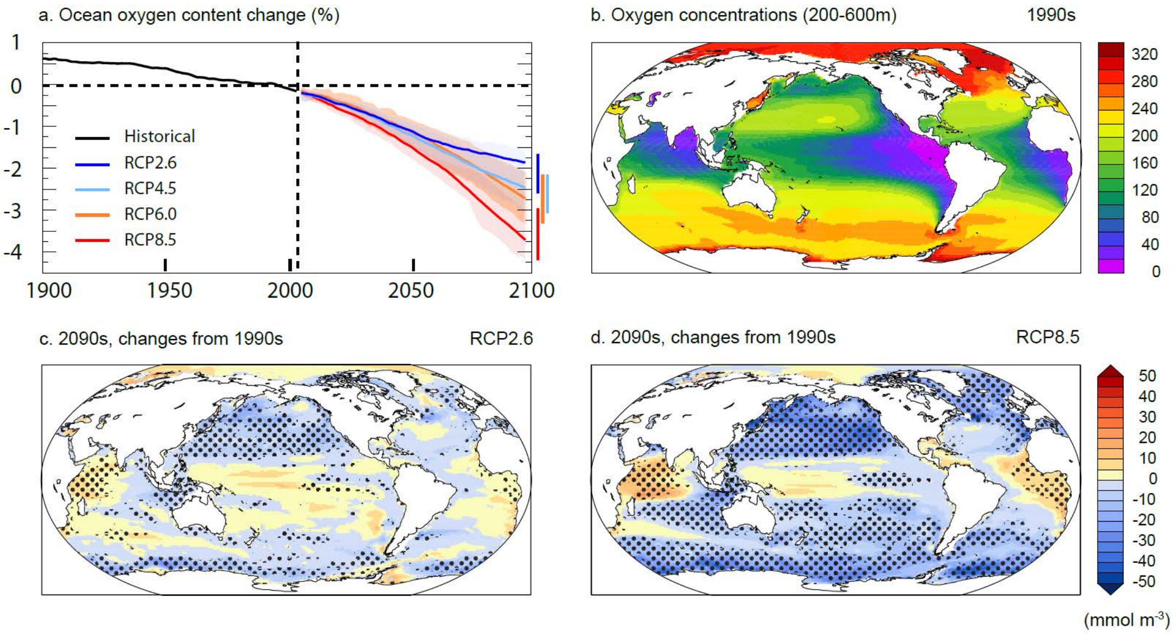

Figure 30-9: (a) Simulated changes in dissolved O_2 (mean and model range as shading) relative to 1990s for RCP2.6, RCP4.5, RCP6.0, and RCP8.5. (b) Multi-model mean dissolved O_2 (μmol m^{-3}) in the main thermocline (200–600 m depth average) for the 1990s, and changes in the 2090s relative to 1990s 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 inter-comparison using the SRES-A2 scenario and have been applied to CMIP5 models here. Models used: CESM1-BGC, GFDL-ESM2G, GFDL-ESM2M, HadGEM2-ES, IPSL-CM5A-LR, IPSL-CM5A-MR, MPI-ESM-LR, MPI-ESM-MR, NorESM1. Figure originally presented in WGI Figure 6.30 in WGI.
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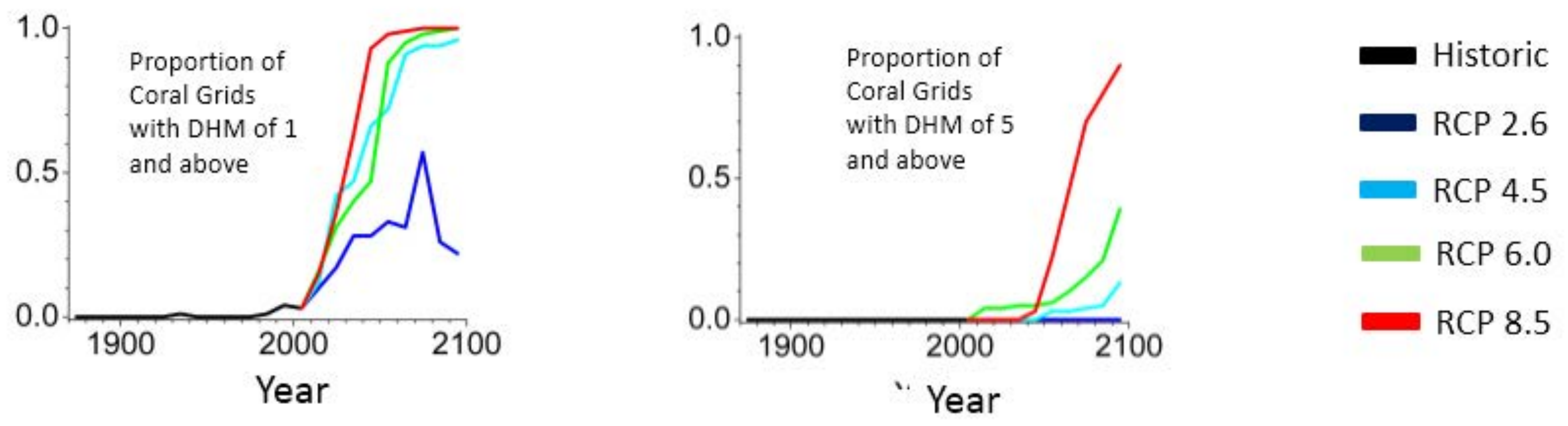
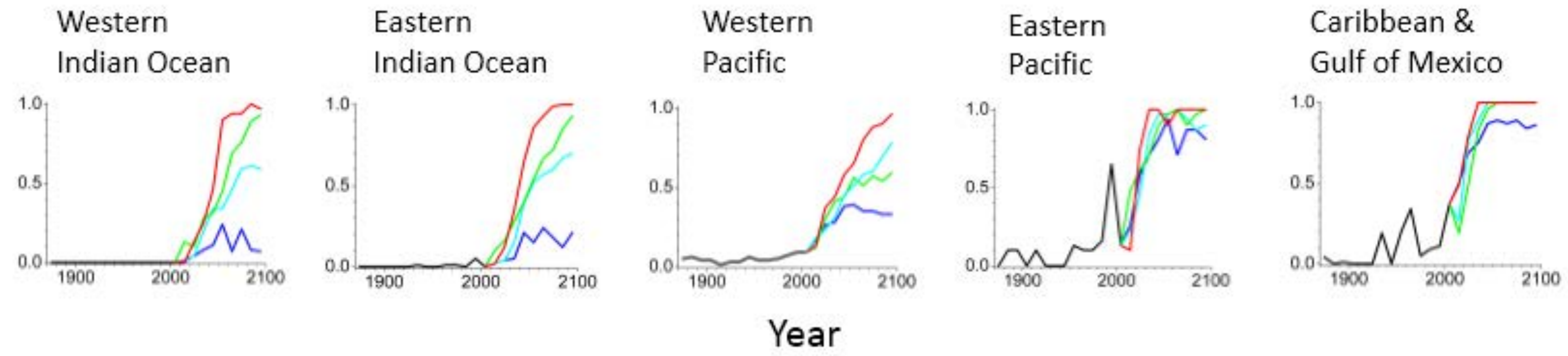
(a) Coral Triangle and SE Asia



(b) Mass Coral Bleaching: DHM > 1



(c) Mass Coral Mortality: DHM > 5

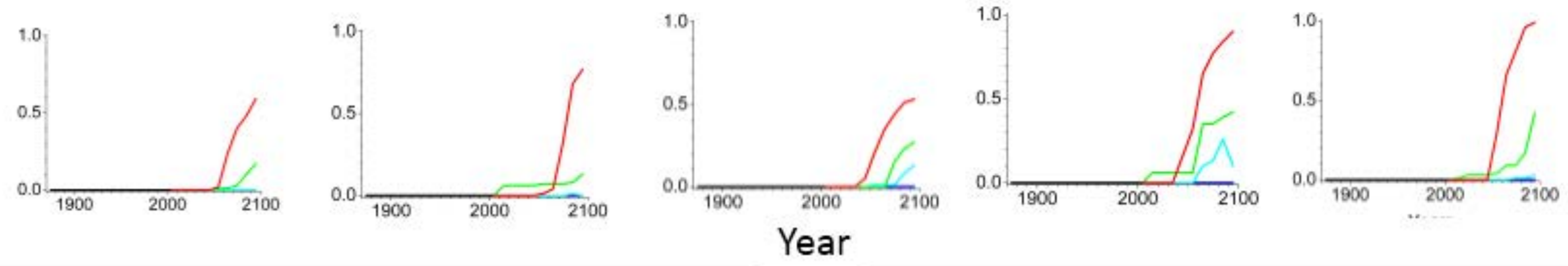
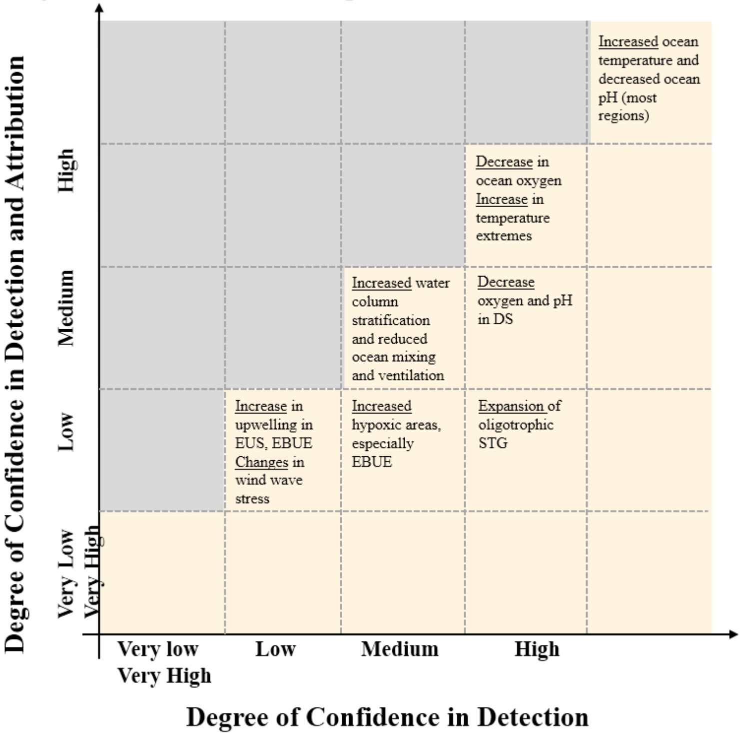


Figure 30-10: Annual maximum proportions of reef pixels with Degree Heating Months [*Donner et al.*, 2007]; DHM)≥1 (used for projecting coral bleaching; [*Strong et al.*, 1997; *Strong et al.*, 2011]) and DHM≥5 (associated with bleaching across 100% of affected areas with significant mortality, [*Eakin et al.*, 2010] for the period 1870–2009 for each of the six coral regions (Figure 30-4d) using the 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 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 SST anomalies were derived using a 1985–2000 maximum monthly mean (MMM) 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.]

A. Physical and Chemical changes



B. Biological and ecological changes

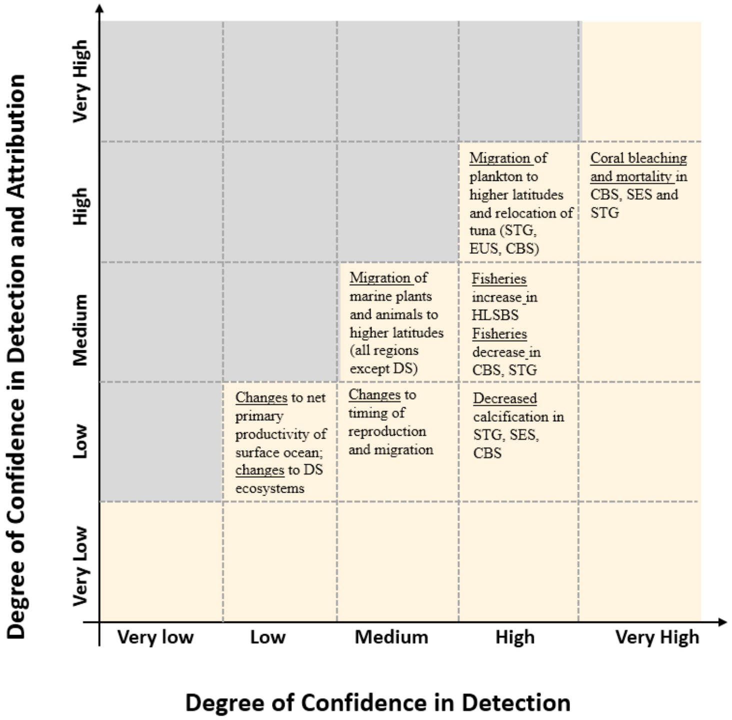
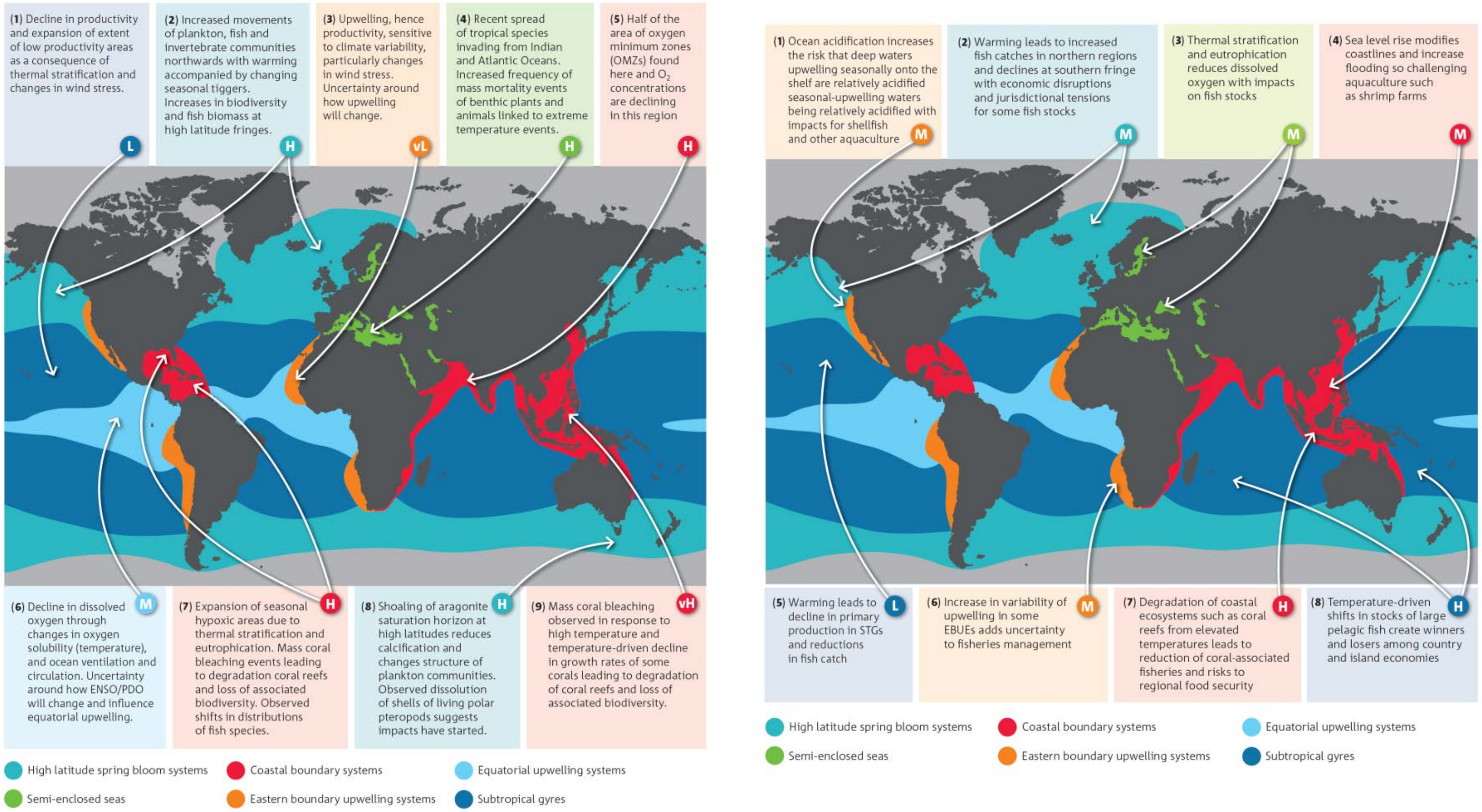


Figure 30-11: Expert assessment of degree of confidence in detection and attribution of physical and chemical changes (a) and ecological changes (b) across sub-regions, as designated in Figure 30-1a, and processes in the Ocean (based on evidence explored throughout Chapter 30 and elsewhere in AR5). Further explanation of this figure is given in 18.3.3–4 and 18.6. **[Illustration to be redrawn to conform to IPCC publication specifications.]**



(1) Decline in productivity and expansion of extent of low productivity areas as a consequence of thermal stratification and changes in wind stress.

(2) Increased movements of plankton, fish and invertebrate communities northwards with warming accompanied by changing seasonal tiggers. Increases in biodiversity and fish biomass at high latitude fringes.

(3) Upwelling, hence productivity, sensitive to climate variability, particularly changes in wind stress. Uncertainty around how upwelling will change.

(4) Recent spread of tropical species invading from Indian and Atlantic Oceans. Increased frequency of mass mortality events of benthic plants and animals linked to extreme temperature events.

(5) Half of the area of oxygen minimum zones (OMZs) found here and O_2 concentrations are declining in this region

(6) Decline in dissolved oxygen through changes in oxygen solubility (temperature), and ocean ventilation and circulation. Uncertainty around how ENSO/PDO will change and influence equatorial upwelling.

(7) Expansion of seasonal hypoxic areas due to thermal stratification and eutrophication. Mass coral bleaching events leading to degradation coral reefs and loss of associated biodiversity. Observed shifts in distributions of fish species.

(8) Shoaling of aragonite saturation horizon at high latitudes reduces calcification and changes structure of plankton communities. Observed dissolution of shells of living polar pteropods suggests impacts have started.

(9) Mass coral bleaching observed in response to high temperature and temperature-driven decline in growth rates of some corals leading to degradation of coral reefs and loss of associated biodiversity.

(1) Ocean acidification increases the risk that deep waters upwelling seasonally onto the shelf are relatively acidified seasonal-upwelling waters being relatively acidified with impacts for shellfish and other aquaculture

(2) Warming leads to increased fish catches in northern regions and declines at southern fringe with economic disruptions and jurisdictional tensions for some fish stocks

(3) Thermal stratification and eutrophication reduces dissolved oxygen with impacts on fish stocks

(4) Sea level rise modifies coastlines and increase flooding so challenging aquaculture such as shrimp farms

(5) Warming leads to decline in primary production in STGs and reductions in fish catch

(6) Increase in variability of upwelling in some EBUEs adds uncertainty to fisheries management

(7) Degradation of coastal ecosystems such as coral reefs from elevated temperatures leads to reduction of coral-associated fisheries and risks to regional food security

(8) Temperature-driven shifts in stocks of large pelagic fish create winners and losers among country and island economies

Figure 30-12: (a) Examples of projected impacts and vulnerabilities associated with climate change in Ocean sub-regions. (b) Examples of risks to fisheries from observed and projected impacts across Ocean sub-regions. Letters indicate level of confidence: (vL): Very low, (L): Low, (M): Medium, (H): High and (vH): Very high. Details of sub-regions are given in Table 30-1a and 30.1.1. **[Illustration to be redrawn to conform to IPCC publication specifications.]**

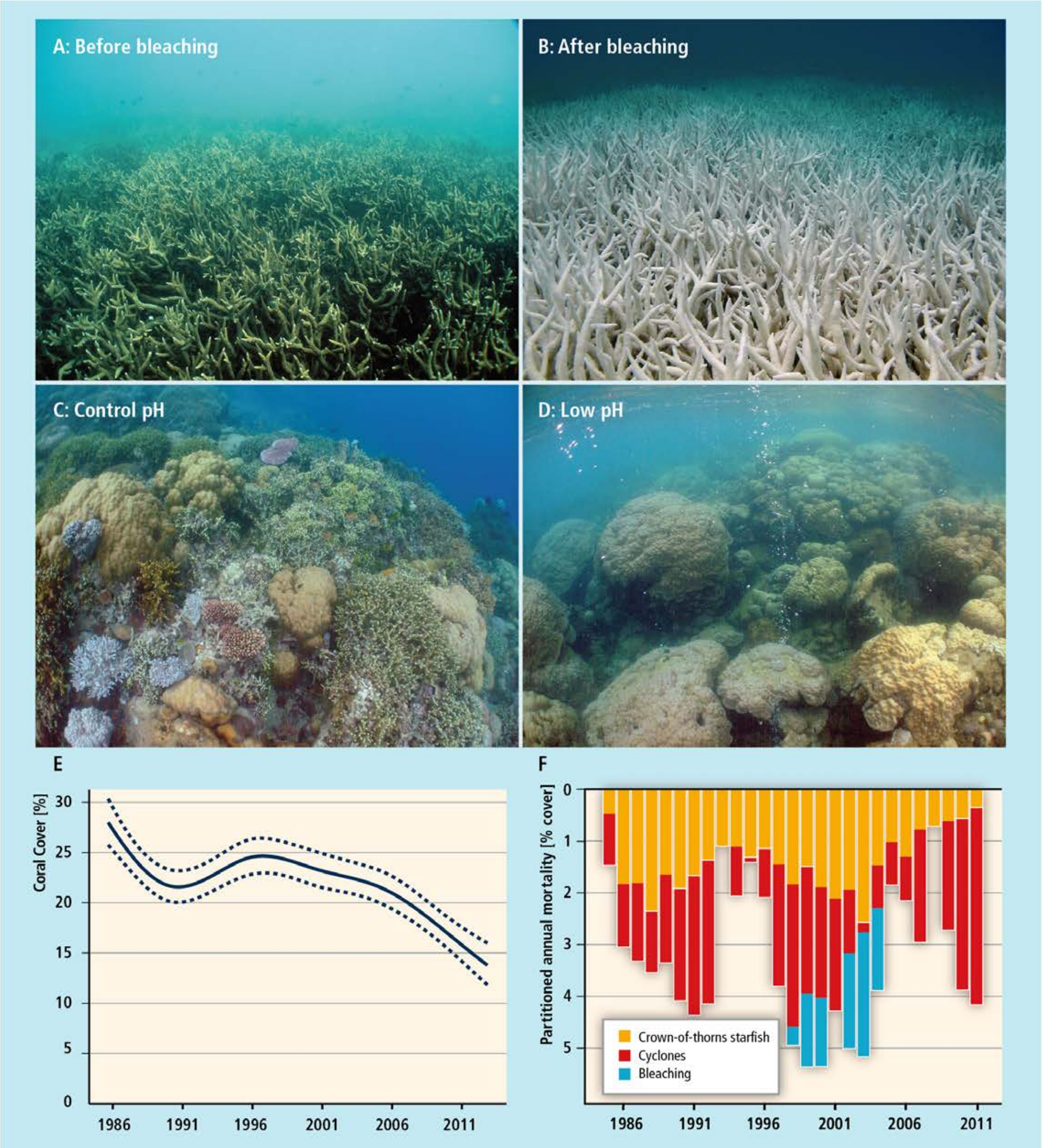


Figure CR-1: A and B: the same coral community before and after a bleaching event in February 2002 at 5 m depth, Halfway Island, Great Barrier Reef. Coral cover at the time of bleaching was 95% bleached almost all of it severely bleached, resulting in mortality of 20.9% (Elvidge *et al*., 2004). Mortality was comparatively low due in part because these coral communities were able to shuffle their symbiont to more thermo-tolerant types (Berkelmans and van Oppen, 2006; Jones *et al*., 2008). C and 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 (panel D; median pH$_T$ ~7.8), reefs are dominated by massive corals while corals with high morphological complexity are underrepresented compared with control sites (D; median pH ~8.0). Reef development ceases at pH$_T$ values below 7.7. pH$_T$: pH on the total scale. E: temporal trend in coral cover for the whole Great Barrier Reef over the period 1985–2012 (N, number of reefs, mean ± 2 standard errors; 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).

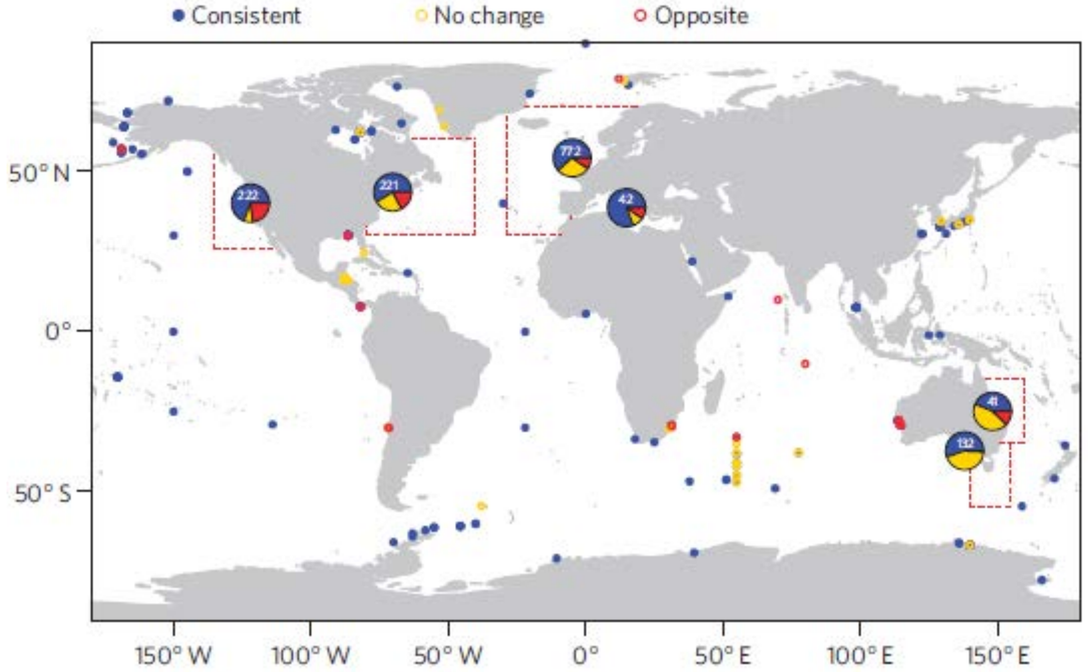


Figure MB-1: 1735 observed responses to climate change from 208 single- and multi-species studies. Changes attributed to climate change (blue), inconsistent with climate change (red) and are equivocal (yellow). Each circle represents the centre of a study area. Where points fall on land, it is because they are centroids of distribution that surround an island or peninsula. Pie charts show the proportions within regions bounded by red squares and in the Mediterranean; numbers indicate the total (consistent, opposite or equivocal) observations within each region. Note: 57% of the studies included were published since AR4 (from Poloczanska *et al.*, 2013).

[Illustration to be redrawn to conform to IPCC publication specifications.]

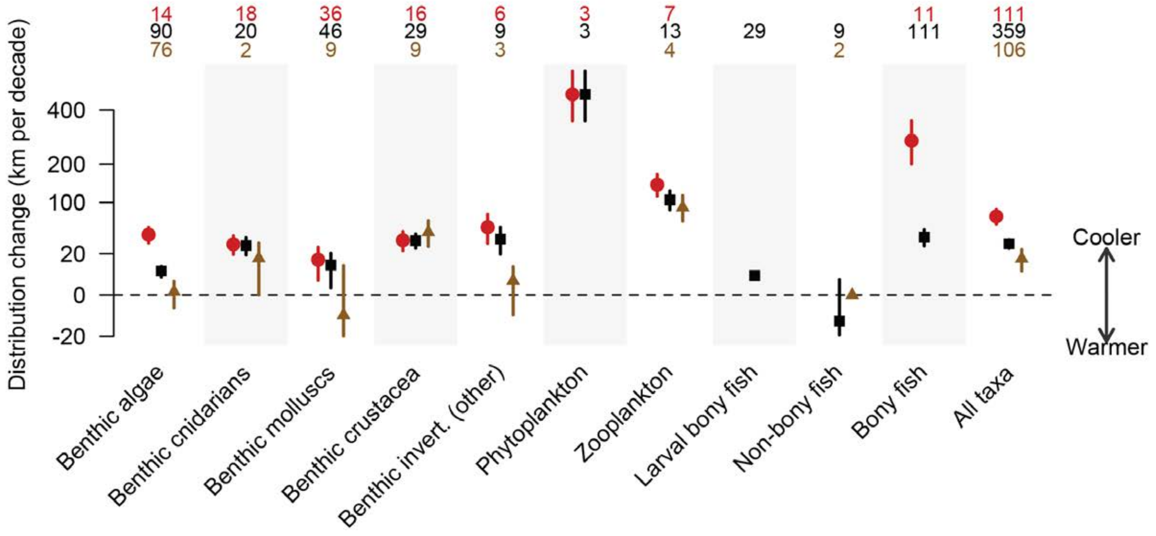


Figure MB-2. Rates of change in distribution (km decade^{-1}) for marine taxonomic groups, measured at the leading edges (red) and trailing edges (brown). Average distribution shifts calculated using all data, regardless of range location, are in black. Distribution rates have been square-root transformed; 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 (from Poloczanska et al., 2013).

[Illustration to be redrawn to conform to IPCC publication specifications.]

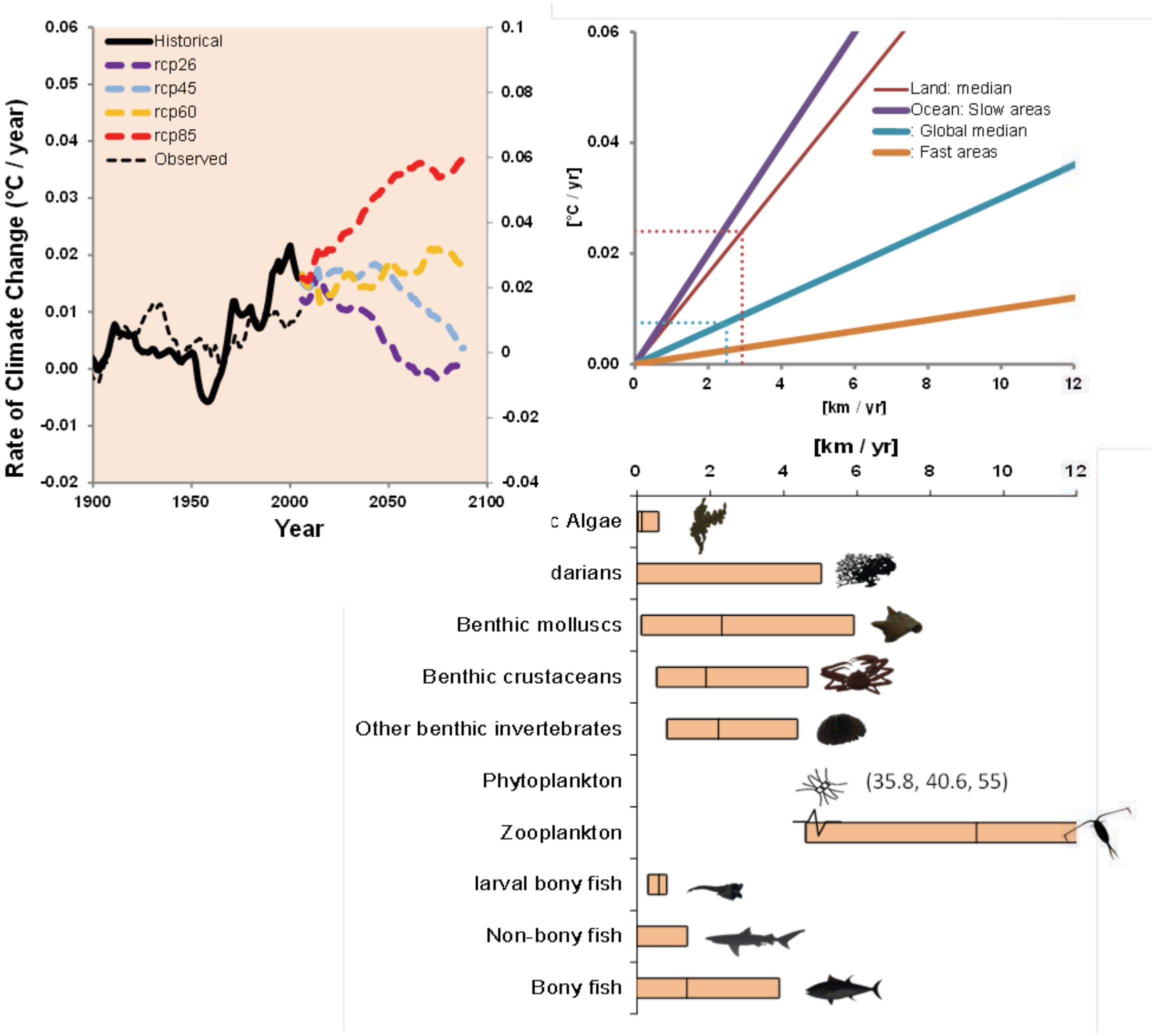


Figure MB-3. A. Rate of climate change for the Ocean (sea surface temperature (SST) °C); B. corresponding climate velocities for the Ocean and median velocity from land (adapted from Burrows et al., 2011); and C. observed rates of displacement of marine taxonomic groups over several decades until 2010. The thin dotted red arrows give an example of interpretation. Rates of climate change of 0.008 °C yr^{-1} correspond to ca. 2.4 km yr^{-1}median climate velocity in the Ocean. When compared to observed rates of displacement, many marine taxonomic groups have been able to track these velocities, except phyto- and zooplankton where rates of displacement greatly exceed climate velocity. 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 (Black dotted line) are derived from HadISST1.1 data set, all other rates are calculated based on the average of the CMIP5 climate model ensembles (Table S30-3) for the historical period and for the future based on the four RCP emissions scenarios. Data were smoothed using a 20-year sliding window. (B) Median climate velocity calculated from HadISST1.1 dataset over 1960–2010 using the methods of Burrows et al., 2011. The three axes represent estimated median climate velocities are representative of areas of slow velocities such as Pacific subtropical gyre (STG) system (Purple line), the global Ocean surface (excluding polar seas, Blue line), and areas of high velocities such as the Coral Triangle and North Sea (Orange line). Figure 30-3 shows climate velocities over the ocean surface calculated over 1960–2010. The Red line corresponds to the median rate over global land surface calculated using historical surface temperatures from the CMIP5 model ensemble (Table S30-3). (C) Rates of displacement for marine taxonomic groups estimated by Poloczanska et al. 2013 using published studies (Figure MB-2 Black data set). Note the displacement rates for phytoplankton exceed the axis, so values are given. **[Illustration to be redrawn to conform to IPCC publication specifications.]**

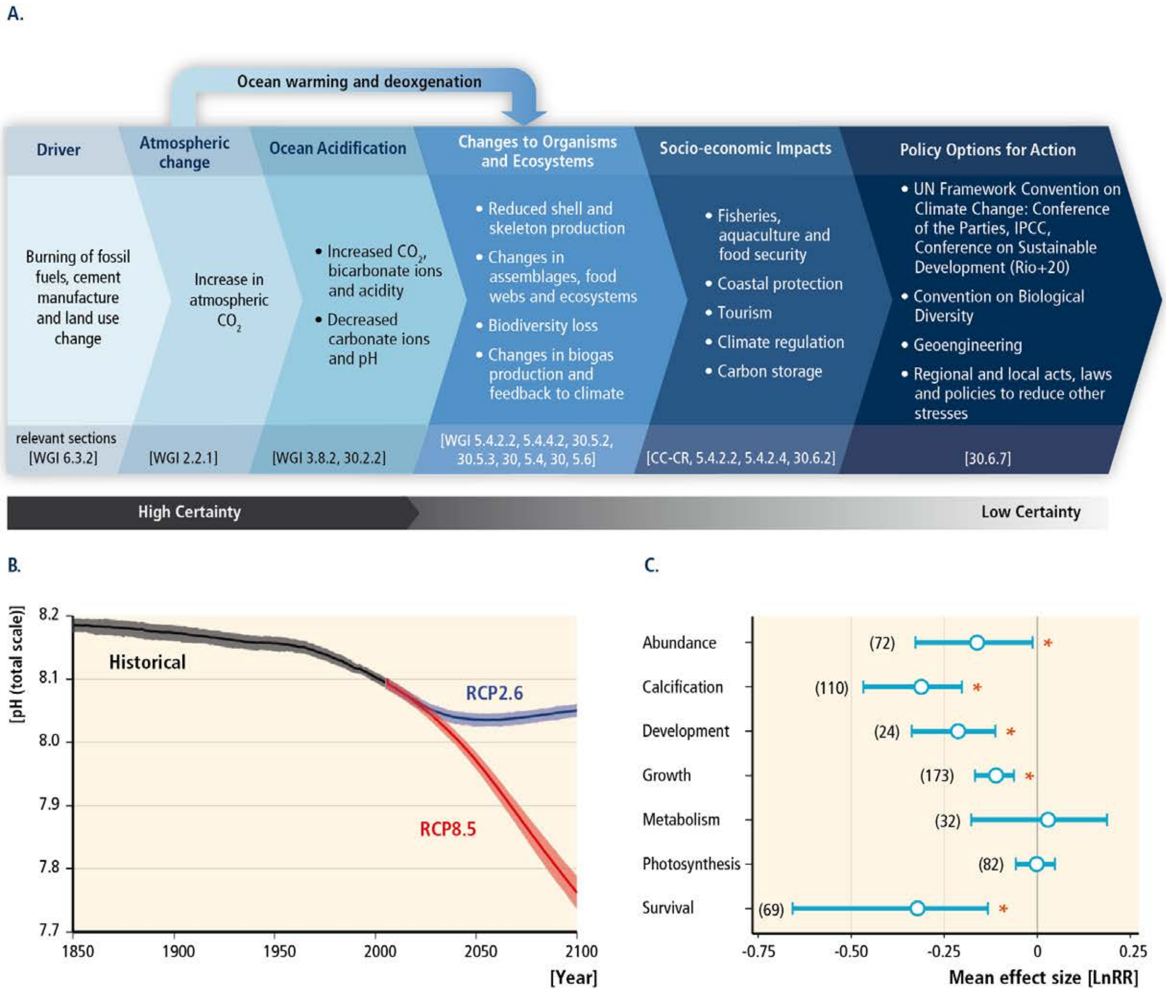


Figure OA-1: A: Overview of the chemical, biological, socio-economic impacts of ocean acidification and of policy options (adapted from Turley and Gattuso, 2012). B: Multi-model simulated time series of global mean ocean surface pH (on the total scale) from CMIP5 climate model simulations from 1850 to 2100. Projections are shown for emission scenarios RCP2.6 (blue) and RCP8.5 (red) for the multi-model mean (solid lines) and range across the distribution of individual model simulations (shading). Black (grey shading) is the modelled 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. The number of CMIP5 models to calculate the multi-model mean is indicated for each time period/scenario (WGI AR5 Figure 6.28). C: Effect of near future acidification (seawater pH reduction of 0.5 unit or less) 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. * denotes a statistically significant effect.

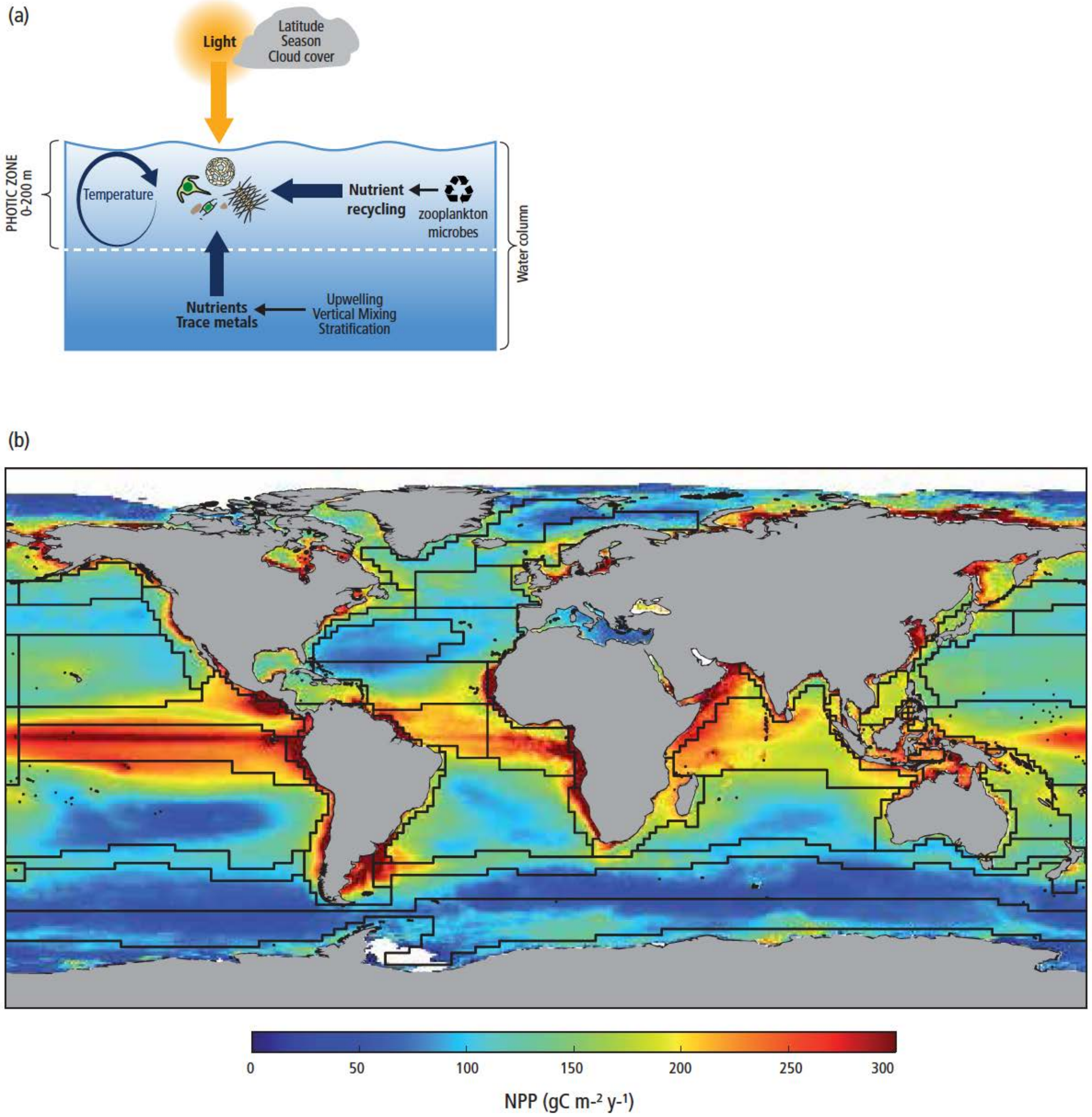


Figure PP-1: A) Environmental factors controlling Net Primary Production (NPP). NPP is mainly controlled by three basic processes: 1) Light conditions in the surface ocean, i.e. the photic zone where photosynthesis occurs, 2) upward flux of nutrients and micronutrients from underlying waters into the photic zone, 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 MODIS Aqua satellite climatology from 2003-2012; NPP was calculated with the Carbon-based Production 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. Figure courtesy of Toby Westberry (OSU) and Ivan Lima (WHOI), satellite data courtesy of NASA Ocean Biology Processing Group.
[Illustration to be redrawn to conform to IPCC publication specifications.]

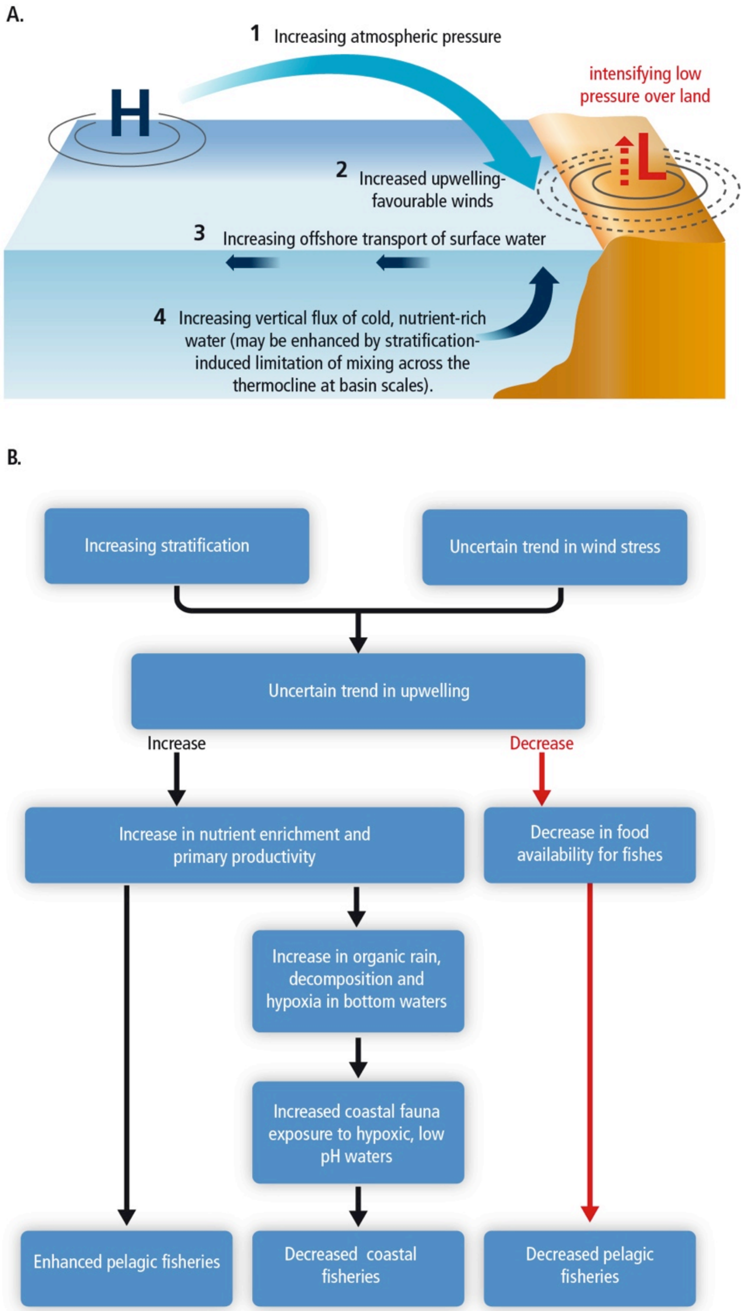


Figure UP-1: Upper panel: Schematic hypothetic mechanism of increasing coastal wind-driven upwelling at eastern boundary systems, where differential warming rates between land and ocean results in increased land-ocean pressure gradients (1) that produce stronger alongshore winds (2) and offshore movement of surface water through Ekman transport (3), and increased upwelling of deep cold nutrient rich waters to replace it (4). Lower panel: 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 decreased coastal fisheries due to an augmented 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.