Chapter 5: Changing Ocean, Marine Ecosystems, and Dependent Communities

Coordinating Lead Authors: Nathaniel L. Bindoff (Australia), William W. L. Cheung (Canada), James G. Kairo (Kenya)

Lead Authors: Javier Arístegui (Spain), Valeria A. Guinder (Argentina), Robert Hallberg (USA), Nathalie Hilmi (Monaco/France), Nianzhi Jiao (China), Md saiful Karim (Australia), Lisa Levin (USA), Sean O'Donoghue (South Africa), Sara R. Purca Cuicapusa (Peru), Baruch Rinkevich (Israel), Toshio Suga (Japan), Alessandro Tagliabue (United Kingdom), Phillip Williamson (United Kingdom)

Contributing Authors: Sevil Acar (Turkey), Juan Jose Alava (Ecuador/Canada), Eddie Allison (United Kingdom), Brian Arbic (USA), Tamatoa Bambridge (French Polynesia), Inka Bartsch (Germany), Laurent Bopp (France), Philip W. Boyd (Australia/ United Kingdom), Thomas Browning (Germany/United Kingdom), Francisco P. Chávez (USA), Mine Cinar (USA), Daniel Costa (USA), Lijing Cheng (China), Omar Defeo (Uruguay), Salpie Djoundourian (Lebanon), Catia Domingues (Australia), Tyler Eddy (Canada), Sonja Endres (Germany), Alan Fox (UK), Christopher Free (USA), Thomas Frölicher (Switzerland), Jean-Pierre Gattuso (France), Gemma Gerber (South Africa), Charles Greene (USA), Nicolas Gruber (Switzerland), Gustaav Hallegraef (Australia), Matthew Harrison (USA), Sebastian Hennige (UK), Mark Hindell (Australia), Andrew Hogg (Australia), Taka Ito (USA), Tiff-Annie Kenny (Canada), Kristy Kroeker (USA), Lester Kwiatkowski (France/UK), Vicky Lam (China/Canada), Philippe LeBillon (Canada), Nadine Le Bris (France), Heike Lotze (Canada), Jennifer MacKinnon (USA), Annick de Marffy-Mantuano (Monaco), Patrick Martell (South Africa), Nadine Marshall (Australia), Kathleen McInnes (Australia), Jose Garcia Molinos (Japan/Spain), Serena Moseman-Valtierra (USA), Andries Motau (South Africa), Sandor Mulsow (Brazil), Kana Mutombo (South Africa), Andreas Oschlies (Germany), Muhammed Oyinlola (Nigeria), Elvira S. Poloczanska (Australia), Nicolas Pascal (France), Maxime Philip (France), Sarah Purkey (USA), Saurabh Rathore (India), Xavier Rebelo (South Africa), Gabrial Reygondeau (France), Jake Rice (Canada), Anthony Richardson (Australia), Ulf Riebesell (Germany), Christopher Roach (Australia), Joacim Rocklöv (Sweden), Murray Roberts (United Kingdom), Alain Safa (France), Sunke Schmidtko (Germany), Gerald Singh (Canada), Bernadette Sloyan (Australia), Karinna von Schuckmann (France), Manal Shehabi (England), Matthew Smith (USA), Amy Shurety (South Africa), Fernando Tuya (Spain), Cristian Vargas (Chile), Colette Wabnitz (France), Caitlin Whalen (USA)

Review Editors: Manuel Barange (South Africa), Brad Seibel (USA)

Chapter Scientist: Axel Durand (Australia)

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Executive Summary

The ocean is essential for all aspects of human well-being and livelihood. It provides key services like climate regulation, through the energy budget, carbon cycle and nutrient cycle. The ocean is the home of biodiversity ranging from microbes to marine mammals that form a wide variety of ecosystems in open pelagic and coastal ocean.

Observations: Climate-related trends, impacts, adaptation

Carbon emissions from human activities are causing ocean warming, acidification, and oxygen loss with some evidence of changes in nutrient cycling and primary production. The warming ocean is affecting marine organisms at multiple trophic levels, impacting fisheries with implications for food production and human communities. Concerns regarding the effectiveness of existing ocean and fisheries governance have already been reported, highlighting the need for timely mitigation and adaptation responses.

The ocean has warmed unabated since 2005, continuing the clear multi-decadal ocean warming trends documented in the IPCC Fifth Assessment Report (AR5). The warming trend is further confirmed by the improved ocean temperature measurements over the last decade. The 0−700 m and 700–2000 m layers of the ocean have warmed at rates of 5.31±0.48 and 4.02±0.97 ZJ yr[1] from 2005 to 2017. The long-term trend for 0-700 m and 700-2000 m layers have warmed 4.35±0.8 and 2.25±0.64 ZJ yr[1] from 1970 to 2017 and is attributed to anthropogenic influences. It is *likely*[1] the ocean warming has continued in the abyssal and deep ocean below 2000m (southern hemisphere and Southern Ocean). {1.8.1, 1.2, 5.2.2}

It is *likely* that the rate of ocean warming has increased since 1993. The 0−700 m and 700–2000 m layers of the ocean have warmed by 3.22±1.61 ZJ and 0.97±0.64 ZJ from 1970 to 1993, and 6.28±0.48 ZJ and 3.86±2.09 ZJ from 1993 to 2017. This represents at least a two-fold increase in heat uptake. {Table 5.1, 5.2.2}

The upper ocean is *very likely* to have been stratifying since 1970. Observed warming and high latitude freshening are making the surface ocean less dense over time relative to the deeper ocean (*high confidence*[2]) and inhibiting the exchange between surface and deep waters. The upper 200m stratification increase is in the *very likely* range of between 2.18% and 2.42% from1970 to 2017. {5.2.2}

Multiple datasets and models show that the rate of ocean uptake of atmospheric CO_2 has continued to strengthen in the recent two decades in response to the increasing concentration of CO_2 in the atmosphere. The *very likely* range for ocean uptake is between 20-30% of total anthropogenic emissions in the recent two decades. Evidence is growing that the ocean carbon sink is dynamic on decadal timescales, especially in the Southern Ocean, which has affected the total global ocean carbon sink (*medium confidence*). {5.2.2.3}

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[1] In this Report, the following terms have been used to indicate the assessed likelihood of an outcome or a result: Virtually certain 99–100% probability, Very likely 90–100%, Likely 66–100%, About as likely as not 33–66%, Unlikely 0–33%, Very unlikely 0–10%, and Exceptionally unlikely 0–1%. Additional terms (Extremely likely: 95–100%, More likely than not >50–100%, and Extremely unlikely 0–5%) may also be used when appropriate. Assessed likelihood is typeset in italics, e.g., *very likely* (see Section 1.9.2 and Figure 1.4 for more details). This Report also uses the term '*likely* range' to indicate that the assessed likelihood of an outcome lies within the 17-83% probability range.

[2] In this Report, the following summary terms are used to describe the available evidence: limited, medium, or robust; and for the degree of agreement: low, medium, or high. A level of confidence is expressed using five qualifiers: very low, low, medium, high, and very high, and typeset in italics, e.g., *medium confidence*. For a given evidence and agreement statement, different confidence levels can be assigned, but increasing levels of evidence and degrees of agreement are correlated with increasing confidence (see Section 1.9.2 and Figure 1.4 for more details).

The ocean is continuing to acidify in response to ongoing ocean carbon uptake. The open ocean surface water pH is observed to be declining (*virtually certain*) by a *very likely* range of 0.017 to 0.027 pH units per decade since the late 1980s across individual time-series observations longer than 15 years. The anthropogenic pH signal is *very likely* to have emerged for three-quarters of the near-surface open ocean prior to 1950 and it is *very likely* that over 95 % of the near-surface open ocean has already been affected. These changes in pH have reduced the stability of mineral forms of calcium carbonate due to a lowering of carbonate ion concentrations, most notably in the upwelling and high latitude regions of the ocean. {5.2.2.3, Box 5.1}

There is a growing consensus that the open ocean is losing oxygen overall with a *very likely* loss of 0.5 to 3.3% between 1970-2010 from the ocean surface to 1000 m (*medium confidence*). Globally, the oxygen loss due to warming is reinforced by other processes associated with ocean physics and biogeochemistry, which cause the majority of the observed oxygen decline (*high confidence*). The oxygen minimum zones are expanding by a *very likely* range of 3-8%, most notably in the tropical oceans, but there is substantial decadal variability that affects the attribution of the overall oxygen declines to human activity in tropical regions (*high confidence*). {5.2.2.4}

In response to ocean warming and increased stratification, open ocean nutrient cycles are being perturbed and there is *high confidence* that this is having a regionally variable impact on primary producers. There is currently *low confidence* in appraising past open ocean productivity trends, including those determined by satellites, due to newly identified region-specific drivers of microbial growth and the lack of corroborating *in situ* time series datasets. {5.2.2.5, 5.2.2.6}

Ocean warming has contributed to observed changes in biogeography of organisms ranging from phytoplankton to marine mammals (*high confidence*), consequently changing community composition (*high confidence*), and in some cases, altering interactions between organisms (*medium confidence*). Observed rate of range shifts since the 1950s and *its very likely range* are estimated to be 51.5±33.3 km per decade and 29.0±15.5 km per decade for organisms in the epipelagic and seafloor ecosystems, respectively. The direction of the majority of the shifts of epipelagic organisms are consistent with a response to warming (*high confidence*). {5.2.3, 5.3}

Warming-induced range expansion of tropical species to higher latitudes has led to increased grazing on some coral reefs, rocky reefs, seagrass meadows and epipelagic ecosystems, leading to altered ecosystem structure (*medium confidence*). Warming, sea level rise, and enhanced loads of nutrients and sediments in deltas have contributed to salinization and deoxygenation in estuaries (*high confidence*), and have caused up-stream redistribution of benthic and pelagic species according to their tolerance limits (*medium confidence*). {5.3.4, 5.3.5, 5.3.6, 5.2.3}

Fisheries catches and their composition in many regions are already impacted by the effects of warming and changing primary production on growth, reproduction and survival of fish stocks (*high confidence*). Ocean warming and changes in primary production in the 20th century are related to changes in productivity of many fish stocks (*high confidence*), with an average decrease of approximately 3% per decade in population replenishment and 4.1% (*very likely range* of 9.0% decline to 0.3% increase) in maximum catch potential (*robust evidence, low agreement* between fish stocks, *medium confidence*). Species composition of fisheries catches since the 1970s in many shelf seas ecosystems of the world is increasing dominated by warm-water species (*medium confidence*). {5.2.3, 5.4.1}

Warming-induced changes in spatial distribution and abundance of fish stocks have already challenged the management of some important fisheries and their economic benefits (*high

confidence). For existing international and national ocean and fisheries governance, there are concerns about the reduced effectiveness to achieve mandated ecological, economic, and social objectives because of observed climate impacts on fisheries resources (*high confidence*). {5.4.2, 5.5.2}

Coastal ecosystems are observed to be under stress from ocean warming and sea level rise that are exacerbated by non-climatic pressures from human activities on ocean and land (*high confidence*). Global wetland area has declined by nearly 50% relative to pre-industrial level as a result of warming, sea level rise, extreme climate events and other human impacts (*medium confidence*). Warming-related mangrove encroachment into subtropical saltmarshes has been observed in the past 50 years (*high confidence*). Distributions of seagrass meadows and kelp forests are contracting at low-latitudes that is attributable to warming (*high confidence*), and in some areas a loss of 36-43% following heat waves (*medium confidence*). Inundation, coastline erosion and salinization are causing inland shifts in plant species distributions, which has been accelerating in the last decades (*medium confidence*). Warming has increased the frequency of large-scale coral bleaching events, causing worldwide reef degradation since 1997-1998 with cases of shifts to algal-dominated reefs (*high confidence*). Sessile calcified organisms (e.g. barnacles, mussels) in intertidal rocky shores are highly sensitive to extreme temperature events and acidification (*high confidence*), a reduction in their biodiversity and abundance have been observed in naturally-acidified rocky reef ecosystems (*medium confidence*). Increased nutrient and organic matter loads in estuaries since the 1970s have exacerbated the effects of warming on bacterial respiration and eutrophication, leading to expansion of hypoxic areas (*high confidence*). {5.3.1, 5.3.2, 5.3.4, 5.3.6}

Coastal and near-shore ecosystems including saltmarshes, mangroves, and vegetated dunes in sandy beaches have a varying capacity to build vertically and expand laterally in response to sea-level rise. These ecosystems provide important services including coastal protection, carbon sequestration and habitat for diverse biota (*high confidence*). The carbon emission associated with the loss of vegetated coastal ecosystems is estimated to be 0.04–1.46 Gt C yr^{-1} (*high confidence*). The natural capacity of ecosystems to adapt to climate impacts may be limited by human activities that fragment wetland habitats and restrict landward migration (*high confidence*). {5.3.2, 5.3.3, 5.4.1, 5.5.1}

Three out of the four major Eastern Boundary Upwelling Ecosystems (EBUS) have shown large-scale wind intensification in the past 60 years (*high confidence*). However, the interaction of coastal warming and local winds may have affected upwelling strength, with the direction of changes varies between and within EBUS (*low confidence*). Increasing trends in ocean acidification and deoxygenation are observed in the two Pacific (California Current and Humboldt Current) EBUS in the last few decades (*high confidence*), although there is *low confidence* to distinguish anthropogenic forcing from internal climate variability. The expanding California EBUS oxygen minimum zone has altered ecosystem structure and fisheries catches (*medium confidence*). {Box 5.3}

Since the early 1980s, the occurrence of harmful algal blooms (HABs) and pathogenic organisms (e.g. *Vibrio*) has increased in coastal areas in response to warming, deoxygenation and eutrophication, with negative impacts on food provisioning, tourism, the economy and human health (*high confidence*). These impacts depend on species-specific responses to the interactive effects of climate change and other human drivers (e.g. pollution). Human communities in poorly monitored areas are among the most vulnerable to these biological hazards (*medium confidence*). {Box 5.4, 5.4.2}

Many frameworks for climate-resilient coastal adaptation have been developed since AR5, with substantial variations in approach between and within countries, and across development status (*high confidence*). Few studies have assessed the success of implementing these frameworks due to the time-lag between implementation, monitoring, evaluation and reporting (*medium confidence*). {5.5.2}

Projections: scenarios and time horizons

Climate models project significant changes in the ocean state over the coming century. Under the high emissions scenario (RCP8.5) the impacts by 2090 are substantially larger and more widespread than for the low emissions scenario (RCP2.6) throughout the surface and deep ocean, including: warming (*virtually certain*); ocean acidification (*virtually certain*); decreased stability of mineral forms of calcite (*virtually certain*); oxygen loss (*very likely*); reduced near-surface nutrients (*likely as not*); decreased net-primary productivity (*high confidence*); reduced fish production (*likely*) and loss of key ecosystems services (*medium confidence*) that are important for human wellbeing and sustainable development. {5.2.2, Box 5.1, 5.2.3, 5.2.4, 5.4}

By 2100 the ocean is *very likely* to warm by 2 to 4 times as much for low emissions (RCP2.6) and 5 to 7 times as much for the high emissions scenario (RCP8.5) compared with the observed changes since 1970. The 0–2000 m layer of the ocean is projected to warm by a further 2150 ZJ (*very likely* range 1710 to 2790 ZJ) between 2017 and 2100 for the RCP8.5 scenario. The 0–2000 m layer is projected to warm by 900 ZJ (*very likely* range 650 to 1340 ZJ) by 2100 for the RCP2.6 scenario, and the overall warming of the ocean will continue this century even after radiative forcing and mean surface temperatures stabilize (*high confidence*). {5.2.2.2}

The upper ocean will continue to stratify. By the end of the century the annual mean stratification of the top 200 m (averaged between 60°S and 60°N relative to the 1986 to 2005 period) is projected to increase in the *very likely* range of 1 to 9% and 12 to 30% for RCP2.6 and RCP8.5 respectively. {5.2.2.2}

It is *very likely* that the majority of coastal regions will experience statistically significant changes in tidal amplitudes over the course of the 21st century. The sign and amplitude of local changes to tides are *very likely* to be impacted by both human coastal adaptation measures and climate drivers. {5.2.2.2.3}

It is *virtually certain* that surface ocean pH will decline, by 0.036-0.042 or 0.287-0.29 pH units by 2081-2100, relative to 2006-2015, for the RCP2.6 or RCP8.5 scenarios, respectively. These pH changes are *very likely* to cause the Arctic and Southern Oceans, as well as the North Pacific and Northwestern Atlantic Oceans to become corrosive for the major mineral forms of calcium carbonate under RCP8.5, but these changes are *virtually certain* to be avoided under the RCP2.6 scenario. There is increasing evidence of an increase in the seasonal exposure to acidified conditions in the future (*high confidence*), with a *very likely* increase in the amplitude of seasonal cycle of hydrogen iron concentrations of 71-90% by 2100, relative to 2000 for the RCP8.5 scenario, especially at high latitudes. {5.2.2.3}

Oxygen is projected to decline further. Globally, the oxygen content of the ocean is *very likely* to decline by 3.2-3.7% by 2081-2100, relative to 2006-2015, for the RCP8.5 scenario or by 1.6-2.0% for the RCP2.6 scenario. The volume of the oceans oxygen minimum zones is projected to grow by a *very likely* range of 7.0±5.6% by 2100 during the RCP8.5 scenario, relative to 1850-1900. The climate signal of oxygen loss will *very likely* emerge from the historical climate by 2050 with a *very likely* range of 59-80% of ocean area being affected by 2031-2050 and rising with a *very likely*

range of 79-91% by 2081-2100 (RCP8.5 emissions scenario). The emergence of oxygen loss is *very likely* smaller in area for the RCP2.6 scenario in the 21st century and by 2090 the emerged area is declining. {5.2.2.4, Box 5.1 Figure 1}

Overall, nitrate concentrations in the upper 100m are *very likely* to decline by 9-14 % across CMIP5 models by 2081-2100, relative to 2006-2015, in response to increased stratification for RCP8.5, with *medium confidence* in these projections due to the *limited evidence* of past changes that can be robustly understood and reproduced by models. There is *low confidence* regarding projected increases in surface ocean iron levels due to systemic uncertainties in these models. {5.2.2.5}

Climate models project that net primary productivity will *very likely* decline by 4-11% for RCP8.5 by 2081-2100, relative to 2006-2015. The decline is due to the combined effects of warming, stratification, light, nutrients and predation and will show regional variations between low and high latitudes (*low confidence*). The tropical ocean NPP will *very likely* decline by 7-16% for RCP8.5, with *medium confidence* as there are improved constraints from historical variability in this region. Globally, the sinking flux of organic matter from the upper ocean into the ocean interior is *very likely* to decrease by 9-16% for RCP8.5 in response to increased stratification and reduced nutrient supply, especially in tropical regions (*medium confidence*), which will reduce organic carbon supply to deep sea ecosystems (*high confidence*). The reduction in food supply to the deep sea is projected to lead to a 5-6% reduction in biomass of benthic biota over more than 97% of the abyssal seafloor by 2100. {5.2.2.6, 5.2.4.2}

New ocean states for a broad suite of climate indices will progressively emerge over a substantial fractions of the ocean in the coming century (relative to past internal ocean variability), with ESMs showing an ordered emergence of first pH, followed by SST, interior oxygen, upper ocean nutrient levels and finally NPP. The anthropogenic pH signal has *very likely* emerged for three quarters of the ocean prior to 1950, with little difference between scenarios. Oxygen changes will *very likely* emerge over 59-80% of the ocean area by 2031-2050 and rises to 79-91% by 2081-2100 (RCP8.5 emissions scenario). The rate and extent of these effects for all variables remain detectable over 30% of the ocean surface in the RCP2.6 scenario, but are much lower than for RCP8.5. {Box 5.1, Box 5.1 Figure 1}

Simulated ocean warming and changes in net primary production during the 21st century are projected to alter community structure of marine organisms (*high confidence*), reduce global marine animal biomass (*medium confidence*) and the maximum potential catches of fish stocks (*medium confidence*) with regional differences in the direction and magnitude of changes (*high confidence*). The global biomass of marine animals, including those that contribute to fisheries, is projected to decrease with a *very likely* range under RCP2.6 and RCP8.5 of 4.3±2.0% and 15.0±5.9%, respectively, by 2080–2099 relative to 1986–2005. The maximum catch potential is projected to decrease by 2.8% to 9.1% (RCP2.6) and 16.2% to 25.5% (RCP8.5) in the 21st century. {5.4.1}

Projected decreases in global marine animal biomass and fish catch potential could elevate the risk of impacts on income, livelihood and food security of the dependent human communities (*medium confidence*). Projected climate change impacts on fisheries also increase the risk of potential conflicts among fishery area users and authorities or between two different communities within the same country (*medium confidence*), exacerbated through competing resource exploitation from international actors and mal-adapted policies (*low confidence*). {5.2.3, 5.4, 5.5.3}

Projected decrease in upper ocean export of organic carbon to the deep seafloor is expected to result in a loss of animal biomass on the deep seafloor by 5.2% to 17.6% by 2090-2100 compared to the present (2006-2015) under RCP8.5 with regional variations (*medium confidence*). Some increases are projected in the polar regions, due to enhanced stratification in the surface ocean, reduced primary production and shifts towards small phytoplankton (*medium confidence*). The projected impacts on biomass in the abyssal seafloor are larger under RCP8.5 than RCP4.5 (*very likely*). The increase in climatic hazards beyond thresholds of tolerance of deep-sea organisms will increase the risk of loss of biodiversity and impacts on functioning of deep water column and seafloor that is important to support ecosystem services, such as carbon sequestration (*medium confidence*). {5.2.4}

Structure and functions of all types of coastal ecosystems will continue to be at moderate to high risk under the RCP2.6 scenario (*medium confidence*) and will face high to very high risk under the RCP8.5 scenario (*high confidence*) by 2100. Seagrass meadows (*high confidence*), kelp forests (*high confidence*) and coral reefs (*very high confidence*) will face high to very high risk already at temperatures 1.5 °C of global warming (*high confidence*). Intertidal rocky shores are also expected to be at very high risk (transition above 3°C) under the RCP8.5 scenario (*medium confidence*). These ecosystems have low to moderate adaptive capacity, as they are highly sensitive to ocean temperatures and acidification. The ecosystems with moderate to high risk (transition above 1.8°C) under future emissions scenarios are mangrove forests, sandy beaches, estuaries and saltmarshes (*medium confidence*). Estuaries and sandy beaches are subject to highly dynamic hydrological and geomorphological processes, giving them more natural adaptive capacity to climate hazards. In these systems, sediment relocation, soil accretion and landward expansion of vegetation may initially mitigate against flooding and habitat loss, but saltmarshes, in particular, will be at very high risk in the context of sea level rise and extreme climate-driven erosion under RCP8.5. {5.3, Figure 5.16}

Expected coastal ecosystem responses over the 21st century are habitat contraction, migration and loss of biodiversity and functionality. Pervasive human coastal disturbances will limit natural ecosystem adaptation to climate hazards (*high confidence*). Global coastal wetlands will lose between 20-90% of their area depending on emissions scenario with impacts on their contributions to carbon sequestration and coastal protection (*high confidence*). Kelp forests at low-latitudes and temperate seagrass meadows will continue to retreat as a result of intensified extreme temperatures, and their low dispersal ability will elevate the risk of local extinction under RCP8.5 (*high confidence*). Intertidal rocky shores will continue to be affected by ocean acidification, warming, and extreme heat exposure during low tide emersion, causing reduction of calcareous species and loss of ecosystem biodiversity and complexity shifting towards algae dominated habitats (*high confidence*). Salinization and expansion of hypoxic conditions will intensify in eutrophic estuaries, especially in mid and high latitudes with microtidal regimes (*high confidence*). Sandy beach ecosystems will increasingly be at risk of eroding, reducing the habitable area for dependent organisms (*high confidence*). {5.3, 5.4.1}

Almost all coral reefs will degrade from their current state, even if global warming remains below 2°C (very *high confidence*), and the remaining shallow coral reef communities will differ in species composition and diversity from present reefs (*very high confidence*). These declines in coral reef health will greatly diminish the services they provide to society, such as food provision (*high confidence*), coastal protection (*high confidence*) and tourism (*medium confidence*). {5.3.4, 5.4.1}

Multiple hazards of warming, deoxygenation, aragonite under-saturation and decrease in flux of organic carbon from the surface ocean will decrease calcification and exacerbate the bioerosion and dissolution of the non-living component of cold-water coral. Habitat-forming,

cold-water corals will be vulnerable where temperature and oxygen exceed the species' thresholds (*medium confidence*). Reduced particulate food supply is projected to be experienced by 95% of cold-water coral ecosystems by 2100 under RCP8.5 relative to the present, leading to a *very likely* range of 8.6±2% biomass loss (*medium confidence*). {5.2.4, Box 5.2}

Anthropogenic changes in Eastern Boundary Upwelling Ecosystems (EBUS) will emerge primarily in the second half of the 21st century (*medium confidence*). EBUS will be impacted by climate change in different ways, with strong regional variability with consequences for fisheries, recreation and climate regulation (*medium confidence*). The Pacific EBUS are projected to have calcium carbonate undersaturation in surface waters within a few decades under RCP8.5 (*high confidence*); combined with warming and decreasing oxygen levels, this will increase the impacts on shellfish larvae, benthic invertebrates and demersal fishes (*high confidence*) and related fisheries and aquaculture (*medium confidence*). The inherent natural variability of EBUS, together with uncertainties in present and future trends in the intensity and seasonality of upwelling, coastal warming and stratification, primary production and biogeochemistry of source waters, poses large challenges in projecting the response of EBUS to climate change and to the adaptation of governance of biodiversity conservation and living marine resources in EBUS (*high confidence*). {Box 5.3}

Climate change impacts on ecosystems and their goods and services threatens key cultural dimensions of lives and livelihoods. These threats include erosion of Indigenous and non-indigenous culture, their knowledge about the ocean and knowledge transmission, reduced access to traditional food, loss of opportunities for aesthetic and spiritual appreciation of the ecosystems, and marine recreational activities (*medium confidence*). Ultimately, these can lead to the loss of part of people's cultural identity and values beyond the rate at which identify and values can be adjusted or substituted (*medium confidence*). {5.4.2}

Climate change increases the exposure and bioaccumulation of contaminants such as persistent organic pollutants and mercury (*medium confidence*), and their risk of impacts on marine ecosystems and seafood safety (*high agreement, medium evidence, medium confidence*). Such risks are particularly large for top predators and for human communities that have high consumption on these organisms, including coastal Indigenous communities (*medium confidence*). {5.4.2}

B1.18 Shifting distributions of fish stocks between governance jurisdictions will increase the risk of potential conflicts among fishery area users and authorities or between two different communities within the same country (*medium confidence*). These fishery governance related risks are widespread under high emissions scenarios with regional hotspots (*medium confidence*), and highlight the limits of existing natural resource management frameworks for addressing ecosystem change (*high confidence*). {5.2.5, 5.4.2.1.3, 5.5, 5.5.2}

Response options to enhance resilience

There is clear evidence for observed climate change impacts throughout the ocean with consequences for human communities and require options to reduce risks and impacts. Coastal blue carbon can contribute to mitigation for many nations but its global scope is modest (offset of < 2% of current emissions) (*likely*). Some ocean indices are expected to emerge earlier than others (e.g., warming, acidification and effects on fish stocks) and could therefore be used to prioritise planning and building resilience. The survival of some keystone ecosystems (e.g., coral reefs) are at risk, while governance structures are not well-matched to the spatial and temporal scale of climate change impacts on ocean systems. Ecosystem restoration may be able to locally reduce climate risks (*medium confidence*) but at

relatively high cost and effectiveness limited to low emissions scenarios and to less sensitive systems (*high confidence*). {5.2, 5.3, 5.4, 5.5}

Coastal blue carbon ecosystems, such as mangroves, salt marshes and seagrasses, can help reduce the risks and impacts of climate change, with multiple co-benefits. Some 151 countries around the world contain at least one of these coastal blue carbon ecosystems and 71 countries contain all three. Below-ground carbon storage in vegetated marine habitats can be up to 1000 tC ha⁻¹, much higher than most terrestrial ecosystems (*high confidence*). Successful implementation of measures to maintain and promote carbon storage in such coastal ecosystems could assist several countries in achieving a balance between emissions and removals of greenhouse gases (*medium confidence*). Conservation of these habitats would also sustain the wide range of ecosystem services they provide and assist with climate adaptation through improving critical habitats for biodiversity, enhancing local fisheries production, and protecting coastal communities from sea level rise and storm events (*high confidence*). The climate mitigation effectiveness of other natural carbon removal processes in coastal waters, such as seaweed ecosystems and proposed non-biological marine CO_2 removal methods, are smaller or currently have higher associated uncertainties. Seaweed aquaculture warrants further research attention. {5.5.1.1, 5.5.1.1, 5.5.1, 5.5.2, 5.5.1.1.3, 5.5.1.1.4}

The potential climatic benefits of blue carbon ecosystems can only be a very modest addition to, and not a replacement for, the very rapid reduction of greenhouse gas emissions. The maximum global mitigation benefits of cost-effective coastal wetland restoration is *unlikely* to be more than 2% of current total emissions from all sources. Nevertheless, the protection and enhancement of coastal blue carbon can be an important contribution to both mitigation and adaptation at the national scale. The feasibility of climate mitigation by open ocean fertilization of productivity is limited to negligible, due to the likely decadal-scale return to the atmosphere of nearly all the extra carbon removed, associated difficulties in carbon accounting, risks of unintended side effects and low acceptability. Other human interventions to enhance marine carbon uptake, e.g., ocean alkalinisation (enhanced weathering), would also have governance challenges, with the increased risk of undesirable ecological consequences (*high confidence*). {5.5.1.2}

Socio-institutional adaptation responses are more frequently reported in the literature than ecosystem-based and built-infrastructure approaches. Hard engineering responses are more effective when supported by ecosystem-based adaptation approaches (*high agreement*), and both approaches are enhanced by combining with socio-institutional approaches for adaptation (*high confidence*). Stakeholder engagement is necessary (*robust evidence, high agreement*). {5.5.2}

Ecosystem Based Adaptation is a cost-effective coastal protection tool that can have many co-benefits, including supporting livelihoods, contributing to carbon sequestration and the provision of a range of other valuable ecosystem services (*high confidence*). Such adaptation does, however, assume that the climate can be stabilised. Under changing climatic conditions there are limits to the effectiveness of ecosystem based adaptation, and these limits are currently difficult to determine. {5.5.2.1}

Socio-institutional adaptation responses, including community-based adaptation, capacity-building, participatory processes, institutional support for adaptation planning and support mechanisms for communities are important tools to address climate change impacts (*high confidence*). For fisheries management, improving coordination of integrated coastal management and marine protected areas have emerged in the literature as important adaptation governance responses (*robust evidence, medium agreement*).{5.5.2.2, 5.5.2.6}

Observed widespread decline in warm water corals has led to the consideration of alternative restoration approaches to enhance climate-resilience. Approaches, such as 'coral reef gardening' have been tested, and ecological engineering and other approaches such as assisted evolution, colonization and chimerism are being researched for reef restoration. However, the effectiveness of these approaches to increase resilience to climate stressors and their large-scale implementation for reef restoration will be limited unless warming and ocean acidification are rapidly controlled (*high confidence*). {Box 5.5, 5.5.2}

Existing ocean governance structures are already facing multi-dimensional, scale-related challenges because of climate change. This trend of increasing complexity will continue (*high confidence*). The mechanisms for the governance of marine Areas Beyond National Jurisdiction, such as ocean acidification, would benefit from further development (*high confidence*). There is also scope to increase the overall effectiveness of international and national ocean governance regimes by increasing cooperation, integration and widening participation (*medium confidence*). Diverse adaptations of ocean-related governance are being tried, and some are producing promising results. However, rigorous evaluation is needed of the effectiveness of these adaptations in achieving their goals. {5.5.3}

There are a broad range of identified barriers and limits for adaptation to climate change in ecosystems and human systems (*high confidence*). Limitations include the space that ecosystems require, non-climatic drivers and human impacts that need to be addressed as part of the adaptation response, the lowering of adaptive capacity of ecosystems because of climate change, and the slower ecosystem recovery rates relative to the recurrence of climate impacts, availability of technology, knowledge and financial support, and existing governance structures (*medium confidence*). {5.5.2}

5.1   Introduction

The ocean is a key component of the Earth System (Chapter 1) as it provides essential life-supporting services (Inniss et al., 2017). For example, it stores heat trapped in the atmosphere caused by increasing concentrations of greenhouse gases, it masks and slows surface warming, it stores excess carbon dioxide and is an important component of global biogeochemical cycles. The ocean is the home to the largest continuous ecosystem, provides habitats for rich marine biodiversity, is an essential source of food and contributes to human health, livelihood and security. The ocean also supports other services to humans, for example, transport and trade, tourism, renewable energy, and cultural services such as aesthetic appeal, local and traditional knowledge and religious practices. Governance of the ocean has a unique set of challenges and opportunities compared with land systems and requiring different treatment under a changing climate.

The IPCC Fifth Assessment Report (AR5) from Working Group I (WGI) showed that there are ongoing changes to the physical and chemical state of the ocean. AR5 WGI report (IPCC, 2013) concluded that (1) 'ocean warming dominates the increased energy stored in the climate system with more than 90% of the energy accumulated since 1971'; (2) 'the ocean has absorbed about 30% of the emitted anthropogenic carbon causing ocean acidification' since pre-industrial times; and (3) it is '*extremely likely* that human influence has been the dominant cause of warming since mid 20th century'.

The IPCC AR5 Working Group II (WGII) concluded that changes in the ocean such as warming, acidification and deoxygenation are affecting marine life from molecular processes to organisms and ecosystems, with major impacts on the use of marine systems by human societies (Pörtner, 2012). IPCC Special Report on the Impacts of Global
Warming of 1.5°C above pre-industrial levels and related global greenhouse gas emission pathways (SR1.5) also concluded that reducing these risks by 'limiting warming to 1.5°C above pre-industrial levels would require transformative systemic change, integrated with sustainable development' and that 'adaptation needs will be lower in a 1.5°C world compared to a 2°C world.' (de Coninck et al., 2018; Hoegh-Guldberg et al., 2018).

This report updates earlier assessments, evaluating new research and knowledge regarding changing ocean climate and ecosystems, risks to ecosystem services, and vulnerability of the dependent communities including governance. It also delves into changes the ocean that were beyond the scope of the previous reports. Radiation management techniques (also known as sunlight reflection methods) are excluded here. Such geo-engineering approaches are addressed in the SR1.5. However, natural carbon uptake and stores in the marine environment are included (Section 5.5.1).

The chapter is structured around three guiding questions:
• What are the key changes in the physical and biogeochemical properties of the ocean (Section 5.2.2)?
• How have these changes impacted key ecosystems, risks to ecosystems services and human wellbeing (Section 5.2.3, 5.2.4, 5.3, 5.4)?
• Are there effective pathways for adaptation and nature-based solutions to risk reduction for marine dependent communities (Section 5.5)?

This chapter covers both regional and global scales and across natural and human systems. Chapter 3 covers the Polar Regions, including their oceans, Chapter 4 covers sea level and its implications, and Chapter 6 covers extremes and abrupt events. This chapter uses IPCC calibrated language around scientific uncertainty, as described in Section 1.8.3. Two emissions scenarios, RCP2.6 and RCP8.5, are used for projections of climate change (RCP means Representative Concentration Pathway; see Cross-Chapter Box 1 in Chapter 1).

5.2   Changing Oceans and Biodiversity

5.2.1   Introduction

This section assesses changes in the ocean. It includes the physical and chemical properties (Section 5.2.2), their impacts on the pelagic ecosystem (Section 5.2.3) and deep seafloor system (Section 5.2.4). In this

assessment, the open ocean and deep seafloor includes areas where the water column is deeper than 200 metres; it is the main subject of Section 5.2. Coastal and shelf seas are primarily discussed in Section 5.3.

5.2.2   Changes in Physical and Biogeochemical Properties

5.2.2.1   Introduction to Changing Open Ocean

The ocean is getting progressively warmer, with parallel changes in ocean chemistry such as acidification and oxygen loss, as documented in the AR5 (Rhein et al., 2013). The global scale warming and acidification trends are readily detectable in oceanic observations, well understood scientifically, and consistently projected by Earth Systems Models (ESMs). Each of these has been directly attributed to anthropogenic forcing from changing concentrations of greenhouse gases and aerosols (Bindoff et al., 2013). These trends in the global average ocean temperature will continue for centuries after the anthropogenic forcing is stabilized (Collins et al., 2013).

The impacts on ocean ecosystems and human societies are primarily driven by regional trends and by the local manifestation of the global-scale changes. At these smaller scales, the temperature, acidification, salinity, nutrient and oxygen concentrations in the ocean are also expected to exhibit basin and local-scale changes. However, the ocean also has significant natural variability at local and basin scales with timescales from minutes to decades and longer (Rhein et al., 2013), which can mask the underlying observed and projected trends (see Box 5.1). The impact of multiple stressors on marine ecosystems is one of the main subjects of this chapter (Section 5.2.3, 5.2.4, 5.3), including new evidence and understanding since the last assessment report (e.g., Gunderson et al., 2016). The most severe impacts of a changing climate will typically be experienced when conditions are driven outside the range of previous experience at rates that are faster than human or ecological systems can adapt (Pörtner et al., 2014; Box 5.1).

This section summarizes our emerging understanding of the primary changes to the ocean, along with an assessment of several key areas of scientific uncertainty about these changes. Because many of these long-term trends have already been extensively discussed in previous assessments (IPCC, 2013), much of this summary of the physical changes is brief except where there are significant new findings.

5.2.2.2   Changing Temperature, Salinity, Circulation

Historically, scientific research expeditions starting in the 19th century have provided occasional sections measuring deep ocean properties (Roemmich et al., 2012). Greater spatial and temporal coverage of temperatures down to about 700 m was obtained using expendable bathythermographs (XBTs) along commercial shipping tracks starting in the 1970s (Abraham et al., 2013). Since the early 2000s, thousands of autonomous profiling floats (Argo floats) have provided high-quality temperature and salinity profiles of the upper 2000 m in ice-free regions of the ocean (Abraham et al., 2013; Riser et al., 2016). Further advances in autonomous floats have been developed that now allow these floats to operate in seasonally ice-covered oceans (Wong and Riser, 2011; Wong and Riser, 2013), and more recently to profile the entire depth of the water column down to 4000 or 6000 m (Johnson et al., 2015; Zilberman, 2017) and to include biogeochemical properties (Johnson et al., 2017). Autonomous floats have revolutionized our sampling and accuracy of the global ocean temperature and salinity records and increased certainty and confidence in global estimates of the earth heat (temperature) budget, particularly since 2004 (Von Schuckmann et al., 2014; Roemmich et al., 2015; Riser et al., 2016), as demonstrated by the convergence of observational estimates of the changes in the heat budget of the upper 2000 m (Figure 5.1). New findings using data collected from such observing platforms mark significant progress since AR5.

To understand the recent and future climate, we use ensembles of coupled ocean-atmosphere-cryosphere-ecosystem models (ESMs) with the full-time history of atmospheric forcing (greenhouse gases, aerosols, solar radiation and volcanic eruptions) for the historical period and projections of the concentrations or emissions of these forcings to 2100. For these projections the RCPs of atmospheric emissions scenarios are used as specified by the Climate Model Inter-comparison Project, Phase 5 (CMIP5) (see Section 1.8.2.3,

Cross-Chapter Box 1, and also IPCC AR5)[3]. This chapter focuses on the low and high emissions scenarios RCP2.6 and RCP8.5, respectively. When these scenarios are used to drive ESMs, it is possible to simulate the recent and future patterns of changes in the ocean temperature, salinity and circulation (and other oceanic properties such as ocean oxygen concentration and acidification, Section 5.2.2.3 and 5.2.2.4). Finally, the projections of ocean changes also informs the detection, attribution and projection of risk and impacts on ecosystems (Sections 5.2.3, 5.2.4 and 5.3), ecosystem services (Section 5.4.1) and human wellbeing (Section 5.4.2) under climate change.

5.2.2.2.1   *Observed and projected global ocean heat uptake*

As AR5 concluded, the ocean is warming as a direct result of anthropogenic changes to the radiative properties of the atmosphere and the heat budget of the Earth (*very likely*) (Bindoff et al., 2013). Over the past few decades our ocean observing system has measured an increase in ocean temperature (Figure 5.1). This temperature increase corresponds to an uptake of over 90% of the excess heat accumulated in the Earth system over this period (Bindoff et al., 2013; Rhein et al., 2013). This heat in the ocean also causes it to expand and has contributed about 43% of the observed global mean sea level rise from 1970-2015 (Section 4.2.2.3.6).

Since AR5, there have been further improvements in our ability to understand and correct instrumental errors and new estimates also attempt to minimize biases in estimating temperature changes arising from traditional data-void filling strategies (Abraham et al., 2013; Durack, 2015; Cheng and Chen, 2017; Cheng et al., 2017). New estimates from ocean observations of ocean heat uptake in the top 2000 m between 1993 and 2017 *very likely* range from 9.2±2.3 ZJ yr^{-1} to 12.1±3.1 ZJ yr^{-1} (Johnson et al., 2018)[4]. Three recent independent estimates do a better job of accounting for instrumental biases and the sparseness of historical ocean temperature measurements than the older studies assessed in AR5, and provide larger and more consistent estimates of heat uptake rates for the 0-2000m layer of 5.8±1.0 ZJ yr^{-1} (Cheng and Chen, 2017; Cheng et al., 2017; Ishii et al., 2017), 6.0±0.8 ZJ yr^{-1} (updated from Domingues et al. (2008)) and 6.3±1.8 ZJ yr^{-1} (Cheng and Chen, 2017; Cheng et al., 2017; Ishii et al., 2017) for the 1971-2010 period assessed by AR5.

Based on these new published methods and revised atlases we update the estimates for ocean heat uptake (Table 5.1, and SM 5.1). For all of the periods assessed in Table 5.1, it is *virtually certain* that the upper ocean (0-700 m) has warmed. These results are consistent with earlier research into the duration of record needed to detect a significant signal in global ocean heat content (Gleckler et al., 2012). Critically, the *high confidence* and *high agreement* in the ocean temperature data means we can detect discernable rates of increase in ocean heat uptake (Gleckler et al., 2012; Cheng et al., 2019). The rate of heat uptake in the upper ocean (0-700m) is *very likely* higher in the 1993-2017 (or 2005-2017) period compared with the 1969-1993 period (see Table 5.1). The deeper layer (700-2000 m) heat uptake rate is *likely* to be higher in the 1993-2017 period compared with the 1969-1993 period. Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)Flato et al. (2013)

Table 5.1: The assessed rate of increase in ocean heat content in the two depth layers 0-700 m and 700-2000 m and their *very likely* ranges. Fluxes in Wm^{-2} are averaged over the Earth's entire surface area. The four periods cover earlier and more recent trends; the 2005-2017 period has the most complete interior ocean data coverage and the greatest consistency between estimates, while longer trends are better for distinguishing between forced changes and internal variability. These observationally-estimated rates come from an assessment of the recent research (See Supplementary

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[3] The 30 CMIP5 ESMs used in here in various contexts were selected based on the availability of ocean data from the historical period, RCP2.6 and RCP8.5 projections, and corresponding control runs to correct for model drift. The models used include: ACCESS1.0, ACCESS1.3, BNU-ESM, BCC-CSM1-1, CCSM4, CESM1, CMCC-CESM, CMCC-CMS, CNRM-CM5, CSIRO-Mk3, CanESM2, FGOALS-S2.0, GFDL-CM3, GFDL-ESM2G, GFDL-ESM2M, GISS-E2-H, GISS-E2-R, HadGEM2-AO, HadGEM2-CC, HadGEM2-ES, INM-CM4, IPSL-CM5A-LR, IPSL-CM5A-MR, IPSL-CM5B-LR, MIROC-ESM, MIROC5, MPI-ESM-LR, MPI-ESM-MR, MRI-CGCM3, and NorESM1-M. Up to 3 ensemble members or variants were included per model, and all changes are relative to a control run with an identical initial condition but with preindustrial forcing. A table with a description and citations for each of these models, along with more detailed discussion of the use of ESM output, can be found in Flato et al. (2013).
[4] ZJ is Zettajoule and is equal to 10^{21} Joules. Warming the entire ocean by 1°C requires about 5500 ZJ; 144 ZJ would warm the top 100 m by about 1°C.

material SM5.1), while the CMIP5 ESM estimates are based on a combined 28-member ensemble of historical, RCP2.6 and RCP8.5 simulations.

Period	Ocean Heat Uptake Rate, ZJ yr^{-1}				Ocean Heat Uptake as Average Fluxes, W m^{-2}			
	1969-1993	1993-2017	1970-2017	2005-2017	1969-1993	1993-2017	1970-2017	2005-2017
Observationally Based Ocean Heat Uptake Estimates:								
0-700m	3.22±1.61	6.28±0.48	4.35±0.80	5.31±0.48	0.20±0.1	0.39±0.03	0.27±0.05	0.33±0.03
700-2000m	0.97±0.64	3.86±2.09	2.25±0.64	4.02±0.97	0.06±0.04	0.24±0.13	0.14±0.04	0.25±0.06
CMIP5 ESM Ensemble-mean Ocean Heat Uptake with 90% Certainty Range from Ensemble Spread:								
0-700m	3.60±1.92	7.37±2.09	5.64±1.90	7.85±2.71	0.22±0.12	0.46±0.13	0.35±0.12	0.49±0.17
700-2000m	1.32±1.49	2.72±1.41	1.99±1.51	3.33±1.75	0.08±0.09	0.17±0.09	0.12±0.09	0.21±0.11

The direct comparison of the observed changes in ocean heat content and the simulated historical changes is undertaken to detect climate change, to attribute the causes of climate change to the forcings in the system, and to evaluate the performance of ESMs. Attribution studies also reject competing hypotheses to explain the global ocean changes such as natural forcing from solar variability or volcanic eruptions (see Section 1.3) (Bindoff et al., 2013). Detection and attribution studies have since been used to detect changes in the rate of ocean heat uptake and to attribute these changes to human activity (Gleckler et al., 2016).

Updated observationally-based estimates of ocean heat uptake are consistent with simulations of equivalent time-periods from an ensemble of CMIP5 ESMs (Table 5.1 and the inset panel in Figure 5.1) (*high confidence*), once the limitations of the historical ocean observing network and the internally generated variability with a single realization of the real world are taken into account (see Section 5.2.2.2). Following the CMIP5 protocol, the ESMs are radiatively forced with observationally derived estimates of greenhouse gas concentrations and aerosols, including natural forcing variations from volcanic eruptions and solar forcing, through 2005; after 2006 each of the ESMs uses either the RCP2.6 or RCP8.5 emissions scenarios. The *very likely* ranges of the observed trends of heat uptake for the four periods and two layers all fall within the *very likely* range of simulated heat uptake from the ESM ensemble (Table 5.1). The difference between observations and average of the simulations in the upper ocean is an overestimate of heat uptake by about 20% and for the deeper layer there an underestimate by a similar amount, but this difference is still well within the *very likely* range from the ensemble of simulations. The overall consistency between observationally-based estimates and ESM simulations of the historical period gives greater confidence in the projections; it is *very likely* that historical simulations agree with observations of the global ocean heat uptake (Table 5.1).

While the collection of the worlds' ESMs have been criticized for having an ensemble mean that does not exhibit the observed 'hiatus' or 'slowdown' of global mean surface temperature increase in the early 21st century (Meehl et al., 2011; Trenberth et al., 2016) , it is increasingly clear that this is at least in part due to the redistribution of heat within the climate system from the surface into the interior ocean and between ocean basins. Individual realizations of ESMs do show decades with slow increases in mean surface temperature change comparable to what was observed, even though these cases exhibit continued interior ocean heat uptake, and every ensemble member exhibits surface warming closer to the ensemble-mean over multidecadal timescales (Meehl et al., 2011; England et al., 2015; Knutson et al., 2016).
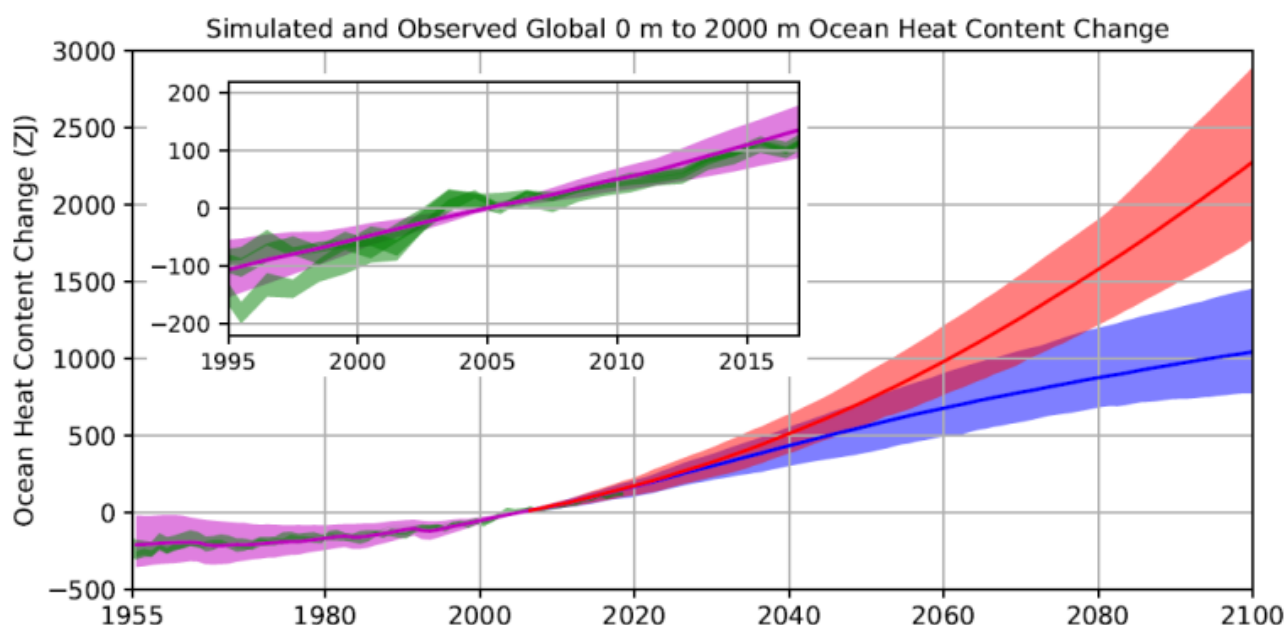
Figure 5.1: Time series of globally integrated upper 2000 m ocean heat content changes in ZJ, relative to the 2000 to 2010 period average, as inferred from observations (green) and as simulated for historical (magenta), RCP2.6 (blue) and RCP8.5 (red) forcing by a 25-member ensemble of CMIP5 ESMs ((Cheng et al., 2019). The shaded green in the outer panel is the *very likely* range determined by combining data from 4 long-term estimates (Green lines; Palmer et al., 2007; Levitus et al., 2012; Lyman and Johnson, 2014; Cheng and Chen, 2017; Cheng et al., 2017; Ishii et al., 2017) processed as in Johnson et al. (2018). The magenta, blue and red lines are the ESM ensemble means, while shading shows each ensemble's 5th to 95th percentile range. In the inset subpanel, the 4 different shaded green areas are the reported *very likely* range of heat content changes as inferred from observations by 4 independent groups (Green lines; Palmer et al., 2007; Lyman and Johnson, 2014; Cheng and Chen, 2017; Cheng et al., 2017; Ishii et al., 2017) processed as in Johnson et al. (2018). In the inset subpanel the RCP2.6 and RCP8.5 projections after 2005 are combined into a single ensemble with the historical simulations.

The ocean will continue to take up heat in the coming decades for all plausible scenarios. As depicting in Figure 5.1, the ensemble of CMIP5 ESMs used by Cheng et al. (2019) project that under RCP2.6, the top 2000 m of the ocean will take up 935 ZJ of heat between 2015 and 2100 (with a *very likely* range of 650 to 1340 ZJ based on the 5th and 95th percentiles of the 25 ESMs used here that have available data from the historical, scenario and control runs for RCP2.6). Under RCP8.5 this ensemble projects heat uptake of 2180 ZJ (with a *very likely* range of 1710 to 2790 ZJ, based on 35 ESMs) between 2015 and 2100. By 2100 the ocean is *very likely* to warm by 2 to 4 times as much for low emissions (RCP2.6) and 5 to 7 times as much for the high emissions scenario (RCP8.5) compared with the observed changes since 1970. With the RCP8.5 scenario, the ocean is *very likely* to take up about twice as much heat as RCP2.6 (Figure. 5.1). Even under RCP2.6 the ocean will continue to warm for several centuries to come (Collins et al., 2013). It is *virtually certain* that the ocean will continue to take up heat throughout the 21st century, and the rate of uptake will depend upon on the emissions scenario we collectively choose to follow.
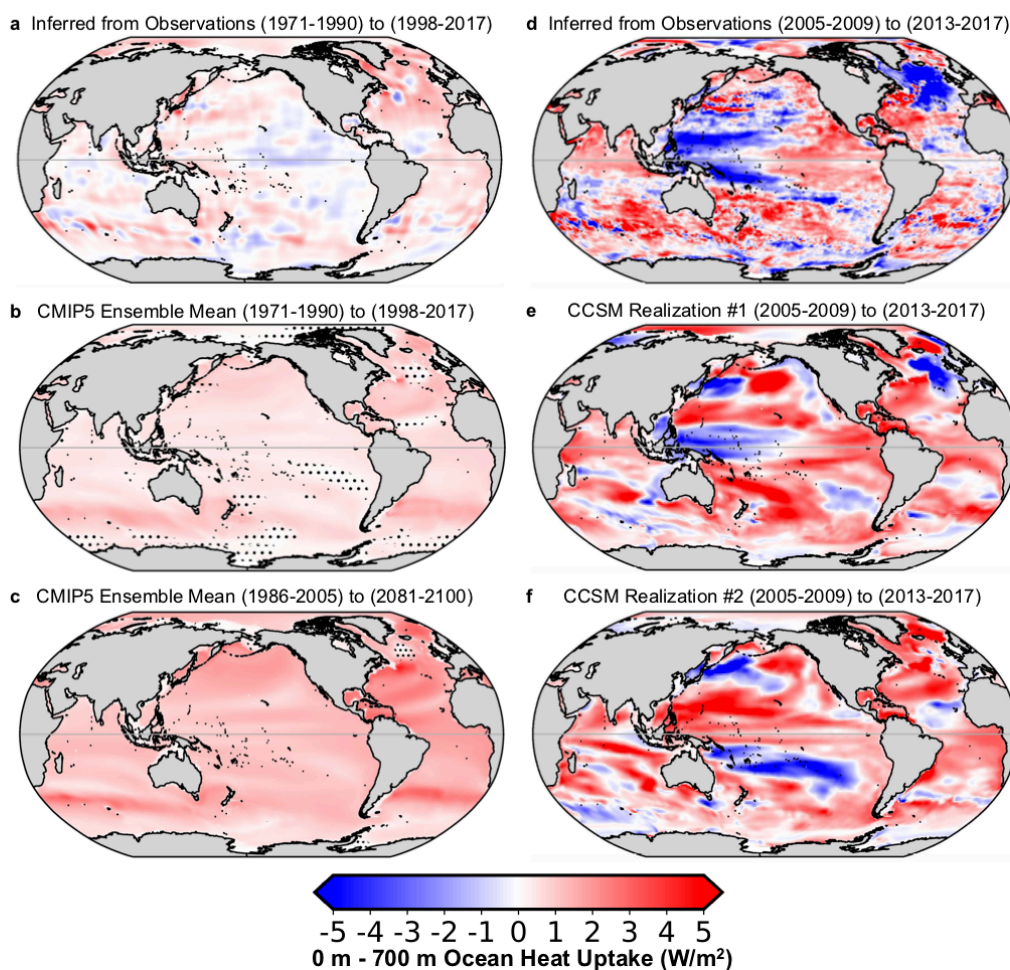
Figure 5.2: Heat uptake by the top 700 m of the ocean, as determined by differences between the averages over two 5- or 20-year intervals converted to a heat flux into the ocean (W m^{-2}), either from observationally-based analyses or a 38-member ensemble of CMIP5 ESMs. (a) Change between (1971-1990) and (1997-2016) as inferred from observations (Good et al., 2013); (b) The ensemble mean change in CMIP5 ESMs for the same time periods as in (a); (c) Projected ensemble mean change in CMIP5 ESMs between (1986-2005) and (2081-2100) for the RCP8.5 forcing scenario. In panels (b) and (c), stippling indicates regions where the ensemble mean change is not significantly different from 0 at the 95% confidence level based on the models' temporal variability. (d) Change between (2004-2008) and (2013-2017) as inferred from observations by the SODA 3.4.2 reanalysis product (Carton et al., 2018); (e) and (f) Estimates of change in heat uptake as in (d) but from two individual realizations of the CCSM ESM. These two realizations are identical apart from their initial conditions, which leads to different timing in their internal modes of variability; they were selected from the full CMIP5 ensemble as examples where one is reminiscent of the recent observed changes while the other has regional changes that have dissimilar timing.

5.2.2.2.2   Structure of anthropogenic climate changes in the ocean
The ensemble average of the CMIP5 ESMs projects widespread ocean warming over the coming century, concentrated in the upper ocean (Figures 5.2c and 5.3) (Kuhlbrodt and Gregory, 2012). The anthropogenic heat will penetrate into the ocean following well-established circulation pathways (Jones et al., 2016a). The greatest vertically integrated heat uptake occurs where there is already the formation of interior waters, such as Antarctic Intermediate Water along the Antarctic Circumpolar Current (Frölicher et al., 2015) or North Atlantic Deep Water precursors in the Nordic Seas (Figure 5.2c), but all water-masses[5] that are subducted

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[5] Following common oceanographic practice dating back to Helland-Hansen (1916) and discussed in detail by Sverdrup et al. (1942), an ocean water-mass is defined as a large volume of seawater with a characteristic range of temperature and salinity properties, typically falling along a line in temperature-salinity space, often with common formation processes and locations.

over decades are expected to experience significant warming (see Figure 5.3). The warming in the subtropical gyres penetrates deeper into the ocean than other gyres (roughly 15°N to 45°N and 15°S to 45°S in Figure 5.3), following the wind-driven bowing down of the density surfaces (the solid lines in Figure 5.3) in these gyres (Terada and Minobe, 2018). The greater warming at 700-2000 m in the Atlantic than the Pacific or Indian Oceans (Figure 5.3) reflects the strong southward transport of recently formed North Atlantic Deep Water at these depths by the Atlantic Meridional Overturning Circulation (AMOC). Two areas that commonly exhibit substantially reduced near-surface warming over the course of the 21st century are the northern North Atlantic, where a slowing AMOC (see Section 6.7.1.1) reduces the northward heat transport and brings the surface temperatures closer to what is found in other ocean basins at these latitudes (Collins et al., 2013), and the southern side of the Southern Ocean, where water upwells that has been submerged for so long that it has not yet experienced significant anthropogenic climate change (Armour et al., 2016). Most of these projected warming patterns are broadly consistent across the current and previous generations of climate models (Mitchell et al., 1995; Collins et al., 2014) as well as observations and theoretical understanding. These multiple lines of evidence give *high confidence* that the projections describe the changes in the real world (*high agreement, robust evidence*).
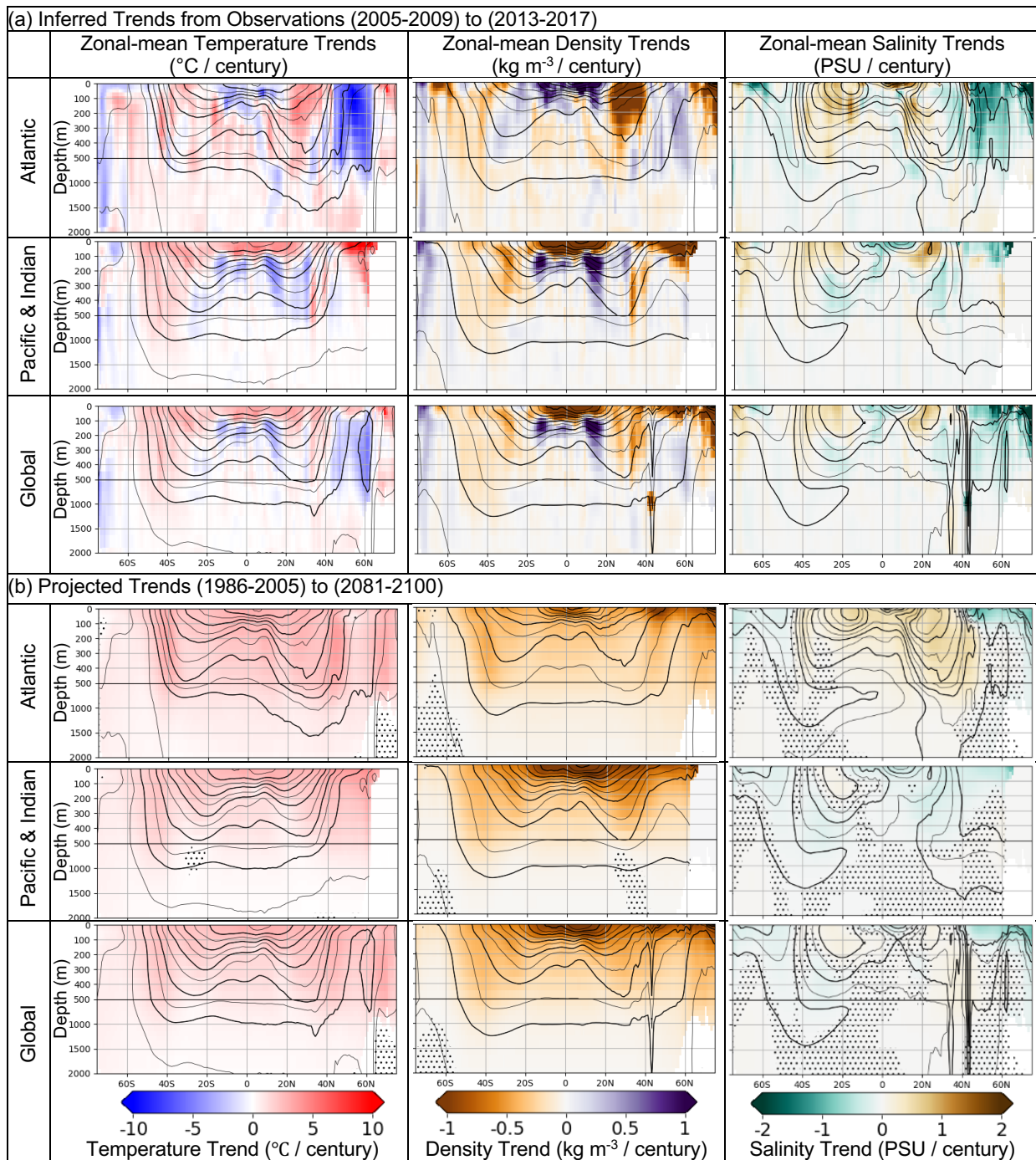
Figure 5.3: Side-view basin-averaged zonal-mean trends (change per century) in water-mass properties in the top 2000 m by basin (a) as inferred from observations (average of 2013 to 2017 minus average of 2005 to 2009) and (b) CMIP5 model projections with RCP8.5 forcing (average of 2081 to 2100 minus average of 1981 to 2000) trends in water-mass changes forcing. Subpanels within each group: top-to-bottom (Atlantic, combined Pacific and Indian, Global); left-to-right (Temperature, *In situ* Density, Salinity). Shaded areas show where the projected changes are not statistically significant at the 95% level. This figure uses the same observationally-derived reanalysis datasets and ensemble of ESMs as in Fig. 5.2c and 5.2d. Solid lines show present contours of these fields; the curious structure in the global-zonal mean contours of density and salinity are due to the relatively salty Mediterranean and fresh Black seas.

The near surface salinity of the ocean is both observed and projected to evolve in ways that reflect the increased intensity of the Earth's hydrologic cycle (Durack, 2015) and the increasing near-surface ocean stratification (Zika et al., 2018). As described in WGI AR5, the ocean surface in areas that currently have net evaporation are expected to become saltier, while areas with net precipitation are expected to get fresher (Rhein et al., 2013), as the patterns of precipitation and evaporation are generally expected to be amplified (Held and Soden, 2006). At longer timescales of decades, the larger scale changes in the ocean circulation

and basin-integrated freshwater imbalances emerge in the near-surface salinity changes, as shown in Figure 5.3b, with an increasingly salty tropical and subtropical Atlantic and Mediterranean contrasting with a freshening Pacific and polar Arctic emerging as robust signals across the suite of ESMs (Collins et al., 2013). The freshening of the high latitudes in the North Atlantic and Arctic basin is consistent with the widely expected weakening of the AMOC (also discussed in Section 6.7), hydrological cycle changes and a decline in the volume of sea ice (discussed in Section 3.2.2).

Projected salinity changes in the subsurface ocean reflect changes in the rates of formation of water-masses or their newly formed properties (Purich et al., 2018). Thus, projected freshening of the Southern Ocean surface leads to a freshening of the Antarctic Intermediate Water that is subducted there, flowing northward from the Southern Ocean as a relatively fresh water-mass at depths of 500-1500 m (Figure 5.3b). Increased surface salinity in the Atlantic subtropical gyres are pumped into the interior by the winds, leading to an increased salinity of the interior subtropical gyres, along with contributions from increasingly salty Mediterranean water (Jordà et al., 2017). Conversely, freshwater capping of the northwestern North Atlantic is projected to inhibit deep convection in the Labrador Sea and the consequent production of Labrador Sea Water in some models (Collins et al., 2013), and contributes to the increased salinity of the North Atlantic between 1000 m and 2000 m depths (Figure 5.3b).

Identifying the specific patterns of anthropogenic climate changes in oceanic observations is complicated by the presence of basin-scale natural variability with timescales ranging from tidal to multi-decadal, and due to the difficulties associated with maintaining high-precision observing systems spanning the ocean basins and limited observational coverage of the extratropical Southern Hemisphere before 2006 (Rhein et al., 2013). Inferences based on oceanographic observations from the 1970s onward show wide-spread warming of the upper 700 m (Figure 5.2a), in broad agreement with the ensemble of historical CMIP5 ESM simulations (Figure 5.2b). These ESMs indicate that anthropogenic regional warming over the past half-century should be discernable at the 95% confidence level in much of the upper oceans (un-stippled areas in Figure 5.2b). Most of the areas where observational analyses (Figure 5.2a) exhibit long-term cooling are either regions where the internally generated variability is large enough to mask the trends (e.g., the Eastern Tropical Pacific, Northwest Atlantic, and Kurushio extension east of Japan, which are stippled in Figure 5.2b), or where the observational coverage early in the record is limited and different analyses can disagree about trends (e.g., the Southern Ocean and extratropical South Pacific). When internal variability is taken into account, the broad consistency in the magnitude and regional distribution of observed and simulated 50-year trends gives confidence to the ESM projections of longer-term oceanic changes described previously.

Detailed regional patterns of trends in temperature and heat content at depths of 0 m to 2000 m during the early 21st century are consistent in various analysis, owing to the improved observing network (Roemmich et al., 2015; Desbruyères et al., 2016a) (Figure 5.2d). At depths of 700–2000 m, observations in all of the ocean basins show broadly warming trends in the well-observed Argo era (2006-present), with particularly significant warming patterns in the Southern Hemisphere extratropics around 40° S and the subpolar North Atlantic (Figure 5.3a). These observed changes support the notion that deep ocean heat content has been continuously increasing. As a result, regional climate change signatures emerge from confounding natural variability sooner in the 700-2000 m depth range than in upper 700 m of the ocean, where interannual modes of variability have a larger influence on the circulation (for a more complete discussion see Johnson et al. (2018)). Despite regional patches of cooling water in the upper 700 m (Figure 5.2d), every one of the world's ocean basins volume averaged over depths of 0 m to 2000 m has experienced significant warming over the last decade (Figure 5.3, and also Desbruyères et al. (2016a)). The greatest warming of the top 2000 m has been in the Southern Ocean (Roemmich et al., 2015; Trenberth et al., 2016), the tropical and subtropical Pacific Ocean (Roemmich et al., 2015), and the tropical and subtropical Atlantic Ocean (Cheng and Chen, 2017). The Southern Hemisphere extratropical oceans accounted for 67%-98% of the total ocean heat increase in the uppermost 2000 m for the period of 2006 to 2013 (Roemmich et al., 2015). Shi et al. (2018) suggest that the dominant ocean heat uptake by the Southern Hemisphere in the early 21st century is expected to become more balanced between the hemispheres as the asymmetric cooling by aerosols decreases.

Large-scale patterns of natural variability at interannual to decadal time scales can mask the long-term warming trend in the upper 700 m, particularly in the tropical Pacific and Indian Oceans (England et al., 2014; Liu et al., 2016) and in the North Atlantic (Buckley and Marshall, 2015). The most significant upper

700 m warming between five-year averages centered on 2007 and 2015 occurred in a large extratropical band of the Southern Hemisphere between 30º S and 60º S, and in the tropical Indian Ocean, the eastern North Pacific and western subtropical North Atlantic (Figure 5.2d). Warming of the southern hemisphere subtropical gyres is driven, in part, by an intensification of Southern Ocean winds in recent decades, facilitating the penetration of heat to deeper depths (Gao et al., 2018). Marginal seas, such as the Mediterranean and Red seas have also exhibited notable warming. Conversely, over this timeframe there were also regions of cooling in the upper 700 m, notably in the North Atlantic around 40º N-60º N and in the western tropical Pacific (Figure 5.2d). Recent relatively cold and fresh surface and subsurface conditions in the North Atlantic have been attributed to anomalous atmospheric forcing (Josey et al., 2018) or weakened transport by the North Atlantic Current and AMOC (Smeed et al., 2018), and in turn may have contributed to an intensification of deep convection in the Labrador Sea since 2012 (Yashayaev and Loder, 2017). All these observed decadal changes can be related to internal decadal variability (Robson et al., 2014; Yeager et al., 2015) even though they resemble expected longer-term anthropogenically forced trends. Substantial decadal-scale warming and cooling trends in the tropical Pacific and Indian oceans can arise from natural El Niño-Southern Oscillation (ENSO) and Indian Ocean Dipole variability (Han et al., 2014). Large ensembles of freely running CMIP5 ESM simulations also show that internal variability can dominate the regional manifestation of the anthropogenic climate signal on decadal timescales (Kay et al., 2014). This is illustrated by the differing warming trends in Figure 5.2e and 5.2f from two identical ESMs that differ only in the weather in their 1850 initial conditions, averaged over the whole 21st century, by contrast, the ensemble of CMIP5 models project statistically significant anthropogenic regional upper 700 m heat content trends almost everywhere (Figure 5.2c).

There are well documented changes in observed ocean temperatures and salinities (Abraham et al., 2013; Ishii et al., 2017). However, attributing these changes in the state of the ocean to anthropogenic causes can be challenging due to the presence of internally generated variability, which can swamp the underlying climate change signal in short records and on regional scales. As can be seen in Figure 5.2, the observed long-term trends (Figure 5.2a) exhibit a striking similarity to the CMIP5 ensemble mean in areas where the models suggest that anthropogenic changes should be statistically significant (Figure 5.2b). However, the trends in the shorter well-observed period covering 2005 to 2017 (Figure 5.2d) exhibits strong trends from internal variability, as illustrated by the differences of two ensemble members of the same ESM with the same forcing but initialized with different weather (Figure 5.2e and 5.2f). Detection and Attribution studies take the internal variability into account and separate the underlying climate signals with the same spatio-temporal sampling as the observations, and apply a range of statistical tests to determine the coherence of the observations with the co-sampled observations (Bindoff et al., 2013; AR5 WG1 Box 10.1).

Since AR5, the use of different and updated oceanographic data sets and increase in the number of ensembles of the CMIP5 simulations (Kay et al., 2014) has improved the overall detection and attribution of human influence. Together these measures increase the coherence of the simulations and reduce noise. For example, an isotherm approach used to reduce the noise from the displacement of isotherms in the upper water column allowing detection in each of the mid-latitude ocean basins was achieved on 60 year time series (Weller et al., 2016). Using all the available ocean temperature and salinity profiles from the Southern Ocean, Swart et al. (2018) show that the warming and freshening patterns were consistent primarily with increased human induced greenhouse gases and secondarily from ozone depletion in the stratosphere, but inconsistent with internal variability. Together the evidence from the AR5, and the discussion above with the new evidence on regional scales across the global oceans, we conclude that the observed long-term upper ocean temperature changes are *very likely* to have a substantial contribution from anthropogenic forcing.

The wind-driven ocean circulation at the end of the 21st century is expected to be qualitatively similar to that in the present-day, even as important buoyancy-loss driven overturning circulations are expected to weaken. ESM projections suggest that some major ocean current transports will exhibit a modest increase (such as the Kuroshio Extension (Terada and Minobe, 2018) or a small decrease (such as for the Indonesian Throughflow (Sen Gupta et al., 2016); many predominantly wind-driven current-system transports are expected to exhibit smaller than 20% changes by 2100 with RCP8.5 forcing. Climate-change induced changes of the circulation in other mid-latitude basins may be difficult to detect or reliably project because of significant natural variability at inter-annual (e.g., El Niño) to decadal (e.g., the Pacific Decadal Oscillation) timescales. The

Antarctic Circumpolar Current is projected to be subject to strengthening westerly winds and substantially reduced rates of Antarctic Bottom Water (AABW) formation, as assessed in the Cross-Chapter Box 7 in Chapter 3. The heat transported by the buoyancy-loss driven AMOC, in particular, contributes to the relatively clement climate of northern Europe and the North Atlantic Basin as a whole, although the wind-driven ocean gyres also contribute to the meridional ocean heat transport (see the review by Buckley and Marshall (2015). As a result, there is a concern that significant changes in ocean circulation could lead to localized climate changes that are much larger than the global mean. Projected and observed changes in the AMOC and the rates of formation of deep water-masses in the North Atlantic are discussed in Chapter 6.7.1, along with the possibility of abrupt or enduring changes resulting from forcing by Greenlandic meltwater. A significant reduction in AMOC would, in turn, modestly weaken the Gulf Stream transport, which also has a substantial wind driven component (Frajka-Williams et al., 2016). Most aspects of the large-scale wind-driven ocean circulation are *very likely* to be qualitatively similar to the circulation in the present-day, with only modest changes in transports and current location.

The global ocean below 2000 m has warmed significantly between the 1980s and 2010s (Figure 5.4), contributing to ocean heat uptake and through thermal expansion to sea level rise (Purkey and Johnson, 2010; Desbruyères et al., 2016b). The observed deep warming rate varies regionally and by depth reflecting differences in the waters influencing particular regions. The deep and abyssal North Atlantic, fed by North Atlantic Deep Water (NADW), has reversed from warming to cooling over the past decade, possibly associated with the North Atlantic Oscillation (e.g., Yashayaev, 2007; Desbruyères et al., 2014) or longer-term weakening in North Atlantic overturning circulation (Caesar et al., 2018; Thornalley et al., 2018). The strongest warming is observed in regions of the deep ocean Antarctic bottom water (AABW) (Purkey et al., 2014). Regions of the ocean fed by AABW from the Weddell Sea have exhibited a possible slowdown in local AABW warming rates (Lyman and Johnson, 2014), while the Pacific, fed by AABW from the shelves along the Ross and Adelie Coast, has continued to warm at an accelerating rate between 1990 and 2018 (Desbruyères et al., 2016b).

To date, assessment of deep ocean (below 2000 m) heat content has mostly been from ship-based data collected along decadal repeats of oceanographic transects (Figure 5.4b) (Talley et al., 2016a). While relatively sparse in space and time compared to the upper ocean, these transects were positioned to optimize sampling of most deep ocean basins and provide the highest quality of salinity, temperature and pressure data. Argo floats capable of sampling to 6000 m have just started to populate select deep ocean basins; this Deep Argo data has just started providing regional deep ocean warming estimates (Johnson et al., 2019). Decadal monitoring by the full global deep Argo array (Johnson et al., 2015), complemented by indirect estimates from space (Llovel et al., 2014; Von Schuckmann et al., 2014), will strongly reduce the currently large uncertainties of deep ocean heat content change estimates in the future.

The spatial and temporal sparseness of observations below 4000 m, along with significant differences between various ESMs, limits our understanding of the exact mechanisms driving the abyssal ocean variability. However, ESMs consistently predict an anthropogenic climate-change induced long-term abyssal warming trend originating in the Southern Ocean due to a reduction in the formation rates of cold Antarctic Bottom Water (Heuzé et al., 2015). Although the abyssal modes of natural variability are not as pronounced as closer to the surface, deep ocean heat content can vary on relatively short time scales through the communication of topographic and planetary waves driven by changes in the rate of deep water formation at high latitudes (Kawase, 1987; Masuda et al., 2010; Spence et al., 2017). Antarctic Bottom Water (AABW) has shown variability in properties and production rates over the past half century (Purkey and Johnson, 2013; Menezes et al., 2017). A slowdown in AABW formation rates may arise from freshening of shelf waters, changes in local winds driving cross shelf mixing, or larger scale dynamics controlling the spin up or down of Southern Ocean gyres influencing the density of outflowing waters over deep sills. Large-scale circulation changes can also alter the properties of the ambient water that is entrained as dense water descends along the Antarctic continental slopes (Spence et al., 2017). Evolving AABW properties may also reflect changes in deep Southern Ocean convection. The Weddell Polynya is a large opening in the wintertime ice of the Weddell Sea that is kept ice-free despite intense heat loss to the atmosphere by convective mixing bringing up warm and salty water from the deep ocean. (See Box 3.2 for a more extensive discussion of polynyas and the Weddell Polynya in particular.) The Weddell Polynya was present in three of the first years of infrared satellite observations of wintertime sea-ice concentrations in the mid-1970s, but it has been closed since 1976, only to reopen in 2016 and 2017. The prominent Weddell Polynya in the mid-

1970s greatly increased the volume of the coldest waters in the deep Weddell Sea. Weddell Polynyas are documented to drive abyssal cold and salty signals and can spread thermal signals as waves further and faster than could be explained by slow advective signals (Martin et al., 2015; Zanowski and Hallberg, 2017); these waves do not directly heat individual water parcels, but instead warm the ocean where they cause the coldest deep layers to spread laterally and thin. However, recovery from the large Weddell polynya of the early 1970s can only explain about 20% of the observed abyssal warming trend (Zanowski et al., 2015).
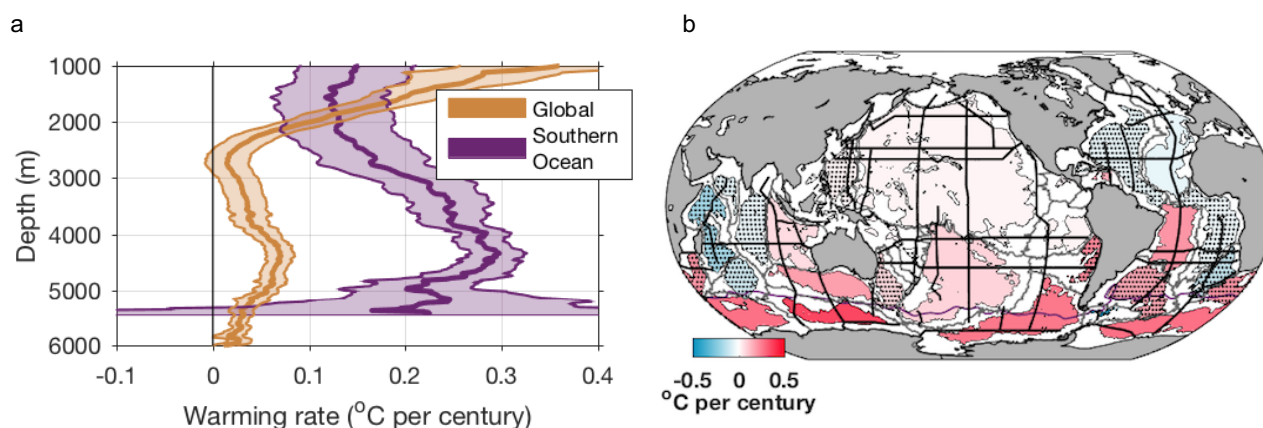
Figure 5.4: Observed rates of warming from 1981 to 2019 (a) as a function of depth globally (orange) and south of the Sub-Antarctic Front (the purple line in (b) at about 55°S) (purple) with 90% confidence intervals and (b) average warming rate (colors) in the abyss (below 4000 m) over various ocean basins (whose boundaries are shown in grey lines), with stippling indicating basins with no significant changes. The black lines show the repeat hydrographic sections used to make these estimates. These figures use updated GoShip data and the techniques of (Purkey and Johnson, 2010).
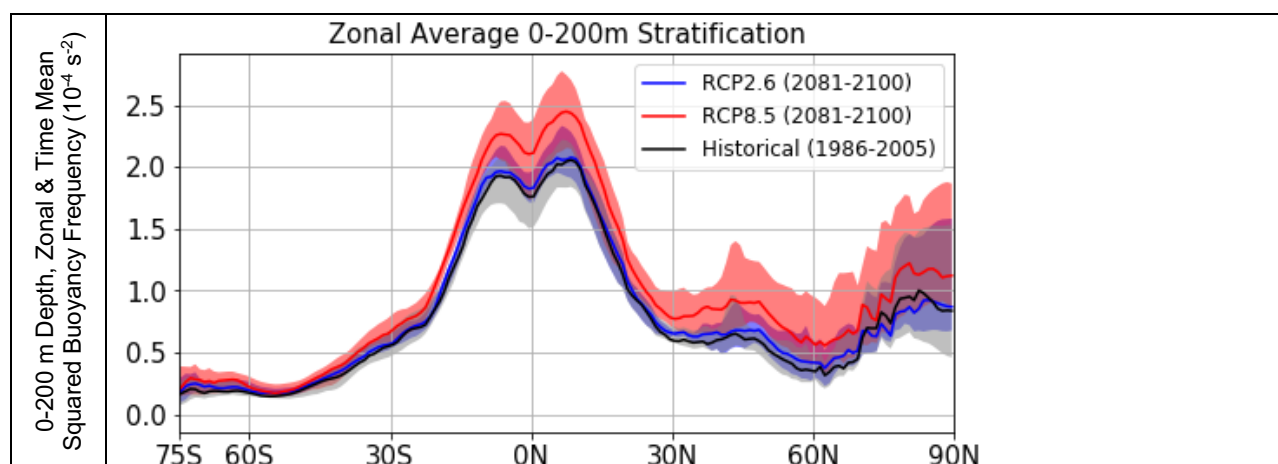
Figure 5.5: Zonal and 20-year mean stratification averaged over the top 200 m of the ocean for the CMIP5 ensemble of simulations at the end of the historical runs (black and grey), and for the end of the 21st century for RCP 2.6 (blue) and RCP8.5 (red) scenarios. The values between the 5th and 95th percentiles of the ensembles are shaded, while the lines are the ensemble mean. These model results are not adjusted by the control-run, so the spread in the various estimates primarily reflect model formulation differences. The average squared buoyancy frequency shown here is nearly linearly proportional to the density difference between the surface and 200 m, and is a measure of the density stratification of the upper ocean.

The ocean's properties are changing most rapidly in the near surface waters that are more immediately exposed to atmospheric forcing. As a result of the surface-intensified warming, the upper few hundred meters of the ocean are becoming more stably stratified (Helm et al., 2011; Talley et al., 2016b). The combination of surface intensified warming and near-surface freshening at high latitudes leading to a projection of more intense near-surface stratification (the downward-increasing vertical gradient of density) across all ocean basins (Figures 5.3 and 5.5) is a robust result with a *high agreement* across successive generations of coupled climate models (Capotondi et al., 2012; Bopp et al., 2013). Based on the projected

changes from individual models between (1986-2005) and (2081-2100), the mean stratification of the upper 200 m averaged between 60°S and 60°N, normalized by the ensemble mean value from (1986-2005) will *very likely* increase by between 1.0% and 9.3% (with 95% confidence, and a CMIP5 median change of 2.6%) for RCP2.6, and by between 12.2% and 30.0% (median value 21.2%) for RCP8.5. Inferences from oceanic observations (Good et al., 2013) suggest that the 20-year mean stratification averaged between 60°S and 60°N and over the top 200 m *very likely* increased by between 2.18% and 2.42% from (1971-1990) to (1998-2017). By contrast, the bottom intensified warming in the abyss (see Figure 5.4) which is consistent with a slowing in the rate of AABW formation, is also associated with a reduction in the abyssal stratification of the ocean (Lyman and Johnson, 2014; Desbruyères et al., 2016b). Both of these changes have consequences for the evolving turbulence and ocean water-mass structure. Based on observational evidence, theoretical understanding and robust ESM projections, it is *very likely* that stratification in the upper few hundred meters of the ocean below the mixed layer will increase significantly in the 21st century over most ocean basins as a result of climate change, and abyssal stratification will *likely* decrease.

Many dynamical consequences of increased stratification are understood with *very high confidence* (see, for instance, Gill (1982) and Vallis (2017)). For the same turbulent kinetic energy dissipation, locally increased stratification reduces the turbulent vertical diffusivity of heat, salinity, oxygen and nutrients (see Section 5.2.2.2.4). Increased stratification in the tropics and subtropical gyres will *likely* lead to a net reduction in the vertical diffusivities of nutrients and other gases within the main thermocline, reducing the flux of nutrients into the euphotic zone and increasing the gradient in oxygen concentrations between the near surface ocean and the interior. Increasing upper ocean stratification (Figure 5.5) acts to restrict the depth of the ocean's surface mixed layer. Increasing stratification increases the buoyancy frequency and the lateral propagation speed of internal gravity waves and boundary waves by about half the percentage change of the stratification itself. Increasing stratification increases both the length of the internal deformation radius (a typical length scale in baroclinic eddy dynamics) and the horizontal scales of internal tides (see Section 5.2.2.2.3) proportionately with the changes in the internal gravity wave speeds. An increase in stratification will increase the lateral propagation of internal Rossby waves (which set up the basin-scale ocean density structure) proportionately. For the same forcing, increasing stratification reduces the geostrophically balanced slope of density surfaces, and hence the vertical extent of basin-scale wind-driven gyres or coastal upwelling circulations. The flattening of density surfaces by increased stratification inhibits advective exchange between the surface and interior ocean (Wang et al., 2015a), with consequences for the uptake of anthropogenic carbon (Section 5.2.2.3), the evolving oxygen distribution (Section 5.2.2.4) and the supply of nutrients to support primary production (Section 5.2.2.5).

5.2.2.2.3   *Tides and coastal physical changes in a changing climate*

Coastal systems are subject to the same large-scale warming trends as the open ocean, but the local response may be dominated by a complex of localized changes in factors such as circulation, mixing, river plumes or the seasonal upwelling of cold water. Using ESMs to project how these factors will interact often requires much finer resolution than is currently affordable in global models, however regional high-resolution models can be effective, especially in marginal seas like the Mediterranean with restricted interactions with the open ocean and that respond primarily to local forcing (Adloff et al., 2015). High resolution regional models have also been used to project robust localized ocean climate changes in wide shelf-seas with more extensive interactions with the open ocean, like those in northwestern Europe (Tinker et al., 2016). The technical difficulties of using nested regional models are much greater in coasts adjacent to energetic large-scale currents like the Gulf Stream, Kuroshio, and Agulhas, and projecting detailed coastal climate change such places may require the use of expensive high resolution global models (Saba et al., 2016). These physical coastal changes have consequences that cascade through ecosystems to people, as is illustrated in detail for eastern boundary upwelling systems in Box 5.2.

Both human structures and ecological systems in the coastal zone are directly impacted by tidal amplitudes, which contribute to high-water levels and the tidal flushing rates of estuaries, embayments, marshes and mangroves. The tides are the response of a forced-damped-resonance system (Arbic et al., 2009). The M_2 tide is the dominant tidal constituent in most places, with a period of ½ lunar day, or 12 hours, 25 minutes; the M_2 tides are created by the differential motion of the solid Earth and oceans in response to the gravitational attraction of the moon (Newton, 1687; Laplace, 1799). The astronomical forcing evolves only slowly, however the tidal damping and basin resonance at tidal frequencies can change in response to changes in sea-level, stratification, and coastal conditions (Müller, 2012; Schindelegger et al., 2018). Several

recent studies have analyzed historical coastal tide gauge data and found amplitude trends of order 1-4% per century (Ray, 2009; Woodworth, 2010; Müller et al., 2011). In some locations, the changes in the tides have been of comparable importance to changes in mean sea level for explaining changes in high water levels (Jay, 2009). For many individual tide gauges, the trends in tidal amplitude are strongly positively or negatively correlated with local time-mean sea level trends (Devlin et al., 2017). Another source of secular tidal changes, changes in oceanic stratification, modifies the rate of energy conversion from the barotropic tides to the internal tides (Jayne and St. Laurent, 2001), the vertical profile of turbulent viscosity on shelves (Müller, 2012), and the propagation speed of the internal tides (Zhao, 2016). For example, Colosi and Munk (2006) found an increase in the amplitude of the principal lunar semidiurnal tide M_2 in Honolulu of about 1 cm over the past 100 years, which they attributed primarily to changes in oceanic stratification bringing about local changes in relative phases of the internal and external M_2 tides, increasing constructive interference. Both sea level and stratification are expected to exhibit robust secular positive trends in the coming century due to climate change, at rates that are significantly larger than historical trends, and people may choose to replace natural beaches and marshes with sea-walls in response to rising sea-levels. As a result, it is *very likely* that the majority of coastal regions will experience statistically significant changes in tidal amplitudes over the course of the 21st century.

Because coastal tides are near resonance in many locations, small changes in sea level and bay shape can change the local tides significantly. For example, the insertion of tidal power plants can have a significant impact on the local tides (Ward et al., 2012). Various observational and modeling studies demonstrate that sea level rise has spatial heterogeneous impacts on the tides, with some locations experiencing decreased tidal amplitudes and others experiencing increased tidal amplitudes (Pickering et al., 2012; Devlin et al., 2017; Pickering et al., 2017). Projections of tidal changes indicate that the patterns and even the sign of changes in tidal amplitudes depend on whether the coastlines are allowed to recede with rising sea levels or are held in place (Pickering et al., 2017; Schindelegger et al., 2018). Pelling et al. (2013) and Hwang et al. (2014) demonstrate that the rapid coastline changes in China's Bohai Sea have already altered the tides in that region and throughout the Yellow Sea (Hwang et al., 2014). Pelling and Green (2014) examine the impact of flood defenses as well as sea level rise on tides on the European Shelf. Such tidal changes have implications for designing flood defenses, for tidal renewable energy, for tidal flushing timescales of estuaries and embayments, and for navigational dredging requirements (Pickering et al., 2012) (Section 5.4.2). The sign and amplitude of local changes to tides are *very likely* to be impacted by both human coastal adaptation measures and climate drivers (listed above).

5.2.2.2.4   Systematic sources of uncertainty in projections of ocean physical changes
ESMs are able to capture the dynamics of the climate system, but all numerical models have approximations and biases. The most commonly used type of ocean component in ESMs is known to exhibit numerically induced vertical mixing that can be a significant fraction of the physical mixing (Ilıcak et al., 2012; Megann, 2018). Because so many ocean models exhibit the same sign of bias, there is a systematic warming of the lower-main thermocline that is not cancelled out when taking the average over the ensemble of all the models in CMIP5. These biases are widely known within the ocean modelling community, and various groups are working to reduce these biases in future ESMs with better ocean model numerics and parameterizations. To correct for model biases, ESM projections are always taken as the difference from a control run without the anomalous forcing. However, some aspects of the ocean response to climate change are nonlinear, and model biases can introduce uncertainties into climate projections. In the case of heat uptake, this is of the order of 10% uncertainty, while for the rate of steric sea level rise (which depends on the nonlinear equation of state of seawater) the uncertainty in CMIP5 models is of the order of 20% (Hallberg et al., 2012).

Mesoscale eddies (geostrophic rotating vortices with spatial scales of 10-100 km that penetrate deeply into the water column, and are often described as the ocean's weather) play an important role in regulating the changes to the larger scale ocean circulation, especially in the Antarctic Circumpolar current, as is discussed in Cross Chapter Box 7. In addition, sub-mesoscale eddies (rotationally influenced motions with smaller horizontal scales of hundreds of metres to about 10 km and intrinsic timescales of a few days that especially arise in association with fronts in the ocean's surface properties) are known to be particularly important in the dynamics of the near-surface ocean boundary layer (see the review by Mahadevan (2016)). Sub-mesoscale instabilities are associated with re-stratifying overturning circulations that can limit the thickness of the well-mixed ocean surface boundary layer near fronts (Bachman et al., 2017). Moreover, sub-

mesoscale motions generate strong vertical velocities that drive fluxes of nutrients from the interior ocean into the euphotic zone or create pockets of reduced mixing with increased phytoplankton residency time within the euphotic zone (Lévy et al., 2012). Intense mesoscale eddies are known to create favourable conditions for sub-mesoscale instabilities as shown in both observational (Bachman et al., 2017) and numerical studies (Brannigan et al., 2017). Intensifying Southern Ocean eddy fields will have a significant local impact on biological productivity, ecosystem structure, and carbon uptake, both directly and via sub-mesoscale processes. At typical CMIP5 ESM resolutions, it is only in the tropics that mesoscale eddies are adequately resolved to explicitly model their effects (Hallberg, 2013), while submesoscale eddies are not resolved anywhere, so eddy effects need to be parameterized in ESMs. Despite great progress over the past 30 years in parameterizing eddy effects, uncertainties in these parameterizations and how eddies will respond to novel conditions continue to contribute to uncertainties in projections of oceanic climate change (*medium confidence*).

Ocean turbulent mixing is a key process regulating the ocean circulation and climate. Turbulent mixing is important for the uptake and redistribution of heat, carbon, nutrients, oxygen and other tracers (properties that are carried along with the flow of water) in the ocean (Schmittner et al., 2009; MacKinnon et al., 2017). Both observations and theory indicate that turbulent mixing in the ocean is not constant in space or time. Global estimates of both the turbulent kinetic energy dissipation rate and the vertical diffusivity, two measures of ocean turbulence, vary over several orders of magnitude throughout the ocean (Figure 5.6) (Polzin et al., 1997; Waterman et al., 2012; Whalen et al., 2012; Alford et al., 2013; Hummels et al., 2013; Sheen et al., 2013; Waterhouse et al., 2014; Kunze, 2017). For a given energy dissipation rate, the turbulent diffusivities of heat, salinity, nutrients and other tracers tend to be smaller with stronger stratification. This dependency on stratification helps explain why the observationally inferred diffusivity in the heavily stratified main thermocline (250 m-1000 m depth) is of similar magnitude to those deeper in the water column, while the turbulent energy density and dissipation rate are much stronger at the shallower depths (Whalen et al., 2012). Oceanic turbulence also fluctuates in time, is modulated by tidal cycles (Klymak et al., 2008), the mesoscale eddy field and seasonal changes (Whalen et al., 2018). In the mixed layer and directly below, turbulence changes according to local conditions, such as the winds, heating rates and local stratification (Sloyan et al., 2010; Moum et al., 2013; D'Asaro, 2014; Tanaka et al., 2015) at diurnal to seasonal and longer timescales. These variations in near-surface turbulence need to be taken into account for ESMs to reproduce more accurately the observed seasonal cycle of surface properties and spatial structure of the depth of the thermally well-mixed near surface layer of the ocean. The spatial and temporal patterns of ocean turbulence help shape ocean tracer distributions (heat, dissolved greenhouse gases, nutrients) and how they will evolve in a changing climate (*high confidence*).

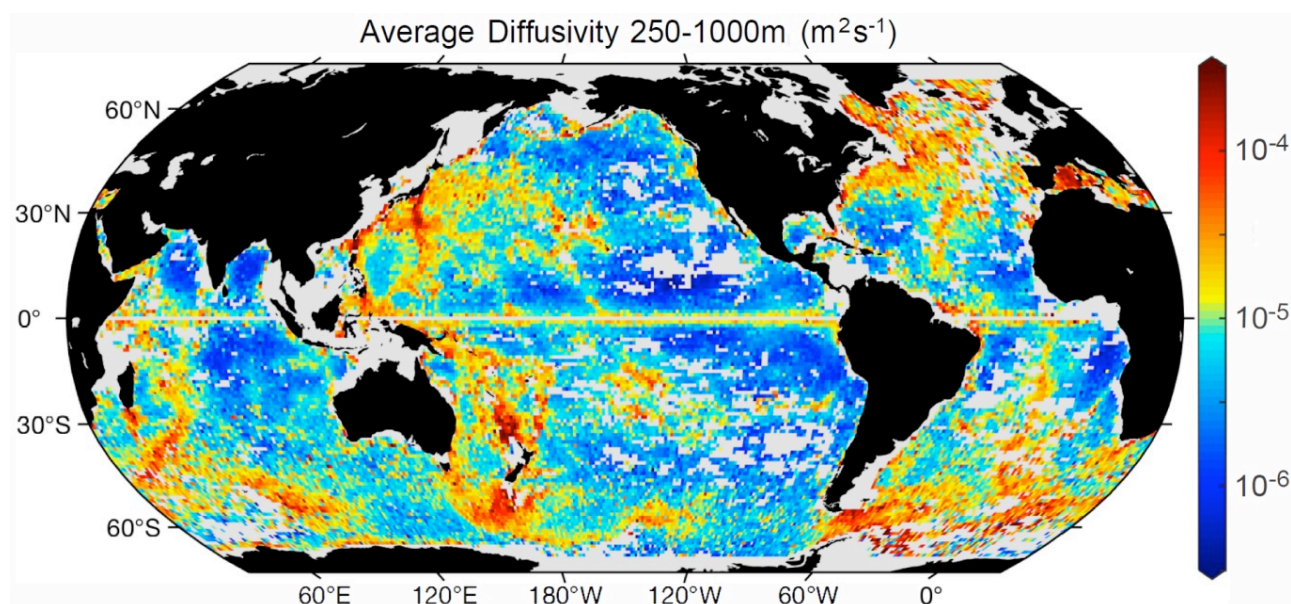**Figure 5.6:** Estimate of the average vertical turbulent diffusivity between 250 and 1000 m calculated by applying fine structure techniques to Argo float data from below the well-mixed near-surface boundary layer. Only bins with at least three estimates are plotted and regions with insufficient data are coloured grey. This figure was created using updated data through April, 2018 with the techniques from Whalen et al. (2012).

Ocean turbulent mixing requires energy sources, many of which are expected to change with a changing climate. Surface wind and buoyancy forcing, the mean and eddying larger-scale ocean circulation itself, and the barotropic tides are all thought to be significant sources of the energy that drives mixing (Wunsch and Ferrari, 2004). Often this energy first passes through the ocean's pervasive field of internal gravity waves that propagate and refract through the varying ocean circulation, often breaking into turbulent mixing far from their sources (Eden and Olbers, 2014; Alford et al., 2016; Melet et al., 2016; Meyer et al., 2016; Zhao et al., 2016b). The energy contributing to the internal waves from the winds and the subsequent turbulence will be altered by changes in tropical storm activity or sea-ice coverage. For example, the increasing extent of ice-free Arctic Ocean has already been observed to lead to increased wind-driven internal waves (Dosser and Rainville, 2016). The Southern Annular Mode is expected to intensify as a result of climate change (Young et al., 2011; Jones et al., 2016b), bringing with it stronger winds, and more wind-energy input over most of the Southern Ocean and a more intense mesoscale eddy field (Hogg et al., 2015). Changes in the near-bottom stratification will alter the rate that the barotropic tides generate internal waves, thereby altering the strength and distribution of the tidally generated mixing. Some of the parameterizations of interior ocean mixing used in CMIP5 ESMs take some changing turbulent energy sources into account (Jayne and St. Laurent, 2001), and more comprehensive mixing treatments are being developed for use in future generations of ESMs (Eden and Olbers, 2014). However, not all of the physical processes leading to the rich structure of mixing shown in Figure 5.6 are well understood or included in ESMs; the prospect of significant changes in the patterns and intensity of ocean turbulent mixing is a potential source of uncertainty (probably at the 10% level) in projections of physical and ecological changes in the ocean, including heat uptake, stratification changes, steric sea-level rise, deoxygenization, and nutrient fluxes (*medium confidence*).

5.2.2.3   Changes in Ocean Carbon

Since AR5, new global-scale data synthesis products, novel methods for their analyses, as well as progress in modeling have substantially increased our quantitative understanding of the role of the ocean in absorbing and storing CO_2 from the atmosphere. The most important progress concerns the data-based quantification of the temporal variability of the ocean carbon sink. While AR5 assessed primarily the climatological mean processes governing the ocean carbon cycle, the most recent work now permits us to assess how these processes have changed in recent decades in response to climate variability and change. Here we focus specifically on the open ocean carbon cycle.

5.2.2.3.1 Ocean carbon fluxes and inventories
The analyses of the steadily growing number of surface ocean CO_2 observations (now more than 20 million observations, SOCATv6 (https://www.socat.info/index.php/2018/06/19/v6-release/) demonstrate that the net ocean uptake of CO_2 from the atmosphere has increased from around 1.2 ± 0.5 Pg C yr^{-1} in the early 1980s to 2.0 ± 0.5 Pg C yr^{-1} in the years 2010-2015 (Rödenbeck et al., 2014; Landschützer et al., 2016). Once new estimates of the outgassing flux stemming from river derived carbon of 0.8 Pg C yr^{-1} (Resplandy et al., 2018) are accounted for, these new observations imply that the rate of global ocean uptake of anthropogenic CO_2 increased from 2.0 ± 0.5 Pg C yr^{-1} to 2.8 ± 0.5 Pg C yr^{-1} between the early 1980s and 2010-2015 (Rödenbeck et al., 2014; Landschützer et al., 2016; Le Quéré et al., 2018). This increase is supported by the current generation of ocean carbon cycle models (Le Quéré et al., 2018), and commensurate with the increase in atmospheric CO_2.

The continuing efforts to re-measure dissolved inorganic carbon (DIC) along many of the repeat hydrographic lines that were occupied during the 1980s and 1990 (Talley et al., 2016a), alongside the preparation of a global quality-controlled database of ocean interior observations (Olsen et al., 2016a), have led to progress since AR5 regarding to the oceanic interior storage of anthropogenic CO_2. Several studies analyzed the changes in the amount of anthropogenic CO_2 that have accumulated between different occupations in the different ocean basins (Wanninkhof et al., 2010; Pérez et al., 2013; Woosley et al., 2016; Carter et al., 2017), confirming that the anthropogenic CO_2 taken up from the atmosphere is transported to depth, where most of it is stored. Using a newly developed reconstruction method, Gruber et al. (2019) extended these results to the globe. They find that between 1994 and 2007, across two standard deviations, that the global ocean has accumulated an additional 30-38 Pg C of anthropogenic CO_2, which is equivalent to an air-sea CO_2 flux of between 2.3-2.9 Pg C yr^{-1} (coherent with surface ocean CO_2 observations), bringing

the total inventory for the year 2007 to 150±20 Pg C. Extrapolating this estimate to the year 2010 gives an inventory of 158±18 Pg C, which is statistically indistinguishable from the 'best' estimate provided by Khatiwala et al. (2013) of 155±31 Pg C and more recently also found from a steady-state ocean model (DeVries, 2014) for this reference year. If the inventory-based estimates are adjusted for the loss of natural carbon, a *very likely* total increase in storage between 1994 and 2007 of 24-34 Pg C, or around 25% of total emissions, is found (Gruber, 2019).

Thus, there is *very high confidence* from surface ocean and ocean interior carbon data that the strength of the ocean sink for anthropogenic carbon has increased in the last two decades in response to the growth of atmospheric CO_2. Multiple lines of evidence indicate that it is *very likely* that the ocean has taken up 20-30% of the global emissions of CO_2 from the burning of fossil fuels, cement production, and land-use change since the mid 1980s. The consistency between independent surface ocean observations and the ocean interior data-based reconstructions supports the assessment of *very high confidence* and provides *robust evidence* that fraction of emissions taken up by the ocean has not changed in a statistically significant manner in the last few decades and remains consistent with AR5

Alongside a globally integrated perspective, these new surface ocean observations also reveal a substantial degree of variability at inter annual and decadal scales (Rödenbeck et al., 2015; Landschützer et al., 2016; Le Quéré et al., 2018). Most notable are the air-sea CO_2 flux variations in the tropics linked to ENSO variations (Rödenbeck et al., 2015; Landschützer et al., 2016), as well as the strong decadal variations in the high latitudes, especially the Southern Ocean (Landschützer et al., 2015; Munro et al., 2015; Ritter et al., 2017), discussed further in Chapter 3 (Section 3.2.1.2.4). Fluctuations in the Southern Ocean CO_2 flux are important as they impart a substantial imprint also on the global uptake fluxes. For instance, reduced Southern Ocean uptake in the 1990 to 2000 period coincided with an exceptionally weak global net uptake of only about 0.8±0.5 Pg C yr $^{-1}$.

Thus, there is growing evidence from multiple datasets that the ocean carbon sink exhibits decadal variability at regional scales that significantly alter the globally integrated sink (*medium confidence*).

Detailed analyses of the spatial structure of the change in storage of anthropogenic CO_2 confirm the variable nature of the ocean carbon sink suggested by the surface observations (Pérez et al., 2013), which are most likely a consequence of changes in ocean circulation (DeVries and Weber, 2017). The increase in anthropogenic CO_2 between 1994 and 2007 occurs throughout the upper 1000 m, but with very different penetration depths, reflecting largely differences in the efficiency, with which the anthropogenic CO_2 is transported from the surface to depth (Gruber et al., 2019) (Figure 5.7). This spatial distribution of how the amount of anthropogenic CO_2 has changed between 1994 and 2007 is similar to the distribution of anthropogenic CO_2 reconstructed for 1994 (Sabine et al., 2004), although the imprint of regional variations in ocean circulation and transport are discernible (Gruber, 2019).
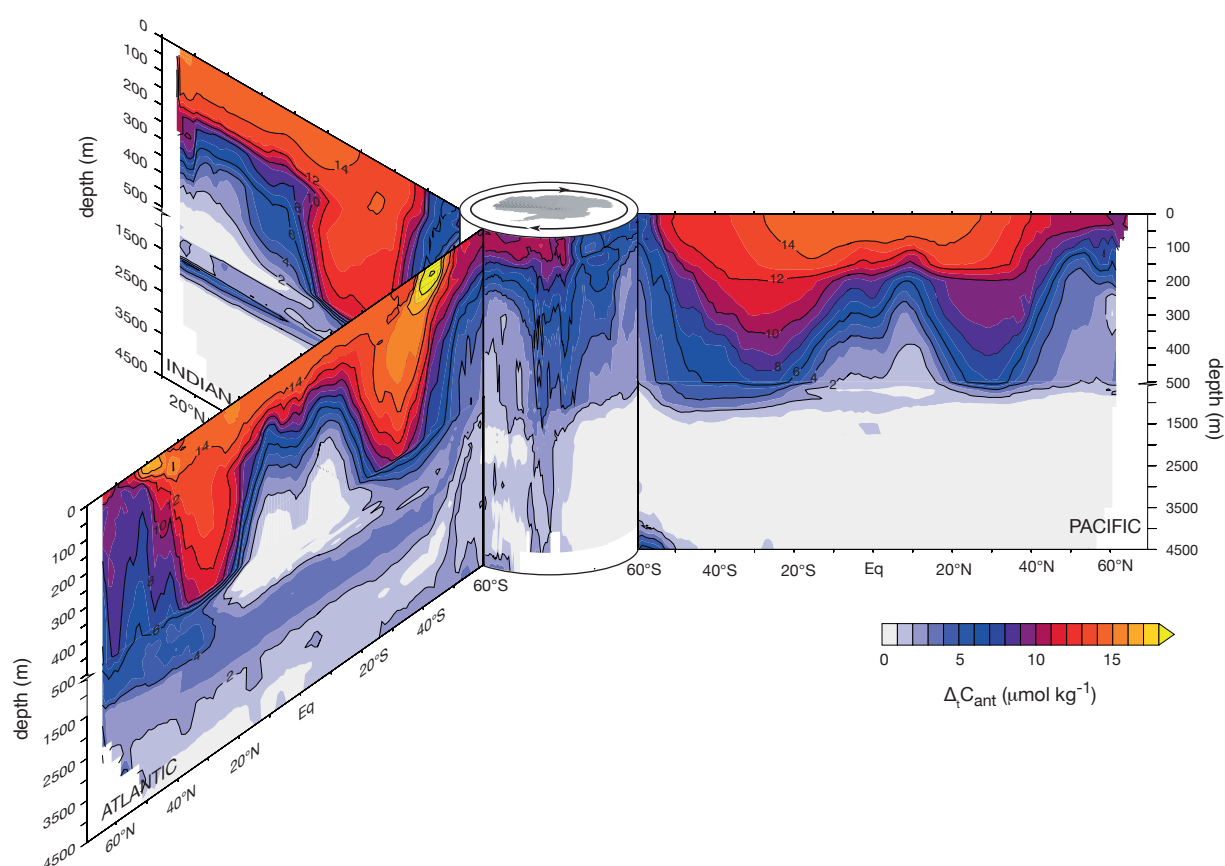
Figure 5.7: Vertical sections of the change in anthropogenic CO_2 from 1994 to 2007 represented by the zonal mean sections in each ocean basin, organized around the Southern Ocean in the center. The upper 500 m are expanded. Contour intervals of anthropogenic CO_2 are 2 μmol kg^{-1} (Gruber, 2019).

5.2.2.3.2 Ocean carbon chemistry

Analyses of direct measurements of ocean chemistry from time-series stations and merged shipboard studies show consistent decreases in surface-ocean pH over the past few decades. Reductions range between 0.013 to 0.03 pH units decade^{-1} over records that span up to 25 years (Table SM5.3). Focusing on the individual time series locations with records longer than 15 years, there is an overall decline of 0.017 to 0.027 (across 99% confidence intervals). Trends calculated from repeat measurements on ocean surveys show a consistent value of around -0.02 pH units decade^{-1} for diverse oceanic regions (Table SM5.3), with greater subsurface than surface trends reported in the subtropical oceans (Dore et al., 2009). At larger spatial scales, surface-ocean pH trends are assessed using shipboard observations of the fugacity of CO_2 and estimates of ocean alkalinity (Takahashi et al., 2014; Lauvset et al., 2015). Between 1991–2011, mean surface-ocean pH has declined by 0.018 ±0.004 units decade^{-1} in 70% of ocean biomes, with the largest declines in the Indian Ocean (–0.027 units decade^{-1}), eastern Equatorial Pacific (–0.026 units decade^{-1}) and the South Pacific subtropical (–0.022 units decade^{-1}) biomes (Lauvset et al., 2015). Due to the close link between carbonate ion concentrations and pH, mean trends in the stability of mineral forms of aragonite and calcite (known as the 'saturation state') that are important for organisms such as coccolithophorids, pteropods and corals follow those of pH, with high latitude regions most vulnerable to under-saturation due to naturally lower mean values.

It is *virtually certain* that ocean pH is declining, and the *virtually likely* range of this decline is 0.017 to 0.027 pH units per decade for the 8 locations where individual time-series observations longer than 15 years exist. This trend is lowering the chemical stability of mineral forms of calcium carbonate and can be attributed to rising atmospheric CO_2 levels.

CMIP5 models are in good agreement with historical observations of declining surface-ocean pH (Figure 5.8a). Models project global surface-ocean declines between 2006-2015 and 2081-2100 of 0.287-0.291 and 0.036-0.042 pH units (both across 99% confidence intervals) for the RCP2.6 and RCP8.5 scenarios,

respectively, with higher reductions in the subsurface of subtropical oceans (Bopp et al., 2013; Gattuso et al., 2015). These changes in pH will be greatest in the Arctic Ocean and the high latitudes of the Atlantic and Pacific Oceans due to their lower buffer capacity and are lowest in contemporary upwelling systems (Figure 5.8b) and will also reduce the stability of calcite minerals (Bopp et al., 2013; Gattuso et al., 2015). The area of the surface ocean (0-10m) characterized by undersaturated conditions in CMIP5 models by 2081-2100 reduces from a *very likely* range of $6.4\text{-}9.5\text{x}10^{12}$ m^2 or $5.5\text{-}7.3\text{x}10^{13}$ m^2 under RCP8.5 (as much as 16-20% of ocean surface area for aragonite), to just $0.01\text{-}0.2$ $\text{x}10^{12}$ m^2 or $0.01\text{-}0.13\text{x}10^{13}$ m^2 under RCP2.6 for either calcite or aragonite minerals, respectively. Under RCP8.5, hotspots for undersaturated waters for calcite remain restricted to the Arctic Ocean, while for aragonite, much of the Southern Ocean and the North Pacific and Northwestern Atlantic Oceans are also projected to become undersaturated (Orr et al., 2005; Hauri et al., 2015; Sasse et al., 2015). These results arise from the very well understood reductions in carbonate ion concentrations at lower pH, the vulnerability of regions with naturally low mean values, and the greater overall sensitivity of aragonite solubility. Regional models, with higher resolution that ESMs, also project year-round corrosive conditions for aragonite in some eastern boundary upwelling systems (Franco et al., 2018a). In the ocean interior, the decline in pH and calcium carbonate saturation state is more uncertain across models (Steiner et al., 2014) as it is modulated by changes to ocean overturning and water mass subduction (Resplandy et al., 2013; Chen et al., 2017). Projected benthic changes in pH over the next century are highly localized and are linked to transport of surface anomalies to depth, with over 20 % of the North Atlantic sea floor deeper than 500 m projected to experience pH reductions greater than 0.2 units by 2100 under the RCP8.5 scenario (Gehlen et al., 2014a). Changes in pH in the abyssal ocean (>3,000 m deep) are greatest in the Atlantic and Arctic Oceans, with lesser impact in the Southern and Pacific Oceans by 2100, mainly due to the circulation timescales (Sweetman et al., 2017).

Overall, it is *virtually certain* that the future surface open ocean will experience pH drops of either 0.036-0.042 (RCP2.6) or 0.287-0.291 (RCP8.5) pH units by 2081-2100, relative to 2006-2105. These pH changes are *very likely* to cause 16-20% of the surface ocean, specifically the Arctic and Southern Oceans, as well as the northern Pacific and north-western Atlantic Oceans, to experience year-round corrosive conditions for aragonite by 2081-2100. It is *virtually certain* these impacts will be avoided under the RCP2.6 scenario. There is *medium confidence*, due to the potential for parallel changes in ocean circulation, that the Arctic and North Atlantic seafloors will experience the largest pH changes over the next century.

Although ocean acidification results in long-term trends in mean ocean chemistry, it can also influence seasonal cycles. Observation-based products indicate that the seasonal cycle of global surface-ocean $p\text{CO}_2$ increased in amplitude by 2.2 ± 0.4 µatm between 1982 and 2014 (Landschützer et al., 2018). CMIP5 models and data-based products similarly project consistent future increases in the seasonal cycle of surface-ocean $p\text{CO}_2$ under the RCP8.5 emissions scenario, with enhanced amplification in high-latitude waters (McNeil and Sasse, 2016). The amplitude of the seasonal cycle of global surface-ocean free acidity ($[\text{H}^+]$) is projected to increase by 71-91% (across 90% confidence intervals) over the twenty-first century under RCP8.5, also with greater amplification in the high-latitudes (Kwiatkowski et al., 2018). Conversely, models project a 12-20% reduction (across 90% confidence intervals) in the seasonal amplitude of surface-ocean pH, as changes in pH represent relative changes in $[\text{H}^+]$ due to their logarithmic relationship, and there are typically greater projected increases in annual mean state $[\text{H}^+]$ than the seasonal amplitude of $[\text{H}^+]$. Models also project a 4-14% (across 90% confidence intervals) reduction in the seasonal amplitude of global mean surface-ocean aragonite saturation state under RCP8.5, with a slight amplification in the subtropics being outweighed by dampening elsewhere. The contrasting changes in the seasonal amplitudes of ocean carbonate chemistry variables derive from different sensitivities to atmospheric CO_2 and climate change and to diverging trends in the seasonal cycles of dissolved inorganic carbon, alkalinity and temperature. Model skill at simulating the seasonal cycles of carbonate chemistry is moderate, with persistent biases in the Southern Ocean, particularly for $p\text{CO}_2$, $[\text{H}^+]$ and pH (Kwiatkowski et al., 2018; Mongwe et al., 2018).

Overall, we assess that alongside the strong mean state changes, it is *very likely* that the amplitude of the seasonal cycle in free acidity will increase by 71-91%, while it is *very likely* that the seasonal cycles of pH and aragonite saturation will decrease by 12-20% and 4-14%, respectively.
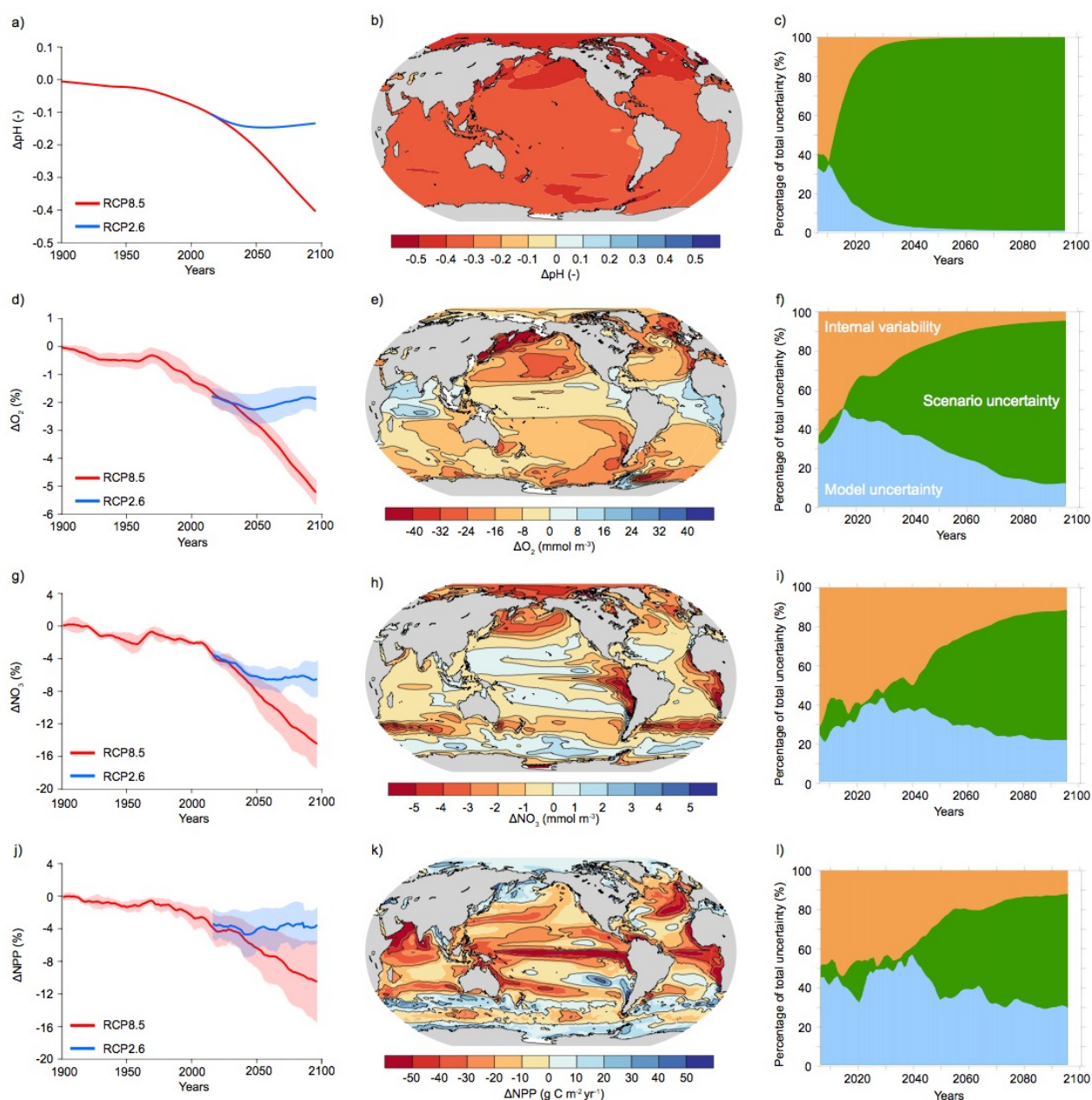
Figure 5.8: Panels a, d, g and j display simulated global changes over the period of 1900 to 2100 (with solid lines representing the multi-model mean and the envelope representing 90% confidence intervals for RCP8.5 and RCP2.6), for surface pH, O_2 concentration averaged over 100 to 600 m depth, upper 100 m nitrate concentrations and NPP integrated over the top 100 m. Differences are calculated relative to the 1850-1900 period. Panels b, e, h and k show spatial patterns of simulated change in surface pH, upper 100 m nitrate concentrations, O_2 concentration averaged over 100 to 600 m depth, and NPP integrated over the top 100 m averaged over 2081-2100, relative to 1850-1900 for RCP8.5. Panels c, f, i and l display time series of the percentage of total uncertainty ascribed to internal variability uncertainty, model uncertainty, and scenario uncertainty in projections of global annual mean changes Figure adapted after (Frölicher et al., 2016). Please note that confidence intervals can be affected by the different number of models available for the RCP8.5 and RCP2.6 scenarios and for different variables. See also Table SM5.4.

5.2.2.4   *Changing Ocean Oxygen*

Ocean oxygen (O_2) levels at the surface are controlled by the balance between oxygen production during photosynthesis, temperature-controlled solubility and air-sea exchange. Deeper in the water column, consumption of oxygen during respiration and redistribution by ocean circulation and mixing are dominant processes. In theory, a warmer more stratified ocean would have a reduced oxygen content, due to the combined influence of lowered gas solubility and a greater interior respiration of organic matter due to enhanced physical isolation of subsurface waters. In accord, global changes in ocean oxygen assessed from

three different analyses of compiled global oxygen datasets going back to the 1960s agree that there is a net loss of oxygen from the ocean over all depths (see Table 5.2). For the 0-1000m depth stratum that contains the most data and is common to all three analyses, oxygen is assessed to have declined by a *very likely* range of 0.5 to 3.3% between 1970 and 2010. For the surface ocean (0-100m) and the thermocline later of 100-600m the *very likely* range of oxygen declines are 0.2-2.1% and 0.7-3.5%, respectively (Table 5.2). Across two studies, global oxygen is assessed to have declined by a *very likely* range of 0.3 to 2.0%, with a similar range of decline for waters deeper than 600m (Table 5.2). The regions of lowest oxygen, known as oxygen minimum zones (OMZs, with oxygen levels lower than 80 μmol L^{-1}), are observed to be expanding by a *very likely* range of 3.0-8.3% across the three studies.

Regionally, all studies agree that the north Pacific and Southern Oceans have shown the largest overall oxygen declines (Figure 5.9), but there is some disagreement regarding the magnitude of the oxygen change in the tropical ocean, with some studies suggesting significant declines (Schmidtko et al., 2017) and other reporting more modest reductions (Helm et al., 2011; Ito et al., 2017) and data coverage is still limited for some regions and deeper than 1000m. Based on the available data, the strongest declines in deep ocean oxygen have occurred in the Equatorial Pacific, North Pacific, Southern Ocean and South Atlantic, with intermediate declines in the Arctic, South Pacific and Equatorial Atlantic, while the North Atlantic has experienced a moderate oxygen increase below 1,200 m (Figure 5.9). A particular difference between parallel oxygen analyses concerns the means of integrating and mapping sparse data across the ocean, both horizontally and vertically, with different studies making specific decisions about averaging grids and integration methods. Moreover, data remains sparse for some ocean regions, depths and periods. Taken together, the challenges of data sparsity, regional differences and the relatively large uncertainties on the oxygen changes across different studies, but also recognizing that oxygen declines are significantly different to zero, leads to *medium confidence* in the observed oxygen decline.

Syntheses of datasets from local time series tend to document stronger trends, with oxygen declines of over 20% at sites in the Northeastern Pacific between 1956-2006 (Whitney et al., 2007), the Northwestern Pacific between 1954-2014 (Sasano et al., 2015) and the California Current between 1984-2011 (Bograd et al., 2015). Despite holding the highest inventory of oxygen in the ocean, oxygen levels in Southern Ocean contributed 25% to the global decline between 1970–1992 (Helm et al., 2011) and have fallen by over 150 Tmol per decade from the 1960s to present (Schmidtko et al., 2017). Observations along ocean cruises as part of the CLIVAR programme have also documented broad thermocline oxygen declines in the northern hemisphere oceans, accompanied by well understood oxygen increases in subtropical and southern hemispheres (Talley et al., 2016b).

Overall there is *medium confidence* that the oxygen content of the upper 1000m has declined with a *very likely* loss of 0.5-3.3% between 1970-2010. Oxygen minimum zones are expanding in volume, by a *very likely* range of 3.0-8.3%. There is *medium confidence* that the largest regional changes have occurred in the Southern Ocean, equatorial regions, North Pacific and South Atlantic due to *medium agreement* among studies.

The role of ocean warming alone in driving the oxygen changes can be appraised using solubility estimates, which vary between around 15-50% for the upper 1000m oxygen trend between studies (Helm et al., 2011; Ito et al., 2017; Schmidtko et al., 2017). The role of other processes, linked to changing ocean ventilation and respiration are challenging to appraise directly, but tend to reinforce the impacts from warming and are probably predominant overall (Oschlies et al., 2018). Indeed, that the observed oxygen decline is negatively correlated with ocean heat content changes (Ito et al., 2017) reflects the overriding role of changing ocean ventilation and associated processes (see also Section 5.2.2). That the ratio of the associated oxygen to heat changes is larger than would be expected from thermal processes alone also highlights the role played by other processes (Oschlies et al., 2018). Local oxygen trends have emphasized the role of changes to ocean physics in western Northern Pacific (Whitney et al., 2013); Sasano et al. (2015), the southern California Current region (Goericke et al., 2015), and the Santa Barbara Basin (Goericke et al., 2015). In regions of high mesoscale activity, such as the tropical north Atlantic, low oxygen eddies can have a significant impact on oxygen dynamics (Karstensen et al., 2015; Grundle et al., 2017). Oxygen fluctuations in the deep ocean have been linked to changes in large scale ocean circulation (Watanabe et al., 2003; Stendardo and Gruber, 2012) and at the global scale, the observed oxygen decline is negatively correlated with ocean heat content changes (Ito et al., 2017). Changes to respiration rates, either due to temperature enhancement or in the

amount/quality of organic material can also be important and the enhanced respiratory demand associated with an intensified monsoon has been invoked as a driver of the expansion of the Arabian Sea OMZ (Lachkar et al., 2018).

Ocean oxygen changes are also affected by climate variability on interannual and decadal timescales, especially for the tropical ocean OMZs (Deutsch et al., 2011). ENSO variability in particular affects the thermocline structure, which then alongside changes in circulation modulates oxygen solubility and respiratory demand in this region (Ito and Deutsch, 2013; Eddebbar et al., 2017). These drivers may then be combined with modifications to overturning and ventilation of OMZs by lateral jets and equatorial current intensity (Duteil et al., 2014). Centennial scale studies based on isotope proxies for low oxygen regions have demonstrated fluctuations in OMZ extent linked to decadal changes in tropical trade winds that affects interior ocean respiratory oxygen demand, which implies that it will be difficult to attribute recent changes in the Pacific OMZ to anthropogenic forcing alone (Deutsch et al., 2015). Parallel work based on oxygen observations (Llanillo et al., 2013), as well as modelling (Duteil et al., 2018) supports the importance of decadal scale variability in the eastern tropical Pacific OMZ. There is some evidence for the potential of a modulating impact on tropical Pacific oxygen at interannual timescales from atmospheric deposition of nitrogen and iron (Ito et al., 2016; Yang and Gruber, 2016).

Table 5.2: Observed oxygen changes for the period 1970 to 2010 for 6 different layers within the ocean. The changes are shown as percentage change of global averages. The layers are depths 0-100, 100-600, 0-1000, and 600-bottom are in metres. The oxygen minimum zone (OMZ) is defined as the ocean volume change that is less than 80 μmol L^{-1}. The estimates and confidence intervals are based published papers (Schmidtko et al. 2018, Ito et al, 2017 and Helm et al. 2011). The assessed change is the average of the available estimates and the 90% Confidence Interval (CI) combines the confidence as their standard deviation with two degrees of freedom.

		Schmidtko		Ito		Helm		Assessed Change	
Layer	Period	Change	90 CI	Change	90 CI	Change	90 CI	Change	90 CI
0-100	1970-2010	-0.38%	±1.06%	-1.65%	±0.63%	-1.30%	±0.54%	-1.11%	±0.95%
100-600	1970-2010	-1.06%	±1.36%	-3.17%	±1.34%	-2.04%	±0.60%	-2.09%	±1.42%
0-1000	1970-2010	-1.35%	±1.38%	-2.70%	±1.30%	-1.74%	±0.54%	-1.93%	±1.39%
600-bottom	1970-2010	-1.51%	±0.62%	NA	NA	-0.81%	±0.57%	-1.16%	±0.84%
OMZ	1970-2010	6.33%	±2.52%	6.10%	1.2%	4.49%	± 2 . 2 5 %	5.64%	±2.66%
Global	1970-2010	-1.43%	±0.70%	NA	NA	-0.87%	±0.53%	-1.15%	±0.88%

At the global scale, there is *high confidence* that the impact of a warmer ocean on oxygen levels is reinforced by other processes associated with ocean physics and biogeochemistry, which cause the majority of the observed oxygen decline. For the tropical Pacific OMZ, there is *medium confidence* arising from *medium agreement* from *medium evidence* that low frequency decadal changes in ocean physics have controlled past fluctuations in OMZ extent.

0 – 1200 m

DO change [μmol kg^{-1} decade^{-1}]

1200 m – sea floor

DO change [μmol kg^{-1} decade^{-1}]
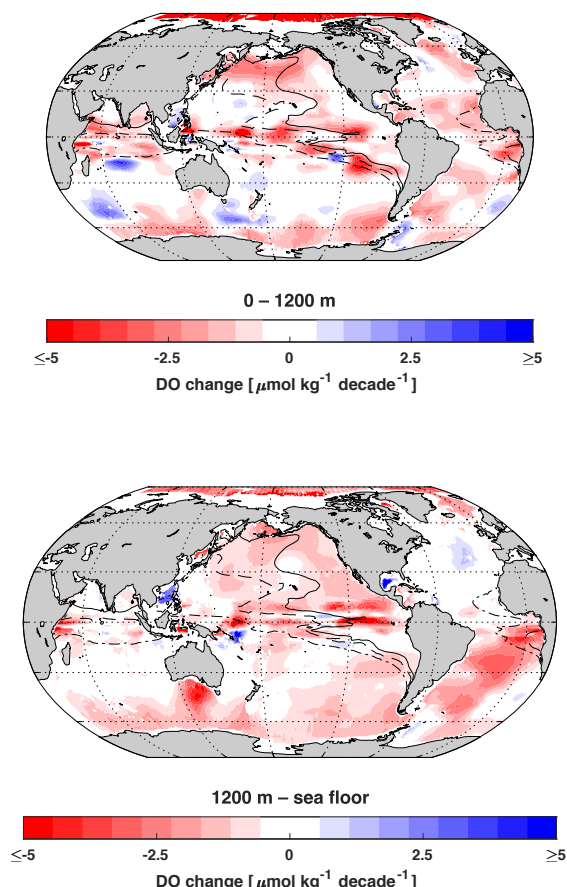
Figure 5.9: Absolute change in dissolved oxygen (umol kg^{-1} per decade) between water depths of (a) 0 and 1,200 m, and (b) 1,200 m and the sea floor over the period 1960-2010. Lines indicate boundaries of OMZs with less than 80 μ mol kg^{-1} oxygen anywhere within the water column (dashed/dotted), less than 40 μ mol kg^{-1} (dashed) and less than 20 μ mol kg^{-1} (solid). Redrawn from Oschlies et al. (2018).

Future changes in oxygen can be appraised from ESMs that account for the combined effects of ocean physics and biogeochemistry. Globally, these models project that it is *very likely* oxygen will decline by 3.2-3.7% or 1.6-2.0% (both across 90% confidence limits) for RCP8.5 or RCP2.6, respectively, relative to 2000 (Bopp et al., 2013). Focussing on the 100-600m depth stratum, O_2 changes by -4 to -3.1% for the RCP8.5 or by -0.5 to +0.1% for the RCP2.6 scenario (relative to 2006-2015, Figure 5.8d). It should be noted that ESMs appear to be underestimating the rate of oxygen change from available datasets from the historical period (Oschlies et al., 2018). Increased tropical ocean stratification reduces interior ocean oxygen by diminishing pathways of ventilation in the subtropical gyres and by inhibiting turbulent mixing with the oxygen-rich surface ocean (see Section 5.2.2.2.4). This relatively robust global modelled trend (Figure 5.8d) however masks important uncertainties in the projection of regional trends (Figure 5.8e), particularly in the tropical ocean OMZs (Bopp et al., 2013; Cocco et al., 2013; Cabré et al., 2015). The uncertainty in the trends in tropical ocean OMZs arises due to the fact that oxygen depletion due to warming induced reductions in oxygen saturation are opposed by oxygen enrichment due to reduced oxygen consumption during respiration in response to predicted declines in marine export production, as well as biases due to model resolution in the tropics and the length of the model spin up (Bopp et al., 2017). The 80 μmol L^{-1} threshold that may be used to define the volume of the oxygen minimum is projected to grow by a *very likely* range of 7.0±5.6% by 2100 during the RCP8.5 scenario or show virtually no change during the RCP2.6 scenario, relative to a 1850-1900 reference period (Figure 5.10). At the seafloor, between 200–3000 m depth strata, the North Pacific, North Atlantic, Arctic and Southern Oceans may see oxygen declines by 0.3% to 3.7% by 2100 (relative to 2005), with abyssal ocean changes being lower and more localised around regions in the North Atlantic and Southern Ocean (Sweetman et al., 2017), but will be modulated by any future changes in overturning strength. There is *high confidence* that the largest changes in deep sea systems will occur after 2100 (Battaglia and Joos, 2018).
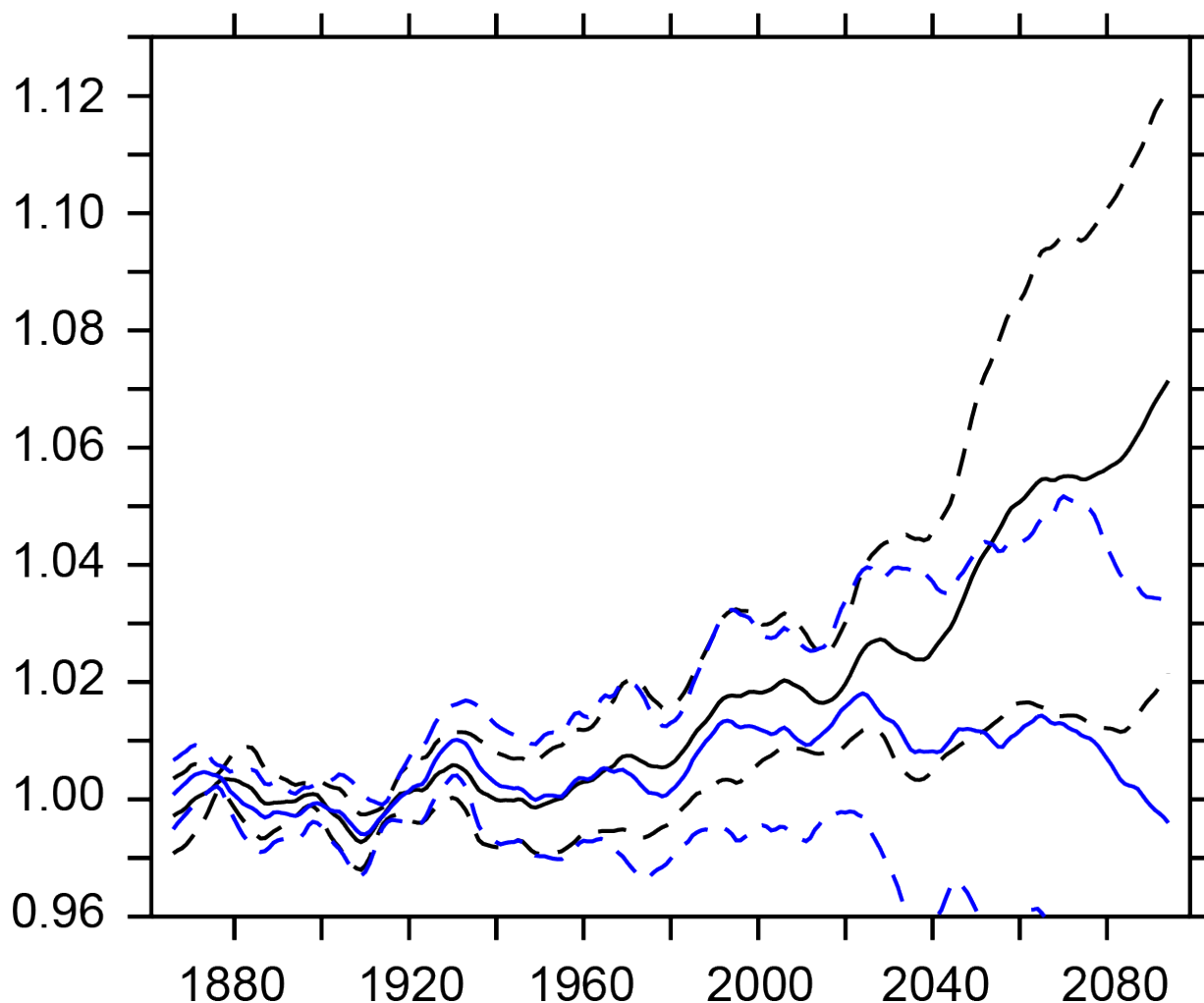
Figure 5.10: The evolution of the volume of the 100-600m layer of the ocean with oxygen concentrations less than 80 μmol L^{-1} for the RCP8.5 (black line) and the RCP2.6 (blue line), normalised to the volume in 1850-1900. Dashed lines indicated the *very likely* range (90% confidence intervals) across the CMIP5 models (CNRM-CM5, GFDL-ESM2M, GFDL-ESM2G, IPSL-CM5A-LR, IPSL-CM5A-MR, MPI-ESM-LR, MPI-ESM-MR and the NCAR-CESM1 models). Models are corrected for drift in O$_2$ using their control simulations.

Simulations extended to 2300 suggest that by 2150 the trend of declining tropical ocean oxygen (both in terms of concentrations and volume of low oxygen waters) may reverse itself, mainly due to the effect of strong declines in primary production and organic matter fluxes to the ocean interior (Fu et al., 2018) or due to enhanced Antarctic ventilation (Yamamoto et al., 2015), but with *low confidence* due to *limited evidence*. At the global scale, 10,000 year intermediate complexity model simulations find that overall ocean oxygen loss shows near linear relationships to equilibrium temperature, itself linearly related to cumulative emissions, and any climate mitigation scenario will reduce peak oxygen loss by 4.4% per degree Celsius of avoided warming (Battaglia and Joos, 2018).

In summary, the total oxygen content of the ocean is *very likely* to decline by 3.2-3.7% by 2100, relative to 2000, for RCP8.5 or by between 1.6-2.0% for RCP2.6 with *medium confidence*. There is *medium confidence* that sea floor changes will be more localised in the North Atlantic and Southern Oceans by 2100, but *high confidence* that the largest deep sea floor changes in oxygen will occur after 2100.

5.2.2.5   *Changing Ocean Nutrients*

Changes to ocean nutrient cycling are driven by modifications to ocean mixing and transport (Section 5.2.2.2.2), internal biogeochemical cycling and fluctuations in external supply, particularly from rivers and

the atmosphere. This assessment will focus on the main nutrients important for driving microbial growth (Section 5.2.2.6), namely nitrogen, phosphorus and iron.

Diverse studies (including shipboard experiments and use of protein biomarkers) have highlighted nitrogen and phosphorus limitation in the stratified tropical ocean regions accompanied by widespread iron limitation at high latitudes and in upwelling regions that typically have elevated levels of productivity (Figure 5.11) (Moore et al., 2013; Saito et al., 2014; Browning et al., 2017; Tagliabue et al., 2017). Moreover, more extensive experimental work has demonstrated overlapping nitrogen-iron co-limitation at the boundaries between gyre and upwelling regimes (Browning et al., 2017). There is *high confidence* arising from *robust evidence* and *high agreement* across different types of studies that the main limiting nutrient is either iron (in most major upwelling regions and the Southern, North Atlantic, and sub-Arctic Pacific Oceans) or nitrogen and phosphorus (in the low productivity tropical ocean gyres).
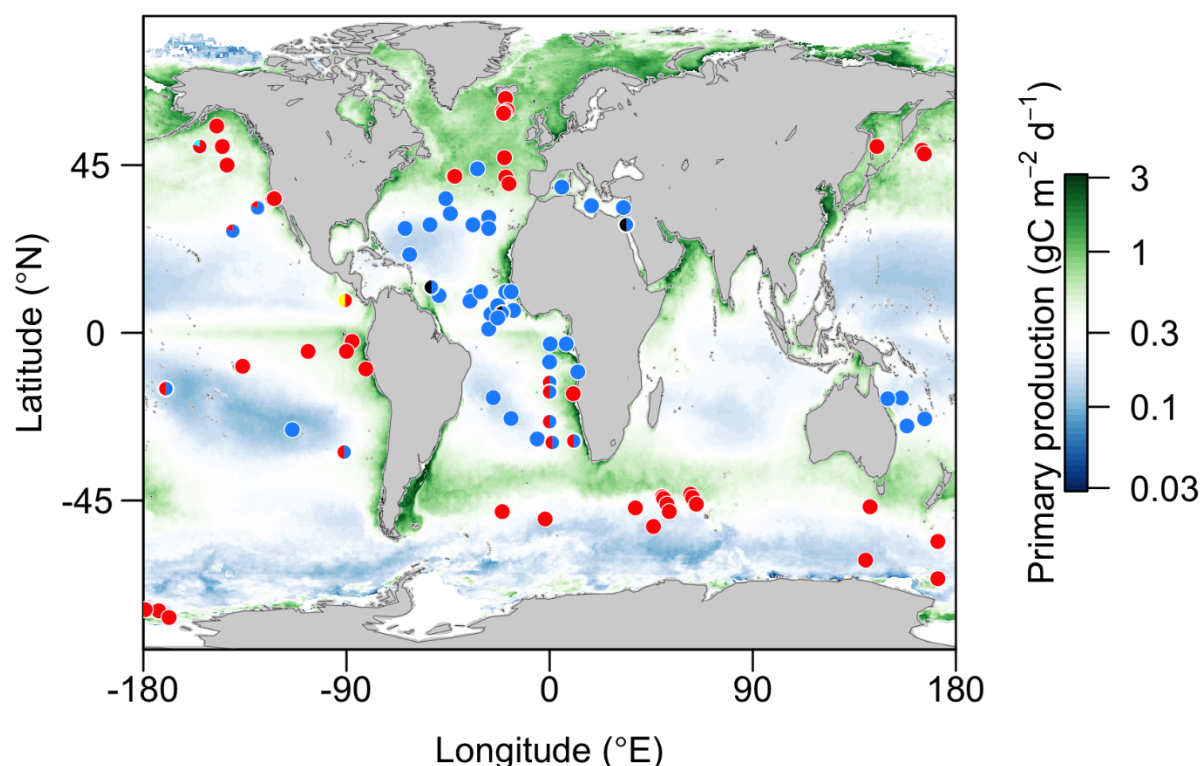
Figure 5.11: Map of the dominant limiting resource (Moore et al., 2013), updated to include new experiments from the north Pacific, tropical Atlantic and south east Atlantic (Browning et al., 2017; Shilova et al., 2017). The background is depth integrated primary productivity using the Vertically Generalized Production Model (VPGM) algorithm. Coloring of the circles indicates the primary limiting nutrients inferred from chlorophyll and/or primary productivity increases following artificial amendment of: N (green), P (black), Fe (red), Co (yellow) and Zn (cyan). Divided circles indicate potentially co-limiting nutrients, e.g., a red-green divided circle indicates Fe-N co-limitation.

There is *limited evidence* on contemporary trends in nutrient levels, either from time-series sites or broader meta-analyses. Increasing inputs of anthropogenic nitrogen from the atmosphere are perturbing ocean nutrient levels (Jickells et al., 2017). In the North Pacific in particular, additional atmospheric nitrogen input has raised the nitrogen to phosphorus ratio between 1988-2011 and induced a progressive shift towards phosphorus limitation in this region (Kim et al., 2011; Kim et al., 2014; Ren et al., 2017). This tendency is supported by modelling experiments that find enhanced atmospheric nitrogen input only has a small influence on productivity due to expanded phosphorus limitation (Yang and Gruber, 2016) and other nitrogen cycle feedbacks (Somes et al., 2016; Landolfi et al., 2017).

In general, future increases in stratification (Dave and Lozier, 2013; Talley et al., 2016b; Kwiatkowski et al., 2017; and see also Section 5.2.2.2) will trap nutrients in the ocean interior and reduce upper ocean nutrient levels, alongside an additional local impact from changes to atmospheric delivery. However, no CMIP5 models accounted for changes in nutrient delivery from dust and anthropogenic aerosols during their experiments, which could be an important component of regional change (Wang et al., 2015b; Somes et al., 2016; Yang and Gruber, 2016). ESMs project a decline in the nitrate content of the upper 100m of 9-14% or 1.5-6% (across 90% confidence intervals) for the RCP8.5 or RCP2.6 scenario, respectively, by 2081-2100 relative to 2006-2015 (Figure 5.8g). The largest absolute declines in nitrate content is projected in the present-day upwelling zones (Figure 5.8h). Projected changes to upper 100m nitrate concentrations are significantly different to zero for both RCP8.5 and RCP2.6 at the 90% confidence level, but are overall lower for the RCP2.6. Scenario, internal variability and inter-model variability contribute roughly equally to the overall projection uncertainty in 2100 (Figure 5.8i) and there is no clear separation of nitrate trends between RCP8.5 and RCP2.6 outside the model uncertainty (Figure 5.8h).

Iron concentrations are projected to increase in the future from ESM simulations, due to enhanced lateral transport into high latitude oceans and reduced biological consumption in regions of declining nitrate (Misumi et al., 2013). Other modelling efforts also suggest greater levels of the more biologically available Fe(II) species in a warmer and more acidic ocean (Tagliabue and Völker, 2011). These modelling studies tend to indicate greater ocean iron availability in the future overall, but the very limited skill of contemporary global ocean iron models in reproducing observations available from the new basin scale datasets from the international GEOTRACES program and neglect for parallel dust supply changes lower the confidence in the models' projected changes (Tagliabue et al., 2016).

Overall, nitrate concentrations in the upper 100m are *very likely* to decline by 9-14% by 2081-2100, relative to 2006-2015 for RCP8.5 or 1.5-6% for RCP2.6, in response to increased stratification, with *medium confidence* in these projections due to the *limited evidence* of past changes that can be robustly understood and reproduced by models. Surface ocean iron levels is projected to increase in the 21st century with *low confidence* due to systemic uncertainties in these models.

5.2.2.6 *Changing Ocean Primary and Export Production*

Ocean primary productivity is a key process in the ocean carbon cycle (see Section 5.2.2.3), as well as for supporting pelagic ocean ecosystems (see Section 5.2.3). Net primary production (NPP) is the product of phytoplankton growth rate and standing stock. Phytoplankton growth is controlled by the combination of temperature, light and nutrients, while the phytoplankton standing stock is modified by both gains from growth and losses due to grazing by zooplankton (Figure 5.12). Export production is here defined as the sinking flux of particulate organic carbon (produced by NPP) across a specified depth horizon. Otherwise known as the biological pump, export production is also a key component of the global carbon cycle (see Section 5.2.2.3) and an essential food supply to benthic organisms (see Section 5.2.3.2). Export production is regulated by the level of primary production and the transfer efficiency with depth, itself controlled by the type of sinking organic carbon, which is affected by the upper ocean food web structure (Boyd et al., 2019).

Satellite datasets that use mathematical algorithms to convert ocean colour, often alongside other remotely sensed information, into chlorophyll or other indexes of phytoplankton biomass and NPP provide the potential to deliver a global meta-analysis of changes in NPP. Since AR5, a variety of studies have reported relatively insignificant changes in overall open ocean chlorophyll levels of $< \pm 1\%$ yr^{-1} for individual time periods (Boyce et al., 2014; Gregg and Rousseaux, 2014; Boyce and Worm, 2015; Hammond et al., 2017). Regionally, trends of ±4% between 2002-2015 for different regions are found when different satellite products are merged, with increases at high latitudes and moderate decreases at low latitudes (Mélin et al., 2017). While some studies report good comparability of merged products (Mélin et al., 2017), others highlight significant mismatches regarding absolute values and decadal trends in NPP between NPP algorithms (Gómez-Letona et al., 2017). Satellite derived NPP shows significant mismatches when compared to *in situ* data and reducing uncertainties in derived NPP is a high priority for the community (Lee et al., 2015), although there is a reasonable correlation in higher biomass coastal regions (Kahru et al., 2009). Importantly, satellite records are not yet long enough to unambiguously isolate long term climate related trends from natural variability (Beaulieu et al., 2013). Overall, there is *low confidence* in satellite-based trends in global ocean NPP due to the time series length and lack of corroborating *in situ* measurements or

other validation time series. This is especially true at regional scales where distinct sets of poorly understood processes dominate.

Future changes in NPP will result from the changing influence from temperature, light, nutrients and grazing (Figure 5.12). Across CMIP models, NPP is predicted to broadly decline or remain constant by 2081-2100, with mean changes by 2100 of – 3.8 to -10.6% and -1.1% to +0.8% across 90% confidence intervals for the RCP85 and RCP26 scenario, respectively (all relative to 2006-2015), with a strong degree of regional symmetry (Figure 5.8k). As seen for nitrate, changes are most marked in low latitude upwelling regions, which are projected to show the largest absolute declines. As for nitrate, projected NPP changes are lower for the RCP26 scenario (Figure 5.8j), but the overall uncertainty is dominated by internal and inter-model variability in 2100 (Figure 5.8l) which results in no clear separation of NPP trends between the RCP85 and RCP26 (Figure 5.8j). Tropical ocean NPP is projected to show a large decline, but is underpinned by substantial intermodal uncertainty, with mean changes of $11 \pm 24\%$ across the suite of CMIP5 models by 2100, relative to 2000 under RCP8.5 (Laufkötter et al., 2015). However, if emergent constraints from the historical record that link the variability of tropical productivity to temperature anomalies then a four-fold decline in inter-model uncertainty results. This leads to a projected tropical ocean decline of $11\pm6\%$, or from 6.8-16.2% across 90% confidence limits, depending on which historical constraint is used (Kwiatkowski et al., 2017). NPP is projected to increases for higher latitude regions, such as the Arctic and Southern Oceans.

Detailed analyses of the interplay between different drivers of NPP, including temperature, light, nutrient levels and grazing from a subset of CMIP5 models, reveals a complex interplay with a strong latitudinal dependence (Laufkötter et al., 2015) summarised in Figure 5.12. Warming acts to enhance growth, most notably at lower latitudes, while light conditions are also predicted to improve, mostly at the poles. Nutrient limitation shows a much more complex response across models, but tends to increase in the tropics and northern high latitudes, with little change in the Southern Ocean. Taken together there is a tendency for reduced growth rates across the entire ocean, but there is a large amount of inter-model variability. The changes in growth are allied to a consistent increase in the grazing loss of biomass to upper trophic levels. Since AR5, we have an increasing body of literature concerning role of biological feedbacks, especially due to interactions between organisms, specific physiological responses and from upper trophic levels on nutrient concentrations, linked to variable food quality (Kwiatkowski et al., 2018), resource recycling (Boyd et al., 2015a; Tagliabue et al., 2017) and interactions between organisms (Lima-Mendez et al., 2015), but their role in shaping the response of NPP to climate change remains a major unknown. Lastly, modelling work suggests that the increasing deposition of anthropogenic aerosols (supplying N and Fe) stimulates biological activity (Wang et al., 2015b) and may compensate for warming driven reductions in primary productivity (Wang et al., 2015b), but these effects do not form part of the CMIP5 projections assessed here.

CMIP5 models show a strong negative relationship between changes in stratification that reduces net nutrient supply and integrated export production (Fu et al., 2016). Export production is projected to decline by 8.9-15.8% or 1.6-4.9% (across 90% confidence intervals) by 2100, relative to 2000 for the RCP8.5 or RCP2.6 scenario, respectively (Bopp et al., 2013; Fu et al., 2016; Laufkötter et al., 2016). The projected changes in export production can be larger than global primary production because they are affected by both the NPP changes, but also how shifts in food web structure modulates the 'transfer efficiency' of particulate organic material (Guidi et al., 2016; Tréguer et al., 2018), which then affects the sinking speed and lability of exported particles through the ocean interior to the sea floor (Bopp et al., 2013; Fu et al., 2016; Laufkötter et al., 2016). Declines in export production over much of the ocean mean that the flux arriving at the sea floor is also predicted to decline, while increases in export production are projected in the polar regions that see enhanced NPP (Sweetman et al., 2017).

The realism in model projections can be appraised via their ability to accurately simulate the limiting nutrient in specific ocean regions (Figure 5.11), with high model skill in reproducing surface distributions of nitrate and phosphate (Laufkötter et al., 2015), raising confidence in projections in nitrogen and phosphorus limited systems, but poor skill in reproducing iron distributions (Tagliabue et al., 2016) lowering confidence in iron limited regions (Figure 5.11). In addition to concentrations of specific nutrients, the response of NPP to environmental change is strongly controlled by accurate representation of the ratio of resources (Moreno et al., 2017). Overall CMIP5 models skill in reproducing patterns of NPP and export production from limited satellite derived estimates range from poor to average (correlation coefficients of 0.1-0.6 across different models (Laufkötter et al., 2016; Moreno et al., 2017)), but it should be noted that complete comprehensive

observational datasets do not exist for these metrics with very few *in situ* observations. As export production is a much better understood net integral of changing net nutrient supply (Sarmiento and Gruber, 2002) and can be constrained by interior ocean nutrient and oxygen levels, there is *medium confidence* in these projections for global changes. Improving the ability of models to reproduce historical NPP is crucial for more accurate projections as model biases in simulating contemporary ocean biogeochemistry play a key role in driving future projections (Fu et al., 2016).

Overall, these assessments balance the range of projections across models alongside the strength of different kinds of observational constraints available, as well as our theoretical or experimental understanding of the impact of a warmer, more stratified ocean on NPP and export production. As for AR5, net primary productivity is *very likely* to decline by 4-11% by 2081-2100, relative to 1850-1900, across CMIP5 models for RCP8.5, but there is *low confidence* for this estimate due to the *medium agreement* among models and the *limited evidence* from observations. It is *very likely* that tropical NPP will decline by 7-16% by 2100 for RCP8.5with *medium confidence*, as there are improved constraints from historical variability in this region. Globally, the increased stratification in the future is *very likely* to reduce export production by 9 to 16% in response to reduced nutrient supply, especially in tropical regions (*medium confidence*).
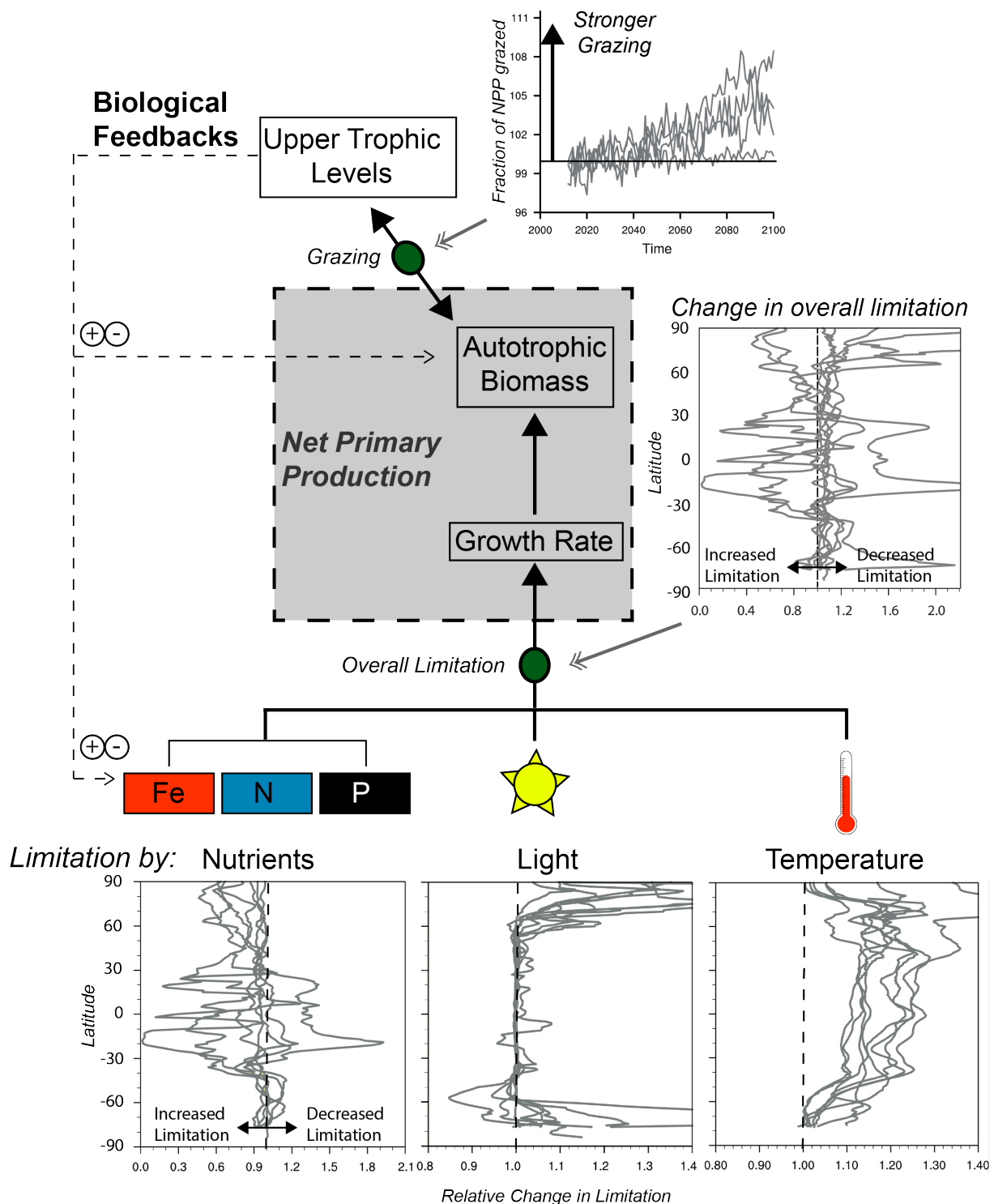
Figure 5.12: A schematic diagram to illustrate how NPP is a combination of microbial growth and biomass. In this context, growth is controlled by three limiting factors (nutrients, light and temperature), while biomass is affected by grazing. The grey lines in the plots represent results from different CMIP5 models as reported by Laufkötter et al. (2015). Poorly understood feedbacks from upper trophic levels on autotroph biomass and nutrients are represented by dashed arrows.

[START BOX 5.1 HERE]

Box 5.1: Time of Emergence and Exposure to Climate Hazards

The concept of time of emergence (ToE) is defined as the time at which the *signal* of climate change in a given variable emerges from a measure of the background variability or *noise* (SROCC Glossary). In

associating a calendar date with the detection, attribution and projection of climate trends, the concept of a ToE has proved useful for policy and planning particularly through informing important climatic thresholds and the uncertainties associated with past and future climate change (Hawkins and Sutton, 2012). However, there is not a single agreed metric and the ToE for a given variable thus depends on choices regarding the space and time scale, the threshold at which emergence is defined and the reference period (WGI AR5 11.3.2.1). Recently, the ToE concept has been expanded to consider variables related to climatic hazards to marine organisms and ecosystems such as pH, carbonate ion concentrations, aragonite and calcite saturation states, nutrient levels and marine primary productivity (Box 5.1, Figure 1) (Ilyina et al., 2009; Friedrich et al., 2012; Keller et al., 2014b; Lovenduski et al., 2015; Rodgers et al., 2015). ToE assessments for the ocean typically quantify the internal variability using the standard deviation of the detrended data over a given time period (Keller et al., 2014b; Rodgers et al., 2015; Henson et al., 2016; Henson et al., 2017), the scenario and model uncertainty associated with different climate scenarios and across available ESMs (Frölicher et al., 2016), and in some cases the autocorrelation of noise (Weatherhead et al., 1998). As more components of 'noise' are accounted for, the ToE lengthens and the ToE is also affected by whether a control simulation or historical variability is used to determine the noise (Hameau et al., 2019).

This assessment considers the ToE of hazards exposed to by marine organisms and ecosystems. These biological components of the ocean respond to climate hazards that emerge locally, rather than to the global and basin-scale averages reported in WGI AR5 (Stocker et al., 2013). Overall, ESMs show that there is an ordered emergence of the climate variables, with pH emerging rapidly across the entire open ocean, followed by SST, interior oxygen, upper ocean nutrient levels and finally NPP under both RCP2.6 and RCP8.5 relative to the 1861-1900 reference period (Box 5.1, Figure 1). Anthropogenic signals remain detectable for over large parts of the ocean even for the RCP2.6 scenario for pH and SST, but are *likely* lowered for nutrients and NPP in the 21st century. For example, for the open ocean, the anthropogenic pH signal in ESM historical simulations is *very likely* to have emerged for three-quarters of the ocean prior to 1950 and it is *very likely* over 95 % of the ocean has already been affected, with little discernable difference between scenarios. The climate signal of oxygen loss will *very likely* emerge from the historical climate by 2050 with a *very likely* range of 59-80% by 2031-2050 and rising with a *very likely* range of 79-91% of the ocean area by 2081-2100 (RCP8.5 emissions scenario). The emergence of oxygen loss is smaller in area under RCP2.6 scenario in the 21st century and by 2090 the emerged area is declining (Henson et al., 2017) (Box 5.1 Figure 1). It has also been shown that changes to oxygen solubility or utilization may emerge earlier than bulk oxygen levels (Hameau et al., 2019).

It must be noted that variability will be greater in the coastal ocean than for the open ocean, which will be important for both hazard exposure for coastal species and the detection of trends. For example, although signals of anthropogenic influences have already emerged from internal variability in the late 20th century for global and basin-scale averaged ocean surface and sub-surface temperature (*very likely*) (AR5 WGI SPM), their ToE and level of confidence vary greatly at local scales and in coastal seas (Frolicher et al., 2016). Pelagic organisms with small range size may thus be more (or less) at risk to warming with earlier (or later) ToE at the scale of the area that they inhabit. From an observational standpoint, analyses that account for autocorrelation of noise suggest time series of around a decade are sufficient to detect a trend in pH or SST, whereas datasets spanning 30 years or longer are typically needed for detection of emergence at local scales for oxygen, nitrate and primary productivity (Henson et al., 2016).
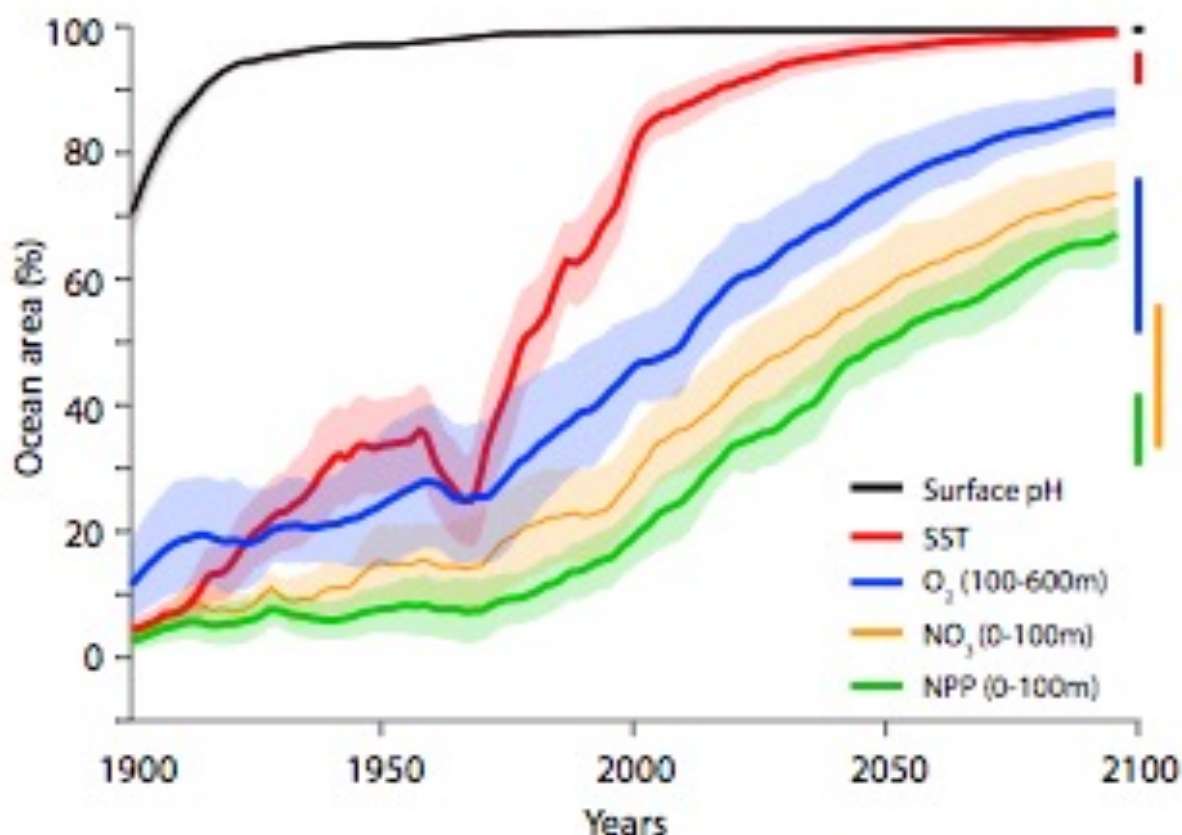
Box 5.1, Figure 1: Time of emergence of key ocean condition variables: sea surface temperature (SST), surface pH, 100-600m oxygen (O_2), 0-100m nitrate (NO_3), and 0-100m integrated net primary production (NPP). The year of emergence represents the year when the mean change relative to the reference period of 1986-2005 is above the standard deviation of each variable over the historical period (Frölicher et al., 2016) and is expressed here in terms of the rate at which different climate signals emerge as a proportion of total ocean area for the RCP8.5 scenario. The final area (and standard deviation) by 2100 under the RCP2.6 scenario is indicated by vertical lines at 2100.

The rapidity of change and its geographic scope, encompassed in the ToE, can be linked to concepts of exposure to hazard and vulnerability of biota. As organisms have evolved to be adaptable to natural variations in the environmental conditions of their habitats, changes to their habitat conditions larger than that typically experienced or specific biological thresholds such as upper temperature or oxygen tolerance may become hazardous (Mora et al., 2013). This would then move from the statistical nature of the 'detection and attribution' nature of the ToE discussed above towards timescales of impacts on organisms useful for ecosystem projections. In doing so, it will be important to think about the differences in habitat suitability between different organisms, including their specific thresholds for specific drivers, e.g. temperature, oxygen or calcium carbonate stability e.g., warming thresholds for coral bleaching (Pendleton et al., 2016) may differ from the temperature and oxygen thresholds for fishes such as Atlantic cod and tunas (Deutsch et al., 2015). Moreover, species with fast generation times relative to the ToE of key habitat conditions (e.g., phytoplankton) may evolve more quickly to environmental change and be less vulnerable to climate change than longer-lived, slower generation time species (e.g., large sharks) (Jones and Cheung, 2018). However, evidence on evolutionary adaptation to expected climate change is limited, thus while shorter generation time may facilitate adaptation to environmental change, it does not necessarily result in successful adaptation of organisms (Section 5.2.3.1).

Earlier ToE and their subsequent biological impacts on organisms and ecosystems increase the urgency of policy responses through both climate mitigation and adaptation (Sections 5.5). However, the rapid emergence of hazards at the local scale in the near-term (already past or in this decade) such as warming and ocean acidification and the resulting impacts on some of the more sensitivity or less adaptive biodiversity and ecosystem services may post challenges for international and regional policies as their often require multiple decades to designate and implement (Box 5.6). In contrast, scope for adaptation for national and

local ocean governance can be more responsive to rapid changes (Sections 5.5.2, 5.5.3). This highlights the opportunities for multi-level adaptation that allows for reducing climate risks that are expected to emergence of stressors and impacts at different time frame (Mackenzie et al., 2014).

[END BOX 5.1 HERE]

5.2.3   Impacts on Pelagic Ecosystems

Marine pelagic ecosystems (the water column extending from the surface ocean down to the deep sea floor) face increasing climate related hazards from the changing environmental conditions (see Section 5.2.2). WGII AR5 (Pörtner et al., 2014) concluded, as also confirmed in Section 5.2.2, that long time-series of more than three or four decades in length are necessary for determining biological trends in the ocean. However, long-term biological observations of pelagic ecosystems are rare and biased toward mid to high latitude systems in the Northern Hemisphere (Edwards et al., 2013; Poloczanska et al., 2013; Poloczanska et al., 2016). This assessment, therefore, combines multiple lines of evidence ranging from experiments, field observations to model simulations to detect and attribute drivers of biological changes in the past, project future climate impacts and risks of pelagic ecosystems. In this section the pelagic ecosystem is subdivided into the surface, epipelagic ocean (<200 m, the uppermost part of the ocean that receives enough sunlight to allow photosynthesis) (Section 5.2.3.1) and the deep pelagic ocean, comprising the twilight, mesopelagic zone (200-1000 m) and the dark, bathypelagic zone (>1000 m deep) (Section 5.2.3.2). Although the WGII AR5 Chapter 30 defined the deep sea as below 1000 m (Hoegh-Guldberg et al., 2014), the absence of photosynthetically useful light and ensuing critical ecological, biogeochemical transformations, and altered human interactions that occur on much of the sea floor below 200 m have led both pelagic and benthic biologists to include the ocean waters and seafloor below 200 m within the definition of the deep sea (Herring and Dixon, 1998; Gage, 2003).

5.2.3.1   The Epipelagic Ocean

This section synthesizes new evidence since AR5 to assess observed changes in relation to the effects of and the interactions between multiple climate and non-climate hazards, and to project future risks of impacts from these hazards on the epipelagic organisms, communities and food web interactions, and their scope and limitation to adapt.

5.2.3.1.1   Detection and attribution of biological changes in the epipelagic ocean
Temperature-driven shifts in distribution and phenology
WGII AR5 concluded that the vulnerability of most organisms to warming is set by their physiology, which defines their limited temperature ranges and thermal sensitivity (Pörtner et al., 2014). Although different hypotheses have been proposed since AR5 to explain the mechanism linking temperature sensitivity of marine organisms and their physiological tolerances (Schulte, 2015; Pörtner et al., 2017; Somero et al., 2017), evidence from physiological experiments and observations from paleo- and contemporary periods continue to support the conclusion from AR5 on the impacts of temperature change beyond thermal tolerance ranges on biological functions such as metabolism, growth and reproduction (Payne et al., 2016; Pörtner and Gutt, 2016; Gunderson et al., 2017), contributing to changes in biogeography and community structure (Beaugrand et al., 2015; Stuart-Smith et al., 2015) (*high agreement, high confidence*). Comparison of biota across land and ocean suggests that marine species are generally inhabiting environment that is closer to their upper temperature limits, explaining the substantially higher rate of local extirpation related to warming relative to those on land (Pinsky et al., 2019). Hypoxia and acidification can also limit the temperature ranges of organisms and exacerbate their sensitivity to warming (Mackenzie et al., 2014; Rosas-Navarro et al., 2016; Pörtner et al., 2017), although interactions vary strongly between species and biological processes (Gobler and Baumann, 2016; Lefevre, 2016).

Shifts in distribution of marine species from phytoplankton to marine mammals continued to be observed since AR5 across all ocean regions (Poloczanska et al., 2016). Recent evidence continues to support that a large proportion of records of observed range shifts in the epipelagic ecosystem (Poloczanska et al., 2016) are correlated with ocean temperature, with an estimated average shift in distribution (including range centroids, northward and southward boundaries) from these records of 51.5 ± 33.3 km per decade since the 1950s (Figure 5.13). Such rate of shift is

significantly faster than those records for organisms in the seafloor; the latter has an average rate of distribution shift of 29.0 ± 15.5 km per decade (44% of the records for seafloor species with range shifts that are consistent with expectation from the observed temperature changes) (Figure 5.13). Comparison of global seafloor-derived planktonic foraminifera from pre-industrial age with recent (from year 1978) communities show that the recent assemblages differ from their pre-industrial with increasing dominance of warmer or cooler species that are mostly consistent with temperature changes (Jonkers et al., 2019). Rate of observed responses also varies between and within animal groups among ocean regions, with zooplankton and fishes having faster recorded range shifts (Pinsky et al., 2013; Asch, 2015; Jones and Cheung, 2015; Poloczanska et al., 2016). For example, analysis of the Continuous Plankton Recorder (CPR) data-series from the North Atlantic in the last decades shows that the range of dinoflagellates tended to closely track the velocity of climate change (the rate of isotherm movement). In contrast, the distribution range of diatoms shifted much more slowly (Chivers et al., 2017) and its distribution seems to be primary influenced by multi-decadal variability rather than from secular temperature trends. The CPR surveys have also provided evidence that some calanoid copepods are expanding poleward in the Northeast Atlantic, at a rate up to 232 km per decade (Beaugrand, 2009; Chivers et al., 2017), although different calanoid species respond differently in the rate and direction of shifts (Philippart et al., 2003; Edwards and Richardson, 2004; Asch, 2015; Crespo et al., 2017). Overall, the observed changes in biogeography are consistent with expected responses to changes in ocean temperature for the majority of marine biota (*high confidence*). This is also consistent with theories and experimental evidence that scale from individual organisms' physiological responses to community level effects (*high confidence*). Sensitivity of organisms' biogeography varies between taxonomic groups (*high confidence*).
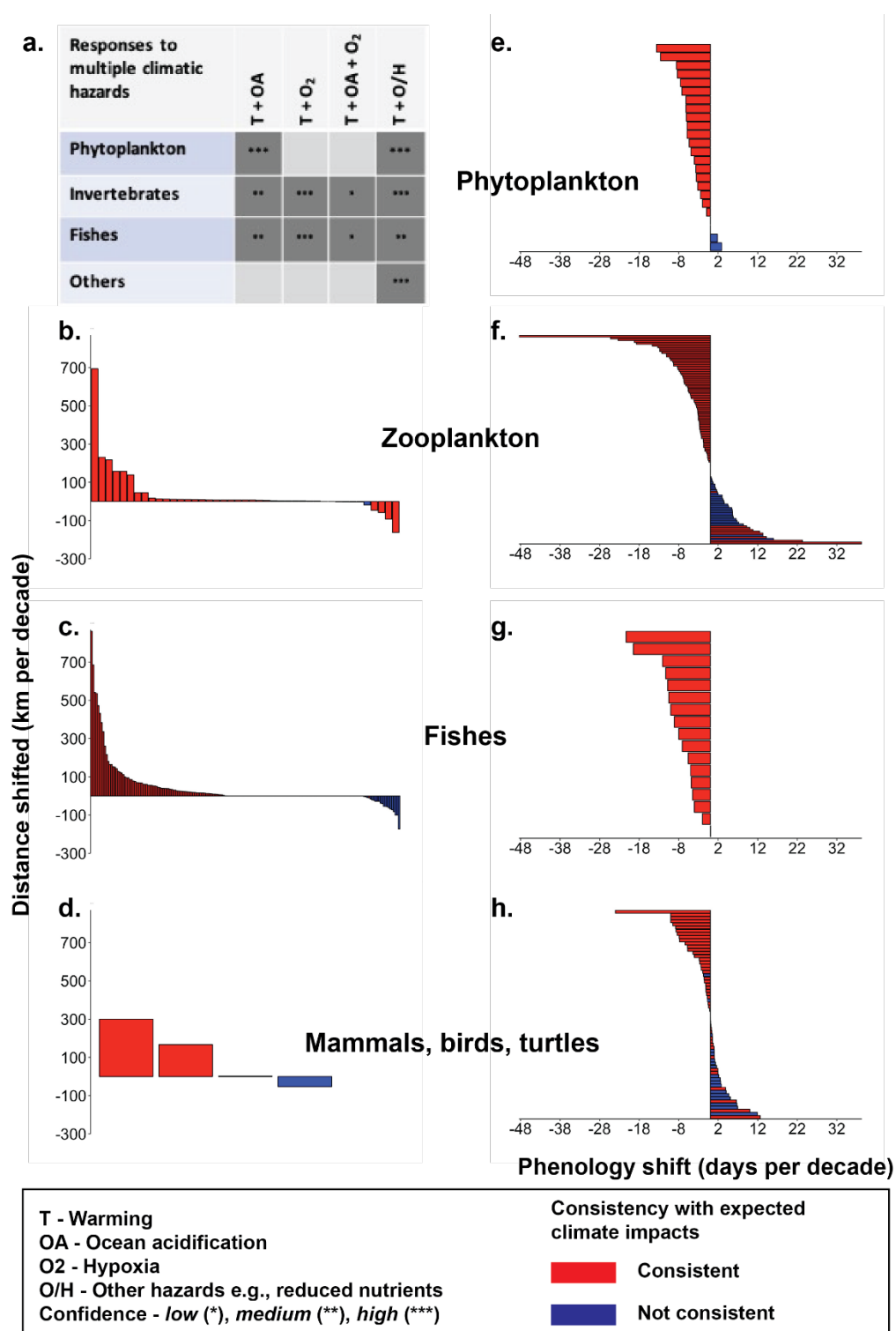
Figure 5.13: Evidence of climate change responses of marine organisms to changes in ocean conditions under climate change. (a) evidence of interactive effects (including synergistic and antagonistic) of multiple climatic hazards (based on Przeslawski et al. (2015); Lefevre (2016); Section 5.2.2, 5.2.3, 5.2.4, 5.3). "Others" mainly include mammals, seabirds and marine reptiles). The lighter-coloured cell represents insufficient information to draw conclusion; (b - d) observations on changes in latitudinal range and (e – h) phenology (based on Poloczanska et al., 2013). For b – h, each bar represents one record.

The rate and direction of observed range shifts are shaped by the interaction between climatic and non-climatic factors (Poloczanska et al., 2013; Sydeman et al., 2015; Poloczanska et al., 2016), such as local temperature and oxygen gradients in the habitat across depth (Cheung et al., 2013; Deutsch et al., 2015), latitude and longitude (Burrows et al., 2014; Barton et al., 2016), ocean currents (Sunday et al., 2015;

Barton et al., 2016; García Molinos et al., 2017), bathymetry in all or part of their life stages (for organisms living on or close to the seafloor) (Pinsky et al., 2013; Kleisner et al., 2015), geographical barriers (Pinsky et al., 2013; Burrows et al., 2014), availability of food and critical habitat (Sydeman et al., 2015), fishing and other non-climatic human impacts (Engelhard et al., 2014; Hoegh-Guldberg et al., 2014). Moreover, observed range shifts in respond to climate change in some regions such as the North Atlantic are strongly influenced by warming due to multi-decadal variability (Edwards et al., 2013; Harris et al., 2014), suggesting that there is a longer time-of-emergence of range shifts from natural variability and a need for longer biological time-series for robust attribution. The rate of shifts in biogeography of organism is influenced by multiple climatic and non-climatic factors (*high confidence*) that can result in non-synchronous shifts in community composition (*high confidence*). There is general under-representation of biogeographical records in low latitudes (Dornelas et al., 2018), rendering detection and attribution of shifts in biogeography in these regions having *medium confidence*. The variation in responses of marine biota to range shifts can cause spatial restructuring of the pelagic ecosystem with consequences for organisms at higher trophic levels (Chivers et al., 2017; Pecl et al., 2017) (*high confidence*). Marine ectotherms have demonstrated some capacity for physiological adjustment and evolutionary adaptation that lowers their sensitivity to warming and decrease in oxygen (Pörtner et al., 2014; Cavallo et al., 2015) (*low confidence*). However, historical responses in abundance and ranges of marine species to ocean warming suggest that adaptation not always suffices to mitigate projected impacts (WGII AR5 Chapter 6) (*high confidence*).

Marine reptiles, seabirds and mammals breathe air, instead of obtaining oxygen from water, and many of them spend some of their life cycle on land, being their abundance and distribution still affected by temperature (Pörtner et al., 2014). Long term population changes and shifts in distribution associated with climate change have been observed for temperate species of seabirds and marine mammals (Henderson et al., 2014; Hiscock and Chilvers, 2014; Ramp et al., 2015) (*high confidence*). For example, Laysan, *Phoebastria immutabilis,* and Wandering, *Diomedea exulans*, albatross have responded positively to climate change as they have been able to take advantage of the increased intensity of winds. This has allowed them to forage farther and faster, making their foraging trips shorter, increasing their foraging efficiency and breeding success (Descamps et al., 2015; Thorne et al., 2016). For reptiles, like sea turtles and snakes, temperature directly affects important life history traits including hatchling size, sex, viability and performance *(high confidence)* (Hays et al., 2003; Pike, 2014; Dudley et al., 2016; Santos et al., 2017). This is particularly important for marine turtles as changing temperatures will affect the hatchling sex ratio because sex is determined by nest site temperature *(high confidence)* (Hatfield et al., 2012; Santidrián Tomillo et al., 2014; Patricio et al., 2017). Loss of breeding substrate, including mostly coastal habitats such as sandy beaches (Section 5.3.3), can reduce the available nesting or pupping habitat for land breeding marine turtles, lizards, seabirds and pinnipeds (Fish et al., 2005; Fuentes et al., 2010; Funayama et al., 2013; Reece et al., 2013; Katselidis et al., 2014; Patino-Martinez et al., 2014; Pike et al., 2015; Reynolds et al., 2015; Marshall et al., 2017) (*high confidence*). Climatic hazards such as sea level rise contributes to the loss of these coastal habitats (see Section 5.3 and Chapter 3). Changes in ocean temperature will also indirectly impact marine mammals, seabirds and reptiles by changing the abundance and distribution of their prey (Polovina, 2005; Polovina et al., 2011; Doney et al., 2012; Sydeman et al., 2015; Briscoe et al., 2017; Woodworth-Jefcoats et al., 2017) (*high confidence*). The distributions of some of these large animals is determined by the occurrence and persistence of oceanic bridges and barriers that are related to climate driven processes (Ascani et al., 2016; McKeon et al., 2016). For example, the decline of Arctic sea-ice is affecting the range and migration patterns of some species and is allowing the exchange of species previously restricted to either the Pacific or Atlantic oceans (Alter et al., 2015; George et al., 2015; Laidre et al., 2015; MacIntyre et al., 2015; McKeon et al., 2016; Breed et al., 2017; Hauser et al., 2017) (Chapter 3). Also, the range expansion of some of these predatory megafauna can affect species endemic to the habitat; for example, while the decrease in summer sea ice in the Arctic may favour the expansion of killer whales (*Orcinus orca*), their occurrence can result in narwhale (*Monodon monoceros*) to avoid the use of key habitats to reduce the risk of killer whales' predation(Bost et al., 2009; Sydeman et al., 2015; Breed et al., 2017) (see Chapter 3; section 3.2.1.4). In addition, marine mammals, seabirds and sea turtles present habitat requirements associated with bathymetric and mesoscale features that facilitate the aggregation of their prey (Bost et al., 2015; Kavanaugh et al., 2015; Hindell et al., 2016; Hunt et al., 2016; Santora et al., 2017). The persistence and location of these features are linked to variations in climate (Crocker et al., 2006; Baez et al., 2011; Dugger et al., 2014; Abrahms et al., 2017; Youngflesh et al., 2017) and to foraging success, juvenile recruitment, breeding phenology, growth rates and population stability (Costa et al., 2010; Ancona and Drummond, 2013; Ducklow et al., 2013; Chambers et al., 2014; Descamps et al., 2015; Abadi

et al., 2017; Bjorndal et al., 2017; Fluhr et al., 2017; Youngflesh et al., 2017) (*high confidence*). Overall, recent evidence further support that impacts of climate change on some marine reptiles, mammals and birds have been observed in recent decades (*high confidence*) and that the direction of impacts vary between species, population and geographic locations (Trivelpiece et al., 2011; Hazen et al., 2013; Clucas et al., 2014; Constable et al., 2014; George et al., 2015) (*high confidence*).

Warming has contributed also to observed changes in phenology (timing of repeated seasonal activities) of marine organisms (Gittings et al., 2018), although observations are biased towards the Northeast Atlantic (Poloczanska et al., 2016; Thackeray et al., 2016). Shifts in the timing of interacting species have occurred in the last decades, eventually leading to uncoupling between prey and predators, with cascading community and ecosystem consequences (Kharouba et al., 2018; Neuheimer et al., 2018). Timing of spring phenology of marine organisms is shifting to earlier in the year under warming, at an average rate of 4.4 ± 1.1 days per decade (Poloczanska et al., 2013), although it is variable among taxonomic groups and among ocean regions (Lindley and Kirby, 2010). This is consistent with the expectations based on the close relationship between temperature and these biological events, supporting evidence from AR5 (Bruge et al., 2016; Poloczanska et al., 2016). Thus, the growing amount of literature and new studies since AR5 WGII and SR1.5 further support that phenology of marine ectotherms in the epipelagic systems are related to ocean warming (*high confidence*) and that the timing of biological events has shifted earlier (*high confidence*).

Observed impacts of multiple climatic hazards

WGII AR5 concludes that multiple climatic hazards from ocean acidification, hypoxia, and decrease in nutrient and food supplies pose risks to marine ecosystems, and the risk can be elevated when combined with warming (Riebesell and Gattuso, 2014; Gattuso et al., 2015). In a recent meta-analysis of 632 published experiments, primary production by temperate non-calcifying plankton increases with elevated temperature and CO_2, whereas tropical plankton decreases productivity because of acidification (Nagelkerken and Connell, 2015). Also, temperature increases consumption and metabolic rates of herbivores but not secondary production; the latter decreases with acidification in calcifying and non-calcifying species. These effects together create a mismatch with carnivores whose metabolic and foraging costs increase with temperature (Nagelkerken and Connell, 2015). Warming may also exacerbate the effects of ocean acidification on the rate of photosynthesis in phytoplankton (Lefevre, 2016). There is some, but limited, reports of observed impacts on calcified pelagic organisms that are attributed to secular trend in ocean acidification and warming (Harvey et al., 2013; Kroeker et al., 2013; Nagelkerken et al., 2015; Boyd et al., 2016). For example, Rivero-Calle et al. (2015) reported, using CPR archives, that stocks of coccolithophores (a group of phytoplankton that forms calcium carbonate platelets) have increased by 2% to over 20% in the North Atlantic over the last five decades, and that this increase is linked to synergistic effects of increasing anthropogenic CO_2 and rising temperatures, as supported by their statistical analysis and a number of experimental studies. Most of the available evidence supports that ocean acidification and hypoxia can act additively or synergistically between each other and with temperature across different groups of biota (Figure 5.13). Limitation of nutrient and food availability and predation pressures can further increase the sensitivity of organismal groups to climate change in specific ecosystems (Riebesell et al., 2017). Climate change also affects organisms indirectly through the impacts on competitiveness between organisms that favour those that are more adaptive to the changing environmental conditions (Alguero-Muniz et al., 2017) and changes in trophic interactions (Seebacher et al., 2014). Overall, direct *in-situ* observations and laboratory experiments show that there are significant responses to the multiple stressors of warming, ocean acidification and low oxygen on phytoplankton, zooplankton and fishes and that these responses can be additive or synergistic (*high confidence,* Figure 5.13).

5.2.3.1.2 Future changes in the epipelagic ocean
WGII AR5 and SR1.5 conclude that projected ocean warming will continue to cause poleward shifts in the distribution and biomass of pelagic species, paralleled by altered seasonal timing of their activities, species abundance, migration pattern and reduction in body size in the 21st century under scenarios of increasing greenhouse gas emission (Pörtner et al., 2014; Hoegh-Guldberg et al., 2018). Simultaneously, projected expansion of oxygen minimum zones and ocean acidification could lead to shifts in community composition toward hypoxia-tolerant and non-calcified organisms, respectively. However, these projected biological changes in the ocean raise questions about how individuals, communities and food webs will respond to the

multiple impacts from climatic and non-climatic stressors in the future, and the feedbacks of the effects of their ecological impacts on modifying the physical and biogeochemical conditions of the ocean (Schaum et al., 2013; Boyd et al., 2016; O'Brien et al., 2016; Moore, 2018). This section focuses on addressing these questions in order to assess the future risk of impacts of climate change on the epipelagic ecosystem.

Future projections on phytoplankton distribution, community structure and biomass

While analysis of outputs from CMIP5 ESMs project that global average net primary production and biomass of phytoplankton community will decrease in the 21st century under RCP2.6 and RCP8.5 (see Section 5.2.2.6). However, the future risk of impacts of epipelagic ecosystem can also depend on changes in community structure of phytoplankton species. Barton et al. (2016) projected the biogeography of 87 taxa of phytoplankton (diatoms and dinoflagellates) in the North Atlantic to 2051-2100 relative to the past (1951-2000) with scenarios of changes in temperature and other ocean conditions such as salinity, density and nutrients under RCP8.5. The study found that 74% of the studied taxa exhibit a poleward shift at a median rate of 12.9 km per decade, but 90% of the taxa shift eastward at a median rate of 42.7 km per decade. Such changes may affect food webs and biogeochemical cycles, and with consequence to the productivity of living marine resources (Stock et al., 2014; Barton et al., 2016).

Outputs from CMIP5 ESMs suggest that projected warming and reduction in nutrient availability in low latitudes, as a result of increasing stratification of the ocean under climate change, will increase the dominance of small-sized phytoplankton, growing more efficiently than larger taxa at low nutrient levels (Dutkiewicz et al., 2013b). Dominant groups in subtropical oceans, like the picoplanktonic cyanobacteria *Synechococcus* and *Prochlorococcus*, are projected to expand their range of distribution towards higher latitudes and increase their abundances by 14–29%, respectively, under a future warmer ocean (Flombaum et al., 2013), although synergistic effects of warming and CO_2 on photosynthetic rates could lead to a dominance of *Synechococcus* over *Prochlorococcus* (Fu et al., 2007) (*low confidence*). Similarly, temperature-driven range shifts towards higher latitudes are also likely for tropical diazotrophic (N_2-fixing) cyanobacteria, although they could disappear from parts of their current tropical ranges where future warming may exceed their maximum thermal tolerance limits (Hutchins and Fu, 2017) (*low confidence*). Modelling experiments show that the effects of warming on phytoplankton community will be exacerbated by ocean acidification at levels expected in the 21st century for RCP8.5, leading to increasing growth rate responses of some phytoplankton groups, such as diazotrophs and *Synechococcus*, with predicted increases in biomass up to 10% in tropical and subtropical waters (Dutkiewicz et al., 2015) (*low confidence*). Furthermore, warming is projected to interact with decreasing oxygen levels and increases in iron in the nutrient-impoverished subtropical waters, favoring the dominance of the diazotrophic colonial cyanobacteria *Trichodesmium* (Sohm et al., 2011; Boyd et al., 2013; Ward et al., 2013; Hutchins and Fu, 2017) (*medium confidence*).

Regional differences in the changes in phytoplankton community and their impacts on epipelagic ecosystem are however complex and depends on multiple interactions of co-varying climate change stressors at regional level (Boyd and Hutchins, 2012). Based on global ocean model simulations, Boyd et al. (2015b) show that the interaction between warming, increased CO_2 and a decline in phosphate and silicate would benefit coccolithophores against diatoms in the northern North Atlantic, despite decreasing rates of calcification. Evidence, based on long-term experiments of acclimation or adaptation to increasing temperatures in combination with elevated CO_2, show that individual growth and carbon fixation rates of coccolithophores at high CO_2 are modulated by temperature, light, nutrients and UV radiation, and could increase calcification while the responses are also species-specific (Lohbeck et al., 2012; Khanna et al., 2013). Calcification of planktonic foraminifera will be however negatively affected by acidification (Roy et al., 2015), and their populations are predicted to experience the greatest decrease in diversity and abundance in sub-polar and tropical areas, under RCP8.5 (Brussaard et al., 2013), however environmental controls of calcite production by foraminifera are still poorly understood (*low confidence*). Boyd et al. (2015b) analysis indicate also that diatoms would benefit from the synergistic effects of increased warming and iron supply in the northern Southern Ocean, as supported by laboratory experiments and field studies with polar diatoms (Rose et al., 2009) (*low confidence)*. At low latitude provinces, projected concurrent increases of CO_2 and iron, and decreases in both nitrate and phosphate supply, may favour nitrogen fixers, but with ocean regional variability, since iron is thought to limit N_2 fixation in the eastern Pacific and phosphorus in the Atlantic Ocean (Gruber, 2019; Wang et al., 2019). However, recent experimental work with the diazotrophic colonial

Trichodesmium and the unicellular *Crocosphaera* have shown a broad range of responses from rising CO_2, with either increases or decreases in N_2 fixation rates, and with mixed evidence on co-limiting processes (Eichner et al., 2014; Garcia et al., 2014; Gradoville et al., 2014; Walworth et al., 2016; Hong et al., 2017; Luo et al., 2019) (*low confidence*).

Overall, the response of phytoplankton to the interactive effects of multiple drivers is complex, and presently ESMs do not resolve the full complexity of their physiological responses (Breitberg et al., 2015; Hutchins and Boyd, 2016; O'Brien et al., 2016), precluding a clear assessment of the effects of these regional distinctive multi-stressor patterns (*high confidence*)

Future projections on zooplankton distribution and biomass

An ensemble of 12 CMIP5 ESMs project average declines of 6.4% ± 0.79% (95% confident limits) and 13.6% ± 1.70%) in zooplankton biomass in the 21st century relative to 1990-1999 historical values under RCP2.6 and RCP8.5 (Kwiatkowski et al., 2019). Also, production of mesozooplankton is projected from a single ESM to decrease by 7.9% between 1951-2000 and 2051-2100 under RCP8.5 (Stock et al., 2014). Such projected decreases in zooplankton biomass and production are partly contributed by climate-induced reduction in phytoplankton production and trophic transfer efficiency particularly in low latitude ecosystems (Stock et al., 2014) (5.2.2.6). The impacts may be larger than these projections if changes in the relative abundance of carbon, nitrogen and phosphorus are considered by the models (Kwiatkowski et al., 2019). The overall projected decrease in zooplankton biomass is characterized by a strong latitudinal differences, with the largest decrease in tropical regions and increase in the polar regions, particularly the Arctic Ocean (Chust et al., 2014; Stock et al., 2014; Kwiatkowski et al., 2019) (Chapter 3) (*high agreement*). However, the projected increase in zooplankton biomass in the polar region may be affected by the seasonality of light cycle at high latitudes that may limit the bloom season at high latitude (Sundby et al., 2016). The projected decrease in zooplankton abundance, particularly in tropical regions, can impact marine organisms higher in the foodweb, including fish populations that are important to fisheries (Woodworth-Jefcoats et al., 2017). Therefore, there is *high agreement* in model projections that global zooplankton biomass will *very likely* reduce in the 21st century, with projected decline under RCP8.5 almost doubled that of RCP2.6 (*very likely*). However, the strong dependence of the projected declines on phytoplankton production (*low confidence*, 5.2.2.6) and simplification in representation of the zooplankton communities and foodweb render their projections having *low confidence*.

Future responses of zooplankton species and communities to climate change are however affected by interactions between multiple climatic drivers. Experiments in laboratory show that acidification could partly counteract some observed effects of increased temperature on zooplankton, although the level and direction of the biological responses vary largely between species (Mayor et al., 2015; Garzke et al., 2016), with results ranging from no effects (Weydmann et al., 2012; McConville et al., 2013; Cripps et al., 2014; Alguero-Muniz et al., 2016; Bailey et al., 2016), to negative effects (Lischka et al., 2011; Cripps et al., 2014; Alguero-Muniz et al., 2017) or positive effects (Alguero-Muniz et al., 2017; Taucher et al., 2017). These differences in response can affect trophic interactions between zooplankton species; for example, some predatory non-calcifying zooplankton may perform better under warmer and lower pH conditions, leading to increased predation on other zooplankton species (Caron and Hutchins, 2012; Winder et al., 2017). Therefore, the large variation in sensitivity between zooplankton to future conditions of warming and ocean acidification suggests elevated risk on community structure and inter-specific interactions of zooplankton in the 21st century (*medium confidence*). Consideration of these species-specific responses may further modify the projected changes in zooplankton biomass by ESMs (Boyd et al., 2015a).

Future projections on fish distribution, size and biomass

Recent model projections since AR5 and SR1.5 continue to support global-scale range shifts of marine fishes at rates of tens to hundreds of km per decade in the 21st century, with rate of shifts being substantially higher under RCP8.5 than RCP2.6 (Jones and Cheung, 2015; Robinson et al., 2015; Morley et al., 2018). Globally, the general direction of range shifts of epipelagic fishes is poleward (Jones and Cheung, 2015; Robinson et al., 2015), while the projected directions of regional and local range shifts generally follow temperature gradients (Morley et al., 2018). Polewards range shifts are projected to result in decreases in species richness in tropical oceans, and increases in mid to high latitude regions leading to global-scale species turnover (sum

of species local extinction and expansion) (Ben Rais Lasram et al., 2010; Jones and Cheung, 2015; Cheung and Pauly, 2016; Molinos et al., 2016) (*medium confidence* on trends, *low confidence* on magnitude because of model uncertainties and limited number of published model simulations). For example, species turnover relative to their present-day richness in the tropical oceans (30ºN to 30ºS) is projected to be 14 to 21% and 37 to 39% by 2031-2050 and 2081-2100 under RCP8.5 (ranges of mean projections from two sets of simulation for marine fish distributions) (Jones and Cheung, 2015; Molinos et al., 2016). In contrast, high latitude regions (> 60ºN and S) is projected to have higher rate of species turnover than the tropics (an average of 48% between the two data sets for region >60ºN). The high species turnover in the Arctic is explained by species' range expansion from lower-latitude and the relatively lower present-day fish species richness in the Arctic. The projected intensity of species turnover is lower under lower emission scenarios (Jones and Cheung, 2015; Molinos et al., 2016) (see also Section 5.4.1) (*high confidence*). Projections from multiple fish species distribution models show hotspots of decrease in species richness in the Indo-Pacific region, and semi-enclosed seas such as the Red Sea and Arabian Gulf (Cheung et al., 2013; Burrows et al., 2014; García Molinos et al., 2015; Jones and Cheung, 2015; Wabnitz et al., 2018) (*medium evidence*, *high agreement*). In addition, geographic barriers such as land boundaries in the poleward species range edge in semi-enclosed seas or lower oxygen water in deeper waters are projected to limit range shifts, resulting in larger relative decrease in species richness (*medium confidence*) (Cheung et al., 2013; Burrows et al., 2014; García Molinos et al., 2015; Jones and Cheung, 2015; Rutterford et al., 2015).

Warming and decrease in oxygen content is projected to impact growth of fishes, leading to reduction in body size and contraction of suitable environmental conditions (Deutsch et al., 2015; Pauly and Cheung, 2017), with the intensity of impacts being directly related to the level of climate change. The projected reduction in abundance of larger-bodied fishes could reduce predation and exacerbate the increase in dominance of smaller-bodied fishes in the epipelagic ecosystem (Lefort et al., 2015). Fishes exposed to ocean acidification level expected under RCP8.5 showed impairments of sensory ability and alteration of behaviour including olfaction, hearing, vision, homing and predator avoidance (Kroeker et al., 2013; Heuer and Grosell, 2014; Nagelkerken et al., 2015). The combined effects of warming, ocean deoxygenation and acidification in the 21st century are projected to exacerbate the impacts on the body size, growth, reproduction and mortality of fishes, and consequently increases their risk of population decline (*medium evidence, high agreement, high confidence*).

Table 5.3: Projected changes in total animal biomass by the mid- and end- of the 21st century under RCP2.6 and RCP8.5. total animal biomass is based on 10 sets of projections for each RCP under the Fisheries and Marine Ecosystems Impact Model Intercomparison Project (FISMIP) (Lotze et al., 2018). The *very likely* ranges of the projections (95% confidence intervals) are provided. Reference period is the present-day (1986-2004).

	Total animal biomass (%)			
	RCP2.6		RCP8.5	
Region	2031-2050	2081-2100	2031-2050	2081-2100
>60º N	8.4 ± 9.3	8.5 ± 13.7	7 ± 9.2	-1.1 ± 20.2
30ºN – 50ºN	-8.1 ± 4	-4.5 ± 3.6	-10.1 ± 4.7	-21.3 ± 9.4
30ºN – 30ºS	-7.2 ± 2.7	-7.3 ± 3.1	-9 ± 3.6	-23.2 ± 9.5
30ºS – 50ºS	-3.3 ± 2.1	-3.5 ± 2.5	-4.2 ± 2.9	-9 ± 9.8
<60º S	1.7 ± 4.5	-0.9 ± 2.9	0.7 ± 3.9	12.4 ±11.9

An ensemble of global-scale marine ecosystem and fisheries models that are part of the Fisheries and Marine Ecosystems Impact Models Intercomparison Project (FishMIP) undertook coordinated simulation experiments and projected future changes in marine animals (mainly invertebrate and fish) globally under climate change (Lotze et al., 2018). These models represent marine biota and ecosystems differently, ranging from population-based to functional traits- and size-based structure and their responses are driven primarily by temperature and net primary production, although oxygen, salinity and ocean advection are considered in a subset of models and play a secondary role in affecting the projected changes in biomass (Blanchard et al., 2012; Fernandes et al., 2013; Carozza et al., 2016; Cheung et al., 2016a). Overall, potential total marine animal biomass is projected to decrease by 4.3 ±2.0% (95% confident intervals) and 15.0±5.9% under RCP2.6 and RCP8.5, respectively, by 2080-2099 relative to 1986-2005, while the decrease is around 4.9% by 2031-2050 across all RCP2.6 and RCP8.5 (*very likely*) (Figure 5.14). Accounting for the removal of

biomass by fishing exacerbates the decrease in biomass for large-bodied animals which are particularly sensitive to fishing (*likely* for the direction of changes). Regionally, total animal biomass decreases largely in tropical and mid-latitude oceans (*very likely*) (Table 5.3, Figure 5.14) (Bryndum-Buchholz et al., 2019). The high uncertainty and the *low confidence* in the projection in the Arctic Ocean (Chapter 3) is because of the large variations in simulation results for this region between the ESMs and between the FishMIP models, as well as the insufficient understanding of the oceanographic changes and their biological implications in the Arctic Ocean. In the Southern Ocean, the decrease in consumer biomass is mainly in the southern Indian Ocean while other parts of the Southern Ocean are projected to have an increase in animal biomass by 2100 under RCP8.5, reflecting mainly the projected pattern of changes in net primary production from the ESMs (see Section 5.2.2.6).
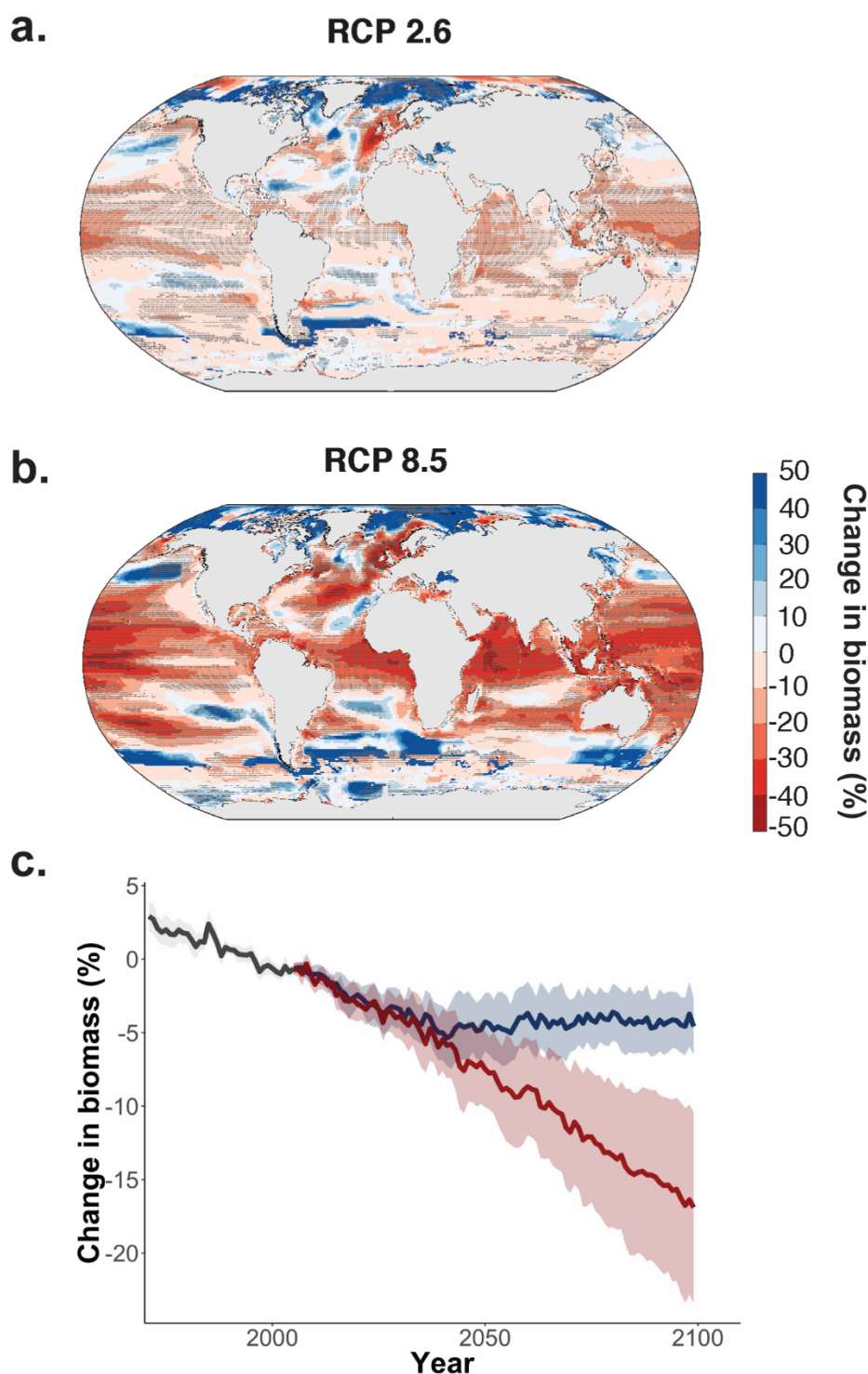
Figure 5.14: Projected changes in total animal biomass (including fishes and invertebrates) based on outputs from 10 sets of projections for each RCP from the Fisheries and Marine Ecosystems Impact Model Intercomparison Project (FISMIP, https://www.isimip.org/gettingstarted/marine-ecosystems-fisheries/) (Lotze et al., 2018); (a, b) multi-model mean change (%) in un-fished total marine animal biomass in 2085-99 relative to 1986-2005 under RCP2.6 and RCP8.5, respectively. Dotted area represents 8 out of 10 sets of model projections agree in the direction of change (c) projected change in global total animal biomass from 1970 to 2099 under RCP2.6 (red) and RCP8.5 (blue). Variability among different ecosystem and Earth-system model combinations (n = 10) expressed as the *very likely* range (95% confidence interval).

Future projections on epipelagic components of the biological pump

A wide range of studies, from laboratory experiments, mesocosm enclosures, synthesis of observations to modeling experiments, provide insights into how the multi-faceted components of the 'biological pump' (the physical and biologically mediated processes responsible for transporting organic carbon from the upper ocean to depth) are projected to be altered in the coming decades. A synthesis of the individual components reported to both influence the performance of the biological pump, and which are sensitive to changing ocean conditions, is presented in Table 5.4. The table lists the putative controlling of each environmental factor, such as warming, that influences the biological pump, and the reported modification (where available) of each individual factor by changing ocean conditions for both the epipelagic ocean and the deep ocean. Analyses of long-term trends in primary production and particle export production, as well as model simulations, reveal that increasing temperatures, leading to enhanced stratification and nutrient limitation, will have the greatest influence on decreasing the flux of particulate organic carbon (POC) to the deep ocean (Bopp et al., 2013; Boyd et al., 2015a; Fu et al., 2016; Laufkötter et al., 2016). However, different lines of evidence (including observation, modeling and experimental studies) provide *low confidence* on the mechanistic understanding of how climatic drivers affect different components of the biological pump in the epipelagic ocean, as well as changes in the efficiency and magnitude of carbon export in the deep ocean (see section below and Table 5.4); this renders the projection of future contribution of the biological carbon pump to the export of POC to the deep ocean having *low confidence*.

Table 5.4: Projected future changes to the ocean biological pump (adapted from Boyd et al. (2015a)). Environmental controls on individual factors that influence downward POC flux are based on published reports from experiments (denoted by **E**), modelling simulations (**M**) and observations (**O**). In some cases, due to the paucity, and regional specificity, of published reports it has been indicated the sign of the projected change on export (in italics), as opposed to the magnitude. NPP: Net Primary Production; POC: Particulate Organic Carbon; DOC: Dissolved Organic Carbon; TEP: Transparent Exopolymer Particles; OA: Ocean Acidification. Climate change denotes multiple controls such as nutrients, temperature and irradiance, as parameterised in coupled ocean atmosphere models. *denotes observed for low latitudes only. ** represents major uncertainty over environmental modulation of this component of the biological pump. ***denotes joint influence of temperature and acidification.

Pump component	Oceanic driver	Projected change (by year 2100)	Confidence	References & Lines of evidence
Epipelagic Ocean				
Phytoplankton growth	Temperature (warming)	~10% Faster (nutrient-replete) no change (nutrient-deplete)	High	(Boyd et al., 2013) **E**; (Maranon et al., 2014) **O***
Net Primary Production (NPP)	Climate change (temperature, nutrients, CO₂)	10-20% decrease (low latitudes); 10-20% increase (high latitudes)	Medium	(Bopp et al., 2013) **M**
Partitioning of NPP (POC, TEP, DOC)	OA	~20% increase in TEP production	Medium	(Engel et al., 2014) **E**; (Riebesell et al., 2007) **E**; (Seebah et al., 2014) **E**
Food web retention of NPP	OA	Enhanced transfer of organic matter to higher trophic levels, reduced N and P sedimentation by 10%	Low	(Boxhammer et al., 2018) **E**
Floristic shifts	Climate change (warming, salinity, OA, iron)	Shift to smaller or larger cells (*less export vs more export; inconclusive*)	Low	(Moràn et al., 2010) **O**; (Li et al., 2009) **O**; (Dutkiewicz et al. 2013a) **M;** (Tréguer et al., 2018) **O;** (Sett et al., 2014) **E**
Differential susceptibility	Temperature (warming)	Growth-rate of grazers more temperature dependent than prey (*less export*)	Low	(Rose and Caron, 2007) **O**
Bacterial hydrolytic effects	Warming, OA	Increase under warming and low pH (variable response in different plankton communities)	Low	(Burrell et al., 2017) **E**
Grazer physiological responses	Warming	Copepods had faster respiration and ingestion rates, but higher mortality (*inconclusive*)	Low	(Isla et al., 2008) **E**
Faunistic shifts	Temperate and subpolar zooplankton species shifts	Temperature (*inconclusive*)	Low	(Edwards et al., 2013) **O**
Food web amplification	Warming	Zooplankton negatively amplify the climate change signal that propagates up from phytoplankton in tropical regions and positively amplify in polar regions	Low	(Chust et al., 2014) **M**; (Stock et al., 2014) **M**
Deep Ocean				
Bacterial hydrolytic enzyme activity	Temperature	20% increase (resource-replete) to no change (resource-deplete)	Low	(Wohlers-Zöllner et al., 2011) **E**; (Endres et al., 2014) **E**; (Bendtsen et al., 2015) **E**; (Piontek et al., 2015) **E****
Particle sinking rates (viscosity)	Warming	5% faster sinking/ °C warming	Low	(Taucher et al., 2014) **M**
Mesozooplankton community composition	Temperature**	Shifts which increase/decrease particle transformations (*less/more export, respectively*)	Low	(**Burd and Jackson, 2002**) **M**; (Ikeda et al., 2001) **O**
Vertical migrators	Climate change (irradiance, temperature)	(*more export*)	Low	(Almén et al., 2014) **O**; (Berge et al., 2014) **O**
Deoxygenation	Climate change	(*more export*)	Low	(Rykaczewski and Dunne, 2010) **M**; (Cocco et al., 2013) **O;** (Hofmann and Schellnhuber, 2009) **M**

1   *5.2.3.2   The Deep Pelagic Ocean*
2
3   *5.2.3.2.1   Detection and attribution of biological changes in the deep ocean*
4   The pelagic realm of the deep ocean represents a key site for remineralisation of organic matter and long-
5   term biological carbon storage and burial in the biosphere (Arístegui et al., 2009), but the observed effects of
6   climate change on deep-sea organisms, communities and biological processes are largely unknown (*high*
7   *confidence*). Observational and model-based methods provide *limited evidence* that the transfer efficiency of
8   organic carbon to the sea floor is partly controlled by temperature and oxygen in the mesopelagic zone,
9   affecting microbial metabolism and zooplankton community structure, with highest efficiencies for high-
10 latitude and oxygen minimum zones (OMZs) (see Section 5.2.2.4 for more detail on OMZs), while below
11 1000 m organic carbon transfer is controlled by particle sinking speed (Boyd et al., 2015a; Marsay et al.,
12 2015; DeVries and Weber, 2017). However, there are contrasting results and *low confidence* on whether
13 transfer efficiencies are highest at low or high latitudes (Boyd et al., 2015a; Marsay et al., 2015; Guidi et al.,
14 2016; DeVries and Weber, 2017; Sweetman et al., 2017). There is also *low confidence* on the effects of
15 increasing temperatures on particulate organic carbon (POC) remineralisation to CO_2 versus POC
16 solubilisation to dissolved organic carbon (DOC) by microbial communities and its storage as refractory
17 DOC (i.e., with life times of >16,000 years) (Legendre et al., 2015).
18
19 *5.2.3.2.2   Future changes in the deep ocean*
20 The global magnitude of the biological pump and how this will be affected by climate change is also
21 uncertain. Model-based studies agree in projecting a global decline in particle gravitational flux to the deep-
22 sea floor, but with regional variability in both the total particle export flux and transfer efficiency (DeVries
23 and Weber, 2017; Sweetman et al., 2017) (see Sections 5.2.2 and 5.2.4). However, recent evidence suggest
24 that other physical and biological processes may contribute nearly as much as the gravitational flux to the
25 carbon transport from the surface to the deep ocean (Boyd et al., 2019), with *low confidence* on the future
26 rate of change in magnitude and direction of these processes. In particular, the 'active flux' of organic
27 carbon due to vertical migration of zooplankton and fishes has been reported to account from 10 to 40% of
28 the gravitational sinking flux (Bianchi et al., 2013; Davison et al., 2013; Hudson et al., 2014; Jónasdóttir et
29 al., 2015; Aumont et al., 2018; Gorgues et al., 2019). Predictions based on model studies suggest that
30 mesopelagic zooplankton and fish communities living at deep scattering layers (DSLs) will increase their
31 biomass by 2100, enhancing their trophic efficiency, because of deep-ocean warming (Section 5.2.2.1;
32 Figures 5.2 and 5.3) and shallowing of DSL (Proud et al., 2017) (*low confidence*). Expansion of OMZs (see
33 Section 5.2.2.4) will also widen the DSL and increase the exposure of mesopelagic organisms to shallower
34 depths (Gilly et al., 2013; Netburn and Anthony Koslow, 2015). In the California Current, the abundance of
35 mesopelagic fishes is closely tied to variations in the OMZ, whose dynamic is linked to the Pacific Decadal
36 Oscillation and El Niño-Southern Oscillation (ENSO) cycles (Koslow et al., 2015). Some large predators,
37 like the Humboldt squid, could indirectly benefit from expanding OMZs due to the aggregation of their
38 primary food source, myctophid fishes (Stewart et al., 2014). However, many non-adapted fish and
39 invertebrates (like diurnal vertical migrators) will have their depth distributions compressed, affecting the
40 carbon transport and trophic efficiency of food webs in the mesopelagic (Stramma et al., 2011; Brown and
41 Thatje, 2014; Rogers, 2015) (*low confidence*). In OMZ waters, where zooplankton is almost absent, like in
42 the Eastern Tropical North Pacific, the microbial remineralisation efficiency of sinking particles would be
43 reduced, eventually increasing the transfer efficiency of organic matter to the deep ocean and thus biological
44 carbon storage (Cavan et al., 2017) (*low confidence;* Table 1). However, increases in ocean temperature may
45 also lead to shallower remineralisation of POC in warm tropical regions, counteracting the storage of carbon
46 in the dark ocean (Marsay et al., 2015). Overall, the direct impacts of climate change on the biological pump
47 are not well understood for the deep pelagic organisms and ecosystems (Pörtner et al., 2014), and there is
48 *low confidence* on the effect of climate change drivers on biological processes in the deep ocean (Table 5.1).
49
50
51 **5.2.4   Impacts on Deep Seafloor Systems**
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53 *5.2.4.1   Changes on the Deep Seafloor*
54
55 The deep seafloor is assessed here as the vast area of the ocean bottom >200 m deep, beyond most
56 continental shelves (Levin and Sibuet, 2012; Boyd et al., 2019) (Figure 5.15). Below 200 m changes in
57 light, food supply, and the physical environment lead to altered benthic (seafloor) animal taxonomic

1   composition, morphologies, lifestyles, and body sizes collectively understood to represent the deep sea
2   (Tyler, 2003).
3
4
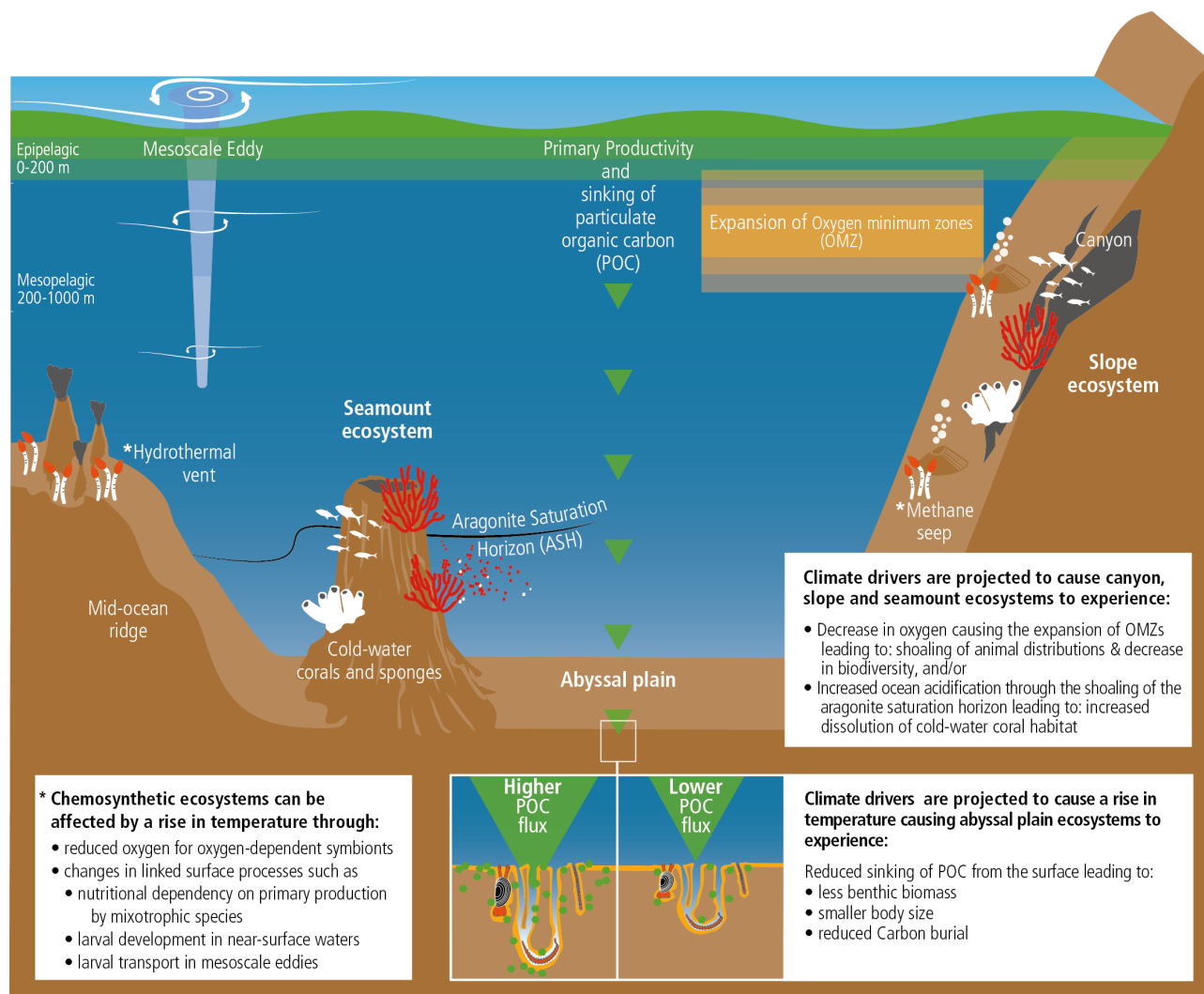
Figure 5.15: A conceptual diagram illustrating how climate drivers are projected to modify deep-sea ecosystems as
discussed in Section 5.2.4.

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8
9
10 Most deep-seafloor ecosystems globally are experiencing rising temperatures, declining oxygen levels, and
11 elevated CO_2, leading to lower pH and carbonate undersaturation (WGII AR5 30.5.7; Section 5.2.2.3). Small
12 changes in exposure to these hazards by deep seafloor ecosystem have been confirmed by observation over
13 the past 50 years. However, analysis using direct seafloor observations of these hazards over the past 15-29
14 years suggest that the environmental conditions are highly variable over time because of the strong and
15 variable influences by ocean conditions from the sea surface (Frigstad et al., 2015; Thomsen et al., 2017).
16 Such high environmental variability makes it difficult to attribute observed trends to anthropogenic drivers
17 using existing datasets (Smith et al., 2013; Hartman et al., 2015; Soltwedel et al., 2016; Thomsen et al.,
18 2017) (*high confidence*). Projections from global Earth system models suggest large changes for
19 temperature by 2100 and beyond under RCP8.5 (relative to present day variation) (Mora et al., 2013;
20 Sweetman et al., 2017; FAO, 2019). The magnitude of the projected changes is lower under RCP 2.6, and in
21 some cases the direction of projected change to 2100 varies regionally under either scenario (FAO, 2019)
22 (*high confidence*).
23
24 *5.2.4.2   Open Ocean Seafloor - Abyssal Plains (3000-6000 m)*
25
26 Abyssal communities (3000-6000 m) cover over 50% of the ocean's surface and are considered to be
27 extremely food limited (Gage and Tyler, 1992; Smith et al., 2018). There is a strong positive relationship

1   between surface primary production, export flux, and organic matter supply to the abyssal seafloor (Smith et
2   al., 2008), with pulses of surface production reflected as carbon input on the deep seafloor in days to months
3   (Thomsen et al., 2017). Both vertical and horizontal transport contribute organic matter to the sea floor
4   (Frischknecht et al., 2018). Food supply to the seafloor regulates faunal biomass, explaining the strong
5   positive relationships documented between surface production and seafloor faunal biomass in the Pacific
6   Ocean (Smith et al., 2013), Gulf of Mexico (Wei et al., 2011) and North Atlantic Ocean (Hartman et al.,
7   2015). Extended time series and broad spatial coverage reveal strong positive relationship between annual
8   POC flux and abyssal sediment community oxygen consumption (Rowe et al., 2008; Smith et al., 2016a).
9   Observed reduction in in POC flux at the abyssal seafloor enhances the relative importance of the microbial
10 loop and reduces the importance of benthic invertebrates in carbon transfer (Dunlop et al., 2016) (single
11 study, *limited evidence*). However, changes in the overlying mesopelagic and bathypelagic communities (see
12 Section 5.2.3.2) will also affect food flux to the deep seafloor, as nekton and zooplankton transfer energy to
13 depth through diel (daily day-night) vertical migrations, ontogenetic (life-staged based) migrations and falls
14 of dead carcasses (Gage, 2003). Therefore, climate change impacts on organic carbon export from the
15 epipelagic (Section 5.2.3.1) and deeper pelagic systems (Section 5.2.3.2) can affect the energy available to
16 support the abyssal seafloor ecosystems (*medium confidence*). However, because observations on historical
17 changes in POC flux in abyssal seafloor ecosystems are limited to a few locations, long-term records show
18 high variability, and mechanistic understanding of factors affecting the biological carbon pump is
19 incomplete, there is *limited evidence* that the abyssal seafloor ecosystem has already been affected by
20 changes in POC flux as a result of climate change. The metabolic rate of deep seafloor ectotherms, and
21 consequently their demand for food, increases with temperature. Thus, observed warming in deep-sea
22 ecosystems (Hoegh-Guldberg et al., 2014) (Section 5.2.2.2.1) is expected to increase the sensitivity of deep
23 seafloor biota to decrease in food supplies associated with a change in POC flux (*high confidence*).
24 However, there is *limited evidence* of observed changes in abyssal biota. Small deep-sea biota demonstrate
25 increased efficiency (effective use of food energy for growth and metabolism with minimal loss) at low food
26 inputs (due to small size and dominance by prokaryotic taxa) (Gambi et al., 2017). Adaptation to low food
27 availability in abyssal ecosystems may confer higher capacity to adjust to reduced food availability than for
28 shallow biota (*limited evidence)*. Overall, the risk of impacts of climate change on abyssal ecosystems
29 through reduction in food supplies from declining POC flux in the present day is low with *low confidence*.
30
31 The globally integrated export flux of carbon is projected to decrease in the open ocean in the 21st century
32 under RCP2.6 (by 1.6-4.9%) and RCP8.5 (by 8.9-15.8%) relative to 2000 (*medium confidence*) (Section
33 5.2.2.6). This change in export flux of carbon is projected to yield declines in POC flux at the abyssal
34 seafloor (representing food supply to benthos) of up to -27% in the Atlantic and up to -31 to -40% in the
35 Pacific and Indian Oceans, with some increases in polar regions (Sweetman et al. 2017). In some models,
36 additional dissolution of calcium carbonate due to ocean acidification further lowers POC flux, causing the
37 projected export production declines to be up to 38% at the northeast Atlantic seafloor (Jones et al., 2014).
38 Lower POC fluxes to the abyss reduce food supply and have been projected to cause a size-shift towards
39 smaller organisms (Jones et al., 2014), resulting in rising respiration rates, lower biomass production
40 efficiency, and lesser energy transfer to higher trophic levels (Brown et al., 2004) (*medium confidence*).
41 Changes are projected to be largest for macrofauna and lesser and similar for megafauna and meiofauna
42 (Jones et al., 2014) (*limited evidence*, *low confidence*). Projections using outputs from seven CMIP5 models
43 suggest that 97.8% ±0.6% (95% CI) of the abyssal seafloor area will experience a biomass decline by 2091-
44 2100 relative to 2006-2015 under RCP 8.5. The projected decreases in overall POC flux to the abyssal
45 seafloor are projected to cause a 5.2% to 17.6 % reduction in seafloor biomass in 2090-2100, relative to
46 2006-2015 under RCP8.5 (Jones et al., 2014). The projected impacts on abyssal seafloor biomass are
47 significantly larger under RCP8.5 than RCP4.5 (Jones et al., 2014). However, existing estimates are based
48 on total POC flux changes and do not account for changes in the type or quality of the sinking material, to
49 which macrofaunal and meiofaunal invertebrates are highly sensitive (Smith et al., 2008; Smith et al., 2009;
50 Tittensor et al., 2011). The projections also do not account for direct faunal responses to changes in
51 temperature, oxygen or the carbonate system, all of which will influence benthic responses to changing food
52 availability (AR5 Chapter 30.5.7), reducing to *medium confidence* the risk assessment that is based on these
53 projections (Fig. 5.16).
54
55 Regionally, while reductions in POC flux are projected at low and mid latitudes in the Pacific, Indian and
56 Atlantic Oceans, increases are projected at high latitudes associated in part with reduction in sea ice cover
57 (Yool et al., 2013; Rogers, 2015; Sweetman et al., 2017; Yool et al., 2017; FAO, 2018a) (see Chapter 3)

(*medium confidence*). Notably, Arctic and Southern Ocean POC fluxes at the abyssal seafloor are projected
to increase by up to 38% and 21%, respectively by 2100 under RCP 8.5 (Sweetman et al., 2017). While an
increase in food supply may yield higher benthic biomass at high latitudes, warmer temperatures and
reduced pH projected for the polar regions (Chapter 3) would elevate faunal metabolic demands, likely
diminishing the benefit of elevated food supply to an unknown extent (Sweetman et al., 2017). Overall,
given the limited food availability for fauna in the abyssal plains and the projected warming (Section
5.2.2.2.2) that increases the demand for food to support the elevated metabolic rates, the projected decrease
in influx of organic matter and seafloor biomass will result in high risks of impacts to abyssal ecosystems by
the end of the 21st century under RCP8.5 (*medium confidence*) (Figure 5.16). The risk of impacts is projected
to be substantially lower under RCP4.5 or RCP2.6 (*high confidence*). The impacts on abyssal seafloor
ecosystems affect functions that are important to support ecosystem services (see Section 5.4.1). For
example, smaller-sized organisms exhibit reduced bioturbation intensity and depth of mixing causing
reduced carbon sequestration (Smith et al., 2008) (Figure 5.15).

5.2.4.3   Bathyal Ecosystems (200 m – 3000 m)

Bathyal ecosystems consist of numerous geomorphic features with steep topography (Figure 5.15). These
include continental slopes covering 5.2% of the seafloor, over 9,400 steep-sided canyons, and > 9,000
conical seamounts (submarine volcanos which are mainly inactive), as well as guyots and ridges which
together cover ~ 6% of the seafloor (Harris et al., 2014). Seamounts and canyons support high animal
densities and biomass including cold-water coral, sponge and bryozoan reefs, exhibit high secondary
production supported by locally enhanced primary production and intensified water flow, function as
diversity hotspots and serve as stepping stones for larval dispersal (Rowden et al., 2010). Canyons transport
particulate organic matter, migrating plankton and coarse material from the shelf, and are sites where
intensified mixing and advection of water masses occurs (De Leo et al., 2010; Levin and Sibuet, 2012;
Fernandez-Arcaya et al., 2017).   Slopes, canyons and seamounts exhibit strong vertical temperature, oxygen
and pH gradients generating sharp ecological zonation (Levin and Sibuet, 2012), thus changes in exposures
are expected to alter the distributions of their communities (Figure 5.15, 5.16*) (medium confidence).*

In some regions, observational records document changing conditions in bathyal ecosystems (Levin, 2018;
Section 5.2.2.4). In the Northeast Pacific continental slopes associated with the California Current
ecosystem, observations over the past 25 years show high variability but an overall trend of decreasing ocean
oxygen and pH levels with oxygen declines of up to 40% and pH declines of 0.08 units in California .
(Goericke et al., 2015) (high agreement, robust evidence, high confidence). Large oxygen declines are linked
to past warming events on continental margins, over multiple time scales from 1-100 ky (Dickson et al.,
2012; Moffitt et al., 2015). Studies across modern oxygen gradients on slopes reveal that suboxic (5-10
μMol kg^{-1} O$_2$) values lead to loss of biodiversity of fish (Gallo and Levin, 2016), invertebrates (Levin, 2003;
Gallo and Levin, 2016; Sperling et al., 2016), and protozoans (Bernhard and Reimers, 1991; Gooday et al.,
2000; Moffitt et al., 2014) (high confidence). Shoaling oxyclines on continental slopes have altered depth
distributions of multiple co-occurring echinoid species over the past 25 y (Sato et al., 2017) and can reduce
the growth rate, and change the skeletal structure and biochemical composition of a common sea urchin
(Sato et al., 2017). In central Pacific oceanic canyons, fish abundance and diversity are reduced at 4-5x
higher oxygen concentrations than on continental slopes (< 31 μMol kg^{-1} O$_2$) (De Leo et al., 2012). Low
oxygen on continental slopes causes reductions in faunal body size and bioturbation (Diaz and Rosenberg,
1995; Levin, 2003; Middelburg and Levin, 2009; Sturdivant et al., 2012), simplification of trophic structure
reducing energy flow to upper trophic levels (Sperling et al., 2013), shifts in carbon processing pathways
from metazoans to protozoans (Woulds et al., 2009), and reduced colonization potential (Levin et al., 2013).
These changes are expected to lead to altered ecosystem structure and function, with lower carbon burial
(Smith et al., 2000; Levin and Dayton, 2009) (medium confidence). Both carbon sequestration and nitrogen
recycling are highly sensitive to small changes in oxygenation within the suboxic zone (Deutsch et al.,
2011).

Bathyal species adapted to OMZs where CO2 levels are characteristically high, appear less vulnerable to the
negative impacts of ocean acidification (Taylor et al., 2014). Benthic foraminifera, which are often the
numerically dominant deep-sea taxon, show no significant effect of short-term exposure to ocean
acidification on survival of multiple species (Dissard et al., 2010; Haynert et al., 2011; Keul et al., 2013;
McIntyre-Wressnig et al., 2014; Wit et al., 2016) and in fact hypoxia in combination with elevated pCO2

favors survival of some foraminifera (Wit et al., 2016). However, lower pH exacerbates shallow foraminiferal sensitivity to warming (Webster et al., 2016). Limited evidence suggests that combined declines in pH and oxygen may lead to increase in some agglutinating taxa and a decrease in carbonate-producing foraminifera, including those using carbonate cement (van Dijk et al., 2017). Exposure to acidification (0.4 unit pH decrease) reduces fecundity and embryo development rate in a bathyal polychaete. Where both oxygen and CO2 stress occur together on bathyal slopes, oxygen can be the primary driver of change (Taylor et al., 2014; Sato et al., 2017). Nematodes are sensitive to changes in temperature (Danovaro et al., 2001; Danovaro et al., 2004; Yodnarasri et al., 2008) and elevated CO2 (Barry et al., 2004; Fleeger et al., 2006; Fleeger et al., 2010). There is low agreement about the direction of meiofaunal responses among studies, reflecting opposing responses in different regions. However, there is high agreement that meiofauna are sensitive to change in environment and food supply (*medium confidence*). Additional research is needed across all taxa on how hypoxia and pH interact (Gobler and Baumann, 2016).

Continental slopes, seamounts and canyons (200-2,500 m) are projected to experience significant warming, pH decline, oxygen loss and decline in POC flux by 2081-2100 (compared to 1951-2000) under RCP 8.5 (Table 5.5). In contrast, the average changes are projected to be 30-50% less under RCP 2.6 (Table 5.5) by 2081-2100. Most ocean regions at bathyal depths (200-2,500 m) except the Southern and Arctic Oceans are predicted to experience on average declining export POC flux under RCP 8.5 by 2081-2100 (Yool et al., 2017; FAO, 2019) with the largest declines of 0.7-8.1 mg C m^{-2} d^{-1} in the Northeast Atlantic (FAO, 2019). There is a strong macroecological relationship between depth, export POC flux, biomass and zonation of macrobenthos on continental slopes (Wei et al., 2011), such that lower POC fluxes will alter seafloor community biomass and structure (*medium confidence*) (See also Section 5.2.4.1). This is modified on the local scale by near-bottom currents, which alter sediment grain size, food availability, and larval dispersal (Wei et al., 2011).

Declines in faunal biomass (6.1 ± 1.6% 95% C.I) are predicted for 96.6% ± 1.2% of seamounts under RCP 8.5 by 2091-2100 relative to 2006-2015, driven by a projected 13.8% ± 3.3% drop in POC flux (Jones et al., 2014). The majority (85%) of mapped canyons are projected to experience comparable benthic biomass declines (Jones et al., 2014). By 2100 under RCP8.5, pH reductions exceeding -0.2 pH units are projected in ~ 23% of North Atlantic deep-sea canyons and 8% of seamounts (Gehlen et al., 2014a), with potential negative consequences for their cold-water coral habitats (See Box 5.2).

Mean temperature (warming) signals are projected to emerge from background variability before 2040 in canyons of the Antarctic, northwest Atlantic, and South Pacific (FAO, 2019). Enhanced stratification and change in the intensity and frequency of downwelling processes under atmospheric forcing (including storms and density-driven cascading events would alter organic matter transported through canyons (Allen and Durrieu de Madron, 2009) (*low confidence*). Changes in the quantity and quality of transferred particulate organic matter, as well as physical disturbance during extreme events cause a complex combination of positive and negative impacts at different depths along the canyon floor (Canals et al., 2006; Pusceddu et al., 2010). Canyons and slopes are recognized as hosting many methane seeps and other chemosynthetic habitats (e.g., whale and wood falls) supported by massive transport of terrestrial organic matter (Pruski et al., 2017); their climate vulnerabilities are discussed below.

Seamounts have been proposed to serve as refugia for cold-water corals facing shoaling aragonite saturation horizons (Tittensor et al., 2011), but could become too warm for deep-water corals in some regions (e.g., projections off Australia) (Thresher et al., 2015) (*one study, low confidence*). Seamounts are major spawning grounds for fishes; reproduction on seamounts may be disrupted by warming (Henry et al., 2016) (*one study, low confidence*). In the North Atlantic, models suggest seamounts are an important source of cold-water coral larvae that maintain resilience under shifting NAO conditions (Fox et al., 2016), thus loss of suitable seamount habitat may have far-reaching consequences (Gehlen et al., 2014b) (*low confidence*) (also see Box 5.2).

Table 5.5: Projected climate changes from the present to 2081-2100 given as mean (min, max) at the deep seafloor for continental slopes, canyons, seamounts and cold-water corals mapped from 200 m to 2500 m under RCP 8.5 and 2.6 Projections are based on three 3-D, fully coupled earth system models (as part of CMIP5): the Geophysical Fluid Dynamics Laboratory's ESM 2G (GFDL-ESM-2G); the Institut Pierre Simon Laplace's CM6-MR (IPSL-CM5A-MR);

1   and (iii) the Max Planck Institute's ESM-MR (MPI-ESM-MR). Export flux at 100 m was converted to export POC flux
2   at the seafloor (*epc*) using the Martin curve following the equation: *epc = epc100 (depth/export depth)$^{-0.858}$*. Projections
3   were made onto the (i) slope from a global ocean basin mask from World Ocean Atlas 2013 V2 (NOAA, 2013), (ii)
4   global distribution of seamounts with summits between 200 and 2 500 m (Kim et al., 2011); (iii) global distribution of
5   submarine canyons with canyon heads shallower than 1, 500 m (Harris and Whiteway, 2011); and (iv) global
6   occurrence of cold-water corals between 200 and 2 500 m.
7

	Temperature (ºC)	pH	DO (μMol kg^{-1})	POC flux (mgC m^{-2} d^{-1})
	RCP 2.6	RCP 2.6	RCP 2.6	RCP 2.6
Continental slopes	+0.30 (-0.44, + 2.30)	-0.06 (-0.19, -0.02)	-3.1 (-49.3, +61.7)	-0.39 (-16.0, +3.9)
Canyons	+0.31 (-0.27, +1.76)	-8.53 (-49.76, +17.82)	-3.5 (-44.7, +29.3)	-0.33 (-10.53, +3.53)
Seamounts	+0.13 (0.01, +0.67)	-4.42 (-27.4, +2.27)	-3.46 (-18.9, +4.1)	-0.15 (-2.20, +1.33)
Cold water corals	+4.3 (-0.29, +1.85)	-0.07 (-0.13, 0.0)	-3.5 (-25.6, +24.7)	-0.7 (-10.5, +3.4)
	RCP 8.5	RCP 8.5	RCP 8.5	RCP 8.5
Continental slopes	0.75 (-8.4, +4.4)	-0.14 (-0.44, -0.02)	-10.2 (-67.8, +53.8)	-0.66 (-33.33, + 10.3)
Canyons	+0.19 (-0.03, +1.14)	-0.11 (-0.35, +0.02)	-0.8 (-28.8, +10.1)	-0.80 (-28.76, +10.07)
Seamounts	+0.66 (-0.75, +3.19)	-0.3 (-0.19, +0.001)	-0.50 (-7.2, +3.0)	-0.50 (-7.18, +2.98)
Cold water corals	+0.96 (-0.42, +3.84)	-0.15 (-0.39, 0.001)	-10.6 (-59.2, + 11.1)	-1.69 (-20.1, + 4.6)

8
9
10   *5.2.4.4   Chemosynthetic Ecosystems*
11
12   Despite having nutrition derived largely from chemosynthetic sources fueled by fluids from the earth's
13   interior, hydrothermal vent and methane seep ecosystems are linked to surface ocean environments and
14   water-column processes in many ways that can expose them to aspects of climate change (*medium*
15   *confidence*). The reliance of vent and seep mussels on surface-derived photosynthetic production to
16   supplement chemosynthetic food sources (Riou et al., 2010; Riekenberg et al., 2016; Demopoulos et al.,
17   2019), and in some cases as a cue for synchronized gametogenesis (sperm and egg production) (Dixon et al.,
18   2006; Tyler et al., 2007) can make them vulnerable to changing amounts or timing of POC flux to the deep
19   seabed in most areas except high latitudes, or to changes in timing of surface production (see Section
20   5.2.2.5) (*limited evidence*) Most of the large, habitat-forming (foundation) species at vents and seeps such as
21   mussels, tubeworms, and clams require oxygen to serve as electron acceptor for aerobic hydrogen-, sulfide-
22   and methane oxidation (Dubilier et al., 2008) and appear unable to grow under dysoxic conditions (< 5-10
23   μmol kg^{-1} O$_2$) (Sweetman et al., 2017) (*medium confidence*). The distributions of these taxa at seeps could
24   be constrained by climate-driven expansion of midwater oxygen minima (Stramma et al., 2008; Schmidtko
25   et al., 2017), which is occurring at water depths where seep ecosystems typically occur on continental
26   margins (200-1000 m). Rising bottom temperatures or shifting of warm currents on continental margins
27   could increase dissociation of buried gas hydrates on margins (Phrampus and Hornbach, 2012) (*low*
28   *confidence*) potentially intensifying anaerobic methane oxidation (which produces hydrogen sulfide)
29   (Boetius and Wenzhoefer, 2013) and expanding cover of methane seep communities (*limited* evidence).
30   Larvae of vent species such as bathymodiolin mussels, alvinocarid shrimp, and some limpets that develop in
31   or near surface waters (Herring and Dixon, 1998; Arellano et al., 2014), are likely to be exposed to warming
32   waters, decreasing pH and carbonate saturation states, and in some places, reduced phytoplankton
33   availability (Section 5.2.2), causing reduced calcification and growth rates (as in shallow water mussel
34   larvae, Frieder et al. (2014)) (*limited evidence, low confidence*). Larvae originating at vents or seeps beneath
35   upwelling regions may also be impaired by effects of hypoxia associated with expanding oxygen minimum
36   zones (Stramma et al., 2008) during migration to the surface (*limited evidence)*. Warming and its effects on
37   climate cycles have the potential to alter patterns of larval transport and population connectivity through
38   changes in circulation (Fox et al., 2016) or surface-generated mesoscale eddies (Adams et al., 2011) (*limited*
39   *evidence*; *low confidence*). Climate-induced changes in the distribution and cover of vent and seep
40   foundation species may involve alteration of attachment substrate, food, and refuge for the many habitat-
41   endemic species that rely on them (Cordes et al., 2010) and for the surrounding deep-sea ecosystems which
42   interact through transport of nutrients and microbes, movement of vagrant predators and scavengers, and
43   plankton interactions (Levin et al., 2016) (*limited evidence*; *low confidence*). There is, however, insufficient

1 analysis of faunal symbiont and nutritional requirements, life histories, larval transport and cross-system
2 interaction to quantify the extent of the consequences described above under future climate conditions.
3
4
5 [START BOX 5.2 HERE]
6
7 **Box 5.2: Cold-Water Corals and Sponges**
8
9 Cold-water corals and sponges form large reefs at the deep seafloor mostly between 200 m-1500 m, creating
10 complex 3-dimensional habitat that supports high biodiversity; they are found at the highest densities on
11 hard substrates of continental slopes, canyons, and seamounts (Buhl-Mortensen et al., 2010). The meta-
12 analysis reported in AR5 Chapter 6 Table 6-3 (Pörtner et al., 2014), identifies 10 studies involving 6 species
13 of cold-water corals that suggest low vulnerability to CO_2 changes at RCP 6.0 and medium vulnerability at
14 RCP 8.5, with negative effects starting at pCO_2 of 445 µatm.
15
16 Scleractinian corals have the capacity to acclimate to high CO_2 conditions due to their capacity to upregulate
17 the pH at the calcification site (Form and Riebesell, 2011; Rodolfo-Metalpa et al., 2015; Gori et al., 2016).
18 The most widely distributed, habitat-forming species in deep water (e.g., *Lophelia pertusa [*renamed
19 *Desmophyllum pertusum*) (Addamo et al., 2016)can continue to calcify at aragonite undersaturation and high
20 CO_2 levels projected for 2100 (750-1100 uatm) based on experiments (Georgian et al., 2016; Kurman et al.,
21 2017) and observations along the natural gradient of carbon chemistry in their distributions (Fillinger and
22 Richter, 2013; Movilla et al., 2014; Baco et al., 2017) (Appendix 1) (*robust evidence*, *medium agreement*,
23 *medium confidence)* and thus appear to be able to acclimate to rising CO_2 levels (Hennige et al., 2015).
24 However, net calcification rates (difference between calcification and dissolution) of *L. pertusa* exposed to
25 aragonite-undersaturated conditions ($\Omega_{arag} < 1$, where Ω_{arag} =aragonite saturation state) often decreases to
26 close to zero or even becomes negative (Lunden et al., 2014; Hennige et al., 2015; Büscher et al., 2017), with
27 genetic variability underpinning ability to calcify at low aragonite saturation states (Kurman et al., 2017).
28 Additionally, skeletons become longer, thinner and weaker (Hennige et al., 2015), and bioerosion is
29 enhanced (e.g., by bacteria, fungi, annelids and sponges) (Schönberg et al., 2017), exacerbating effects of
30 dissolution of the skeleton. *L. pertusa* can calcify when exposed to multiple environmental stresses in the
31 laboratory (Hennige et al., 2015; Büscher et al., 2017), but cannot survive with warming above water
32 temperatures of 14°C to 15ºC or oxygen concentrations below 1.6 ml l^{-1} in the Gulf of Mexico, 3.3 ml l^{-1} in
33 the North Atlantic, 2 ml l^{-1} in the Mediterranean, and 0.5 to 1. 5 ml l^{-1} in the SE Atlantic (Brooke et al., 2013;
34 Lunden et al., 2014; Hanz et al., 2019), highlighting the existence of critical thresholds for cold water coral
35 populations living at the edge of their tolerance. The role of temporal dynamics, species-specific thermal
36 tolerances, and food availability in mediating the response to combinations of stressors is recognized but is
37 still poorly studied under *in situ* conditions (Lartaud et al., 2014; Naumann et al., 2014; Baco et al., 2017).
38
39 Sponges also form critical habitat in the deep ocean but are much less well studied than CWC with respect to
40 climate change. The geologic record, modern distributions and evolutionary and metabolic pathways suggest
41 that sponges are more tolerant to warm temperatures, high CO_2 and low oxygen than are CWC (Schulz et al.,
42 2013). One habitat forming, deep-sea sponge along with its microbiome (microbial inhabitants) has been
43 shown in laboratory experiments to tolerate a 5°C increase in temperature, albeit with evidence of stress
44 (Strand et al., 2017), while ocean acidification (pH 7.5) reduces the feeding of two deep-sea demosponge
45 taxa (Robertson et al., 2017).
46
47 Generally, the deep-sea areas where cold-water corals may be found are projected to be exposed to multiple
48 climate hazards in the 21st century because of the projected ocean warming, oxygen loss, and decrease in
49 POC flux (Table 5.5) under scenarios of greenhouse gas emissions. The average changes in these climate
50 hazards for coral-water corals are projected to be almost halved under RCP2.6 relative to RCP 8.5 (Table
51 5.5). Under RCP 8.5, 95% ±2% (95% CI) of coral-water coral habitats are projected to experience animal
52 biomass decline (-8.6% ± 2.0%) globally by 2091-2100 relative to 2006-2015, driven by a projected 21% ±
53 9% drop in POC flux (Jones et al., 2014). However, nutritional co-reliance of coral-water corals on
54 zooplankton (Höfer et al., 2018) and carbon fixation by symbiotic microbes (Middelburg et al., 2015), is not
55 incorporated into the models, adding uncertainty to these estimates. Regionally, suitable habitat for coral-
56 water corals in the NE Atlantic is projected to decrease with multiple climatic hazards (warming,
57 acidification, decreases in oxygen and POC flux) under RCP 8.5 for 2081-2100 (FAO, 2019), with up to

1   98% loss of suitable habitat by 2099 due to shoaling aragonite saturation horizons. In the Southern
2   hemisphere, a tolerance threshold of 7°C and decline of aragonite saturation below that required for survival
3   (Ω_{arag} < 0.84) can cause large loss of cold-water corals habitat (*Solenosmilia variabilis*) on seamounts off
4   Australia and New Zealand under future projections of warming and acidification to 2099 at RCP 4.5 and
5   nearly complete loss under RCP 8.5 (Thresher et al., 2015).
6
7   Overall, cold-water corals can survive conditions of aragonite-undersaturation associated with ocean
8   acidification but sensitivity varies among species and skeletons will be weakened (*medium confidence*). The
9   largest impacts on calcification and growth will occur when aragonite saturation is accompanied by warming
10 and/or decrease in oxygen concentration beyond the tolerance limits of these corals (*medium confidence*).
11 Given present day occurrence of 95% of cold-water corals above the aragonite saturation horizon (Guinotte
12 et al., 2006) and that no adaptation has been detected with regard to increased dissolution of exposed
13 aragonite (Eyre et al., 2014), there is limited scope for the non-living components of cold water corals and
14 for the large, non-living reef framework that comprises deep water reefs to avoid dissolution under RCP 8.5
15 in the 21st century (*high confidence*). Multiple climatic hazards of warming, deoxygenation, aragonite under-
16 saturation and decrease in POC flux are projected to negatively affect cold-water corals worldwide from the
17 present day by 2100 (*high confidence*). Uncertainty remains in the adaptive capacity of living cold-water
18 corals to cope with these changes and in the influence of altered regional current patterns on connectivity
19 (Fox et al., 2016; Roberts et al., 2016). Sponges and the habitat they form may be less vulnerable than cold-
20 water corals to warming, acidification and deoxygenation that will occur under RCP 8.5 in 2100 (*low
21 confidence*).
22
23 [END BOX 5.2 HERE]
24
25
26 ***5.2.5   Risk Assessment of Open Ocean Ecosystems***
27
28 This section synthesizes the assessment of climate impacts on open ocean and deep-seafloor ecosystem
29 structure and functioning and the levels of risk under future conditions of global warming (see SM5.2). The
30 format for Figure 5.16 matches that of Figure 19.4 of AR5 (Pörtner et al., 2014) and Figure 3.20 of SR1.5
31 (Hoegh-Guldberg et al., 2018), indicating the levels of additional risk as colors (white, yellow, red and
32 purple). Each column in Figure 5.16 indicates how risks increase with ocean warming, acidification (OA),
33 deoxygenation, and POC flux with a focus on present day conditions (2000s) and future conditions by the
34 year 2100 under low (RCP 2.6) and high (RCP 8.5) CO_2 emission scenarios. The transition between the
35 levels of risk to each type of ecosystem is estimated from key evidence assessed in earlier parts of this
36 Chapter (Sections 5.2.2, 5.2.3, 5.2.4). Sea surface temperature is chosen to provide an indication of the
37 changes in all these variables because it is closely related to cumulative carbon emission (Gattuso et al.,
38 2015) which is the main climatic driver of the hazards. Sea surface temperature slightly (<0.1 °C) from
39 global atmospheric temperature (Karl et al., 2015), while the transition values may have an error of ±0.3 °C
40 depending on the consensus of expert judgment. The deep seafloor embers are generated based on earth
41 system model projection of climate variables to the seafloor under RCP 2.6 and RCP 8.5 scenarios, and then
42 translated to RCP- associated change in SST. The assessed confidence in assigning the levels of risk at
43 present day and future scenarios are *low, medium, high*, and *very high* levels of confidence. A detailed
44 account of the procedures involved in the ember for each type of ecosystem, such as their exposure to
45 climate hazards, sensitivity of key biotic and abiotic components, natural adaptive capacity, observed
46 impacts and projected risks, and regional hotspots of vulnerability is provided in the SM5.2 and Table 5.5.
47 The risk assessment for cold-water corals is in agreement with the conclusions in AR5 Ch6.3.1.4.1, although
48 more recent literature is assessed in Box 5.2 and Table SM5.5.
49
50 Overall, the upper ocean (0−700 m) and 700-2000m layers have both warmed from 2004 to 2016 (*virtually
51 certain*) and the abyssal ocean continues to warm in the Southern Hemisphere (*high confidence*). The ocean
52 is stratifying; observed warming and high latitude freshening are both surface intensified trends making the
53 surface ocean lighter at a faster rate than deeper in the ocean (*high confidence*) (Section 5.2.2.2). It is *very
54 likely* that stratification in the upper few hundred meters of the ocean will increase significantly in the 21st
55 century. It is *virtually certain* that ocean pH is declining by ~0.02 pH units per decade where time-series
    observations exist (Section 5.2.2.3). The anthropogenic pH signal has already emerged over the entire

1   surface ocean (*high confidence*) and emission scenarios are the most important control of surface ocean pH
2   relative to internal variability for most of the 21st century at both global and local scale (*virtually certain*).
3   The oxygen content of the global ocean has declined by about 0.5 to 3.3% in 0-1000m meter layer (Section
4   5.2.2.4). Over the next century oxygen declines of 3.5% by 2100 are predicted by CMIP5 models globally
5   (*medium confidence*), with *low confidence* at regional scales, especially in the tropics. The largest changes in
6   the deep sea will occur after 2100 (Section 5.2.2.3). CMIP5 models project a decrease in global net primary
7   production (*medium confidence*) with increases in high latitude (*low confidence*) and decreases in low
8   latitude (*medium confidence*) (Section 5.2.2.6) in response to changes in ocean nutrient supply (Section
9   5.2.2.5). These models also project reductions by 8.9-15.8% in the globally integrated POC flux for RCP8.5,
10   with decreases in tropical regions and increases at higher latitudes *(medium confidence)*, affecting the
11   organic carbon supply to the deep-sea floor ecosystems (*high confidence*) (Section 5.2.2.6). However, there
12   is *low confidence* on the mechanistic understanding of how climatic drivers will affect the different
13   components of the biological pump in the epipelagic ocean (Table 5.4). Therefore, the exposure to hazard for
14   epipelagic ecosystems ranges from moderate (RCP2.6) to high (RCP8.5), with uncertain effects and
15   tolerance of planktonic organisms, fishes and large vertebrates to interactive climate stressors. Major risks
16   are predicted for declining productivity and fish biomass in tropical and subtropical waters (RCP8.5)
17   (SM5.2).
18
19   The climatic hazards for pelagic organisms from plankton to mammals are driving changes in eco-
20   physiology, biogeography and ecology and biodiversity (*high confidence*) (Section 5.2.3.1). Observed and
21   projected population declines in the equator-ward range boundary (*medium confidence*), expansion in the
22   poleward boundary (*high confidence*), earlier timing of biological events (*high confidence*), overall shift
23   species composition (*high confidence*) and decreases in animal biomass (*medium confidence*), are consistent
24   with expected responses to climate change (Section 5.2.3; Figure 5.13). It is *likely* that increased OA has not
25   yet caused sufficient reduction in fitness to decrease abundances of calcifying phytoplankton and
26   zooplankton, but *is very likely* (*high confidence*) that calcifying planktonic organisms will experience great
27   decreases in abundance and diversity under high emission scenarios by the end of the century. Therefore,
28   impacts to the epipelagic ecosystems are already observed in the present day (Figure 5.16). Based on
29   simulation modelling and experimental findings, the combined effects of warming, ocean deoxygenation,
30   OA and changes in net primary production in the 21st century are projected to exacerbate the impacts on the
31   growth, reproduction and mortality of fishes, and consequently increase the risk of population decline (*high
32   confidence*) (Section 5.2.3.1). There may be some capacity for adjustment and evolutionary adaptation that
33   lowers their sensitivity to warming and decrease in oxygen (*low confidence*). However, historical responses
34   in abundance and ranges of marine fishes to ocean warming and decrease in oxygen in the past suggest that
35   adaptation is not always sufficient to mitigate the observed impacts (*medium confidence*) (Section 5.2.3)
36   (SM5.2).
37
38   Despite its remoteness, most of the deep seafloor ecosystems already have or are projected to experience
39   rising temperatures and declining oxygen, pH and POC flux beyond natural variability within the next half
40   century (See Section 5.2.4). On slopes, seamounts and canyons these changes are projected to be much
41   larger under RCP8.5 than under RCP2.6 (*high confidence*), with greatest effects on seafloor community
42   diversity and function from expansion of low oxygen zones and aragonite undersaturation (*medium
43   confidence*). As critical thresholds of temperature, oxygen and CO_2 are exceeded, coral species will alter
44   their depth distributions, non-living carbonate will experience dissolution and bioerosion, and stress will be
45   exacerbated by lower food supply. These changes are projected to cause loss of cold water coral habitat with
46   highest climate hazard in the Arctic and North Atlantic Ocean (*medium confidence*), while sponges may be
47   more tolerant (Box 5.2) (*low confidence*). Projected changes in food supply to the seafloor at abyssal depths
48   combined with warmer temperatures are anticipated to cause reductions in biomass and body size (*medium
49   confidence*) that could affect the carbon cycle in this century under RCP 8.5 (*low confidence*). Even at
50   hydrothermal vents and methane seeps, some dominant species such as mussels may be vulnerable to
51   reduced photosynthetically-based food supply or have planktonic larvae or oxidizing symbionts that are
52   negatively affected by warming, acidification and oxygen loss (*low confidence*).
53
54   Widespread attributes of deep seafloor fauna (e.g., great longevity, high levels of habitat specialization
55   including well-defined physiological tolerances and thresholds, dependence on environmental triggers for
56   reproduction, and highly developed mutualistic interactions) can increase the vulnerability of selected taxa to
57   changing conditions (FAO, 2019) (*medium confidence*). However, some deep-sea taxa (e.g., foraminifera

1 and nematodes) may be more resilient to environmental change than their shallow-water counterparts (*low*
2 *confidence*). Observations, experiments and model projections indicate that impacts of climate change have
3 or are expected to take place in this century, indicating a transition from undetectable risk to moderate risk at
4 <1.5ºC for continental slope, canyon and seamount habitats, and for cold-water corals (Figure 5.16).
5 Emergence of risk is expected to occur later at around the mid-21st century under RCP8.5 for abyssal plain
6 and chemosynthetic ecosystems (vents and seeps) (Figure 5.16). All deep-seafloor ecosystems are expected
7 to be subject to at least moderate risk under RCP 8.5 by the end of the 21st century, with cold-water corals
8 experiencing a transition from moderate to high risk below 3ºC (SM5.2).
9
10
11 **5.3   Changing Coastal Ecosystems and Biodiversity**
12
13 The world's shelf seas and coastal waters (hereafter 'coastal seas') extend from the coastline to the 200 m
14 water depth contour. They encompass diverse ecosystems, including estuaries, sandy beaches, kelp forests,
15 mangroves and coral reefs. Although they occupy a small part of the global ocean (7.6%), coastal seas
16 provide up to 30% of global marine primary production and about 50% of the organic carbon supplied to the
17 deep ocean (Chen, 2003; Bauer et al., 2013) (Sections 5.2.4.1 and 5.4.1.1). Coastal seas include several
18 frontal and upwelling areas (Box 5.3) that support high fisheries yields (Scales et al., 2014), and productive
19 coastal ecosystems, such as wetlands (McLeod et al., 2011). Mangrove forests, seagrass meadows and kelp
20 forests form important habitats supporting high biodiversity while offering opportunities for climate change
21 mitigation and adaptation (Section 5.5.1.2) (Duarte et al., 2013), with mangrove forests providing physical
22 protection against extreme events such as storms and floods (Kelleway et al., 2017a) (Sections 5.4.1.2 and
23 4.3.3.5.4). The regional characteristics and habitat heterogeneity of many coastal seas support endemic fauna
24 and flora (e.g., seagrass meadows in the Mediterranean), which makes them particularly vulnerable to
25 climate change impacts with high risk of diversity loss and alterations in ecosystem structure and functioning
26 (Rilov, 2016; Chefaoui et al., 2018).
27
28 Near-shore coastal ecosystems are classified by their geomorphological structure (e.g., estuaries, sandy
29 beaches, and rocky shores) or foundation species (e.g., saltmarshes, seagrass meadows, mangrove forests,
30 coral reefs, and kelp forests). All these coastal ecosystems are threatened to a varying degree by sea level
31 rise (SLR), warming, acidification, deoxygenation and extreme weather events (Sections 5.3.1 to 5.3.7).
32 Unlike the open ocean where detection and attribution of climate driven-physical and chemical changes are
33 robust (Section 5.2.2), coastal ecosystems display regional complexity that can render the conclusive
34 detection and attribution of climate effects uncertain. The hydrological complexity of coastal ecosystems that
35 affects their biota is driven by the interactions between the land (e.g. river and groundwater discharges), the
36 sea (e.g. circulation, tides) (Section 5.2.2.2.3) and seabed structures and substrates (Sharples et al., 2017;
37 Chen et al., 2018; Laurent et al., 2018; Zahid et al., 2018).
38
39 Additionally, the high density of human populations on coastal land causes most of the adjacent marine
40 ecosystems to be impacted by local anthropogenic disturbances such as eutrophication, coastline
41 modifications, pollution and overfishing (Levin et al., 2015; Diop and Scheren, 2016; Maavara et al., 2017;
42 Dunn et al., 2018) (Section 4.3.2.2, Integrative Cross-Chapter Box 9). Climate driven impacts interact with
43 such human disturbances and pose a serious risk to ecosystems structure and functioning (Gattuso et al.,
44 2015). Projections of the ecological impacts of climate change in coastal ecosystems must therefore deal
45 with many emerging complexities such as the differentiation between the long-term climate trends (e.g.,
46 progressive ocean acidification) and the short-term natural fluctuations (Boyd et al., 2018), ranging from the
47 seasons to interannual climate oscillations like El Niño. The 'time of emergence' for specific climate drivers
48 to exceed background variability varies between ecosystems and is strongly sensitive to projected emission
49 scenarios (Hammond et al., 2017; Reusch et al., 2018) (Box 5.1).
50
51 This section summarizes our updated understanding of ecological and functional changes that coastal
52 ecosystems are experiencing due to multiple climate and non-climatic human drivers, and their synergies.
53 Additional experimental and long-term observational evidence since AR5 WGII (Wong et al., 2014a) and
54 SR1.5 (Hoegh-Guldberg et al., 2018) improves the attribution of impacts on all the types of coastal
55 ecosystems assessed here to climate trends (Sections 5.3.1 to 5.3.6). Moreover, the emergent impacts
56 detected in the present strengthen the projection of risk of each ecosystem under future emission scenarios
57 by 2100, depending on their exposure to different climate hazards (Section 5.3.7).

1
5.3.1   *Estuaries*
3
4 Estuarine ecosystems are defined by the river-sea interface that provides high habitat heterogeneity and
5 supports high biodiversity across freshwater and subtidal zones (Basset et al., 2013). AR5 WGII (Wong et
6 al., 2014a) and SR1.5 (Hoegh-Guldberg et al., 2018) concluded that estuarine ecosystems have been
7 impacted by sea level rise (SLR) and human influences that drive salinization, resulting in increased
8 flooding, land degradation and erosion of coastal areas around estuaries.
9
10 Observations since AR5 provide further evidence that SLR increases seawater intrusions and raises salinity
11 in estuaries. Salinization of estuaries can be exacerbated by droughts and modifications of drainage area by
12 human activities (Ross et al., 2015; Cardoso-Mohedano et al., 2018; Hallett et al., 2018; Zahid et al., 2018).
13 The changing salinity gradients in estuaries have been linked to the observed upstream expansion of brackish
14 and marine benthic and pelagic communities, and a reduction in the diversity and richness of freshwater
15 fauna (Robins et al., 2016b; Raimonet and Cloern, 2017; Hallett et al., 2018; Addino et al., 2019) (*medium
16 confidence*). However, because the distribution of benthic species in estuaries is strongly determined by
17 sediment properties like grain size, the gradient of sediment types in estuaries can be a barrier to upstream
18 shifts of brackish and marine benthic biota, leading to a reduction in species richness in mid- to upper-
19 estuarine areas and altering food webs (Little et al., 2017; Hudson et al., 2018; Addino et al., 2019).
20 Similarly, estuarine wetlands (Section 5.3.2) have reduced their extent and productivity in response to
21 increased salinity, inundation and wave exposure, especially in areas with limited capacity for soil accretion
22 or inland migration due to coastal squeezing (Sections 4.3.2.3, 5.3.2) (*high confidence*). Poleward migration
23 of tropical and sub-tropical biota between estuaries has been observed in response to warming (Hallett et al.,
24 2018) (*medium confidence*), in agreement with the global trend of biogeographic shifts of marine organisms
25 (Sections 5.2.3.1.1; 5.3.2-5.3.6).
26
27 Intensive human activities around estuaries and river deltas worldwide has substantially increased nutrient
28 and organic matter inputs into such systems since the 1970s (Maavara et al., 2017). Increased organic matter
29 accumulation has been shown to interact with warming, resulting in intensification of bacterial degradation
30 and eutrophication (Maavara et al., 2017; Chen et al., 2018; Fennel and Testa, 2019), contributing to an
31 increase in the frequency and extent of hypoxic zones (Breitberg et al., 2015; Gobler and Baumann, 2016).
32 The interaction between warming, increased nutrient loading, and hypoxia has shown to be related to the
33 increased occurrences of harmful algal blooms (Anderson et al., 2015; Paerl et al., 2018b) (Box 5.4) (*high
34 confidence*), pathogenic bacteria such as *Vibrio* species (Baker-Austin et al., 2017; Kopprio et al., 2017)
35 (Section 5.4.2) (*low confidence*), and mortalities of invertebrates and fish communities (Jeppesen et al.,
36 2018; Warwick et al., 2018) (*medium confidence*).
37
38 Fluctuations in estuarine salinity, turbidity and nutrient gradients are influenced by changes in precipitation
39 and wind-stress caused by large-scale climatic variations such as the El Niño/Southern Oscillation (ENSO),
40 the North Atlantic Oscillation (NAO) and the *South Atlantic Meridional Overturning Circulation (SAMOC)*
41 which have shown persistent anomalies associated with climate change since the 1970s (Wang and Cai,
42 2013; Delworth and Zeng, 2016; García-Moreiras et al., 2018). Similarly, storm surges and heat waves have
43 increased nutrients and sediment loads in estuaries (Tweedley et al., 2016; Arias-Ortiz et al., 2018; Chen et
44 al., 2018). Sustained long-term observations (15-40 years) provide evidence that large-scale climatic
45 variations and extreme events affect plankton phenology and composition in estuaries worldwide with
46 regional differences in the characteristics of the responses (Thompson et al., 2015; Abreu et al., 2017;
47 Marques et al., 2017; Arias-Ortiz et al., 2018; López-Abbate et al., 2019) (*high confidence*). Although these
48 changes in ecosystem components may be attributed to climate variability (Box 5.1), they demonstrate the
49 sensitivity of estuarine ecosystems to climate change. Also, these large-scale climate events are *likely* to be
50 intensified in the 21st century (Stocker, 2014) (Section 6.5.1).
51
52 Salinization in estuaries is projected to continue in response to SLR, warming and droughts under global
53 warming greater than 1.5°C (*high confidence*), and will pose further risks to ecosystems biodiversity and
54 functioning (Zhou et al., 2017; Hallett et al., 2018; Zahid et al., 2018; Elliott et al., 2019) (Section 4.3.3.4,
55 Cross-Chapter Box 7) (*medium confidence*). Estuarine wetlands are resilient to modest rates of SLR due to
56 their sediment relocation capacity, but such adaptation is not expected to keep pace with projected rates of
57 SLR under the RCP8.5 climate scenario (Section 5.3.2) (*high confidence*). Moreover, human activities that

1 inhibit sediment movement and deposition in coastal deltas increase the likelihood of their shrinking as a
2 result of SLR (Brown et al., 2018b; Schuerch et al., 2018b) (*medium confidence*).
3
4 Oxygen-depleted dead zones in coastal areas are already a problem; they are projected to increase under the
5 co-occurrence and intensification of climate threats and eutrophication (Breitburg et al., 2018; Laurent et al.,
6 2018) (Section 5.2.2.4). While warming is the primary climate driver of deoxygenation in the open ocean,
7 eutrophication is projected to increase in estuaries due to human activities and intensified precipitation
8 increasing riverine nitrogen loads under both RCP2.6 and RCP8.5 scenarios, both mid-century (2031–2060)
9 and later (2071–2100) (Sinha et al., 2017). Moreover, enhanced stratification in estuaries in response to
10 warming is also expected to increase the risk of hypoxia through reduced vertical mixing (Du et al., 2018;
11 Hallett et al., 2018; Warwick et al., 2018).The effects of warming will be more pronounced on high latitude
12 and temperate shallow estuaries with limited exchange with the open ocean (e.g., Río de La Plata Estuary,
13 Baltic Sea and Chesapeake Bay) and seasonality that already leads to dead zone development when
14 summertime temperatures reach critical values (e.g., Black Sea) (Altieri and Gedan, 2015) (*medium
15 confidence*). The coastal acidification related to this expansion of hypoxic zones (Zhang and Gao, 2016; Cai
16 et al., 2017; Laurent et al., 2017) imposes risk for sensitive organisms (Beck et al., 2011; Duarte et al., 2013;
17 Feely et al., 2016; Carstensen et al., 2018).
18
19 The interaction of SLR and changes in precipitation will have a more severe impact on shallow estuaries
20 (<10 m) than on deep basin estuaries (>10 m) (Hallett et al., 2018; Elliott et al., 2019) (*medium confidence*).
21 For a projected SLR of 1 m, climate-related risks for shallow estuaries ecosystems are estimated to increase
22 through increased tidal current amplitudes (by 5% on average), energy dissipation, vertical mixing and
23 salinity intrusion (Prandle and Lane, 2015). Estuaries with high tidal exchanges and associated well-
24 developed sediment areas are more resilient to global climate changes than estuaries with low tidal
25 exchanges and sediment supply, since the latter are more vulnerable to SLR and changes in river flow
26 (Brown et al., 2018b; Warwick et al., 2018) (*medium confidence*).
27
28 Overall, this assessment concludes that there is evidence of upstream redistribution of marine biotic
29 communities in estuaries driven by increased sea water intrusion (*medium confidence*). Such distribution
30 shifts are limited by physical barriers such as the availability of benthic substrates leading to reduction of
31 suitable habitats for estuarine communities (*medium confidence*). Warming has led to poleward range shifts
32 of biota between estuaries (*medium confidence*). Increased nutrient inputs from intensive human
33 development in deltas increases bacterial respiration, which in turn is exacerbated by warming, leading to an
34 expansion of suboxic and anoxic areas (*high confidence*). These changes reduce the survival of estuarine
35 animals (*medium confidence*), and increase the occurrence of harmful algal blooms (HABs) and pathogenic
36 microbes (*medium confidence*). Projected warming, SLR and tidal changes in the 21st century will continue
37 to expand salinization and hypoxia in estuaries (*medium confidence*). These impacts will be more
38 pronounced under higher emission scenarios, and in temperate and high latitude estuaries that are eutrophic,
39 shallow and that naturally have low sediment supply.
40
41 ### 5.3.2   *Coastal Wetlands (Saltmarshes, Seagrass Meadows and Mangrove Forests)*
42
43 Coastal vegetated wetlands include saltmarshes, mangrove forests and subtidal seagrass meadows
44 ecosystems, considered to be the main 'blue carbon' habitats (Sections 5.4.1 and 5.5.1.1) (McLeod et al.,
45 2011). IPCC AR5 WGII and SR1.5 concluded that wetland salinization is occurring at a large geographic
46 scale (*high confidence*); that rising water temperatures has led to shifts in plant species distribution (*medium
47 confidence*)(Wong et al., 2014b); and that SLR and storms are causing wetland erosion and habitat loss,
48 enhanced by human disturbances (*high confidence*) (Section 4.3.3.5.1) (Wong et al., 2014b). This section
49 assesses new evidence since AR5 and SR1.5 of observed climate impacts and future risks of these vegetated
50 wetlands in terms of their role in supporting biodiversity and key ecosystem functions. The recent literature
51 confirms and strengthens the SR1.5 conclusions (Section 5.3.7 and Figure 5.16).
52
53 Nearly 50% of the pre-industrial, natural extent of global coastal wetlands have been lost since the 19th
54 century (Li et al., 2018a). Such a reduction in wetlands is primarily caused by non-climatic drivers such as
55 alteration of drainage, agriculture development, coastal settlement, hydrological alterations and reductions in
56 sediment supply (Adam, 2002; Wang et al., 2014; Kroeger et al., 2017; Thomas et al., 2017; Li et al.,
57 2018a). However, large-scale mortality events of mangroves from 'natural causes' has also occurred globally

1   since the 1960s; ~70% of this loss has resulted from low frequency, high intensity weather events, such as
2   tropical cyclones (45%) and climatic extremes such as droughts, SLR variations and heat waves (Sippo et
3   al., 2018) (*high confidence*). In Australia, the mangrove loss due to heat waves accounted for 22% of global
4   mangrove forests (Sippo et al., 2018), with negative impacts on ecosystem biodiversity and the provisioning
5   of services (Carugati et al., 2018; Saintilan et al., 2018) (Section 5.4). In coastal areas with sufficient
6   sediment supply across the Indo-Pacific region, inland expansion of mangroves is occurring as a result of
7   vertical accretion and root growth, allowing them to keep pace with current SLR (Lovelock et al., 2015). In
8   seagrass meadows, temperature is the main limiting range factor, and over the past decades there have been
9   several global die-off events (Hoegh-Guldberg et al., 2018). The vulnerability of seagrasses to warming
10   varies locally depending on soil accretion and herbivory (El-Hacen et al., 2018; Marbà et al., 2018; Vergés et
11   al., 2018) and on the population assemblages (e.g. expansion at high latitudes) (Beca-Carretero et al., 2018;
12   Duarte et al., 2018). The compounding effects of heat waves, hypersaline conditions and increased turbidity
13   and nutrient levels associated with floods have been shown to cause negative changes in the composition and
14   biomass of co-occurring seagrass species (Nowicki et al., 2017; Arias-Ortiz et al., 2018; Lin et al., 2018)
15   (*high confidence*). For example, in Shark Bay, Western Australia, a marine heat wave in austral summer
16   2010/2011 caused widespread losses (36 % of area) of seagrass meadows, with negative implications for
17   carbon storage (Arias-Ortiz et al., 2018). The poleward expansion of tropical mangroves into subtropical
18   saltmarshes as a result of increase in temperature has been also observed over the past half century on five
19   continents (Saintilan et al., 2014; Saintilan et al., 2018) (*high confidence*); for example, in the Texas Gulf
20   Coast (Armitage et al., 2015). The loss of open areas with herbaceous plants (saltmarshes) reduces food and
21   habitat availability for resident and migratory animals (Kelleway et al., 2017a; Lin et al., 2018) (Section
22   5.4.1.2).
23
24   The ability of salt marshes to increase their elevation and withstand erosion under SLR depends on the
25   development of new soil by the external supply of mineral sediments and organic accretion by local biota
26   (Section 5.4.1, Figure 5.19) (Bouma et al., 2016). In some places, critical organic accretion rates are
27   declining due to reduced plant productivity from stress by more frequent inundation, and increased plant and
28   microbial respiration rates as a result of warming; consequently, the elevation of marshes from soil accretion
29   is slower than the rate of rising sea level, resulting in reduction of saltmarsh area (Carey et al., 2017; Watson
30   et al., 2017b). Vegetation loss rates were significantly negatively correlated with marsh elevation, suggesting
31   inundation due to SLR since 1970 as the main driver, enhanced by storms and increased tidal range in back
32   barrier marshes (Watson et al., 2017b). Plant species that are more sensitive to higher temperatures and
33   increases in saltwater intrusion were found to be less abundant and in some cases replaced by salinity-
34   tolerant species (Janousek et al., 2017; Piovan et al., 2019). Plant community restructuring has resulted in
35   biodiversity loss (Pratolongo et al., 2013; Raposa et al., 2017) and reduced above- and below-ground
36   productivity (McLeod et al., 2011; Watson et al., 2017b). As a result of tidal flooding, saltmarsh soils do not
37   dry out and high levels of carbon can accumulate under anaerobic conditions. This is coupled with generally
38   low rates of methane emission which is strongly limited in saline marshes (Poffenbarger et al., 2011; Martin
39   and Moseman-Valtierra, 2015; Kroeger et al., 2017; Tong et al., 2018) (*high confidence*).
40
41   Non-climatic human pressures on wetland ecosystems, including overfishing (Crotty et al., 2017),
42   eutrophication (Legault II et al., 2018), and invasive species (Zhang et al., 2016), interact with climate
43   change drivers and affect wetlands composition and structure, with the impacts varying between regions and
44   species (Tomas et al., 2015; O'Brien et al., 2017; Pagès et al., 2017; York et al., 2017). The intensity of
45   herbivory on seagrasses is expected to increase with global warming, particularly in temperate areas, because
46   of the migration of tropical herbivores into temperate seagrass meadows (Hyndes et al., 2016; Vergés et al.,
47   2018) (*medium confidence*, Section 5.2.3.1.1). Warming also reduces the fitness of seedlings by increasing
48   necrosis and susceptibility to consumers and pathogenic pressure while reducing establishment potential and
49   nutritional (Olsen et al., 2016b; Hernán et al., 2017). Because herbivores play a key role in modulating the
50   biomass of plant communities, their more intense activity affects the provision of services in these
51   ecosystems (Scott et al., 2018) (Section 5.4).
52
53   Globally, between 20-90% of existing coastal wetland area is projected to be lost by 2100 (Blankespoor et
54   al., 2014; Crosby et al., 2016; Spencer et al., 2016), depending on different SLR projections under future
55   emission scenarios. These projected changes vary regionally and between different types of wetlands.
56   Gaining area may be possible, at least locally, if vertical sediment accretion occurs together with lateral re-
57   accommodation (Brown et al., 2018b; Schuerch et al., 2018b) (Section 4.3.3.5.1). Local losses may also be

1   higher; for example, in New England, where regional rates of SLR have been as much as 50% greater than
2   the global average (from 1 to 5.83 mm yr^{-1}; 1979-2015) (Watson et al., 2017a) and where projections suggest
3   that 40%–95% of saltmarshes will be submerged by the end of this century (Valiela et al., 2018). In some
4   species of seagrasses, enhanced temperature-driven flowering (Ruiz-Frau et al., 2017) and greater biomass
5   production in response to elevated CO_2 (Campbell and Fourqurean, 2018) may increase resilience to
6   warming. Nevertheless, severe habitat loss (70%) of endemic species such as *Posidonia oceanica* is
7   projected by 2050 with the potential for functional extinction by 2100 under RCP8.5 climate scenario. For
8   *Cymodosea nodosa*, the species with the highest thermal optima (Savva et al., 2018), warming is expected to
9   lead to significant reduction of meadows (46% under RCP8.5) in the Mediterranean, although potentially
10 compensated in part by future expansion into the Atlantic (Chefaoui et al., 2018).
11
12 The mangrove habitats of small islands, with lack of rivers, steep topography, sediment-starved areas,
13 groundwater extraction and coastal development, are particularly vulnerable to SLR. Although mangrove
14 ecosystems may survive the increased storm intensity and sea levels projected until 2100 under RCP2.6
15 (Ward et al., 2016), for RCP 8.5 they are only resilient up to 2050 conditions (Sasmito et al., 2016). Negative
16 climate impacts will be exacerbated in cases where anthropogenic barriers cause further 'coastal squeeze'
17 that prevents inland movement of plants and limits relocation of sediment (*medium confidence*) (Enwright et
18 al., 2016; Borchert et al., 2018).
19
20 In conclusion, substantial evidence supports with *high confidence* that warming and salinization of wetlands
21 caused by SLR are causing shifts in the distribution of plant species inland and poleward, such as mangrove
22 encroachment into subtropical saltmarshes (*high confidence*) or seagrass meadows contraction at low
23 latitudes (*high confidence*). Plants with low tolerance to flooding and extreme temperatures are particularly
24 vulnerable and may be locally extirpated (*medium confidence*). The flooded area of saltmarshes can become
25 a mudflat or be colonized by more tolerant, invasive species, whose expansion is favoured by combined
26 effects of warming, rising CO_2 and nutrient enrichment (*medium confidence*). The loss of vegetated coastal
27 ecosystems causes a reduction in carbon storage with positive feedbacks to the climate system (*high
28 confidence*) (Section 5.4.1.2). SLR and warming are expected to continue to reduce the area of coastal
29 wetlands, with a projected global loss of 20-90% by the end of the century depending on emission scenarios.
30 High risk of total local loss is projected under the RCP8.5 emission scenario by 2100 (*medium confidence*),
31 especially if landward migration and sediment supply is constrained by human modification of shorelines
32 and river flows (*medium confidence*).
33
34 ### 5.3.3   Sandy Beaches
35
36 Sandy beaches represent 31% of the world's ice-free shoreline (Luijendijk et al., 2018). They provide
37 habitat for dune vegetation, benthic fauna and sea birds; also nesting area for marine turtles (Defeo et al.,
38 2009), as well as several key ecosystem services (Drius et al., 2019) (Section 5.4.1.2). Sandy beach
39 ecosystems are physically dynamic, where sediment movement is a key driver of benthic flora and fauna
40 zonation (Schlacher and Thompson, 2013; van Puijenbroek et al., 2017). In AR5 WGII (Wong et al., 2014b)
41 and SR1.5 (Hoegh-Guldberg et al., 2018), climate impacts on sandy beach ecosystems were not assessed
42 individually but together with other coastal systems that included beaches, barriers, sand dunes, rocky coasts,
43 aquifers and lagoons. Those assessments concluded with *high confidence* that SLR, storminess, wave
44 energy, and weathering regimes will continue to erode coastal shorelines and affect the soil accretion and
45 land-based ecosystems, with highly site-specific effects (*high confidence*). Infrastructure and geological
46 constraints reduce shoreline movement and cause coastal squeeze (*high confidence*). Assessment in Section
47 4.3.3.3 supports the conclusions in AR5 and SR1.5 regarding the erosion of sandy coastlines. This section
48 specifically assesses the combined climate and non-climatic impacts on sandy beach biodiversity, ecosystem
49 structure and functioning.
50
51 Worldwide, sandy beaches show vegetation transformations caused by erosion following locally severe
52 wave events (Castelle et al., 2017; Delgado-Fernandez et al., 2019; Zinnert et al., 2019) (Table SM5.7). The
53 original dense vegetation is replaced by sparser vegetation (Zinnert et al., 2019) and has a generally slow
54 recovery (multiple years to decades) (Castelle et al., 2017). In some instances, the changes persist over
55 decades, resulting in a regime shift in the beach morphology (Kuriyama and Yanagishima, 2018). Such
56 changes in vegetation and beach morphology in response to local disturbances were also related to shifts in
57 the associated fauna composition (Carcedo et al., 2017; Delgado-Fernandez et al., 2019). Direct attribution

1   of these observed events to climate change is not available despite early evidence (since the 1970) and an
2   emerging literature (Section 4.3.3.1) (Table SM5.7).
3
4   Sandy beaches show similar patterns of biogeographical shifts following warming, with increased
5   dominance of species more tolerant to higher temperatures, as observed in other ocean ecosystems (5.2.3.1.1,
6   Table SM5.7). Examples of these observed shifts in abundance and distribution of benthic fauna in sandy
7   beaches are found in the Pacific and Atlantic coasts of North and South America, and in Australia, including
8   increased mortality of clam populations close to their upper temperature limits with low population recovery
9   (Orlando et al., 2019), and poleward expansion of crabs since the 1980s that were related to warming
10   (Schoeman et al., 2015) (Table SM5.7). Also, mass mortalities of beach clams have occurred during warm
11   phases of El Niño events (Orlando et al., 2019)(and Table SM5.7), parasite infestations on dense
12   populations (Vázquez et al., 2016) and high wave exposure (Turra et al., 2016).
13
14   Human disturbances have caused coastal squeeze and morphological changes in sandy beaches (Martínez et
15   al., 2017; Rêgo et al., 2018; Delgado-Fernandez et al., 2019). Along with SLR and climate-driven
16   intensification of waves and offshore winds, these hazards have increased erosion rates suggesting a reduced
17   resilience due to insufficient sediment supply and accretion capacity (Castelle et al., 2017; Houser et al.,
18   2018; Kuriyama and Yanagishima, 2018). Narrow sandy beaches such as those in south California (Vitousek
19   et al., 2017) or central Chile (Martínez et al., 2017) are particularly vulnerable to climate hazards when
20   combined with human disturbances and where landward retreat of beach profile and benthic organisms is
21   constrained due to increasing urbanization (Hubbard et al., 2014) (Section 4.3.2.3).
22
23   Notwithstanding the uncertainty in projecting future interactions of SLR with other natural and human
24   impacts on sandy shorelines (Le Cozannet et al., 2019; Orlando et al., 2019), they are expected to continue to
25   reduce their area and change their topography due to SLR and increased extreme climatic erosive events.
26   This will be especially important in low-lying coastal areas with high population and building densities
27   (*medium confidence*, SM 4.2). Megafauna that use sandy beaches during vulnerable parts of their life cycles
28   could be particularly impacted (Laloë et al., 2017). For example, the modelled incubation temperatures of
29   green turtles have increased by 1°C since the mid-1970s, resulting in an average 20% increase in the
30   proportion of female hatchlings over this period (Patrício et al., 2019). By 2100, global temperatures will
31   approach lethal levels for incubation in existing nesting sites, and hatchling success is expected to drop to
32   32% under RCP8.5 scenario, with 93% of the hatchlings expected to be female (76% under RCP4.5). A
33   possible microhabitat adaptation such as shadowed vegetated areas, however, could allow for continued
34   male production throughout the 21st century (Patrício et al., 2019). In addition, a projected global mean SLR
35   of ~1.2 m under the upper likely range of RCP8.5 by 2100 implies a loss of 59% and 67% in the present
36   nesting area of the green turtle and the loggerhead respectively in the Mediterranean (Varela et al., 2019),
37   and a loss of 43% in the nesting area of green turtles in West Africa (Patrício et al., 2019). Moreover, benthic
38   crustaceans of sandy beaches, including isopods, crabs and amphipods, generally follow the temperature-
39   body size gradient in which body size decreases towards warmer lower-latitude regions (Jaramillo et al.,
40   2017). Assuming that the physiological underpinning of the relationship between body size and temperature
41   can be applied to warming (see 5.2.2, *medium confidence*), the body size of sandy beach crustaceans is
42   expected to decrease under warming (*low evidence, medium agreement*).
43
44   Overall, changes in sandy beach morphology have been observed from climate related events, such as storm
45   surges, intensified offshore winds, and from coastal degradation caused by humans (*high confidence*), with
46   impacts on beach habitats (e.g. benthic megafauna) (*medium confidence*). The direct influence of
47   contemporary SLR on shoreline behaviour is emerging, but attribution of such changes to SLR remains
48   difficult (Section 4.3.3.1). Projected changes in mean and extreme sea levels (Section 4.2.3) and warming
49   (Section 5.2.1) under RCP8.5 are expected to result in high risk of impacts on sandy beach ecosystems by
50   the end of the 21st century (*medium confidence*, Figure 5.16), taking account of the slow recovery rate of
51   sandy beach vegetation, the direct loss of habitats and the high climatic sensitivity of some fauna. Under
52   RCP2.6, the risk of impacts on sandy beaches is expected to be only slightly higher than the present-day
53   level (*low confidence*, Figure 5.16). However, pervasive coastal urbanization lowers the buffering capacity
54   and recovery potential of sandy beach ecosystems to impacts from SLR and warming and thus is expected to
55   limit their resilience to climate change (*high confidence*).
56

1 ### 5.3.4   Coral Reefs
2
3 Human activities and warming have already led to major impacts on shallow water tropical coral reefs
4 caused by species replacement, bleaching, and decreased coral cover while warming, ocean acidification and
5 climate hazards will put warm-water corals at very high risk even if global warming can be limited to 1.5°C
6 above pre-industrial level (Hoegh-Guldberg et al., 2018; Kubicek et al., 2019; Sully et al., 2019). While
7 providing new evidence to support these previous assessments (Kleypas, 2019), this assessment focuses on
8 evaluating the variations in sensitivities and responses of coral reefs and their associated biota to highlight
9 comparative risks and resiliences.
10
11 New evidence since AR5 and SR1.5 confirms the impacts of ocean warming (Kao et al., 2018; Jury and
12 Toonen, 2019) and acidification (Jiang et al., 2018; Mollica et al., 2018; Bove et al., 2019) on coral reefs
13 (*high confidence*), enhancing reef dissolution and bioerosion (*high confidence*), affecting coral species
14 distribution, and leading to community changes (Agostini et al., 2018) (*high confidence*). The rate of sea
15 level rise (primarily noticed in small reef islands) may outpace the growth of reefs to keep up although there
16 is *low agreement* in the literature (Brown et al., 2011; Perry et al., 2018) (*low confidence*). Reefs are further
17 exposed to other increased impacts, such as enhanced storm intensity (Lavender et al., 2018), turbidity and
18 increased runoff from the land (Kleypas, 2019) (*high confidence*). Recovery of coral reefs resulting from
19 repeated disturbance events is slow (Hughes et al., 2019a; Ingeman et al., 2019) (*high confidence*). Only few
20 coral reef areas show some resilience to global change drivers (Fine et al., 2019) (*low confidence*).
21
22 Globally, coral reefs and their associated communities are projected to change their species composition and
23 biodiversity as a result of future interactions of multiple climatic and non-climatic hazards (Kleypas, 2019;
24 Kubicek et al., 2019; Rinkevich, 2019) (*high evidence, very high agreement, very high confidence*). Multiple
25 stressors act together to increase the risk of population declines or local extinction of reef-associated species
26 through impacts of warming and ocean acidification on physiology and behaviours (Gunderson et al., 2017)
27 (*high confidence*). Alteration of composition of coral reef-associated biota is exacerbated by changes in
28 habitat conditions through increased sedimentation and nutrient concentrations from human coastal activities
29 (Fabricius, 2005) (*high confidence*). Coral ecosystems in tropical small islands are also at high risk of being
30 affected by extreme events, including storms, with their impacts exacerbated by sea level rise (Duvat et al.,
31 2017; Harborne et al., 2017) (*high confidence*). Such risks on coral reef-associated communities are
32 substantially elevated when the level of these climatic and non-climatic hazards are above thresholds that
33 may cause phase shifts in reef communities (McCook, 1999; Hughes et al., 2010; Graham et al., 2013;
34 Hughes et al., 2018) (*high confidence*). A phase shift is characterized by an abrupt decrease in coral
35 abundance or cover, with concurrent increase in the dominance of non-reef-building organisms, such as
36 algae and soft corals (Kleypas, 2019). Such phase shifts have already been observed in many coral reefs
37 worldwide (Wernberg et al., 2016; Kleypas, 2019).
38
39 Notwithstanding the conclusion that coral reefs globally are projected to greatly decline at 2°C warming
40 relative to pre-industrial level (Cacciapaglia and van Woesik, 2018; Dietz et al., 2018; Hoegh-Guldberg et
41 al., 2018), climate impacts can be affected by variations in the sensitivity and adaptive capacity across coral
42 species and coral reef ecosystems. Laboratory experiments show that some warm water corals possess the
43 cellular, physiological or molecular machineries that could help them acclimatize or adapt to the effects of
44 global change (*medium confidence*) (DeBiasse and Kelly, 2016; Gibbin et al., 2017; Wall et al., 2017; Camp
45 et al., 2018; Donelson et al., 2018; Drake et al., 2018; Veilleux and Donelson, 2018; Hughes et al., 2019b).
46 For example, there are species or genotypes that show less impacts by either ocean acidification or increased
47 temperatures (Cornwall et al., 2018; Gintert et al., 2018). Some corals and their symbionts might be able to
48 use epigenetic (heritable phenotype changes that do not involve alterations in the DNA sequences)
49 mechanisms to reduce their sensitivity to temperature changes in their environment and to pass such traits to
50 their offspring (Liew et al., 2017; Torda et al., 2017; Li et al., 2018b; Liew et al., 2018). The variations in
51 sensitivity and adaptive capacity of coral species to warming and ocean acidification contribute to changes in
52 species composition of coral reefs as they are exposed to climatic and non-climatic hazards (Ingeman et al.,
53 2019; Kleypas, 2019; Kubicek et al., 2019) (*high confidence*). However, it has not yet been established
54 whether coral and coral-associated biota adaptation may hold beyond 1.5°C warming. The onset of coral
55 bleaching in the last decade has occurred at higher sea surface temperatures (~0.5°C) than in the previous
56 decade, suggesting that coral populations that remain after preceding bleaching events may have a higher
57 thermal threshold (Sully et al., 2019) (*medium confidence*), potentially as a result of the increased dominance

1 of species with lower sensitivity or higher adaptive capacity (Schulz et al., 2013; McClanahan et al., 2014;
2 Mumby and van Woesik, 2014; Pandolfi, 2015; Folkersen, 2018) (*medium confidence*).
3
4 Coral reefs in deeper or mesophotic waters (found in tropical/subtropical regions at 30-150 m depth) may
5 serve as refuges and sources for larval supply to those reefs exposed to disturbances (e.g., bleaching, storms,
6 floods from land, sedimentation, tourism impacts) (Bridge et al., 2013; Thomas et al., 2015; Lindfield et al.,
7 2016; Smith et al., 2016b; Bongaerts et al., 2017). Reefs exposed to local oceanographic characteristics that
8 reduce warming, such as upwelling, may similarly provide refuges and larval sources (Tkachenko and
9 Soong, 2017). However, recent evidence suggests that mesophotic coral reefs are at higher risk than
10 previously indicated (Rocha et al., 2018). Monitoring of coral reefs worldwide shows that some areas in the
11 eastern tropical Pacific Ocean (Smith et al., 2017), the Caribbean (Chollett and Mumby, 2013), the Red Sea
12 (Fine et al., 2013; Osman et al., 2017), the Persian Gulf (Coles and Riegl, 2013) and the Great Barrier Reef,
13 Australia (Hughes et al., 2010; Morgan et al., 2017) have recovered more rapidly after bleaching than the
14 larger-scale average (*medium confidence*). There are regional differences in reef vulnerability when
15 considering scales larger than 100 km or over latitudinal gradients (van Hooidonk et al., 2013; Heron et al.,
16 2016; Langlais et al., 2017; McClenachan et al., 2017) (*high confidence*).
17
18 Based on findings from simulation modelling, SR1.5 concluded that "coral reefs are projected to decline by a
19 further 70–90% at 1.5°C (very *high confidence*) with larger losses (>99%) at 2°C (*very high confidence*)".
20 The variations in exposure, sensitivity and adaptive capacity between coral populations and regions are
21 further projected to cause large changes in the composition and structure of the remaining coral reefs, with
22 large regional differences (van Hooidonk et al., 2016; Hoegh-Guldberg et al., 2018; Kleypas, 2019; Kubicek
23 et al., 2019; Sully et al., 2019).
24
25 ### *5.3.5 Rocky Shores*
26
27 Rocky shore ecosystems span the intertidal and shallow subtidal zones of the world's temperate coasts and
28 are typically dominated by calcareous mussels or seaweeds (macro-algae). Other organisms that inhabit
29 rocky shores are coralline algae (i.e., maerl beds), polychaetes, molluscs, bryozoans and sponges. Intertidal
30 habitats are characterized by strong environmental gradients, and are exposed to marine and atmospheric
31 climate regimes (Hawkins et al., 2016). IPCC AR5 (Wong et al., 2014a) concluded that rocky shores are
32 among the better-understood coastal ecosystems in terms of potential impacts of climate variability and
33 change. The high sensitivity of sessile organisms (e.g. barnacles, mussels) to extreme temperature events
34 (e.g. mass mortality and drastic biodiversity loss of mussels beds), and to acidification (widely observed in
35 manipulative experiments) gives *high confidence* that rocky shore species are at high risk of changes in
36 distribution and abundance from these two drivers. SR1.5 (Hoegh-Guldberg et al., 2018) concluded that
37 rocky coasts are already experiencing large-scale changes, and critical thresholds are expected to be reached
38 at warming of 1.5°C and above (*high confidence*).
39
40 More observational and empirical evidence since AR5 and SR1.5 confirms that climate change poses high
41 risk to rocky shore ecosystems' biodiversity, structure and functioning through warming, acidification, SLR
42 and extreme events (Agostini et al., 2018; Duarte and Krause-Jensen, 2018; Ullah et al., 2018; Milazzo et al.,
43 2019). Immobile intertidal organisms are especially vulnerable to warming, due to the potential for extreme
44 heat exposure during low tide emersion and prolonged desiccation events (Hawkins et al., 2016; Zamir et al.,
45 2018) (*high confidence*). This effect is expected to lower the upper vertical limit of intertidal communities
46 (Hawkins et al., 2016), reducing their suitable habitat (Harley, 2011), and accompanied by temperature-
47 induced increases in predation by consumers (Sanford, 1999). While previous studies have documented a
48 poleward shift in species distributions of rocky intertidal and reef algae (Duarte et al., 2013; Nicastro et al.,
49 2013) and faunal species (Barry et al., 1995; Mieszkowska et al., 2006; Lima et al., 2007), local extinctions
50 at the equatorial or warm edge of species ranges are increasingly being attributed to climate change
51 (Yeruham et al., 2015; Sorte et al., 2017) (*high confidence*). Extreme heat waves are expected to cause
52 mortality among rocky shore species (Gazeau et al., 2014; Jurgens et al., 2015) and subsequent declines or
53 losses in important species can have cascading effects on the whole intertidal community and the services it
54 provides (Gatti et al., 2017; Sorte et al., 2017; Sunday et al., 2017). Coralline fauna adapted to narrow
55 environmental conditions seem especially vulnerable to heat waves, with observed mass mortalities in the
56 Adriatic Sea in response to extreme summer temperatures (Kružić et al., 2016). The loss of thermal refugia
57 associated with continued warming could exacerbate the impacts of heat stress on rocky intertidal

1   communities (Lima et al., 2016). Nevertheless, experimental data indicate that some coralline algae that are
2   well adapted to highly variable transitional environments can tolerate the warming projected for 2100 under
3   RCP8.5; for these species, ocean acidification will constitute the main hazard (Nannini et al., 2015).
4
5   Ocean acidification is expected to decrease the net calcification (*high confidence*) and abundance (*medium*
6   *confidence*) of rocky intertidal and reef-associated species (Kroeker et al., 2013), and the dissolution of
7   calcareous species has already been documented in tide-pool communities (Kwiatkowski et al., 2016; Duarte
8   and Krause-Jensen, 2018). Recent experimental and field studies, however, have demonstrated the
9   importance of food resources in mediating the effects of ocean acidification on vulnerable rocky shores
10   species (Ciais et al., 2013; Ramajo et al., 2016), suggesting that species' vulnerability to ocean acidification
11   may be most pronounced in areas of high heat stress and low food availability (*medium confidence*) (Kroeker
12   et al., 2017). There is increasing evidence that the interactions between multiple climate drivers will
13   determine species vulnerability and the ecosystem impacts of climate change (Hewitt et al., 2016).
14
15   Studies on naturally-acidified rocky reef ecosystems suggest ocean acidification will simplify rocky shore
16   ecosystems, due to an overgrowth by macroalgae, a reduction in biodiversity, and a reduction in the
17   abundance of calcareous species (*medium confidence*) (Kroeker et al., 2013; Linares et al., 2015). These
18   shifts in community structure and function have been observed in CO_2 seep communities (Hall-Spencer et
19   al., 2008), already exposed to levels of pCO_2 expected to generally occur by the end of the century (Agostini
20   et al., 2018). Reductions in the abundance of calcareous herbivores that can create space for rarer species by
21   grazing the dominant algae, are expected to contribute to the overgrowth of fleshy macroalgae on rocky
22   shores (Baggini et al., 2015). This shift towards macroalgae is associated with a simplification of the food
23   web at lower trophic levels (Kroeker et al., 2011).
24
25   At the local scale, warming and ocean acidification are expected to change energy flows within rocky shores
26   ecosystems (*medium confidence*). Experiments indicate that both climate drivers may boost primary
27   productivity in some cases (Goldenberg et al., 2017); however, increased metabolic demands and greater
28   consumption by predators under warmer temperature increase the strength of top-down control (predation
29   mortalities of herbivores) and thus counteracts the effects of increased bottom-up productivity (Goldenberg
30   et al., 2017; Kordas et al., 2017). Ocean acidification could also increase species energetic costs and the
31   grazing rate of herbivores, affecting ecosystem responses to increased primary productivity (Ghedini et al.,
32   2015). Although these increasingly complex experiments have highlighted the potential for species
33   interactions to mediate the effects of climate change, our understanding of the effects on intact, functioning
34   ecosystems is limited. Despite predictions for increased production and herbivory with warming and
35   acidification, an experimental study of a more complex food web revealed an overall reduction in the energy
36   flow to higher trophic levels and a shift towards detritus-based food webs (Ullah et al., 2018).
37
38   Overall, intertidal rocky shores ecosystems are highly sensitive to ocean warming, acidification and extreme
39   heat exposure during low tide emersion (*high confidence*). More field and experimental evidence shows that
40   these ecosystems are at a moderate risk at present and this level is expected to rise to very high under the
41   RCP8.5 scenario by the end of the century (see Section 5.3.7). Benthic species will continue to relocate in
42   the intertidal zones and experience mass mortality events due to warming (*high confidence*). Interactive
43   effects between acidification and warming will exacerbate the negative impacts on rocky shore communities,
44   causing a shift towards a less diverse ecosystem in terms of species richness and complexity, increasingly
45   dominated by macroalgae (*high confidence*).
46
47   ### 5.3.6   Kelp Forests
48
49   Kelp forests are three-dimensional, highly productive coastal ecosystems with a reported global net primary
50   production between 1.02 –1.96 GtC yr^{-1} (Krause-Jensen and Duarte, 2016). They cover about 25% of the
51   world's coastline (Filbee-Dexter et al., 2016), mostly temperate and polar (Steneck et al., 2003). Canopy-
52   forming macroalgae provide habitat for very many associated invertebrates and fish communities
53   (Pessarrodona et al., 2019). This assessment synthesizes new evidence since SR1.5 on climate risks and
54   impacts, and their interactions with non-climatic drivers on ecosystem biodiversity, structure and
55   functioning.
56

1   Observational and experimental evidence since SR1.5 (Hoegh-Guldberg et al., 2018) supports that report's
2   conclusions that kelp forests are already experiencing large-scale changes, and that critical thresholds occur
3   for some forests at 1.5°C of global warming (*high confidence*). Due to their low capacity to relocate and high
4   sensitivity to warming, kelp forests are projected to experience higher frequency of mass mortality events as
5   the exposure to extreme temperature rises (*very high confidence*). Moreover, changes in ocean currents have
6   facilitated the entry of tropical herbivorous fish into temperate kelp forests decreasing their distribution and
7   abundance (*medium confidence*). More evidence from model projections in the 21st century supports this
8   observed range contraction of kelp forests at the warm end of their distributional margins and expansion at
9   the poleward end with the rate being faster for high emission scenarios (*high confidence*).
10
11   New global estimates show that the abundance of kelp forests has decreased at a rate of ~2 % per year over
12   the past half century (Wernberg et al., 2019), mainly due to ocean warming and marine heat waves (e.g. in
13   Western Australia a mean loss of 43 % in area followed a marine heat weave in summer 2010/2011
14   (Wernberg et al., 2016), Section 6.4.2.1), as well as from other human stressors (*high confidence*) (Filbee-
15   Dexter and Wernberg, 2018). At some localities, human-driven environmental changes such as coastal
16   eutrophication and pollution is causing severe deterioration of kelp forests adding to the loss of these
17   ecosystems from warming, storms and heat weaves (Andersen et al., 2013; Filbee-Dexter and Wernberg,
18   2018).
19
20   Two global datasets and one dataset covering European coastlines (Araujo et al., 2016; Krumhans et al.,
21   2016; Poloczanska et al., 2016) identify large local and regional variations in kelp abundance over the past
22   half century with 38% of these ecoregions showing a decline, 27% an increase and 35% no change
23   (Krumhans et al., 2016). These data reflect the high spatio-temporal variability and resilience of kelp forests
24   (Reed et al., 2016; Wernberg et al., 2018). For example, a 34 year dataset of kelp canopy biomass along the
25   California coastline does not yet show a significant response to global warming because this ecosystem
26   responds to low frequency marine climate oscillations (Bell et al., 2018c). However, between 1950 – 2010
27   regional warming caused consistent negative responses in abundance, phenology, demography and
28   calcification of macroalgae for the northeast Atlantic and southeast Indian Ocean (Poloczanska et al., 2016).
29   Declines in kelp forest abundance attributed to climate change and not related to sea urchin overgrazing
30   (which is a major driver of decline and regime shift; Ling et al. (2014)) have been documented since the
31   1970s and evidence has increased within the last two decades (Filbee-Dexter and Wernberg, 2018). Despite
32   a lack of data from some regions such as South America (Pérez-Matus et al., 2017), observational evidence
33   since SR1.5 supports with *very high confidence* that warming is driving a contraction of kelp forests at low
34   latitudes (Franco et al., 2018b; Casado-Amezúa et al., 2019; Pessarrodona et al., 2019) and expansion in
35   polar regions (*medium confidence*) (Section 3.2.3.1.2) (Bartsch et al., 2016; Paar et al., 2016).
36
37   In many areas worldwide where the distribution range of kelp has contracted due to climatic and non-
38   climatic drivers, it has been replaced by a less diverse and less complex turf-dominated ecosystem (Filbee-
39   Dexter and Wernberg, 2018) (*high confidence*). Kelp supports other ecosystem components by providing
40   food, substrate for spawning and habitat that mediate trophic interactions (O'Brien et al., 2018); its
41   degradation therefore reduces species richness, biomass production and dependent flora and fauna species
42   (Teagle and Smale, 2018; Pessarrodona et al., 2019). In the north-east Atlantic, the warm water species
43   *Laminaria ochroleuca* is expanding poleward into regions previously dominated by the cold-water species *L.*
44   *hyperborea* which is retreating at its southern edge. These two kelp species are similar in morphology, but
45   the cold water *L. hyperborea* hosts sessile communities of algae and invertebrates 12 times more diverse and
46   richer in biomass than the warm water kelp species (Teagle and Smale, 2018). Climate-driven shifts in the
47   species composition also affect carbon cycling, because warm-temperate kelps produce larger pools of
48   organic matter than cold-temperate species, and their detritus is degraded faster (Pessarrodona et al., 2019).
49
50   New empirical eco-physiological studies in combination with field surveys support the evidence for climate
51   change causing kelp forest degradation and range-shifts (Franco et al., 2018b; Wernberg et al., 2018). For
52   example, interactive effects of ocean warming and acidification cause kelp degradation and disease-like
53   symptoms, with detrimental effects on photosynthetic efficiency (Qiu et al., 2019). Enhanced herbivory due
54   to warming and the establishment of herbivorous fish species in temperate kelp forest has been observed to
55   enhance ecosystem degradation (Vergés et al., 2016). However, invader seaweed species driven by warming
56   can create more complex trophic interactions, reducing the consumption by herbivorous gastropods
57   (Miranda et al., 2019). Increased physical stress by storm events also alters the kelps community, affecting

1   the recruitment time of kelp species. The resulting dominance of younger stages favors species with a year-
2   round spore production or an opportunistic life strategy, reducing the kelp canopy (Pereira et al., 2017).
3
4   Projections of future distribution of kelp species based on their physiological thresholds show major species-
5   specific range shifts under different emission scenarios. For example, under RCP2.6, laminaria and other
6   canopy-forming seaweed species in the Northwest Atlantic are projected to show northward range shifts at
7   their southern (warm) edge of ≤40 km, with some equatorial range expansion from 2050 to 2100. That
8   northward range shift increases to 406 km under RCP8.5 (at 13–19 km per decade, including contractions of
9   their warmer edges) (Wilson et al., 2019). Whilst no changes in species richness are projected under RCP2.6,
10 more than 50% richness loss is projected under RCP8.5 in some areas (Wilson et al., 2019). Overall, model
11 projections show that worldwide range contractions of kelps can be expected to continue at the warm end of
12 distributional margins and range expansions at their poleward end (*high confidence*) ((Raybaud et al., 2013;
13 Assis et al., 2016; Assis et al., 2018; Wilson et al., 2019).
14
15 In summary, kelp forests have experienced large-scale habitat loss and degradation of ecosystem structure
16 and functioning over the past half century, implying a moderate to high level of risk at present conditions of
17 global warming (*high confidence*) (Section 5.3.7). The loss of kelp forests is followed by the colonization of
18 turfs, which contributes to the reduction in habitat complexity, carbon storage and diversity (*high
19 confidence*). Kelp ecosystems are expected to continue to decline in temperate regions driven by ocean
20 warming and intensification of extreme climate events (*high confidence*). The level of risk for the ecosystem
21 is projected to rise to very high under RCP8.5 scenario by 2100 (*high confidence*).
22
23 ### 5.3.7   *Risk Assessment for Coastal Ecosystems*
24
25 This section synthesizes the assessment of climate impacts on coastal ecosystems' biodiversity, structure and
26 functioning and the levels of risk under contrasting future conditions of global warming. As described in
27 Section 5.2.5, the format for Figure 5.16 matches that of Figure 19.4 of IPCC AR5 (Oppenheimer et al.,
28 2015) and Figure 3.20 of SR1.5 (Hoegh-Guldberg et al., 2018), indicating the levels of additional risk as
29 colours (white, yellow, red and purple). The elements or burning embers for coastal ecosystems (Figure
30 5.16) indicate how risks increase with ocean warming, acidification, deoxygenation, sea level rise and
31 extreme events with a comparison between present day conditions (2000s) and future conditions by the year
32 2100 under low (RCP2.6) and high (RCP8.5) CO_2 emission scenarios. The transition between the levels of
33 risk for each type of coastal ecosystem is estimated from key evidence assessed in Sections 5.3.1 to 5.3.6.
34 The embers are based on sea surface temperature which differs by 0.2 °C from global atmospheric
35 temperature (Karl et al., 2015). The transition-values may have an error of ±0.3 °C depending on the
36 consensus of expert judgment. The assessed confidence in assigning the levels of risk at present day and
37 future scenarios are *low, medium, high*, and *very high* levels of confidence. A detailed account of the
38 procedures involved in developing the ember for each type of coastal ecosystem is given in the
39 Supplementary Material (SM5.3). This Supplementary Material includes the description of climate hazards,
40 sensitivity of key biotic and abiotic components, natural adaptive capacity, and observed impacts and
41 projected risks. The burning embers for seagrass meadows, warm water corals and mangrove forests are in
42 agreement with the conclusions in SR1.5 (Hoegh-Guldberg et al., 2018). The more recent literature assessed
43 here strengthens the overall confidence in the assignment of transition and the level of risk for each
44 ecosystem.
45
46 Detection and attribution studies show that climate change impacts began over the past 50 years in coastal
47 ecosystems, indicating a transition from undetectable risk (white areas in Figure 5.16) to moderate risk
48 below recent temperatures (*high confidence*). This transition occurs at lower global levels of warming for
49 coral reefs (0.2–0.4°C) (*high confidence*), seagrass meadows (0.5–0.8°C) (*very high confidence*) and kelp
50 forests (0.6–0.8°C) (*high confidence*), with coral reefs already at high risk (0.4–0.6°C) for the present day
51 (*very high confidence*). Global common responses include large-scale coral bleaching events (Section 5.3.4)
52 and contraction of seagrass meadows (Section 5.3.2) and kelp forests (Section 5.3.6) at low-latitudes (*high
53 confidence*), in response to warming and marine heat waves. Degraded coral reefs and kelp forests have
54 shifted to algal and turf-dominated ecosystem at several regions worldwide, causing loss of habitat
    complexity and biodiversity.

1   The transition from undetectable to moderate risk in saltmarshes (Section 5.3.2) and rocky shores (Section
2   5.3.5) takes place between 0.7–1.3°C of global warming (*medium/high confidence*), and between 0.9–1.8°C
3   (*medium confidence)* in sandy beaches (Section 5.3.3), estuaries (Section 5.3.1) and mangrove forests
4   (Section 5.3.2) (Figure 5.16). In all these coastal ecosystems, the detection and attribution of changes in
5   biodiversity, structure and functioning are not as robust as in coral, seagrass and kelp ecosystems that have
6   been extensively studied over the past decades and are highly sensitive to extreme climate events. Estuaries
7   and sandy beaches are highly dynamic in terms of hydrological and geomorphological processes, giving
8   them more natural adaptive capacity to climate impacts. In these systems, sediment relocation, soil accretion
9   and landward expansion of vegetation may mitigate against flooding and habitat loss in the context of sea
10 level rise and extreme climate-driven erosion. Common global responses observed since 1970 include
11 poleward expansion of mangrove forests due to warming; transformation of saltmarshes into mudflats; shifts
12 in species composition in response to flooding and salinization; upstream migration of estuarine biota; and
13 redistribution of macrobenthic communities in sandy beaches. Calcified organisms in intertidal rocky shores
14 are highly sensitive to ocean warming and acidification, marine heat waves and heat exposure during low
15 tide, with observed mass mortality events and reduced calcification.
16
17 In all coastal ecosystems, multiple climate hazards will emerge from historical variability in the 21st century
18 under RCP8.5 (Box 5.1), while the time of emergence will be later and with less climate hazard under
19 RCP2.6. Non-climatic human impacts such as eutrophication add to, and in some cases, exacerbate these
20 large-scale slow climate drivers beyond biological thresholds at local scale (e.g., deoxygenation).
21
22 All coastal ecosystems will experience high to very high risk under RCP8.5 by the end of the 21st century. In
23 particular, coral reefs, seagrass meadows and kelp forests will face high to very high risk even at 1.5°C
24 above pre-industrial temperature (*high confidence*). The ecosystems expected to be at very high risk under
25 the high emission scenario are coral reefs (transition from high to very high risk 0.6–1.2°C) (*very high
26 confidence*), seagrasses meadows (2.2–3.0°C) (*high confidence*), kelp forests (2.5–2.7°C) (*high confidence*)
27 and rocky shores (2.9–3.4°C) (*medium confidence*). These ecosystems have low to moderate adaptive
28 capacity, as they are highly sensitive to ocean warming, marine heat waves and acidification. For example,
29 kelp forests at low-latitudes and temperate seagrass meadows with endemic species will continue to retreat
30 with more frequent extreme temperatures, and their low dispersal ability will elevate the risk of local
31 extinction. Biogenic shallow reefs with calcified organisms (e.g., corals, mussels, calcified algae) are
32 particularly sensitive to ocean acidification and compound effects with rising temperatures, deoxygenation,
33 sea level rise and increasing extreme events, making these ecosystems highly vulnerable (with low
34 resilience) to future emission scenarios. Furthermore, almost all coral reefs will greatly decline from their
35 current levels, even if global warming remains below 2°C (*very high confidence*). Any coral reefs that do
36 survive to the end of the century will not be the same because of irreversible changes in habitat structure and
37 functioning, including species extinctions and food web disruptions; these changes are already taking place
38 (e.g. the Caribbean reefs). The transition to new ecosystem states driven by unpredictable pulses of
39 disturbance and progressive climate hazards will have negative impacts on ecosystem services (Section 5.4).
40
41 The ecosystems at moderate to high risk under future emission scenarios (Figure 5.16) are mangrove forests
42 (transition from moderate to high risk at 2.5–2.7°C of global warming), estuaries and sandy beaches (2.3–
43 3.0°C) and saltmarshes (transition from moderate to high risk at 1.8–2.7°C and from high to very high risk at
44 3.0–3.4°C) (*medium confidence*). Mangrove forests and saltmarshes can initially cope with sea level rise by
45 plant biomass accumulation, soil accretion and sediment relocation, but the evidence shows they are unlikely
46 to withstand the sea level rise projected under RCP8.5. Moreover, pervasive coastal squeeze and human-
47 driven habitat deterioration will reduce the natural capacity of these ecosystems to adapt to climate impacts
48 (*high confidence*). Projected warming and sea level rise by the end of the century will continue to expand
49 salinization and hypoxia in estuaries with high risk of impacts for benthic and pelagic biota. These impacts
50 will be more pronounced under RCP8.5 in more vulnerable eutrophic, shallow and microtidal estuaries in
51 temperate and high latitudes. Erosion in sandy beach ecosystems will continue with global warming, rising
52 sea level and more intense and frequent storm surges and marine heat waves. The risk of losing habitats for
53 flora and fauna is expected to rise to high level under the high emission scenario by the end of the 21st
54 century (*medium confidence*, Figure 5.16). By contrast, the risk of impacts is expected to be only slightly
55 higher than present for a low emission scenario than today (*medium confidence,* Figure 5.16).
56

1   All types of ecosystems that have been assessed in the open ocean (Sections 5.2.3 and 5.2.4) and coastal
2   areas (Sections 5.3.1 to 5.3.6) show increased risk under both the low and the high emission scenarios
3   (RCP2.6 and RCP8.5) compared with the present level of change (Figure 5.16). In all assessed cases with all
4   of the factors considered (climate drivers and physiological understanding), RCP2.6 has a lower level of risk
5   than RCP8.5 (*very high confidence*).
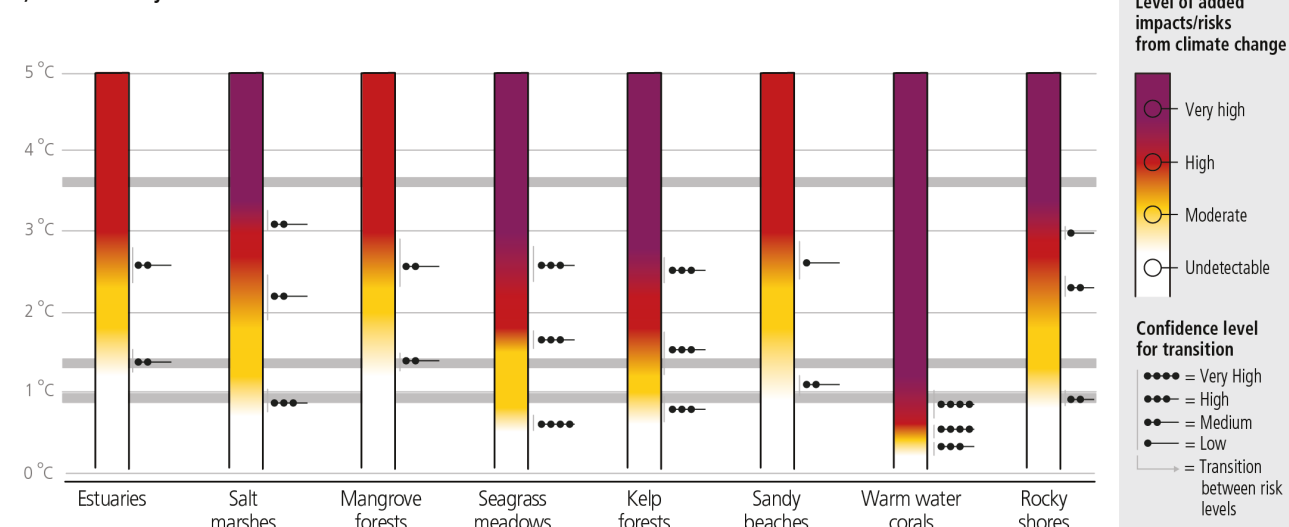6
7
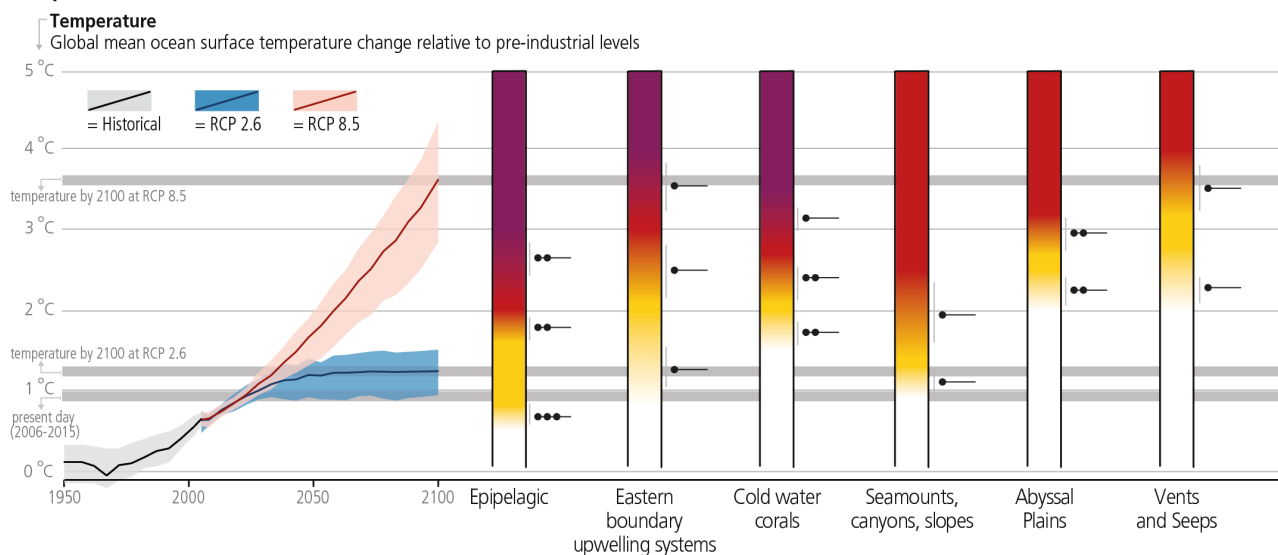
8
9   **Figure 5.16:** Risk scenarios for open ocean (upper panel) and coastal (lower panel) ecosystems based on observed and
10   projected climate impacts. 'Present day' corresponds to the 2000s, whereas the different greenhouse emissions
11   scenarios: RCP2.6 and RCP8.5 correspond to year 2100. Multiple climatic hazards are considered, including ocean
12   warming, deoxygenation, acidification, changes in nutrient, particulate organic carbon flux and sea level rise (see
13   sections 5.2 and 5.3). The projected changes in sea surface temperature (SST) from an ensemble of general circulation
14   models (left panels) indicates the level of ocean changes under RCP2.6 and RCP8.5 (see Cross Chapter Box 1 Table
15   CB1 for the projected global average changes in average air temperature, SST and other selected ocean variables).
16   Global average impacts/risks are represented. Regional variations of risks/impacts are described in Section 5.2.5, 5.3.7
17   and the Supplementary Material (SM5.2 and SM5.5). Impact/risk levels do not consider human risk reduction strategies
18   such as societal adaptation, or future changes in non-climatic hazards. The grey vertical bars indicate the transition
19   between the levels of risks, with their confidence level based on expert judgment. Note: The figure depicts climate
20   change impacts and risks on warm water corals taken from SR1.5, based on global models. Observed impacts on coral
21   reefs ecosystems outlined in Section 5.3.4 and Box 5.5 reveal a more complex situation that may result in regional
22   differences in confidence levels.
23
24

5.4 Changing Marine Ecosystem Services and Human Wellbeing

Ecosystem services are the environmental processes and functions that have monetary or intrinsic value to human society; they render benefits to people and support human wellbeing (Tallis et al., 2010; Costanza et al., 2014). Marine ecosystem services are generated throughout the ocean, from shallow water to the deep sea (Armstrong et al., 2012; Thurber et al., 2014). Although all ecosystem services are interconnected (Leadley et al., 2014), they can be broadly divided into provisioning services, regulating services, supporting services and cultural services (Sandifer and Sutton-Grier, 2014), as considered below (Section 5.4.1) together with their implications for human wellbeing (Section 5.4.2). Ecosystem services have also been described as 'nature's contribution to people' (Díaz et al., 2018).

5.4.1 Changes in Key Ecosystem Services

WGII AR5 concluded that climate change increases the risk of impacts on the goods and services derived from marine biodiversity and ecosystems (Pörtner et al., 2014). SR1.5 concluded that current ecosystem services from the ocean are expected to be reduced at 1.5°C of global warming, with losses being even greater at 2°C of global warming. These reductions in services are driven by decreasing ocean productivity, biogeographic shifts, damage to ecosystems, loss of fisheries productivity and changes to ocean chemistry (*high confidence*) (Hoegh-Guldberg et al., 2018). Building on these previous assessments, this section assesses new evidence on observed impacts and future risk of climate change on ecosystem goods and services from the open ocean (Section 5.2) and coastal ecosystems (Section 5.3). Chapter 3 assesses ecosystem services in polar oceans.

5.4.1.1 Provisioning Services

Fisheries are an important provisioning service from marine ecosystems, providing food, nutrition, income and livelihoods for many millions of people around the world (FAO, 2018b). Globally, total fish catches amount to 80–105 MT annually in the 2000s (FAO, 2016; Pauly and Zeller, 2016; FAO, 2018a), directly generating over $80 billion USD of revenue (Sumaila et al., 2015). Most global fisheries are considered to be fully- to over-exploited (FAO, 2018b). Over 80% of the global fish catch is estimated to be from coastal and shelf seas with less than 20% from the high seas (Sumaila et al., 2015) (Figure 5.17).
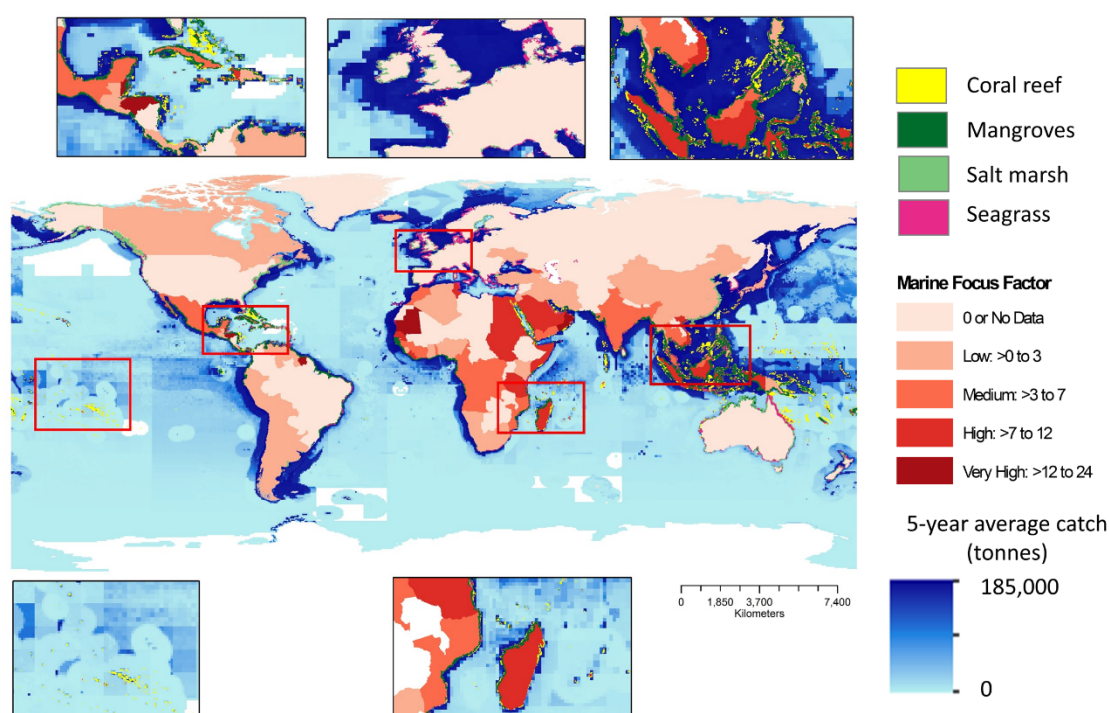

Figure 5.17: Global distribution of fish catches (average 2010-2014, based on (Pauly and Zeller, 2016)), coastal habitats including seagrasses (UNEP-WCMC and FT, 2017) saltmarshes (Mcowen et al., 2017), mangroves (Spalding, 2010), coral reefs (UNEP-WCMC and WRI, 2010) and an index (called Marine Focus Factor) for the inclusion of the

ocean in the Nationally Determined Contributions (NDCs) published by each country (Gallo et al., 2017). The higher the Marine Focus Factor, the more frequent use of ocean in the country's NDCs.
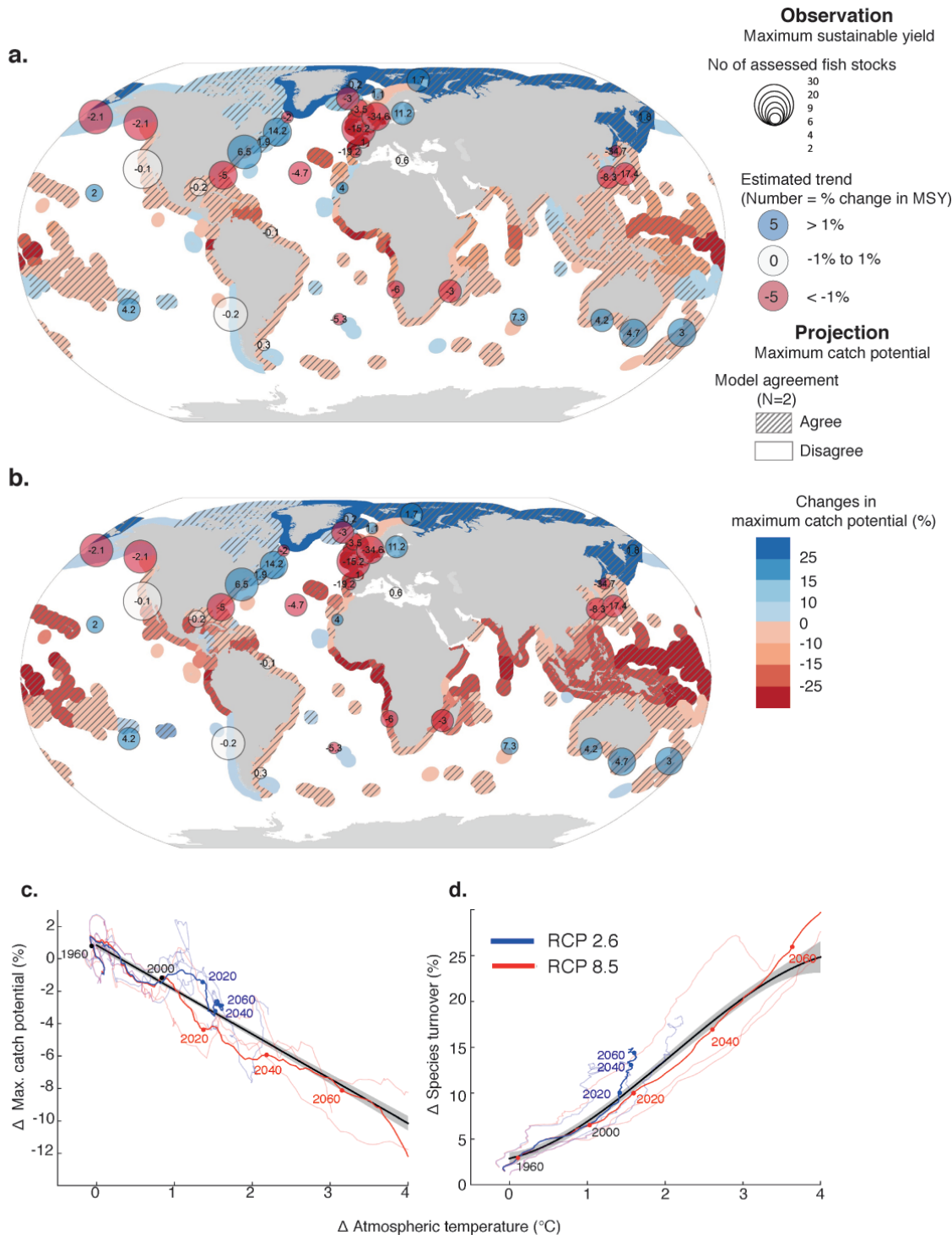
Figure 5.18: Historical and projected maximum sustainable yield (MSY) and maximum fish catch potential by region. Historical trends in MSY is based on time-series of fish stock assessment data (Free et al., 2019) represented as circles in panels (a) and (b). The size of the circle represents the number of assessed fish stocks while the number in the circle represents the estimated percent change in MSY since the 1930s. Projected changes in maximum catch potential by 2050 (average between 2041–2060) relative to 2000 (1991–2010) under (a) RCP2.6 and (b) RCP8.5 scenarios from two models: Dynamic Bioclimate Envelope Model and dynamic size-spectrum foodweb model with the color in each ocean region representing the projected level of change and the shading representing where both models agree in the direction of change (Cheung et al., 2018a). Also presented is the scaling between projected global atmospheric warming (relative to 1950-1961) and (c) changes in maximum fish catch potential and (d) species turnover using the Dynamic Bioclimate Envelope Model and outputs from three CMIP5 ESMs (Cheung and Pauly, 2016).

Observed fish catches have been related to net primary production (NPP) and water temperature, with the direction and magnitude of the relationship varying between regions and fish stocks (Cheung et al., 2008; McOwen et al., 2015; Britten et al., 2016; Stock et al., 2017). The maximum catch potential of large marine ecosystems generally increases with their (NPP) and their energy transfer efficiency, but the relative importance of total NPP to fisheries production is lower in nutrient-poor systems with microbially-dominated foodwebs (Section 5.2) and empirical relationships between NPP and fisheries production over-estimate potential catches in polar regions (Stock et al., 2017) (Chapter 3). Here, potential fish catch or maximum catch potential refers to the potential of the fish stocks to provide long-term fish catches; it is considered a proxy of maximum sustainable yield. However, the actual catches realized by fisheries will depend strongly on past and present fishing effort and the exploitation status of the resources (Cheung et al., 2018a; Barange, 2019). Observed variations between regions suggest that changes in temperature and NPP in the past (Section 5.2.2, 5.2.3) may have also affected maximum catch potential (*medium evidence, high agreement, medium confidence*).

Changes in fish catches from 1998 to 2006 in 47 large marine ecosystems around the world were found to be significantly related to: changes in estimated cholorophyll *a* (a proxy for phytoplankton biomass) in 18 of these ecosystems (mostly tropical and eastern boundary upwelling systems); changes in sea surface temperature in 12 of these ecosystems (mostly mid-latitude); and changes in fishing intensity in 16 of these ecosystems (widely spread) (McOwen et al., 2015). Analysis of population data since the 1950s for 262 fish stocks across 39 large marine ecosystems and the high seas suggest that average recruitment to the stocks has declined by around 3% of the historical maximum per decade with variations between regions and stocks (Britten et al., 2016). The declines (69% of the studied stocks, 31 of the 39 assessed large marine ecosystems) are significantly related to estimated chlorophyll *a* concentration and the intensity of fishing, with the North Atlantic showing the steepest declines (Britten et al., 2016). In addition, recent meta-analysis of population data from 235 fish stocks worldwide from 1930 to 2010 suggest that the maximum catch potential from these populations decreased by 4.1% (95% confidence span 9.0% decline to 0.3% increase) during this period with variations between fish stocks and regions (Free et al., 2019). Specifically, temperature is a significant factor explaining changes in catch potential of 12% of the fish stocks, with East Asian regions having the largest stock declines related to warming. In intermediate latitudes across the Atlantic, Indian and Pacific Oceans, catches of tropical tunas, including skipjack and yellowfin tuna, are significantly and positively related to increases in sea surface temperature, although the overall catches across latitudinal zones do not show significant change (Monllor-Hurtado et al., 2017). Observational evidence from spatial and temporal linkages between catches and oceanographic variables therefore supports the conclusions from WGII AR5 and SR1.5 that potential fisheries catches have already been impacted by the effects of warming and changing primary production on growth, reproduction and survival of fish stocks (*robust evidence*, *high agreement*, *high confidence*).

There are substantial variations in the direction of changes and the attribution of climatic drivers between regions and fish stocks, and the availability of datasets is biased towards mid-latitude areas and epipelagic and coastal ecosystems. As a result, quantitative attribution of climate impacts on the productivity of specific fish stocks has *low confidence*. Changes in catch potential for fish stocks and regions worldwide that were considered overfished were most sensitive to warming (Essington et al., 2015; Britten et al., 2016; Free et al., 2019). This suggest that climatic drivers and overfishing have interacted synergistically in impacting some fish stocks and their catches (*high confidence*). In addition, analysis of historical catch records since AR5 show further warming-related changes in species composition, with an increased dominance of warm-water species in coastal and shelf seas since the 1970s (Cheung et al., 2013; Keskin and Pauly, 2014; Tsikliras et al., 2014; Maharaj et al., 2018). Many marine ecosystems worldwide have shown an increased dominance of warm-water species following increases in sea water temperature (5.2.3, 5.3), with parallel changes in the species composition of fish catches since the 1970s in many of the studied shelf seas (*high confidence*).

Based on CMIP5 ESM projections of changes in temperature, net primary production, oxygen, salinity and sea ice extent, two marine ecosystem and fisheries models project a decrease in maximum catch potential under RCP 2.6 of 2.8–8.7% by 2050 and 2.8–9.1% by 2095 relative to 2000 (FAO, 2018a). Under RCP 8.5, the projected decrease was larger: 7.0–11.8% by 2050 and 16.2–25.5% by 2095 relative to 2000 (Figure

1   5.18). The trends agree with the projected changes in total marine animal biomass for the 21st century
2   (Blanchard et al., 2017; Lotze et al., 2018) (Section 5.2.3). A single fisheries model with atmospheric
3   warming projected a potential catch loss of 3.4 million tonnes and decreases of 6.4% of catch potential of the
4   exploited species per degree Celsius atmospheric warming relative to 1951–1960 level (Cheung et al.,
5   2016b) (Figure 5.18). Interactions between temperature, net primary production and transfer efficiency of
6   energy across the foodweb are projected to amplify these trends, with projected decreases greater than 50%
7   in some regions by 2100 under high emissions scenarios (Stock et al., 2017). Thus, there is high model
8   agreement that ocean warming and changes in net primary production in the 21st century will reduce the
9   global maximum catch potential, particularly in tropical oceans (*high confidence*) and alter the distribution
10   and composition of exploited species (*high confidence*). The projected risk of these fisheries impacts
11   increases with increasing greenhouse gas emissions (*high confidence*). However, given the uncertainties of
12   projected changes in ocean conditions from ESMs (Section 5.2.2), and that most global scale fisheries
13   models are largely driven by changes in temperature and primary production while other changes in ocean
14   biogeochemical changes are not explicitly considered (Tittensor et al., 2018), the quantitative magnitude of
15   the projected changes in maximum catch potential is considered to have *medium* and *low* confidence at
16   global and regional scales, respectively. Given the significant interactions between catch potential and level
17   of fisheries exploitation, the realized catches in the 21st century would depend on future scenarios of fishing
18   and fisheries governance (Section 5.4.2, 5.5). As a result, projections of realized catches have *low*
19   *confidence*.
20
21   Tropical oceans are projected to experience much larger impacts (three times or more decrease in catch
22   potential) than the global average, particularly the western central Pacific Ocean, eastern central Atlantic
23   Ocean and the western Indian Ocean, by the end of the 21st century under RCP 8.5 (Blanchard et al., 2017).
24   For example, around the exclusive economic zones of the Pacific Islands states, more than 50% of exploited
25   fishes and invertebrates are projected to become locally-extinct in many regions by 2100 relative to the
26   recent past under RCP 8.5 (Asch et al., 2017). These factors cause 74% of the area to experience a projected
27   loss in catch potential of more than 50%. Under RCP 2.6, the area of large projected catch loss is projected
28   to be halved (Asch et al., 2017). However, while temperate commercially-important tunas species such as
29   albacore, Atlantic and southern bluefin) are projected to shift poleward and decrease in abundance in the
30   tropics, some tropical species such as skipjack tuna are projected to remain abundant, but with changes in
31   distribution patterns in low-latitude regions by the mid-21st century, with some models projecting subsequent
32   decrease under RCP8.5 (Lehodey et al., 2013; Dueri et al., 2014; Erauskin-Extramiana et al., 2019). Recent
33   evidence therefore supports the conclusion from previous assessments (Pörtner et al., 2014; Hoegh-
34   Guldberg et al., 2018) that low latitude fish catch potential are projected to have a high risk of climate
35   impacts, which will be exacerbated by higher greenhouse gas emissions (*medium evidence, high agreement,*
36   *high confidence*). Tropical fish catch potential of some species resilient to the changing environment may
37   have lower climate risk in the near-term although their risk increases substantially further into the 21st
38   century under RCP8.5 (*medium confidence*). In contrast, the catch potential in the Arctic is projected to
39   increase, although with high inter-model variability (*medium evidence, low agreement, low confidence*)
40   (Cheung and Pauly, 2016; Blanchard et al., 2017) (Chapter 3).
41
42   Although demersal fisheries in the deep ocean represent a small proportion of global fisheries catches, they
43   are economically valuable for some countries, and there is increasing commercial interest in mesopelagic
44   (deep pelagic ocean) fisheries (St. John et al., 2016). Commercially-exploited fish and shellfish from deep
45   sea ecosystems will be exposed to climate risks from physical and chemical changes in ocean conditions
46   including warming, decreased oxygen, reduced aragonite saturation state, and decreased supply of
47   particulate organic matter from the upper ocean (Section 5.2.3, 5.2.4) (FAO, 2019). These biogeochemical
48   changes may reduce the growth, reproduction and survivorship of deep-ocean fish stocks, which will alter
49   their distributions, in similar ways to those in the surface ocean, impacting their fish catch potential (FAO,
50   2019). For example, in the eastern Pacific near-bottom oxygen concentration is positively correlated with
51   biomass of commercially harvested species (Keller et al., 2010) and catch per unit effort (Banse, 1968;
52   Rosenberg et al., 1983; Keller et al., 2015); some commercially harvested species only appear during
53   oxygenation events associated with El Niño (Arntz et al., 2006). In the mesopelagic zone, expansion of the
54   oxygen minimum zone (OMZ) results in habitat compression that can increase catchability of fish stocks
55   such as tunas (Prince et al., 2010; Stramma et al., 2011). Also, as OMZ expands, the potential may exist for
56   increased availability and harvest of hypoxia-tolerant species such as Humboldt squid (*Dosidicus gigas*),
57   thornyheads (*Sebastolobus* spp.) or dover sole (*Microstomus pacificus*) (Gilly et al., 2013; Gallo and Levin,

1   2016). However, any expansion of the OMZ will interact with other climatic hazards such as warming,
2   which then adds to the overall risk of impacts on fish stocks and their catches (Breitburg et al., 2018).
3   Overall, the abundance of fisheries resources and potential catches from the deep sea will be at high risk of
4   impacts in the 21st century under RCP 8.5 (*low confidence*), with reduced risk under RCP 2.6 (*medium*
5   *confidence*).
6
7   In addition to capture fisheries, mariculture (marine aquaculture) is also an important marine ecosystem
8   provisioning service, contributing about 27.7 million tonnes of seafood in 2016 (FAO, 2018b). Recent
9   projections of climate change impacts on mariculture, based on thermal tolerance and the effects of changing
10   temperature, primary production and ocean acidification, suggest an overall decline in mariculture potential
11   by 2100 under RCP 8.5 with large regional variations (Froehlich et al., 2018). Modelling analyses for farmed
12   Atlantic salmon, cobia and seabream also suggest that climate change would reduce their growth potential in
13   ocean areas where temperature is projected to increase to levels outside the thermal tolerance ranges of these
14   species (Klinger et al., 2017). This decrease in growth could therefore translate into a decrease in the general
15   productivity of the sector (*limited evidence, low confidence*); however, new potential areas and the use of
16   more climate-resilient strains or species for mariculture may emerge that could reduce the risk of impacts on
17   potential mariculture production (*limited evidence, low confidence*).
18
19
20   [START OF BOX 5.3 HERE]
21
22   **Box 5.3: Responses of Coupled Human-Natural Eastern Boundary Upwelling Systems to Climate**
23   **   Change**
24
25   Eastern Boundary Upwelling Systems (EBUS) are among the world's most productive ocean ecosystems
26   (Kämpf and Chapman, 2016). They directly support livelihoods in coastal communities and provide many
27   wider benefits to human society (García-Reyes et al., 2015; Levin and Le Bris, 2015). The high productivity
28   of EBUS is supported by the upwelling of cold and nutrient-rich waters, itself driven by equator-ward
29   alongshore winds that cause the displacement of surface waters offshore and their replacement by deeper
30   waters. Total annual fish catches from the four main EBUS (California Current, Humboldt Current, Canary
31   Current and Benguela Current) were 16-24 tonnes per year in the 2000s, providing around 17% of the global
32   catch (Pauly and Zeller, 2016). These catches are consumed locally, as well as being processed and exported
33   as seafood, fish meals and oils to support aquaculture and livestock production. Upwelling of cold deeper
34   water also increases the condensation of humid air in coastal areas, benefitting coastal vegetation and
35   agriculture and suppressing forest fires (Black et al., 2014). The high concentration of marine mammals
36   attracted by the productive upwelling ecosystem support lucrative eco-tourism, such as whale-watching in
37   the California Current (Kämpf and Chapman, 2016). The total economic value of the goods and services
38   provided by the Humboldt Current alone is estimated to be $19.45 billion USD per year (Gutiérrez et al.).
39   Thus, although their area is small compared to other pelagic ecosystems, climate change impacts on EBUS
40   will have disproportionately large consequences for human society (*very high confidence*).
41
42   The coupled human-natural EBUS are vulnerable to the multiple effects of climate change with large
43   regional variation (Blasiak et al., 2017). Observations and modelling analyses suggest that winds have
44   intensified in most EBUS (except the Canary Current) during the last 60 years, with several hypotheses
45   proposed to explain the mechanisms (Sydeman et al., 2014; García-Reyes et al., 2015; Rykaczewski et al.,
46   2015; Varela et al., 2015). ESMs predict reduction of wind and upwelling intensity in EBUS at low latitudes
47   and enhancement at high latitudes for RCP8.5, with an overall reduction in either upwelling intensity or
48   extension (Belmadani et al., 2014; Rykaczewski et al., 2015; Sousa et al., 2017). However, coastal warming
49   and wind intensification may lead to variable countervailing responses to upwelling intensification at local
50   scales (García-Reyes et al., 2015; Wang et al., 2015a; Oyarzún and Brierley, 2018; Xiu et al., 2018). Local
51   winds and mesoscale oceanographic features (not resolved in most global ESMs) are thought to have a
52   greater impact on regional productivity than large-scale wind patterns (Renault et al., 2016; Xiu et al., 2018).
53
54   There is conflicting evidence in SST trends in recent decades, even among the same EBUS, due to varying
55   spatio-temporal resolution of SST data and the superimposed effects of interannual to multi-decadal
56   variability (García-Reyes et al., 2015). Some EBUS are close to important thresholds in terms of
57   oxygenation and ocean acidification (Gruber et al., 2012; Franco et al., 2018a; Levin, 2018). Large-scale

coastal and offshore data for the California Current indicate that there have been decadal decreases in pH and dissolved oxygen affecting organisms and ecosystems (Alin et al., 2012; Breitburg et al., 2018; Levin, 2018). Model projections for 2100 suggest strong effects of deoxygenation and reduced pH in the Humboldt Current and the California Current under RCP 8.5 (Gruber et al., 2012; García-Reyes et al., 2015), affecting seafloor habitats and invertebrate fisheries (Marshall et al., 2017; Hodgson et al., 2018). For instance, the Humboldt Current is projected to experience widespread aragonite undersaturation within a few decades (Franco et al., 2018a), with strong impacts on calcified organisms. Such ocean acidification could be worsened by synergistic effects of ocean warming and deoxygenation (Lachkar, 2014).

The climate change impacts on ecosystem services from EBUS vary according to the biophysical and the socio-economic characteristics of the upwelling systems (García-Reyes et al., 2015) and supplementary material (SM5.4). The fisheries are not only highly sensitive to upwelling conditions but also by fishing effects on the exploited populations. For example, the anchoveta population collapsed in the Humboldt Current after an El Niño in the 1970s (Gutiérrez et al., 2017). Because small pelagic fisheries from upwelling regions are the main source of the global fishmeal market, decreases in their catches increase the international fishmeal price, increasing the price of other food commodities (like aquaculture-derived fish) that rely on fishmeal for their production (Merino et al., 2010; Carlson et al., 2017).

Any decrease in fish catches in EBUS will affect regional food security. For example, coastal fisheries in the Canary Current are an important source of micronutrients to nearby West African countries (Golden et al., 2016) that have particularly high susceptibility to climate change impacts and low adaptive capacity, because of their strong dependence on the fisheries resources, a rapidly growing population and regional conflicts. Decreased small pelagic fish stocks also increase the mortality and reduce reproduction of larger vertebrates such as hake (Guevara-Carrasco and Lleonart, 2008), whales and seabirds (Essington et al., 2015). Impacts on these organisms affect other non-fishing sectors that are dependent on EBUS, such as whale-watching in the California Current, and generally degrade their intrinsic value.

Overall, EBUS have been changing with intensification of winds that drives the upwelling, leading to changes in water temperature and other ocean biogeochemistry (*medium confidence*). Three out of the four major EBUS have shown upwelling intensification in the past 60 years, with strongly increasing trends in ocean acidification and deoxygenation in the two Pacific EBUS in the last few decades (*high confidence*). The expanding oxygen minimum zone in the California EBUS has altered ecosystem structure and and fisheries catches (*medium confidence*). However, the direction and magnitude of observed changes vary among and within EBUS, with uncertainties regarding the driving mechanisms behind this variability. Moreover, the high natural variability of EBUS and their insufficient representation by global ESMs gives *low confidence* that these observed changes can be attributed to anthropogenic causes, which are predicted to emerge primarily in the second half of the 21st century (*medium confidence*) (Brady et al., 2017). Given the high sensitivity of the coupled human-natural EBUS to oceanographic changes, the future sustainable delivery of key ecosystem services from EBUS is at risk under climate change; those that are most at risk in the 21st century include fisheries (*high confidence*), aquaculture (*medium confidence*), coastal tourism (*low confidence*) and climate regulation (*low confidence*). For vulnerable human communities with a strong dependence on EBUS services and low adaptive capacity, such as those along the Canary Current system (Belhabib et al., 2016; Blasiak et al., 2017), unmitigated climate change effects on EBUS (complicated by other non-climatic stresses such as social unrest) have a high risk of altering their development pathways (*high confidence*).

[END OF BOX 5.3 HERE]

5.4.1.2   Regulating Services

Regulating services are those ecosystem functions, like climate regulation, that allow the environment to be in conditions conducive to human wellbeing and development (Costanza et al., 2017). WGII AR5 concluded that climate change will alter biological, chemical and physical processes in the ocean that provide feedback on the climate system through their effects on atmospheric composition (*high confidence*) (Pörtner et al., 2014). Sections 5.2 and 5.3 consider new evidence since AR5 regarding climate impacts on marine ecosystems and associated risks; their implications for regulating services are examined here.

1
2   A major regulating service provided by marine ecosystems is carbon sequestration. The observed net carbon
3   uptake from the atmosphere to the global ocean varied between 1.0–2.5 GtC yr^{-1} between 2000 and 2012,
4   with a *very likely* uptake of 30-38 Gt of anthropogenic C over the period 1994–2007 (Section 5.2.2.3,
5   Gruber et al., 2019). Estimates of carbon sequestered in the deep ocean range from 0.4 GtC yr^{-1} (Rogers,
6   2015) to 1.6 GtC yr^{-1} (Armstrong et al., 2010) with the annual burial rate (permanent removal to sediment)
7   around 0.2 GtC yr^{-1} (Armstrong et al., 2010).
8
9   Deep-sea ecosystems also contribute to the removal of methane released from the beneath the seabed
10   through microbial anaerobic oxidation and the sequestration of methane-derived carbon in carbonate
11   (Marlow et al., 2014; Thurber et al., 2014). In coastal ecosystems, carbon is biologically sequestered in
12   coastal sediments, commonly known as 'blue carbon' (Section 5.5.1). Tidal wetlands play disproportionately
13   important roles in coastal carbon budgets, forming critical linkages between rivers, estuaries, and oceans
14   (Najjar et al., 2018). Mean carbon storage in the top meter of soil is estimated at 280 MgC ha^{-1} for
15   mangroves, 250 MgC ha^{-1} for salt marshes, and 140 MgC ha^{-1} for seagrass meadows, with long-term rates of
16   carbon accumulation in sediments of salt marshes, mangroves, and seagrasses ranging from 18 to 1713 gC
17   m^{-2} yr^{-1} (Pendleton et al., 2012). These values are, however, highly variable (Section 5.5.1.2). The large
18   space and time scales mean that there is a long time-lag between seafloor change and detectable changes in
19   carbon sequestration. These large lags, in turn render assessment of climate impacts on regulatory services in
20   the deep ocean having *low confidence*.
21
22   Under RCP 2.6, CMIP5 ESMs project a reduced net ocean carbon uptake by 2080, to around 1.0 GtC yr^{-1}.
23   Under RCP 8.5, net ocean carbon uptake increases to a net sink of around 5.5 GtC yr^{-1}, but with variability
24   between models (Lovenduski et al., 2016). Although the open ocean biological pump contributes only part of
25   current carbon uptake (Boyd et al. 2019), the downward carbon flux at 1000 m is projected to decrease by 9-
26   16% globally under RCP 8.5 by 2100. A projected decrease in carbon sequestration in the North Atlantic by
27   27-41% has been estimated to represent a loss of $170–3000 billion USD in abatement (mitigation) costs and
28   $23–401 billion USD in social costs (Barange et al., 2017). Others have highlighted the declining value of
29   open ocean carbon sequestration in the eastern tropical Pacific (Martin et al., 2016b) and the Mediterranean
30   (Melaku Canu et al., 2015). The open ocean therefore seems *very likely* to reduce its carbon uptake by the
31   end of the 21st century, with the reduction *very likely* being greater under RCP 8.5 than for RCP 2.6;
32   however, specific projections only have *medium confidence* due to uncertainties associated with the structure
33   of the models and with the future behaviour of the biological carbon pump (Section 5.2.2.3.1, 5.2.3).
34
35   Coastal blue carbon ecosystems provide climate regulatory services through their carbon removal and
36   storage (Section 5.3.3). The current rates of loss of blue carbon ecosystems, partly due to climate change
37   (Section 5.3) results in release of their stored CO_2 to the atmosphere (Section 5.5.1.2.2). However, increases
38   in carbon sequestration are also possible; for example, temperature-driven displacement of saltmarsh plants
39   by mangrove trees may increase carbon uptake in coastal wetlands (Megonigal et al., 2016). Different rates
40   of sea level rise may have opposite effects, with potential increases in net carbon uptake for slowly rising sea
41   levels (assuming inland habitat migration is possible), but net carbon release for more rapid sea level rise
42   (Figure 5.19). Such contrasting feedbacks between scenarios arise from the different responses of plant
43   biomass, sediment accretion and inundation that control the overall response of vegetated coastal ecosystems
44   to rising sea level (Gonneea et al., 2019). Thus, under high emission scenarios, sea level rise and warming
45   are expected to reduce carbon sequestration by vegetated coastal ecosystems (*medium confidence*); however,
46   under conditions of slow sea level rise, there may be net increase in carbon uptake by some coastal wetlands
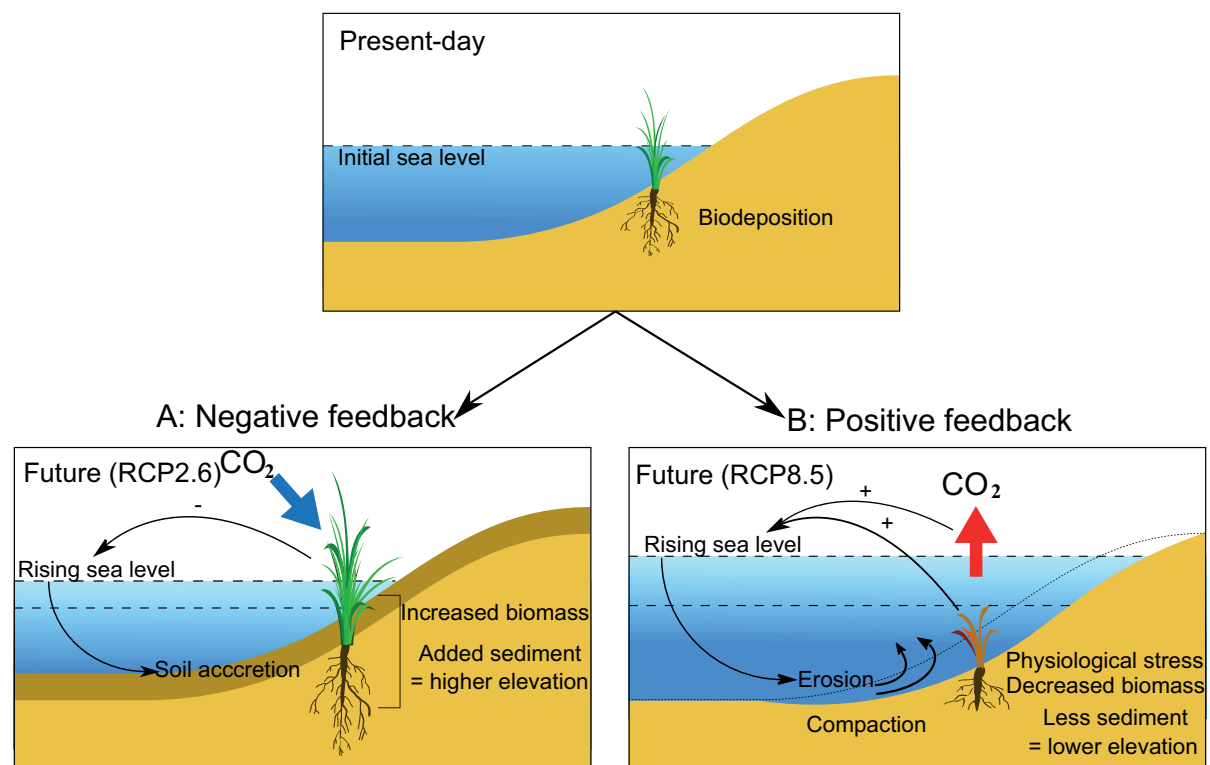47   (*medium confidence*).
48
49

Figure 5.19: Biogeomorphic climate feedbacks involving plant biomass, sediment accretion and inundation that control the response of vegetated coastal ecosystems to rising sea levels. (A) Under high rates of soil formation plants are able to offset gradual sea level rise and may produce a negative feedback by increasing the uptake of atmospheric CO_2. In addition, below ground root production contributes to the formation of new soils and consolidates the seabed substrates. (B) Under low rate of soil formation, and when sea level rises and exceeds critical thresholds, plants become severely stressed by inundation leading to less organic accretion and below ground subsidence and decay producing a positive feedback by net CO_2 outgassing. This figure does not consider landward movements, controlled by topography and human land-use.

Coastal vegetation-rich ecosystems such as mangroves, coral reefs and saltmarshes reduce storm impacts, protect the coastline from erosion, and help buffer the impacts of sea level rise, wave action, and even moderate-sized tsunamis (Orth et al., 2006; Ferrario et al., 2014; Rao et al., 2015) (Section 5.5.2.2). Their loss or degradation under climate change (Sections 5.3) would therefore reduce the benefits of these regulatory services to coastal human communities (Perry et al., 2018), increasing the risk of damage and mortality from natural disasters (Rao et al., 2015) (*high confidence*). In some locations where climate-induced range expansion of coastal wetlands occurs, regulatory services such as storm protection and nutrient storage may be enhanced; however, the replacement of an existing ecosystem by others (e.g. saltmarshes replaced by mangroves) may reduce habitat availability for fauna requiring specific vegetation structure and consequently other types of ecosystem services (Kelleway et al., 2017b; Sheng and Zou, 2017).

5.4.1.3   Supporting Services

Supporting ecosystem services are structures and processes, such as habitats, biodiversity and productivity, that maintain the ecosystem functions that deliver other services (Costanza et al., 2017). Marine supporting services include: primary and secondary production; habitat provision for feeding, spawning or nursery grounds, and refugia; and biodiversity. All these provide essential support for provisioning, regulating or cultural services (Haines-Young and Potschin, 2013; Bopp et al., 2017). Therefore, climate change impacts on supporting services provided by marine ecosystems are directly dependent on the risks and impacts on their biodiversity and ecosystem functions, which are assessed in Sections 5.2.3, 5.2.4 and 5.3. Previously, WGII AR5 highlighted the importance of the potential loss or degradation of habitat forming calcifying algae and corals, and the projected changes in waterways for Arctic shipping (Pörtner et al., 2014). The latter topic is considered in Chapter 3 and Section 5.4.2.4.

1 Publications since AR5 provide further evidence that coastal habitats are at risk from sea level rise, warming
2 and other climate-related hazards (see Section 5.3). All these changes to supporting services have
3 implications for other ecosystem services (Costanza et al., 2014), such as altering fish catches and their
4 composition (Pratchett et al., 2014; Carrasquilla-Henao and Juanes, 2017; Maharaj et al., 2018) (Section
5 5.4.1.1) and carbon sequestration (Section 5.4.1.2). In the epipelagic ocean, climate change affects the
6 pattern and magnitude of global net primary production (Section 5.2.2.6) and the export of organic matter;
7 both these processes support ecosystem services in the deep ocean (Section 5.2.4) and elsewhere. Projected
8 ocean acidification and oxygen loss will also affect deep ocean biodiversity and habitat support linked to
9 provisioning services in the deep ocean (Section 5.2.3.2, 5.2.4). Overall, there is *high confidence* that marine
10 habitat loss and degradation have already impacted supporting services from many marine ecosystems
11 worldwide. The confidence on the attribution of those impacts to climate change depends on the assessment
12 of the ocean and coastal ecosystems (Section 5.2.3, 5.2.4, 5.3). Projected climate-driven alterations of marine
13 habitats will increase the future risks of impacts on supporting services (*high confidence*).

15 *5.4.1.4   Cultural Services*

17 Cultural ecosystem services include recreation, tourism, aesthetic, cultural identity and spiritual experiences.
18 These services are a product of humans experiencing nature and the availability of nature to provide the
19 experiences (Chan et al., 2012). There is increasing evidence to support the conclusion in WGII AR5 that
20 the intrinsic values and cultural importance of marine ecosystems, such as indigenous culture, recreational
21 fishing and tourism, that are dependent on biodiversity and other ecosystem functions, are at risk from
22 climate change. Since marine cultural services are inherently integrated with human wellbeing, their
23 assessment is provided in Section 5.4.2.

25 **5.4.2   Climate Risk, Vulnerability and Exposure of Human Communities and their Wellbeing**

27 Human communities heavily depend on the ocean through the goods and services provided by marine
28 ecosystems (Section 5.4.1) (Hilmi et al., 2015). The values of ocean-based economic activities are estimated
29 to be trillions of USD, generating hundreds of millions of jobs (Hoegh-Guldberg, 2015; Spalding, 2016). As
30 climate change is impacting marine biodiversity and ecosystem services (Section 5.3.1), human communities
31 and their wellbeing will also be affected. This section is based on diverse types of information, from
32 quantitative modelling to qualitative studies, using expert opinion, local knowledge and indigenous
33 knowledge (Cross-Chapter Box 4 in Chapter 1). Projection and assessment of risk and vulnerabilities not
34 only depend on climate change scenarios but are also strongly dependent on scenarios of future social-
35 economic development (Cross-Chapter Box 1 in Chapter 1).

37 This assessment divides the linkages between ecosystem services and human communities and their
38 wellbeing into the three pillars of sustainable development, as used by the World Commission on
39 Environment and Development (WCED). The three pillars are social and cultural, economic and
40 environmental. Table 5.6 lists the specific dimensions under these pillars that are assessed in this section.
41 Synthesis of risks and opportunities of climate change on human communities and wellbeing is at the end of
42 this section through the lens of ocean economy and the United Nations' Sustainable Development Goals.

45 **Table 5.6**: The social, cultural and economic dimensions assessed in Section 5.4.2.

Dimensions	Sections under 5.4.2
Human and environmental health	Water-borne diseases (5.4.2.1.1)
	Harmful algal blooms (Box 5.4)
	Interactions with contaminants (5.4.2.1.2)
	Food security (5.4.2.1.3)
Culture and other social dimensions	Cultural and aesthetic values (5.4.2.2.1)
	Potential conflicts in resource utilization (5.4.2.2.2)
Monetary and material wealth	Fisheries (5.4.2.3.1)
	Coastal and marine tourism (5.4.2.3.2)
	Property values and coastal infrastructure (5.4.2.3.3)

1
2
3   *5.4.2.1   Human Health and Environmental Health*
4
5   *5.4.2.1.1   Water-borne diseases*
6   SR1.5 concluded that climate change will result in an aerial expansion and increased risk of water-borne
7   disease with regional differences (*high to very high confidence*)(Hoegh-Guldberg et al., 2018). WGII AR5
8   concluded that warming, excessive nutrient and seawater inundation due to sea level rise are projected to
9   exacerbate the expansion and threat of cholera (Pörtner et al., 2014) (*medium confidence*). This assessment
10   focuses on health risks caused by *Vibrio* bacteria and Harmful Algal Blooms (HABs). *Vibrio cholerae*
11   (causing cholera) is estimated to be responsible for around 760,000 and 650,000 cases of human illness and
12   death respectively in the world in 2010 (Kirk et al., 2015). An assessment of HABs is given in Box 5.4.

13   *Vibrio* species naturally occur in warm, nutrient-rich and low salinity coastal waters. Since IPCC AR5,
14   analysis of the the Continuous Plankton Recorder dataset (Section 5.2.3) has shown a significant increase in
15   *Vibrio* abundance in the North Sea over the period 1958–2011 related to sea surface warming (Vezzulli et
16   al., 2016). Other time-series data have confirmed a poleward expansion of *Vibrio* pathogens in mid- to high-
17   latitude regions, ascribed at least partly to climate change (Baker-Austin et al., 2013; Baker-Austin et al.,
18   2017). Extreme weather events such as flooding and tropical cyclones are also linked to increased incidences
19   of *Vibrio*-related disease, suggested to be caused by the increased exposure of human populations to the
20   pathogens during these extreme events (Baker-Austin et al., 2017). New evidence since AR5 therefore
21   increases support for the linkages between warming, extreme weather events and increased risk of diseases
22   caused by *Vibrio* bacteria (*very high confidence*).
23
24   Extrapolating from the observed relationship between environmental conditions and current *Vibrio*
25   distributions, coastal areas that experience future warming, changes in precipitation and increases in nutrient
26   inputs can be expected to see an increase in prevalence of *Vibrio* pathogens. These effects have been
27   simulated in a global-scale model that relates occurrences of *Vibrio* with sea surface temperature, pH,
28   dissolved oxygen and chlorophyll *a* concentration under the SRES B1 scenario (Escobar et al., 2015). In the
29   Baltic Sea, a nearly two-fold increase in the area suitable for *Vibrio* is projected between 2015 and 2050 for
30   both RCP 4.5 and RCP 8.5 scenarios (relating to projected SST increase of 4°C to 5°C), resulting in an
31   elevated risk of *Vibrio* infections (Semenza et al., 2017). Projected conditions of increased coastal flooding
32   from storm surges and sea level rise (Section 5.2.2) will also increase exposure to waterborne disease
33   (Ashbolt, 2019), such as *Vibrio* (*medium confidence*). However, uncertainty in the socio-economic factors
34   affecting the future vulnerabilities of human populations render quantitative projections of the magnitude of
35   health impacts uncertain (Lloyd et al., 2016).
36
37
38   [START BOX 5.4 HERE]
39
40   **Box 5.4: Harmful Algal Blooms and Climate Change**
41
42   Harmful algal blooms (HABs) are proliferations of phytoplankton (mostly dinoflagellates, diatoms and
43   cyanobacteria) and macroaglae that have negative effects on marine environments and associated biota.
44   Impacts include water discoloration and foam accumulation, anoxia, contamination of seafood with toxins,
45   disruption of food webs, and massive large-scale mortality of marine biota (Hallegraeff, 2010; Quillien et al.,
46   2012; Quillien et al., 2015; Amaya et al., 2018; García-Mendoza et al., 2018; Álvarez et al., 2019). WGII
47   AR5 concluded that harmful algal outbreaks had increased in frequency and intensity, caused partly by
48   warming, nutrient fluctuations in upwelling areas, and coastal eutrophication (*medium confidence*); however,
49   there was *limited evidence* and *low confidence* for future climate change effects on HABs (WGII AR5
50   Chapters 5, 6) (Pörtner et al., 2014; Wong et al., 2014b). Since AR5, HABs have increasingly affected
51   human society, with negative impacts on food provisioning, tourism, the economy and human health
52   (Anderson et al., 2015; Berdalet et al., 2017). For example, HABs caused an estimated loss of $42 million
53   USD for the tuna industry in Baja California, Mexico (García-Mendoza et al., 2018) and mortality of more
54   than 40,000 tonnes of cultivated salmon in Chile (Díaz et al., 2019). This additional observational and
55   experimental evidence has improved detection and attribution of HABs to climate change, demonstrating

1   that shifts in biogeography, increased abundance and increased toxicity of HABs in recent years have been
2   partly or wholly caused by warming and by other, more direct human drivers.
3
4   New studies since AR5 show range expansion of warm-water HAB species, such as *Gambierdiscus* that
5   causes ciguatera fish poisoning (Kohli et al., 2014; Bravo et al., 2015; Sparrow et al., 2017); contraction of
6   cold-water species (Tester et al., 2010; Rodríguez et al., 2017); the detection of novel phycotoxins and toxic
7   species (Akselman et al., 2015; Guinder et al., 2018; Paredes et al., 2019; Tillmann et al., 2019); and
8   regional increases in the occurrence and intensity of toxic phytoplankton blooms (McKibben et al., 2017;
9   Díaz et al., 2019) in relation to ocean warming. For example, growth of the toxic dinoflagellates
10   *Alexandrium* and *Dinophysis,* producers of paralytic shellfish poisoning and okadaic acid, respectively, is
11   enhanced by warmer conditions in the North Atlantic and North Pacific (Gobler et al., 2017), whilst
12   environmental conditions linked with warm phases of ENSO are associated with blooms of toxic *Pseudo-*
13   *nitzschia* species in the Northern California Current (McKibben et al., 2017), with devastating effects on
14   coastal ecosystems (McCabe et al., 2016; Ritzman et al., 2018). Regional variations of trends in HAB
15   occurrences can be explained by spatial differences in climate drivers (temperature, water column
16   stratification, ocean acidification, precipitation and extreme weather events), as well as non-climatic drivers,
17   such as eutrophication and pollution (Hallegraeff, 2010; Hallegraeff, 2016; Glibert et al., 2018; Paerl et al.,
18   2018a).
19
20   Experimental studies have provided additional evidence for the role of environmental drivers in inducing
21   HABs and their degree of impact. These studies include those showing that toxin production can be affected
22   by grazers (Tammilehto et al., 2015; Xu and Kiørboe, 2018) and changing nutrient levels (Van de Waal et
23   al., 2013; Brunson et al., 2018). The biosynthesis of domoic acid by some *Pseudo-nitschia* species is induced
24   by combined phosphate limitation and high CO_2 conditions (Brunson et al., 2018), with their growth and
25   toxicity enhanced by warming in incubation experiments (Zhu et al., 2017). Recent mesocosm experiments
26   using natural subtropical planktonic communities found that simulated CO_2 emission scenarios (between
27   RCP 2.6 and RCP 8.5 by 2100) improved the competitive fitness of the toxic microalgae *Vicicitus globosus*
28   for CO_2 treatments above 600 μatm, and induced blooms above 800 μatm, with severe negative impacts for
29   other components of the planktonic food web (Riebesell et al., 2018). Experiments with the toxic
30   dinoflagellate *Akashiwo sanguinea* (hemolytic activity) have also shown that a combination of high CO_2
31   levels, warming and high irradiance stimulate the growth and toxicity of this HAB species (Ou et al., 2017).
32
33   Given the worldwide distribution of the key toxic species of *Alexandrium*, *Pseudo-nitzschia* and *Dinophysis*,
34   if the current relationship between warming and the occurrences of HABs associated with these species
35   persists in the future (Gobler et al., 2017; Townhill et al., 2018) (*medium confidence*), the projected changes
36   in ocean conditions can be expected to intensify HAB-related risks for coastal biodiversity and ecosystems
37   services *(high confidence)*. The greatest risk is expected for estuarine organisms (Section 5.3.1) because
38   HABs occurrences are stimulated by riverine nutrient loads, and exacerbated by warming and the lower
39   dissolved oxygen and pH in estuarine environments (Gobler and Baumann, 2016; Paredes-Banda et al.,
40   2018).
41
42   Local scale sustained monitoring programs and early warning systems for HABs can alert resource managers
43   and stakeholders of their potential occurrences so that they can take actions (e.g., toxic seafood alerts or
44   relocation of activities) to reduce the impacts of HABs (Anderson et al., 2015; Wells et al., 2015) (*high*
45   *confidence*). There is *limited evidence* in determining the degree to which reduction of non-climatic
46   anthropogenic stressors can reduce risk of HABs (Section 5.5.2), although this approach may be effective in
47   some areas (*low confidence*); for example, controlling nutrient inputs from human sources may reduce the
48   risk of occurrence of HABs in the Baltic Sea. Other techniques such as active chemical and biological
49   interventions are at experimental stage.
50
51   Overall, the occurrence of HABs, their toxicity and risk on natural and human systems are projected to
52   continue to increase with warming and rising CO_2 in the 21st century(Glibert et al., 2014; Martín-García et
53   al., 2014; McCabe et al., 2016; Paerl et al., 2016; Gobler et al., 2017; McKibben et al., 2017; Rodríguez et
54   al., 2017; Paerl et al., 2018a; Riebesell et al., 2018) *(high confidence)*. Moreover, poleward distributional
55   shifts of HAB species are expected to continue as a result of warming (Townhill et al., 2018). The increasing
56   likelihood of occurrences of HABs under climate change also elevates their risks on ecosystem services such

1   as fisheries, aquaculture and tourism as well as public health (Section 5.4.2, *high confidence*). Such risks will
2   be greatest in poorly monitored areas (Borbor-Córdova et al., 2018; Cuellar-Martinez et al., 2018)
3
4   [END BOX 5.4 HERE]
5
6
7   *5.4.2.1.2   Interactions between climate change and contaminants*
8   Climate change–contaminant interactions can alter the bioaccumulation and amplify biomagnification of
9   several contaminant classes (Boxall et al., 2009; Alava et al., 2018). This section assesses two types of
10   contaminants that are of concern to environmental and human health as examples of other contaminants with
11   similar properties (Alava et al., 2017). These two types of contaminants are the toxic and fat-soluble
12   persistent organic pollutants (POPs), such as polychlorinated biphenyls (PCBs), as well as the neurotoxic
13   and protein-binding organic form of mercury, methylmercury (MeHg) (Alava et al., 2017). POPs and MeHg
14   are bioaccumulated by marine organisms and biomagnified in food webs, reaching exposure concentrations
15   that become harmful and toxic to populations of apex predators such as marine mammals (Desforges et al.,
16   2017; Desforges et al., 2018) (Figure 5.20). Human exposure to POPs and MeHg can lead to serious health
17   effects (Ishikawa and Ikegaki, 1980; UNEP, 2013; Fort et al., 2015; Scheuhammer et al., 2015).
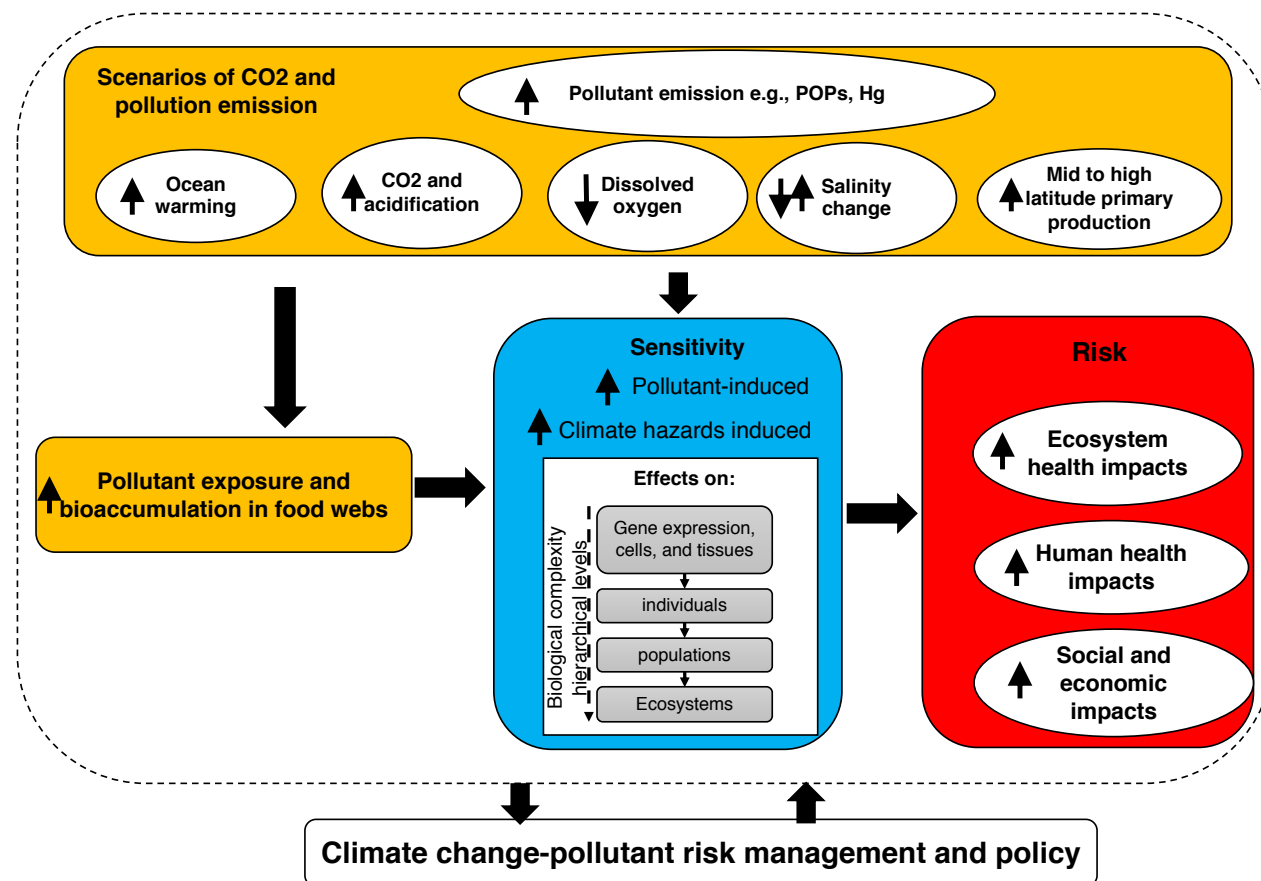18
19

20
21   **Figure 5.20**: The pathways through which scenario of climatic and pollutant hazards (orange boxes) and their
22   interactions can lead to increases in exposure to hazards by the biota, ecosystems and people, sensitivity, their
23   sensitivity (blue box) and the risk of impacts to ecosystem and human health and societies (red box). Such risks will
24   interact with climate-pollutant risk management and policy. The synthesis is based on literature review presented in
25   Alava et al. (2017). Adapted from Alava et al. (2017).
26
27
28   Inorganic forms of mercury are more soluble in low pH water, while higher temperature increases mercury
29   uptake and the metabolic activity of bacteria, thereby increasing mercury methylation, uptake by organisms
30   and bioaccumulation rates (Scheuhammer, 1991; Celo et al., 2006; López et al., 2010; Macdonald and
31   Loseto, 2010; Riget et al., 2010; Corbitt et al., 2011; Krabbenhoft and Sunderland, 2013; Roberts et al.,
32   2013; de Orte et al., 2014; McKinney et al., 2015), although there is *limited evidence* on the extent of

1   exacerbation by ocean acidification expected in the 21st century. Increased melting of snow and ice from
2   alpine ecosystems and mountains (Chapter 2) can also increase the release of POPs and MeHg from land-
3   based sources into coastal ecosystems (Morrissey et al., 2005). Modelling projections for the Faroe Islands
4   region suggest increased bioaccumulation of methyl mercury under climate change, with an average
5   increases in MeHg concentrations in marine species of 1.6% to 1.8% and 4.1% to 4.7% under ocean
6   warming scenarios of +0.8°C and +2.0°C, respectively, with an associated increase in potential human intake
7   of mercury beyond levels recommended by the World Health Organization (Booth and Zeller, 2005).
8   Foodweb modeling for the northeastern Pacific projects that concentrations of MeHg and PCBs in top
9   predators could increase by 8% and 3%, respectively, by 2100 under RCP 8.5 relative to current levels
10   (Alava et al., 2018). Climate-related pollution risks are of particular concern in Arctic ecosystems and their
11   associated indigenous communities because of the bioaccumulation of POPs and MeHg, causing long-term
12   contamination of traditional seafoods (Marques et al., 2010; Tirado et al., 2010; Alava et al., 2017) of high
13   dietary importance (Cisneros-Montemayor et al., 2016).
14
15   Overall, climate change can increase the exposure and bioaccumulation of contaminants and thus the risk of
16   impacts of POPs and MeHg on marine ecosystems and their dependent human communities as suggested by
17   indirect evidence and model simulations (Marques et al., 2010; Tirado et al., 2010; Alava et al., 2017) (*high
18   agreement*). However, there is *limited evidence* on observed increase in POPs and MeHg due to climate
19   change. Apex predators and human communities that consume them, including Arctic communities and
20   other coastal indigenous populations, are thus vulnerable to increase in exposure to these contaminants and
21   the resulting health effects (*medium evidence, medium agreement*).
22
23   The risk of microplastics has become a major concern for the ocean as they are highly persistent and have
24   accumulated in many different marine environments, including the deep sea (Woodall et al., 2014a; Woodall
25   et al., 2014b; GESAMP, 2015; van Sebille et al., 2015; Waller et al., 2017; de Sá et al., 2018; Everaert et al.,
26   2018; Botterell et al., 2019). There is *limited evidence* at present to assess their risk to marine ecosystems,
27   wildlife and potentially humans through human consumption of seafood under climate change.
28
29   *5.4.2.1.3   Food security*
30   Seafood provides protein, fatty acids, vitamins and other micronutrients essential for human health such as
31   iodine and selenium (Golden et al., 2016). Over 4.5 billion people in the world obtain more than 15% of
32   their protein intake from seafood, including algae and marine mammals as well as fish and shellfish (Béné et
33   al., 2015; FAO, 2017)). Around 1.39 billion people obtain at least 20% of their supply of essential
34   micronutrients from fish (Golden et al., 2016). IPCC SR1.5 concluded that global warming poses large risks
35   to food security globally and regionally, especially in low-latitude areas, including fisheries (*medium
36   confidence*) (Hoegh-Guldberg et al., 2018). This section builds on the assessment on observed and projected
37   climate impacts on fish catches (Section 5.4.1.1) and further assess how such impacts interact with other
38   climatic and non-climatic drivers in affecting food security through fisheries.
39
40   Many of populations that are already facing challenges in food insecurity reside in low latitude regions such
41   as in the Pacific Islands and West Africa where maximum fisheries catch potential is projected to decrease
42   under climate change security (Golden et al., 2016; Hilmi et al., 2017) (Section 5.4.1; Figure 5.21) and
43   where land-based food production is also at risk (Blanchard et al., 2017) (*medium confidence*). Populations
44   in these regions are also estimated to have the highest proportion of their micronutrient intake relative to the
45   total animal sourced food (Golden et al., 2016). (Figure 5.21). This highlights their strong dependence on
46   seafood as a source of nutrition that further elevates their vulnerability to food security from climate change
47   impacts on seafood supply (*high confidence*). Modeling of seafood trade networks suggests that Central and
48   West African nations are particularly vulnerable to shocks from decrease in seafood supply from
49   international imports; thus their climate risks of seafood insecurity could be exacerbated by climate impacts
50   on catches and seafood supply elsewhere (Gephart et al., 2016). In addition, experimental studies suggest
51   that warming and ocean acidification reduce the nutritional quality of some seafood by reducing levels of
52   protein, lipid and omega-3 fatty acids (Tate et al., 2017; Ab Lah et al., 2018; Lemasson et al., 2019).
53
54   Non-climatic factors may exacerbate climate effects on seafood security. Over-exploitation of fish stocks
55   reduces fish catches (Section 5.4.1.1) (Golden et al., 2016), whilst strong cultural dependence on seafood in
56   many coastal communities may pose constraints in their adaptive capacity to changing fish availability
57   (Marushka et al., 2019). The shift from traditional nutritious wild caught seafood-based diets of coastal

1 indigenous communities, towards increased consumption of processed energy-dense foods high in fat,
2 refined sugar, and sodium, due to social and economic changes (Kuhnlein and Receveur, 1996; Shannon,
3 2002; Charlton et al., 2016; Batal et al., 2017), has important consequences on diet quality and nutritional
4 status (Thaman, 1982; Quinn et al., 2012; Luick et al., 2014). This has led to an increased prevalence of
5 obesity, diabetes, and other diet-related chronic diseases (Gracey, 2007; Sheikh et al., 2011) as well as the
6 related decrease in access to culturally or religiously significant food items. The risk of climate change on
7 coastal communities through the ocean could therefore be increased by non-climatic factors such as
8 economic development, trade, effectiveness of resource governance and cultural changes (*high confidence*).
9
10 In summary, the food security of many coastal communities, particularly in low latitude developing regions,
11 is vulnerable to decreases in seafood supply (*medium confidence*) because of their strong dependence on
12 seafood to meet their basic nutritional requirements (*medium confidence*), limited alternative sources of
13 some of the essential nutrients obtained from seafood (*medium confidence*), and exposure to multiple
14 hazards on their food security (*high confidence*). Although direct evidence from attribution analysis is not
15 available, climate change may have already contributed to malnutrition by decreasing seafood supply in
16 these vulnerable communities (*low confidence*) and reduce coastal Indigenous communities' reliance on
17 seafood-based diets (*low confidence*). Projected decreases in potential fish catches in tropical areas (*high*
18 *confidence*) and a possible decrease in the nutritional content of seafood (*low confidence*) will further
19 increase the risk of impacts on food security in low latitude developing regions, with that risk being greater
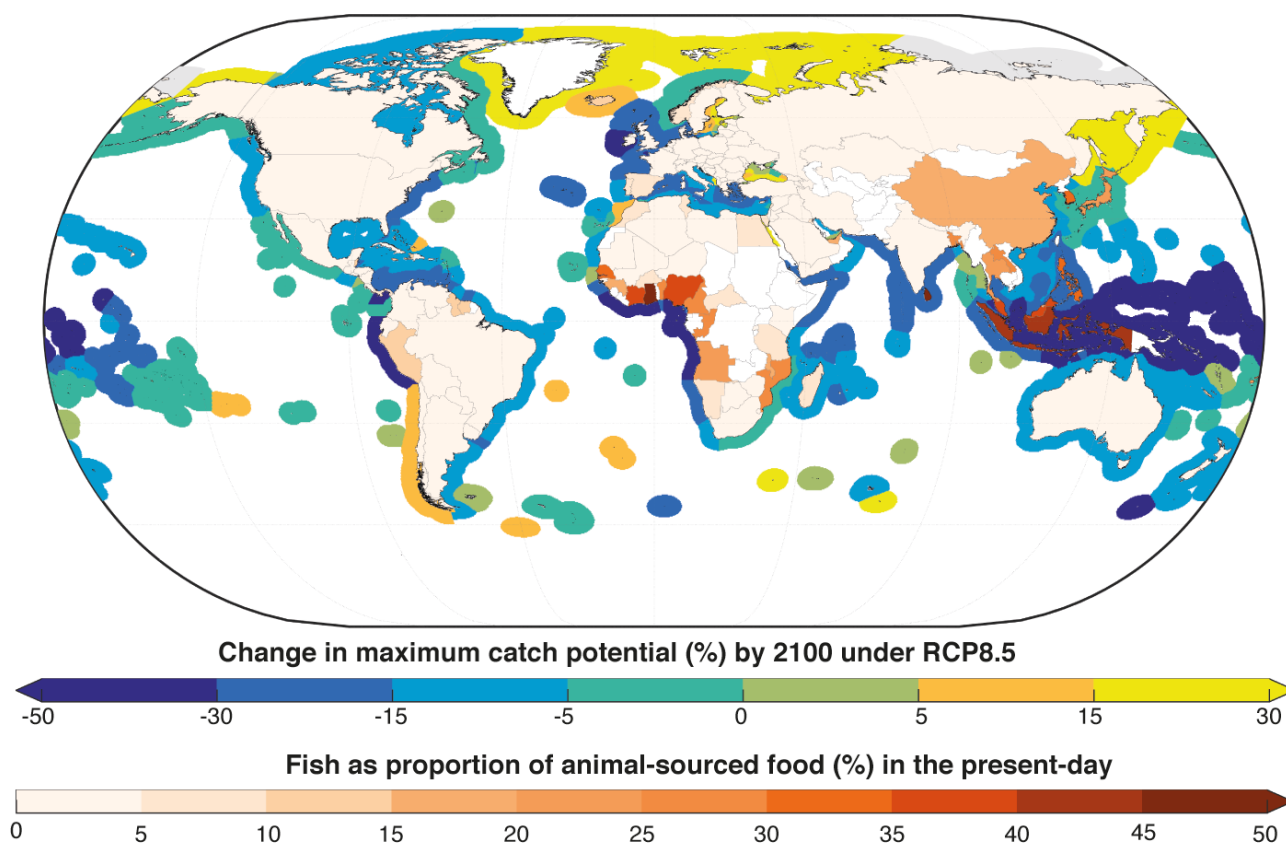20 under high emission scenarios (*medium confidence*).
21
22

Change in maximum catch potential (%) by 2100 under RCP8.5

| -50 | -30 | -15 | -5 | 0 | 5 | 15 | 30 |

Fish as proportion of animal-sourced food (%) in the present-day

| 0 | 5 | 10 | 15 | 20 | 25 | 30 | 35 | 40 | 45 | 50 |

Figure 5.21: Over the ocean the projected changes in catch potential (Section 5.4.1.1), and on land, each countries current proportion of fish micronutrient intake relative to the total animal sourced food (ASF) (Golden et al., 2016). The colour scale on land is the proportion of fish micronutrient intake relative to the total animal sourced food (ASF); the scale on the ocean is projected change in maximum catch potential under RCP8.5 by 2100 relative to the 2000s.

30 *5.4.2.2   Cultural and Other Social Dimensions*

32 *5.4.2.2.1   Cultural and aesthetic values*
33 Climate change threatens key cultural dimensions of lives and livelihoods (Adger et al., 2012), because
34 people develop strong cultural ties and associate distinctive meanings with many natural places and biota in

1   the form of traditions, customs and ways of life (Marshall et al., 2018). These impacts have been felt both by
2   indigenous and non-indigenous peoples. Recent estimates suggest that there are more than 1900 indigenous
3   groups along the coastline with around 27 million people across 87 countries (Cisneros-Montemayor et al.,
4   2016). WGII AR5 concluded that climate change will affect the harvests of marine species with spiritual and
5   aesthetic importance to indigenous cultures (Pörtner et al., 2014). This section further assesses the effects of
6   climate change on indigenous knowledge and local knowledge and their transmission and the implication for
7   well-being of people, complementing the assessment for Arctic indigenous people in Chapter 3.
8
9   Indigenous knowledge is passed and appreciated over timeframes ranging from several generations to a few
10 centuries (Cross Chapter Box 4 in Chapter 1). The adjustment of the transmission and the network of
11 indigenous knowledge on the ocean and coasts, and related perceptions and practice, implies a reworking of
12 these knowledge systems where the individuals and the groups are actors in a narrative and historical
13 construction (Roué, 2012; Alderson-Day et al., 2015). Sea level rise is already transforming the seascape,
14 such as the shape of shores in many low-lying islands in the Pacific, leading to modification or
15 disappearance of geomorphological features that represent gods and mythological ancestors (Camus, 2017;
16 Kench et al., 2018). These changing seascape also affects the mobility of people and residence patterns, and
17 consequently, the structure and transmission of indigenous knowledge (Camus, 2017). The fear of sea level
18 rise and climate change encourage security measures and the grouping of local people to the safest places,
19 contributing to the erosion of indigenous culture and their knowledge about the ocean (Bambridge and Le
20 Meur, 2018), and impairment of opportunity for social elevation for some Pacific indigenous communities
21 (Borthwick, 2016).
22
23 Climate change is also projected to shift the biogeography and potential catches of fishes and invertebrates
24 (5.2.3.1, 5.3, 5.4.1.1) that form an integral part of the culture, economy and diet of many indigenous
25 communities, such as those situated along the Pacific Coast of North America (Lynn et al., 2013).
26 Indigenous fishing communities that depend on traditional marine resources for food and economic security
27 are particularly vulnerable to climate change through reduced capacity to conduct traditional harvests
28 because of reduced access to, or availability of, resources (Larsen et al., 2014; Weatherdon et al., 2016).
29 Overall, the transmission of indigenous culture and knowledge is at risk because of sea level rise affecting
30 sea- and land-scapes, the availability and access to culturally important marine species, and communities'
31 reliance on the ocean for their livelihood and their cultural beliefs (*low confidence*). Strong attachment to
32 traditional marine-based livelihoods has also been reported for non-indigenous communities in Canada
33 (Davis, 2015), the USA (Paolisso et al., 2012), Spain (Ruiz et al., 2012) and Australia (Metcalf et al., 2015).
34 Reduction in populations of fish species that have supported livelihoods for generations, and deteriorations
35 of iconic elements of seascapes are putting the well-being of these communities at risk (*high confidence*).
36
37 Other cultural values supported by the ocean are diverse. They include education, based on knowledge of
38 marine environments. Such education can increase knowledge and awareness of climate change impacts and
39 the efficacy of their mitigation (Meadows, 2011); it can also influence the extent to which stewardship
40 activities are adopted (von Heland et al., 2014; Wynveen and Sutton, 2015; Bennett et al., 2018), and can
41 help develop new networks between coastal people and environmental managers for the purposes of
42 planning and implementing new adaptation strategies (Wynveen and Sutton, 2015). A critical element in
43 reducing vulnerability to climate change is to educate people that they are an integral part of the Earth
44 system and have a huge influence on the balance of the system. An important marine ecosystem service is to
45 support such education (Malone, 2016). Thus, education can play a pivotal role in how climate change is
46 perceived and experienced, and marine biodiversity and ecosystems play an important role in this. At the
47 same time, climate change impacts on marine ecosystems (Sections 5.2.3, 5.2.4) can affect the role of the
48 ocean in supporting such public education (*medium evidence, high agreement, medium confidence*).
49
50 The aesthetic appreciation of natural places is one of the fundamental ways in which people relate to their
51 environment. IPCC AR5 WGII noted that climate change may impact marine species with aesthetic
52 importance that affect local and indigenous cultures, local economies and challenge cultural preservation
53 (Pörtner et al., 2014). Evidence since AR5 confirms that aesthetically-appreciated aspects of marine
54 ecosystems are important for supporting local and international economies (especially through tourism),
55 human wellbeing, and stewardship. For example, Marshall et al. (2018) found that aesthetic values are a
56 critically important cultural value for all cultural groups, and are important for maintaining sense of place,
57 pride, identity and opportunities for inspiration, spirituality, recreation and wellbeing. However, climate

1   change induced degradation and loss of biodiversity and habitats (Section 5.2.3, 5.2.4, 5.3) can also
2   negatively impact the ecosystem features that are currently appreciated by human communities, such as coral
3   reefs, mangroves, charismatic species (such as some marine mammals and seabirds) and geomorphological
4   features (e.g., sandy beaches). There are also aesthetic and inspirational values of marine biodiversity and
5   ecosystems that are important to the psychological and spiritual well-being of people, including film,
6   literature and art, and recreation (Pescaroli and Magni, 2015). Other cultural dimensions that are becoming
7   more widely acknowledged as potentially disturbed by climate change include the appreciation of scientific,
8   artistic, spiritual, and health opportunities, as well as appreciation of biodiversity, lifestyle and aesthetics
9   (Marshall et al., 2018). Thus, climate change may also affect the way in which marine ecosystems support
10 human well-being through cultural dimensions. However, the difficulties in evaluating the importance of
11 aesthetic aspects of marine ecosystems, and in detecting and attributing of climate change impacts, result in
12 such assessment having *low confidence*.
13
14 Climate change affects human cultures and well-being differently. For example, (Marshall et al., 2018)
15 assessed the importance of identity, pride, place, aesthetics, biodiversity, lifestyle, scientific value and
16 wellbeing within the Great Barrier Reef region by 8,300 people across multiple cultural groups. These
17 groups included indigenous and non-indigenous local residents, Australians (non-local), international and
18 domestic tourists, tourism operators, and commercial fishers. They found that all groups highly rated all
19 (listed) cultural values, suggesting that these values are critically associated with iconic ecosystems. Climate
20 change impacts upon the Great Barrier Reef, through increased temperatures, cyclones and sea level rise that
21 cumulatively degrade the quality of the Reef, are therefore liable to result in cultural impacts for all groups.
22 However, survey that assess the emotional responses to degradation of the Great Barrier Reef by similar
23 stakeholder groups reported different levels of impacts among these groups (Marshall et al., 2019).
24 Therefore, many ocean and coastal dependent communities value marine ecosystems highly and climate
25 impacts can affect their well-being, although the sensitivity to such impacts can vary among stakeholder
26 groups (Marshall et al., 2019) (*low confidence*).
27
28 Climate change may alter the environment too rapidly for cultural adaptation to keep pace. This is because
29 the culture that forms around a natural environment can be so integral to people's lives that disassociation
30 from that environment can induce a sense of disorientation and disempowerment (Fisher and Brown, 2015).
31 The adaptive capacity of people to moderate or influence cultural impacts, and thereby reduce vulnerability
32 to such impacts, is also culturally determined (Cinner et al., 2018). For example, when a resource user such
33 as a fisher, farmer, or forester is suddenly faced with the prospect that their resource-based occupation is no
34 longer viable, they lose not only a means of earning an income but also an important part of their identity
35 (Marshall et al., 2012; Tidball, 2012). Loss of identity can, in turn, have severe economic, psychological, and
36 cultural impacts (Turner et al., 2008). Climate change can quickly alter the quality of, or access to, a natural
37 resource through degradation or coastal inundation, so that livelihoods and lifestyles are no longer able to be
38 supported by that resource. When people are displaced from places that they value, there is strong evidence
39 that their cultures are diminished, and in many cases endangered. There are no effective substitutions for, or
40 adequate compensation for, lost sites of significance (Adger et al., 2012). As sensitivity marine ecosystems
41 such as coral reefs and kelp forest are impacted by climate change at rapid rate (Section 5.3), these can lead
42 to the loss of part of people's cultural identity and values beyond the rate at which identify and values can be
43 adjusted or substituted (*medium confidence*).
44
45 *5.4.2.2.2   Potential conflicts in resource utilization*
46 Redistribution of marine species in response to direct and indirect effects of climate change may also disrupt
47 existing marine resource sharing and associated governance (Miller and Russ, 2014; Pinsky et al., 2018).
48 These effects have contributed to disputes in international fisheries management for North Atlantic mackerel
49 (Spijkers and Boonstra, 2017) and Pacific salmon (Miller and Russ, 2014). These disagreements have
50 stressed diplomatic relations in some cases (Pinsky et al., 2018). Decreases and fluctuations in fish stock
51 abundance and fish catches have also contributed to past disputes (Belhabib et al., 2016; Pomeroy et al.,
52 2016; Blasiak et al., 2017). Under climate change, shifts in abundance and distribution of fish stocks are
53 projected to intensify in the 21st century (Sections 5.2.3, 5.3, 5.4.1.1). Stocks may locally increase and
54 decrease elsewhere. New or increased fishing opportunities may be created when exploited fish stocks shift
55 their distribution into a country's waters where their abundance was previously too low to support viable
56 fisheries (Pinsky et al., 2018). The number of new transboundary stocks occurring in exclusive economic
57 zones worldwide was projected to be around 46 and 60 under RCP 2.6 and RCP 8.5, respectively, by 2060

1   relative to 1950-2014 (Pinsky et al., 2018). However, such alteration of the sharing of resources between
2   countries would challenge existing international fisheries governance regimes and, without sufficient
3   adaptation responses, increase the potential for disputes in resource allocation and management (Belhabib et
4   al., 2018; Pinsky et al., 2018). Overall, projected climate change impacts on fisheries in the 21st century
5   increase the risk of potential conflicts among fishery area users and authorities or between two different
6   communities within the same country (Ndhlovu et al., 2017; Shaffril et al., 2017; Spijkers and Boonstra,
7   2017) (*medium confidence*), exacerbated through competing resource exploitation from international actors
8   and mal-adapted policies (*low confidence*). Such risks can be reduced by appropriate fisheries governance
9   responses that are discussed in Sections 5.5.2 and 5.5.3.
10
11   *5.4.2.3   Monetary and Material Wealth*
12
13   *5.4.2.3.1   Wealth generated from fisheries*
14   Global gross revenues from marine fisheries were around $150 billion in 2010 USD (Swartz et al., 2013; Tai
15   et al., 2017). Capture fisheries provide full-time and part-time jobs for an estimated 260 ± 6 million people
16   in the 2000s period, of whom 22 ± 0.45 million are small-scale fishers (Teh and Sumaila, 2013). Small-scale
17   fisheries are important for the livelihood and viability of coastal communities worldwide (Chuenpagdee,
18   2011). AR5 WGII concluded with *low confidence* that climate change will lead to a global decrease in
19   revenue with regional differences that are driven by spatial variations of climate impacts on and the
20   flexibility and capacities of food production systems (Pörtner et al., 2014). AR5 also highlighted the high
21   vulnerability of mollusc aquaculture to ocean acidification. For example, the oyster industry in the Pacific
22   has lost nearly $110 million USD in annual revenue due to ocean acidification (Ekstrom et al., 2015). This
23   section examines the rapidly growing literature assessing the risks of climate change on fisheries and
24   aquaculture sectors, and the potential interaction between climatic and non-climatic drivers on the economics
25   of fisheries. However, new evidence on observed economic impacts of climate change on fisheries since
26   AR5 is limited.
27
28   Since AR5, projections on climate change impacts on the economics of marine fisheries have incorporated a
29   broader range of social-economic considerations. Driven by shifts in species distributions and maximum
30   catch potential of fish stocks (Section 5.4.1), if the ex-vessel price of catches remains the same, marine
31   fisheries maximum revenue potential are projected to be negatively impacted in 89% of the world's fishing
32   countries under the RCP 8.5 scenario by the 2050s relative to the current status, with projected global
33   decreases of 10.4% ±4.2% and 7.1% ±3.5% under RCP 8.5 and RCP 2.6, respectively, by 2050 relative to
34   2000 (Lam et al., 2016). While the projected changes in revenues are sensitive to price scenarios (Lam et al.,
35   2016), future maximum revenue potential is reduced under high emission scenarios (Sumaila et al., 2019).
36   For example, when the elasticity of seafood price in relation to their supply was modelled explicitly,
37   fisheries maximum revenue potential under a 1.5°C atmospheric warming scenario was projected to be
38   higher than for 3.5°C warming by 7.4% ($13.1 billion USD) ± 2.3%, across projections from three CMIP5
39   models (Sumaila et al., 2019). Accounting for the subsequent impacts on the dependent communities and
40   relative to the 1.5°C warming scenario, that study also projected a decrease in seafood workers' incomes of
41   7.8% (US $3.7 billion) ± 2.3% and an increase in households' seafood expenditure by the global population
42   of 3.2% (US $6.3 billion) ± 3.9% annually under a 3.5°C warming scenario (Sumaila et al., 2019).
43
44   Fisheries management strategies and fishing effort affect the realised catch and economic benefits of fishing
45   (Barange, 2019). Modelling analysis of fish stocks with available data worldwide showed that for RCP 6.0,
46   adaptation of fisheries by accommodating shifts in species distribution and abundance, as well as rebuilding
47   existing overexploited or depleted fish stocks, is projected to lead to substantially higher global profits
48   (154%), harvest (34%), and biomass (60%) in the future, relative to a no adaptation scenario. However, the
49   total profit, harvest and biomass are negatively affected even with the full adaptation scenario under RCP 8.5
50   (Gaines et al., 2018). Overall, climate change impacts on the abundance, distribution and potential catches of
51   fish stocks (see Section 5.3.1) are expected to reduce the maximum potential revenues of global fisheries
52   (*high agreement, medium evidence, medium confidence*). These impacts on fisheries will increase the risk of
53   impacts on the income and livelihoods of people working in these economic sectors by 2050 under high
54   greenhouse gas emission scenarios relative to low emission scenario (*high confidence*). Rebuilding
55   overexploited or depleted fisheries can help improve economic efficiency and reduce climate risk, provided
56   that emissions are greatly reduced (*medium confidence*).
57

1   The economic implications of climate change on fisheries vary between regions and countries because of the
2   differences in exposure to revenue changes and the sensitivity and adaptive capacity of the fishing
3   communities to these changes (Hilmi et al., 2015). Regions where the maximum potential revenue is
4   projected to decrease coincide with areas where indicators such as human development index suggest high
5   economic vulnerability to climate change (Barbier, 2015; Lam et al., 2016). Many coastal communities in
6   these regions rely heavily on fish and fisheries as a major source of animal proteins, nutritional needs,
7   income and job opportunities (FAO, 2018a). Negative impacts on the catch and total fisheries revenues for
8   these countries are expected to have greater implications for jobs, economies, food and nutritional security
9   than the impacts on regions with high Human Development Index (Allison et al., 2009; Srinivasan et al.,
10 2010; Golden et al., 2016; Blasiak et al., 2017). Climate change impacts to coral reefs and other fish habitats,
11 as well as to targeted fish and invertebrate species themselves are expected to reduce harvests from small-
12 scale, coastal fisheries by up to 20% by 2050, and by up to 50% by 2100, under RCP8.5 (Bell et al., 2018a).
13 Therefore, climate risk to communities that are strongly dependent on fisheries associated with ecosystems
14 that are particularly sensitive to climate change such as coral reefs will have be particularly high (Cinner et
15 al., 2016) (*high confidence*).
16
17 Climate change may also worsen non-climate-related socio-economic shocks and stresses, and hence is an
18 obstacle to economic developments (Hallegatte et al., 2015). Climate risk on the economics of fishing is
19 projected to be higher for tropical developing countries where existing adaptive capacity to the risk is lower,
20 thereby challenging their sustainable economic development (*high confidence*). However, observed impacts
21 are not yet well documented (Lacoue-Labarthe et al., 2016) , and there are many uncertainties relating to
22 how climate change would affect the dynamics of fishing costs, with consequent adjustment of fishing effort
23 that might intensify or lessen the overcapacity issue. Studies have attempted to project how fishers may
24 respond to changes in fish distribution and abundance by incorporating different management systems
25 (Haynie and Pfeiffer, 2012; Galbraith et al., 2017). However, the impacts of climate change on management
26 effectiveness and trade practices is still inadequately understood (Galbraith et al., 2017).
27
28 *5.4.2.3.2   Wealth generated from coastal and marine tourism sector*
29 Tourism is one of the largest sectors in the global economy. Between 1995-1998 and 2011-2014, the average
30 total contribution of tourism to global GDP increased from $69 billion USD (6.8%) to $166 billion USD
31 (8.5%) respectively, and generated more than 21 million jobs between 2011–2014 (UNCTAD, 2018).
32 Coastal tourism and other marine-related recreational activities contributes substantially to the tourism sector
33 (Cisneros-Montemayor et al., 2013; O'Malley et al., 2013; Spalding et al., 2017; Giorgio et al., 2018;
34 UNWTO, 2018). For example, it is estimated that around 121 million people a year participated in marine-
35 based recreational activities, generating $47 billion 2003USD in expenditures and supporting one million
36 jobs (Cisneros-Montemayor and Sumaila, 2010). Tourism is one of the main industries that provides
37 opportunities for social and economic development (Jiang and DeLacy, 2014), and marine tourism is
38 particularly important for many coastal developing countries and Small Island Developing States (SIDS).
39 WGII AR5 identified the tourism sector in the Caribbean region as particularly vulnerable to climate change
40 effects, due to hurricanes, whilst IPCC SR1.5 concluded that warming will directly affect climate-dependent
41 tourism markets on a worldwide basis (*medium confidence*) (Hoegh-Guldberg et al., 2018). This assessment
42 provides updates since AR5 and SR1.5.
43
44 Empirical modelling of future risks to tourism is based on projected climate impacts (Section 5.3) for
45 relevant coastal ecosystems, including degradation or loss of beach and coral reef assets (Weatherdon et al.,
46 2016) (Section 4.3.3.6.2). These projections are developed from the relationship between the economic
47 benefits generated from coral reef related tourism with observed characteristics of coral reefs, the
48 characteristics of tourism activities. Based on scenarios of projected future warming and decreases in coral
49 reef coverage, a global loss of tourism and recreation value in the near-future (2031-2050) of $2.57–2.95
50 billion yr^{-1} 2000USD is projected under RCP 2.6, and of $3.88–5.80 billion yr^{-1} 2000USD under RCP 8.5
51 (Chen et al., 2015). Opinion surveys in four countries suggest that if severe coral bleaching persists in the
52 Great Barrier Reef, tourism in adjacent areas could greatly decline, from 2.8 million to around 1.7 million
53 visitors per year, equivalent to more than $1 billion AUS (~$0.69 billion USD using exchange rate in 2019)
54 i.e., in tourism expenditure and with potential loss of around 10,000 jobs (Swann and Campbell, 2016).
55
56 Many coastal tourism destinations are exposed to risks of flooding, sea level rise and coastal squeeze on
57 coastal ecosystems (Lithgow et al., 2019) (Section 5.3); there are also other climate related-risks. Droughts,

1   which are projected to be more frequent, will also impact the tourism industry (and local food security)
2   through water and food shortages (Pearce et al., 2018). If climate change and ocean acidification reduce the
3   seafood supply, the attractiveness of coastal regions for tourists will also decrease (Wabnitz et al., 2017).
4   North Atlantic hurricanes and tropical storms have increased in intensity over the last 30 years, with climate
5   projections indicating an increasing trend in hurricane intensity (Chapter 6). Three major Caribbean storms,
6   Harvey, Irma and Maria, occurred in 2017, with loss and damage to the tourism industries of Dominica, the
7   British Virgin Islands, and Antigua and Barbuda estimated at US $2.2 billion, and environmental recovery
8   costs estimated at US $6.8 million (UNDP, 2017). Pacific tourist destinations, which tend to focus on
9   nature-based and marine activities, are also at high risk of extreme events and other climate change impacts
10 (Klint et al., 2015). However, global tourism has a high carbon footprint (flights, cruises, etc.) (Lenzen et al.,
11 2018), so any reduction in the intensity of this sector would help mitigate climate change.
12
13 Evidence from recent studies on projected climate risks on recreational fishing is equivocal, with the
14 direction of impacts depending on the location, species targeted and societal context. For example: poleward
15 range shifts of marine fish (Section 5.2.3) could yield new opportunities for recreational fishing in mid- to
16 high latitude regions (DiSegni and Shechter, 2013); projected increases in air temperature may enable longer
17 fishing days in some area (Dundas and von Haefen, 2015); and extreme events may alter the composition of
18 recreational fishing catches (Santos et al., 2016). Since climate risks to recreational fishing vary largely
19 depending on the responses of the targeted species to climate-related pressures, there is *low confidence* in the
20 overall risk to the activity.
21
22 Overall, evidence since AR5 and SR1.5 confirms that climate impacts to coastal ecosystems would increase
23 risks to coastal tourism, particularly under high emission scenarios (*medium confidence*). Economic impacts
24 will be greatest for those developing countries where tourism is the main source of foreign revenue (*medium*
25 to *high evidence*).
26
27 *5.4.2.3.3   Property values*
28 The integrity of ecosystems and their services can affect the value of human assets, particularly coastal
29 properties and infrastructure (Hoegh-Guldberg et al., 2018). Climate change is expected to have negative
30 impacts on coastal properties and their value through the loss and damage caused by sea level rise, increased
31 storm intensity (hurricanes and cyclones), heat waves, floods, droughts and other extreme events,
32 particularly in tropical SIDS (Chapter 4). Natural disasters already cost Pacific Island Countries and
33 Territories between 0.5– 6.6 % of GDP per year (World Bank, 2017), with localized damages and losses
34 from individual storms far exceeding these estimates (e.g., 64% of Vanuatu's GDP for Cyclone Pam in
35 2015). The impacts of natural disasters on Jamaica's coastal transport infrastructure are currently estimated
36 to be a significant proportion of their GDP, and such costs are projected to increase substantially in the next
37 few decades under climate change (UNCTAD, 2017; Monioudi et al., 2018). In 2015, tropical storm Erika
38 devastated Dominica causing $483 million USD in damages and losses (mostly related to transport, housing
39 and agriculture), equivalent to 90% of Dominica's GDP (World Bank, 2017). For the USA, Ackerman and
40 Stanton (2007) forecast that annual real estate losses due to climate change could increase from 0.17% of
41 GDP in 2025 to 0.36% in 2100, with Atlantic and Gulf Coast states being the most vulnerable. Other North
42 American studies have shown that informed coastal property owners are willing to initially invest in
43 infrastructure to counter climate change impacts (McNamara and Keeler, 2013); however, they would avoid
44 further investment if adaptation costs increase substantially and there are greater risks of long-term impacts
45 (Putra et al., 2015).
46
47 The impacts of changing marine ecosystems and ecosystem services on the value of human assets need to
48 consider the risk perception, future development and adaptation responses of human communities (Section
49 5.5.2, Chapter 4) (Bunten and Kahn, 2014). For example, the potential for climate impacts on the value of
50 coastal real estate will depend on the changing insurance market or the cost of adaptation measures, which in
51 turn depend on the willingness to pay by asset holders and wider society, including local and national
52 governments. Further research is needed to discount valuations for potential losses that may occur in the
53 future but with uncertain occurrence, and to improve real estate loss estimates over local to regional scales.
54
55 Marine ecosystem services contribute to climate moderation and coastal defenses (5.4.1.2). However, while
56 the above studies in this section acknowledge the contribution of many climate impacts on real estate and
57 infrastructure through ecosystem losses and degradation, often they are not accounted for in quantitative

1   economic impact assessments. Overall, there is *high confidence* that sea level rise, increases in storm
2   intensity and other extreme events will impact the values of coastal real estates and infrastructure,
3   particularly in tropical SIDS, through the risk and impacts of direct physical damages. However, there is *low*
4   *confidence* that impacts due to underlying loss and damage of ecosystems and their services are being
5   similarly accounted for.
6
7   *5.4.2.4   Risk and Opportunities for Ocean Economy*
8
9   The 'ocean economy' refers to the sustainable use of ocean resources for economic growth, improved
10   livelihoods and jobs, and ocean ecosystem health (World Bank, 2017). In SR1.5 (Hoegh-Guldberg et al.,
11   2018) and elsewhere here (Chapters 3 and 5), the risks and opportunities of specific sectors that contribute to
12   the ocean economy under climate change are assessed. The fishing industry is particularly important in this
13   context. As previously noted, warming has already directly impacted coastal and open ocean fishing
14   activities in some regions (5.4.1.1, 5.4.2.3.1); the risk of fishery impacts is exacerbated by the observed
15   climate-driven changes to coral reefs and other coastal ecosystems that contribute to the productivity of
16   exploited fish species (5.4.1.3, 5.4.2.3.1); and there are challenges to sustainable management of
17   transboundary fisheries resources caused by species' range shifts and associated governance challenges
18   (5.4.2.2.2)..
19
20   Fisheries-related national and local economies of many tropical developing countries are exposed high
21   climate risks (5.4.2.3.1) (Blasiak et al., 2017), as a result of the projected large decrease in maximum catch
22   and revenue potential under RCP 8.5 in the 21st century (5.4.1.1). Historical examples from fishery over-
23   exploitation indicate that a large decrease in catches for specific fish stocks have had substantial negative
24   effects for dependent economies and communities (Brierley and Kingsford, 2009; Davis, 2015). Moreover,
25   coastal economies that are dependent on marine tourism and recreational activities are also exposed to
26   elevated risks from impacts on biota that are important for these sectors (5.4.2.3.2). Nevertheless, new
27   opportunities for coastal tourism may occur in future for some regions as a result of species' biogeographic
28   shifts (5.4.2.3.2) and increased accessibility, such as in the Arctic (Chapter 3).
29
30   Decrease in sea ice in the Arctic is opening up economic opportunities for the oil and gas exploration,
31   mining industries and shipping that are currently important economic sectors in the ocean (Pelletier and Guy,
32   2012; George, 2013) (Section 3.4.3; 3.5.3). Although the Arctic region has oil and gas reserves estimated to
33   account for one-tenth of world oil and a quarter of global gas (*U.S. Geological Survey released on 24 July*
34   *2008*), offshore oil and gas exploration with poor regulation or as a result of accidents poses additional risk
35   of impacts on species, populations, assemblages, to ecosystems by modifying a variety of ecological
36   parameters (e.g., biodiversity, biomass, and productivity) (Cordes et al., 2016) threatening the sensitive
37   Arctic ecosystems and the livelihood of dependent communities (Section 3.5.3.3).
38
39   Similarly, global warming and changing weather patterns may have a substantial impact on global trade and
40   transport pathways (Koetse and Rietveld, 2009); for example, the reduction in sea ice in the Arctic Ocean
41   during summer opens up the possibility for sea transport on the Northwest or Northeast Passage for several
42   months per year (Ng et al., 2018) (Section 3.5.3.2). Both routes may provide opportunities for more efficient
43   transport between North America, Europe, Russia and China for fleets with established Arctic equipment,
44   and may open up access to known natural resources which have so far been covered by ice (Guy and
45   Lasserre, 2016). However, whether the Arctic shipping routes will be a realistic alternative depends not only
46   on regulatory frameworks and economic aspects (such as infrastructure and reliability of the routes) but also
47   on societal trends and values, demographics, and tourism demand (Prowse et al., 2009; Wassmann et al.,
48   2010; Pelletier and Guy, 2012; George, 2013; Hodgson et al., 2016; Pizzolato et al., 2016; Dawson, 2017)
49   (Section 3.2.4.2, 3.4.3.3). Simultaneously, shipping routes through the Arctic pose additional risk from
50   human impact such as pollution, introduction of invasive species and collision with marine mammals, and
51   emission of short-lived climate forcers that can amplify warming in the region and accelerate localized
52   warming (Wan et al., 2016) (Section 3.5.3.2).
53
54   Existing governance may not be sufficient to limit the elevated risk on Arctic ecosystems and their
55   dependent economies from increased shipping activities (Section 3.4.3, 3.5.3). Climate change may bring
56   new economic opportunities, particularly for polar oil and gas development (*medium confidence*), shipping
57   (*medium confidence*) and tourism (*low confidence*) although realization of these opportunities will pose

1   uncertain ecological risks to sensitive ecosystems and biota, and the dependent human communities in the
2   region (*high confidence*).
3
4   Ocean renewable energy provides an emerging alternative to fossil fuels and comprises energy extraction
5   from offshore winds, tides, waves, ocean thermal gradients, currents and salinity gradients (Harrison and
6   Wallace, 2005; Koetse and Rietveld, 2009; Bae et al., 2010; Jaroszweski et al., 2010; O Rourke et al., 2010;
7   Hooper and Austen, 2013; Kempener and Neumann, 2014b; Kempener and Neumann, 2014a; Abanades et
8   al., 2015; Astariz et al., 2015; Borthwick, 2016; Foteinis and Tsoutsos, 2017; Manasseh et al., 2017; Becker
9   et al., 2018; Gattuso et al., 2018; Hemer et al., 2018; Dinh and McKeogh, 2019b; Dinh and McKeogh,
10   2019a). Other potential sources of marine renewable energy include algal biofuels (Greene et al., 2010;
11   Greene et al., 2016). While such approaches offers a way to mitigate climate change, changes in climatic
12   conditions (such as waves and winds) may impact marine renewable energy installations and their
13   effectiveness (Harrison and Wallace, 2005). A more comprehensive assessment of these issues is expected
14   to be provided by IPCC WGIII in the AR6 full report.
15
16   Overall, some major existing ocean economy sectors such as fishing, coastal tourism and recreation are
17   already at risk by climate change (*medium confidence*), and all sectors are expected to have elevated risks
18   with high future emission scenarios (*high confidence*). The emerging demand for alternative energy sources
19   is expected to generate economic opportunities for the ocean renewable energy section (*high confidence*),
20   although their potential may also be affected by climate change (*low confidence*).
21
22   *5.4.2.5   Impacts of Changing Ocean on Sustainable Development Goals*
23
24   Climate change impacts will have consequences for the ability of human society to achieve sustainable
25   development. IPCC SR1.5 concludes that "Limiting global warming to 1.5°C rather than 2°C would make it
26   markedly easier to achieve many aspects of sustainable development, with greater potential to eradicate
27   poverty and reduce inequalities (*medium evidence, high agreement*)". This assessment focuses on how
28   climate change impacts on marine ecosystems would challenge sustainable development, using the United
29   Nations Sustainable Development Goals (SDGs) as a framework to discuss the linkages between those
30   issues.
31
32   Climate impacts on marine ecosystems affect their ability to provide seafood and raw materials, and to
33   support biodiversity, habitats and other regulating processes (Section 5.4.1), and these impacts on the ocean
34   affect people directly and indirectly (Sections 5.4.2.1, 5.4.2.2, 5.4.2.3). SDG 14 is the goal that is most
35   directly relevant: "Life below water: including indicators for marine pollution, habitat restoration and
36   protected areas, ocean acidification, fisheries, and coastal development."
37
38   Climate impacts in the ocean to other SDGs are mediated through social and economic factors when the
39   SDG targets are affected (Singh et al. 2019). For example, climate impacts on marine ecosystem services
40   related to primary industries that provide food, income and livelihood to people have direct implications for
41   a range of SDGs. These SDGs include "no poverty" (SDG 1), "zero hunger" (SDG 2), "decent work and
42   economic growth" (SDG 8), "reduced inequalities" (SDG 10) and "responsible consumption and
43   production" (SGD 12) (Singh et al. 2019, Figure 5.22). These impacts relate to changing ocean under climate
44   change that affect the pathways to build sustainable economies and eliminate poverty (Sections 5.4.2.4),
45   eliminate hunger and achieve food security (Section 5.4.2.1.3), reduce inequalities (Sections 5.4.2.2) and
46   achieve responsible consumption and production (Sections 5.4.2.3.1) (Carvalho et al., 2017; Castells-
47   Quintana et al., 2017). Climate change is also creating living conditions in coastal areas that are less suitable
48   to human settlement and changing distributions of marine disease vectors (5.4.2.1.1, 5.4.2.3.3), reducing our
49   chances of achieving the goal for good health and wellbeing (SDG 3) (Pearse, 2017; Wouters et al., 2017).
50   Women are often engaged in jobs and livelihood sources that are more exposed to climate change impacts
51   from the ocean such as impacts on fisheries (Section 5.4.2.3.1) and impacts of sea level rise on coastal
52   regions (Chapter 6). For example, in Senegal, women disproportionately engage in rice crop cultivation in
53   coastal flood plain (Linares, 2009), and are thus exposed to the risks on their livelihood from rising sea
54   levels and resulting salinization (Dennis et al., 1995). Flooding in Bangladesh has increased the vulnerability
55   of women to harassment and abuse as the flooding upends normal life and increases crime rates (Azad et al.,
56   2013). As such, climate change may negatively affect our ability to achieve "gender equality" (SDG 5)
57   (Salehyan, 2008). Impacts on living conditions as well as changing recreational, aesthetic, and spiritual

1 experiences also affect our ability to achieve "sustainable cities and communities" (SDG 11) (Section
2 5.4.2.2.1). The consequences of climate change in the ocean to achieving the remaining SDGs are less clear.
3 However, the SDGs are interlinked, and achieving SDG 14, and especially the targets of increasing
4 economic benefits to Small Island Developing States and Least Developed Countries, as well as eliminating
5 illegal fishing and overfishing, will benefits all other SDGs (Singh et al., 2017). The interlinkages among
6 SDGS mean climate change impact on the ocean will affect all other SDGs beside SDG14 in various ways,
7 some possible direct and many indirect (*low confidence*).
8
9 Overall, climate change impacts on the ocean will negatively affect the chance of achieving the Sustainable
10 Development Goals and sustaining their benefits (*medium confidence*).
11
12

Climate Impacts on Ecosystem Services

13
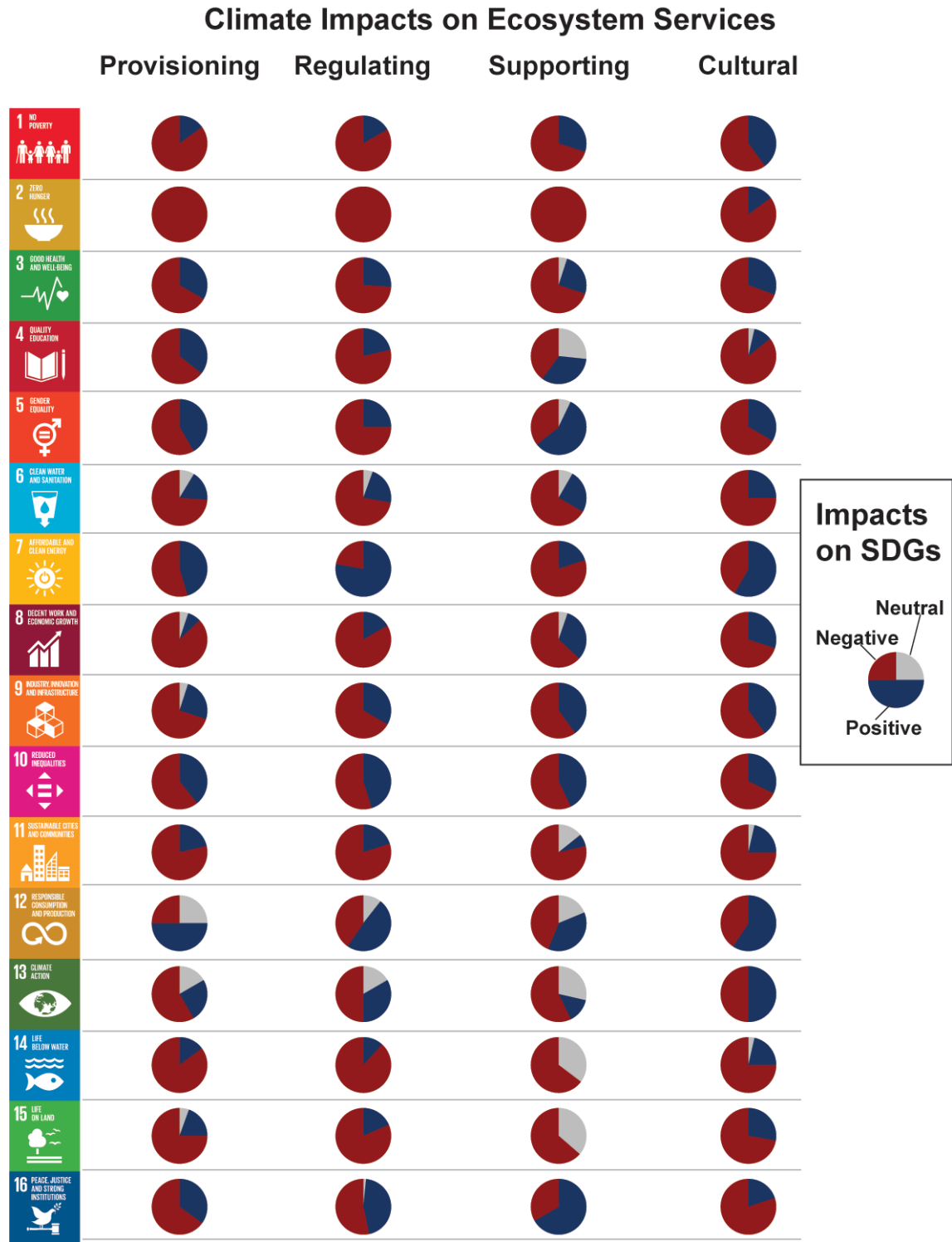
1 **Figure 5.22**: Summary of the types of relationships (negative, neutral and positive) between impacted marine
2 ecosystem services (Provisioning, Regulating, Supporting and Cultural) and the Sustainable Development Goals based
3 on literature review and expert-based analysis (Singh et al., 2019). Pie charts represent the proportion of targets within
4 SDGs that a particular ocean SDG target contributes to according to the literature reviewed and expert-based analysis
5 presented in Singh et al. (2019).
6
7
8 ## 5.5   Risk-reduction Responses and their Governance
9
10 ### 5.5.1   Ocean-based Mitigation
11
12 #### 5.5.1.1   Context for Blue Carbon and Overview Assessment
13
14 There is political and scientific agreement on the need for a wide range of mitigation actions to avoid
15 dangerous climate change (UNEP, 2017; IPCC, 2018). Opportunities to reduce emissions by the greater use
16 of ocean renewable energy are identified in Section 5.4.2.3.2. Here, in accordance with the approved
17 scoping of this report, the assessment of mitigation options is limited to the management of natural ocean
18 processes, i.e. requiring policy intervention, with a focus on 'blue carbon'. Natural processes *per se*,
19 although important to the climate system and the global carbon cycle, are not a mitigation response. Two
20 management approaches are possible: first, actions to maintain the integrity of natural carbon stores, thereby
21 decreasing their potential release of greenhouse gases, whether caused by human or climate-drivers; and
22 second, through actions that enhance the longterm (century-scale) removal of greenhouse gases from the
23 atmosphere by marine systems, primarily by biological means.
24
25 These mitigation approaches match those proposed using terrestrial natural processes (Griscom et al., 2017),
26 with extensive afforestation and reforestation included in all climate models that limit future warming to
27 $1.5°C$ (de Coninck et al., 2018). As on land, reliable carbon accounting is a critical consideration (Grassi et
28 al., 2017), together with confidence in the longterm security of carbon storage. The feasibility of
29 climatically-significant (and societally-acceptable) mitigation using marine natural processes therefore
30 depends on a robust quantitative understanding of how human actions can affect the uptake and release of
31 greenhouse gases from different marine environments, interacting with natural biological, physical and
32 chemical processes. Whilst CO_2 is the most important greenhouse gas, marine fluxes of methane and nitrous
33 oxide can also be important, for both coastal regions and the open ocean (Arévalo-Martínez et al., 2015;
34 Borges et al., 2016; Hamdan and Wickland, 2016).
35
36 The term 'blue carbon' was originally used to cover biological carbon in all marine ecosystems (Nellemann
37 et al., 2009). Subsequent use of the term has focused on carbon-accumulating coastal habitats structured by
38 rooted plants, such as mangroves, tidal saltmarshes and seagrass meadows, that are relatively amenable to
39 management (McLeod et al., 2011; Pendleton et al., 2012; Thomas, 2014; Macreadie et al., 2017a; Alongi,
40 2018; Windham-Myers et al., 2018; Lovelock and Duarte, 2019). Comparisons across the full range of
41 freshwater and saline wetland types are assisted by standardised approaches (Nahlik and Fennessy, 2016;
42 Vázquez-González et al., 2017). Seaweeds (macroalgae) can also be considered as coastal blue carbon
43 (Krause-Jensen and Duarte, 2016; Krause-Jensen et al., 2018; Raven, 2018), however, because of differences
44 in their carbon processing, their climate mitigation potential is assessed separately within Section 5.5.1.2
45 below.
46
47 In the open ocean, the biological carbon pump is driven by the combination of photosynthesis by
48 phytoplankton and downward transfer of particulate carbon by a variety of processes (Henson et al., 2010;
49 DeVries et al., 2017); it results in large-scale transfer of around 10 GtC yr^{-1} carbon from near-surface waters
50 to the ocean interior (Boyd et al., 2019). Most of this carbon is respired in the mesopelagic and contributes to
51 the 37,000 GtC inventory of dissolved inorganic carbon, with around ~0.1 GtC yr^{-1} eventually being
52 permanently removed in deep sea sediments (Cartapanis et al., 2018). In addition, the microbial carbon
53 pump (Jiao et al., 2010) produces refractory dissolved organic molecules throughout the water column at a
54 rate of around 0.4 GtC yr^{-1} (Jiao et al., 2014b), which due to their residence time of hundreds to thousands
55 of years maintain the 700 GtC inventory of dissolved organic carbon in the ocean (Jiao et al., 2010; Jiao et
56 al., 2014a; Legendre et al., 2015; Jiao et al., 2018a). The natural removal of carbon by the various carbon
57 pumps is closely balanced by upwelling and outgassing, with the ocean a moderate source of CO_2 under pre-

1 industrial conditions (Ciais et al., 2013). The mitigation potential of managing natural processes in the open
2 ocean is only briefly assessed here (Section 5.5.1.3).
3
4 Gattuso et al. (2018) provide an overview assessment of the environmental, technical and societal
5 feasibilities of using a range of ocean management actions to reduce climate change and its impacts. Their
6 results for nine actions based on natural processes are summarised in Figure 5.23, also including marine
7 renewable energy (wind, wave and tidal) for comparison. Eight semi-quantitative criteria were used to assess
8 each action: maximum potential effectiveness by 2100 in reducing climatic drivers (ocean warming, ocean
9 acidification and sea level rise), assuming full theoretical implementation; technological readiness and lead
10 time to full potential effectiveness (subsequently combined as technical feasibility); duration of benefits; co-
11 benefits; trade-offs (originally described as dis-benefits); cost-effectiveness; and governability (capability of
12 implementation, and management of any associated conflicts). Here, governability is considered as a
13 constraint (governability challenges) reversing the scoring scale used by Gattuso et al. (2018)
14
15 Global measures (circles in Figure 5.23) can be regarded as mitigation, reducing drivers; local measures
16 (rectangles), are primarily ecosystem-based adaptation, reducing impacts (Section 5.5.2), although they may
17 also contribute to mitigation; two actions were considered at both scales. Gattuso et al. (2018) did not
18 consider the effects of actions on ocean oxygenation, notwithstanding the importance of deoxygenation as a
19 component of climate change. Additional detail is given in Supplementary Material SM5.4.
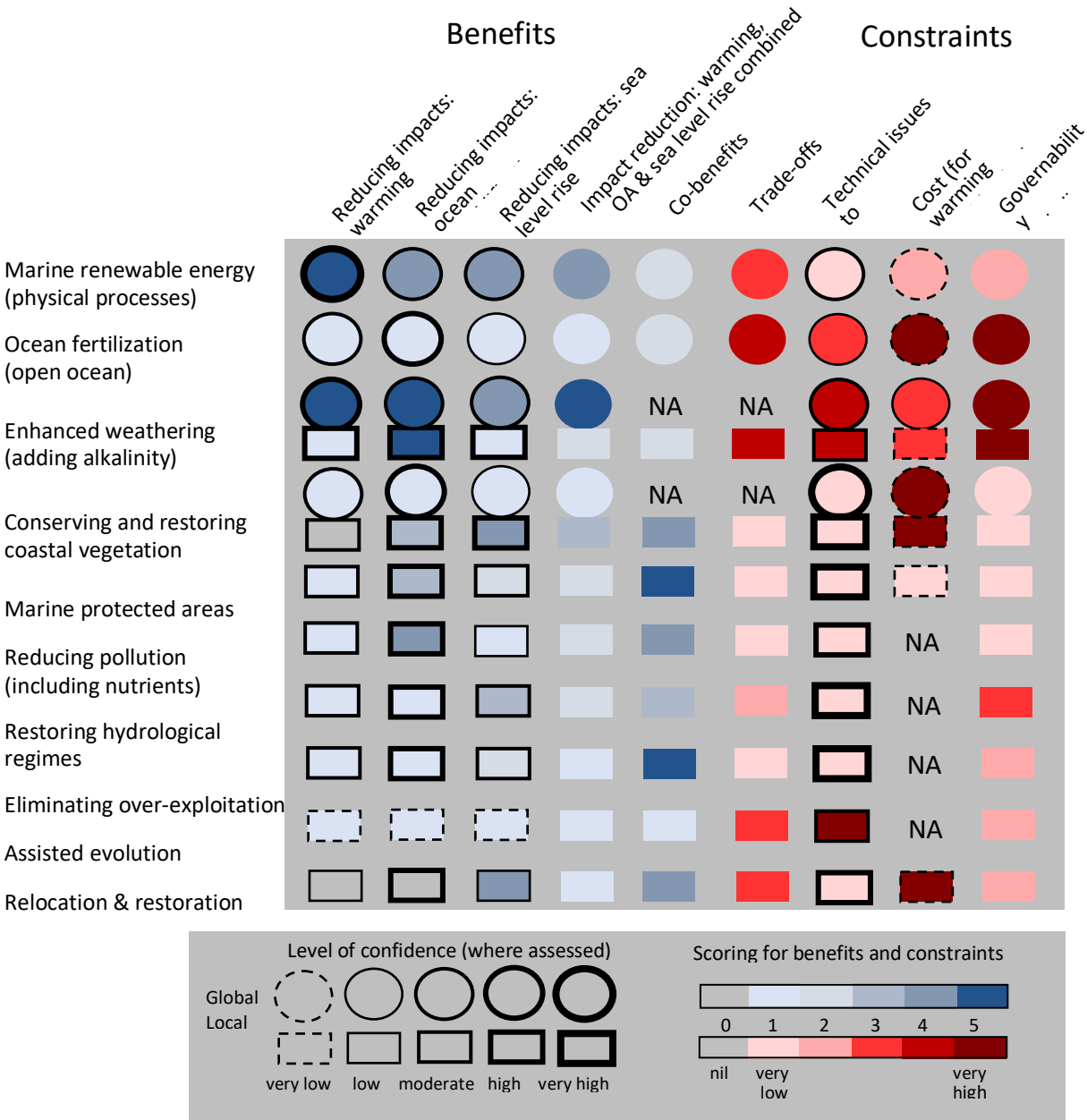20

Figure 5.23: Summary of potential benefits and constraints of ocean-based risk-reduction options using natural processes, from literature-based expert assessments by Gattuso et al. (2018) . Mitigation effectiveness was quantified relative to RCP 8.5, assuming maximum theoretical implementation, with reduction of climate-related drivers considered at either global or local (< 100 km²) scale, shown as circles or rectangles respectively. Impact reduction, co-benefits and trade-offs are in the context of eight sensitive marine ecosystems and ecosystem services. 'Technical issues to overcome' is based on scores for technological readiness, lead time for full implementation and duration of effects. Cost is based on USD per tonne of CO_2 either not released or removed from the atmosphere (for global measures) or per hectare of coastal area with action implemented (for local measures). 'Governance challenges' shows the potential difficulty of implementation by the international community. NA, not assessed. Additional information on scoring methods is given in Supplementary Material SM5.4, Tables SM5.9a and SM5.9b.

5.5.1.2   Climate Mitigation in the Coastal Ocean

5.5.1.2.1   Opportunities and challenges relating to coastal carbon

Estuaries, shelf seas and a wide range of other intertidal and shallow-water habitats (Section 5.3) play an important role in the global carbon cycle through their primary production by rooted plants, seaweeds (macroalgae) and phytoplankton, and also by processing riverine organic carbon. However, the natural carbon dynamics of these systems have been greatly changed by human activities (Regnier et al., 2013;

1   Cloern et al., 2016; Day and Rybczyk, 2019) (*high confidence*). Direct anthropogenic impacts include
2   coastal land-use change (Ramesh et al., 2015; Li et al., 2018a); indirect effects include increased nutrient
3   delivery and other changes in river catchments (Jiao et al., 2011; Regnier et al., 2013), and marine resource
4   exploitation in shelf seas (Bauer et al., 2013). There is *high confidence* that these human-driven changes will
5   continue, reflecting coastal settlement trends and global population growth (Barragán and de Andrés, 2015).
6
7   Policy recognition of the mitigation benefits of coastal ecosystems requires quantitative information on their
8   actual and potential carbon uptake and storage at the local and national scale, within an international
9   framework for carbon accounting (Crooks et al., 2011; Hejnowicz et al., 2015). Such methods are being
10 developed for coastal habitats structured by rooted plants (Needelman et al., 2018; Troxler et al., 2018;
11 Needelman et al., 2019), considered here as 'coastal vegetation', linked to protocols for verification of
12 longterm carbon removal and financial incentives (Crooks et al., 2011; Hejnowicz et al., 2015) and building
13 on techniques used for managing terrestrial carbon sinks (Ahmed and Glaser, 2016b; Aziz et al., 2016).
14 Proposals to apply carbon accounting to seaweeds, the water column and shelf sea sediments (Krause-Jensen
15 and Duarte, 2016; Zhang et al., 2017) are less well-developed.
16
17 *5.5.1.2.2   Coastal vegetation: mangrove, saltmarsh and seagrass ecosystems*
18 Mangrove, saltmarsh and seagrass habitats are widely recognized as blue carbon ecosystems with mitigation
19 potential (Chmura et al., 2003; Duarte et al., 2005; Kennedy et al., 2010; McLeod et al., 2011). Although
20 covering only ~0.1% of the Earth's surface, these three ecosystems together have been estimated to support
21 1–10% of global marine primary production (Duarte et al., 2017). More than 150 countries contain at least
22 one of these ecosystems; 71 countries contain all three (Herr and Landis, 2016), and 74 countries mention
23 such coastal wetlands (five specifically as blue carbon) in their Nationally Determined Contributions
24 (NDCs) to the Paris Agreement (Martin et al., 2016a; Gallo et al., 2017).
25
26 These three vegetated coastal habitats are characterized by high, yet variable, organic carbon storage in their
27 soils and sediments on a per unit area basis (*high confidence*). In the humid tropics, mangrove below-ground
28 organic carbon is typically 500–1000 tC ha^{-1} (Donato et al., 2011; Alongi and Mukhopadhyay, 2015;
29 Howard et al., 2017)), although only ~50 tC ha^{-1} in arid regions (Almahasheer et al., 2017). Australian
30 saltmarshes show particularly wide variation in organic carbon storage, ranging from 15 to 1000 tC ha^{-1} (top
31 1 m) with mean of 165 tC ha^{-1} (Kelleway et al., 2016; Macreadie et al., 2017b). For seagrass meadows,
32 storage values are typically 400–1600 tC ha^{-1} but can exceed 2000 tC ha^{-1} (Serrano et al., 2014). These
33 accumulations have occurred over decadal to millennial timescales (McKee et al., 2007; Lo Iacono et al.,
34 2008). Such blue carbon stock values are similar to freshwater wetlands and peat, but higher than for most
35 forest soils (Laffoley and Grimsditch, 2009; Pan et al., 2011) (*high confidence*).
36
37 When vegetated coastal ecosystems are disturbed, a proportion of their stored carbon is released back to the
38 atmosphere, along with other greenhouse gases (Marba and Duarte, 2009; Duarte et al., 2010; Pendleton et
39 al., 2012; Lovelock et al., 2017). Globally, around 25–50% of vegetated coastal habitats have already been
40 lost or degraded due to coastal agricultural developments, urbanization and other human disturbance during
41 the past 100 years (McLeod et al., 2011). The highest historical losses (60-90%) have occurred in Europe
42 and China (Jickells et al., 2015; Gu et al., 2018; Li et al., 2018a). Current losses are estimated at 0.2–3.0%
43 yr^{-1}, depending on vegetation type and location (Hiraishi et al., 2014; Alongi and Mukhopadhyay, 2015;
44 Atwood et al., 2017) (*medium confidence*). Associated global carbon emissions are estimated at 0.04– 0.28
45 GtC yr^{-1} (Pendleton et al., 2012); ; 0.06–0.61 GtC yr^{-1} (Howard et al., 2017); 0.10–1.46 GtC yr^{-1} (Lovelock
46 et al., 2017); and 0.007 GtC yr^{-1} (mangroves only) (Taillardat et al., 2018). This range of values reflects
47 uncertainties regarding the global rate of habitat loss, and the proportion of carbon remineralised to CO_2.
48
49 Mitigation through emission reduction can therefore be achieved by habitat protection, to greatly reduce or
50 end the human-driven loss of mangrove, saltmarsh and seagrass ecosystems. Such action could potentially
51 produce nationally-significant mitigation (>1% of fossil fuel emissions) for several countries (Taillardat et
52 al., 2018). However, there are still many uncertainties in quantifying carbon release due to habitat
53 degradation and loss (Lovelock et al., 2017), and hence in determining emission reductions. Furthermore,
54 this mitigation option is not available to those countries where habitat loss is not currently occurring, e.g.
55 Bangladesh (Taillardat et al., 2018). Since legal structures already exist in many countries to protect coastal
56 wetlands, the main policy need may be the enforcement of national regulation and site-specific marine
57 protected areas (Miteva et al., 2015; Herr et al., 2017; Howard et al., 2017).

The alternative mitigation approach using coastal blue carbon ecosystems is to enhance the natural carbon uptake of such habitats, not only by increasing their spatial coverage through habitat restoration and new habitat creation, but also by taking management measures to maximise the carbon uptake and storage for existing coastal ecosystems. Such measures include reducing anthropogenic nutrient inputs and other pollutants; restoring hydrology, by removing barriers to tidal flow and sediment delivery; and reinstating predators (to reduce carbon loss caused by some bioturbators) (Macreadie et al., 2017a). Per unit area of habitat created, restored or rehabilitated, such actions may offer high rates of carbon removal: widely-quoted values are 226±39 gC m^{-1} yr^{-1} for mangroves, 218±24 gC m^{-1} yr^{-1} for saltmarsh and 138±38 gC m^{-1} yr^{-1} for seagrass ecosystems (McLeod et al., 2011; Isensee et al., 2019).

Around 90 restoration and rehabilitation projects for mangroves have been documented (López-Portillo et al., 2017), with associated development of a range of restoration evaluation methods (Zhao et al., 2016a). Saltmarsh restoration is reviewed by Adam (2019) and seagrass restoration by van Katwijk et al. (2016). Consistent conclusions, supported by other studies (Bayraktarov et al., 2016; Wylie et al., 2016) are that: natural regeneration increases the likelihood of longterm survival; higher success rates are achieved with strong stakeholder engagement; and it is critical that the (human) factors causing original loss and degradation have been properly addressed (*high confidence*).

Quantification of the climatic benefits of such actions is, however, not straightforward. Measurements of carbon burial rates show high site-specific variability, being strongly affected by a wide range of environmental factors for mangroves (Adame et al., 2017; Schile et al., 2017), seagrasses (Lavery et al., 2013) and salt marshes (Kelleway et al., 2017b). The reliable determination of sediment accumulation rates is a key consideration, with associated uncertainties not fully reflected in the (McLeod et al., 2011) estimates given above. In particular, geochemical-based studies have indicated that seagrass carbon burial may have been greatly overestimated (Johannessen and Macdonald, 2016). These issues are contentious (Johannessen and Macdonald, 2018a; Johannessen and Macdonald, 2018b; Macreadie et al., 2018; Oreska et al., 2018); their scientific resolution is highly desirable. Additional complexities relating to the mitigation role of coastal blue carbon ecosystems include the following:

- Emissions of other greenhouse gases also need to be taken into account (Keller, 2019b). Methane release from mangrove habitats can reduce the scale of their climatic benefits by 18–22% (Adams et al., 2012; Chen and Ganapin, 2016; Chmura et al., 2016; Rosentreter et al., 2018; Cameron et al., 2019) and nitrous oxide and methane together may offset saltmarsh CO_2 uptake by 24–31% (Adams et al., 2012). Nitrous oxide emissions are strongly affected by nutrient loading (Chmura et al., 2016); under pristine conditions, mangroves can provide a sink rather than a source (Maher et al., 2016). Note that values of the 'offset' depend on the metrics used for determining CO_2 equivalents.
- Carbonate formation, releasing CO_2, may also reduce the benefits of carbon storage by similar proportions (Howard et al., 2017; Macreadie et al., 2017a; Kennedy et al., 2018; Saderne et al., 2019).
- Lateral transfers are not well-quantified. Whilst some of the carbon stored in coastal marine sediments may be recalcitrant carbon from terrestrial or atmospheric sources (and should therefore be excluded) (Chew and Gallagher, 2018), export of dissolved organic carbon, inorganic carbon and alkalinity may be considered as additional sequestration (Maher et al., 2018; Santos et al., 2019).
- The permanence of vegetated coastal systems, even if well-protected, cannot be assumed under future temperature regimes (Ward et al., 2016; Duke et al., 2017; Jennerjahn et al., 2017; Nowicki et al., 2017)
- Responses to future sea level rise are also uncertain and complex (Kirwan and Megonigal, 2013; Spencer et al., 2016) . However, impacts are not necessarily negative: carbon sequestration capacity may increase where totally new habitats are created (Barnes, 2017), or if mangroves replace salt marshes (Kelleway et al., 2016).

In summary, a combination of both conservation and restoration of mangrove, saltmarsh and seagrass habitats can contribute to in national mitigation effort for those countries with relatively large coastlines where such ecosystems naturally occur (Murdiyarso et al., 2015; Atwood et al., 2017). However, the associated current uncertainties in quantifying relevant carbon storage and flows are expected to be problematic for reliable measurement, reporting and verification (*high confidence*).

1   At the global scale, synthesis studies have estimated the potential additional sequestration achieved by
2   coastal blue carbon management as ~0.05 GtC yr^{-1} (Griscom et al., 2017) and 0.04 GtC yr^{-1} (National
3   Academies of Sciences, Engineering, and Medicine, 2019), assuming that a relatively high proportion of
4   vegetated ecosystems can be re-instated to their 1980–90 extents. These values compare to current net
5   anthropogenic emissions from all sources of 10.0 GtC yr^{-1} (Le Quéré et al., 2018), and are consistent with
6   the 'very low' scores by (Gattuso et al., 2018) for the climate mitigation benefits of conserving and restoring
7   coastal vegetation (Figure 5.23). This assumed scale of restoration would be challenging, because of the
8   semi-permanent and on-going nature of most coastal land-use change, such as human settlement, conversion
9   to agriculture and aquaculture, shoreline hardening, and port development (Gittman et al., 2015; Li et al.,
10   2018a).
11
12   Restoration costs could also be an important constraint for large-scale application. Based on published data
13   from 246 observations, Bayraktarov et al. (2016) estimated median total costs for restoration of one hectare
14   of mangrove, saltmarsh and seagrass habitat to be ~USD 2,508, USD 151,129 and USD 383,672
15   respectively, in 2010 prices. For each ecosystem, there was high variability in costs according to the
16   economy of the country where the restoration projects were carried out, and the restoration technique
17   applied. Assessment of coastal conservation and restoration costs is also given in Section 4.4.2.3, in Box 5.5
18   (in the context of coral reef restoration costs) and Section 5.5.2.5.
19
20   Measures to protect and restore coastal blue carbon habitats provide many other societal benefits in addition
21   to climate regulation (Section 5.4.1). In particular, there is *high confidence* that coastal wetlands benefit
22   local fisheries, enhance biodiversity, give storm protection, reduce coastal erosion, improve water quality,
23   and support local livelihoods (Costanza et al., 2008; Spalding et al., 2014). Coastal ecosystems may keep
24   pace with sufficiently gradual sea level rise, and may be more cost-effective in flood protection than hard
25   infrastructure like seawalls (Temmerman et al., 2013; Möller, 2019). Coastal blue carbon can therefore be
26   considered as a 'no regrets' mitigation option at the national level in many countries, in addition to (not a
27   replacement for) more effective mitigation measures. Additional research is needed over the full range of
28   environmental conditions to improve knowledge and understanding of the complex carbon dynamics of
29   coastal vegetation and associated systems, to enable well-quantified and cost-effective carbon sequestration
30   enhancement (Vázquez-González et al., 2017; Windham-Myers et al., 2018).
31
32   *5.5.1.2.3   Seaweeds (macroalgae)*
33   Seaweeds do not directly transfer carbon to marine sediments, unlike the rooted coastal vegetation
34   considered above (Howard et al., 2017). Nevertheless, seaweed detritus can deliver carbon to sedimentary
35   sites (Hill et al., 2015) and may provide a source of refractory dissolved organic (Krause-Jensen and Duarte,
36   2016). Recent studies indicate that globally-important amounts of carbon may be involved in these processes
37   (Krause-Jensen and Duarte, 2016; Krause-Jensen et al., 2018; Smale et al., 2018). There is, however,
38   currently *low confidence* that enhancement of natural seaweed production can provide a significant
39   mitigation response, due to large uncertainties relating to sequestration duration and effectiveness. Such
40   considerations relate to transport pathways, the fate of material transported to deeper water, and the
41   timescales of its subsequent return to the atmosphere over decadal to century timescales.
42
43   Seaweed aquaculture is inherently more manageable as a mitigation response (N'Yeurt et al., 2012; Chung et
44   al., 2013; Chung et al., 2017; Duarte et al., 2017). If linked to biofuel or biogas production (N'Yeurt and
45   Iese, 2014; Moreira and Pires, 2016; Sondak et al., 2017), there would be potential to reduce emissions (as
46   an alternative to fossil fuels); if also linked to carbon capture and storage (Hughes et al., 2012), it may be
47   possible to achieve negative emissions (net CO_2 removal from the atmosphere). Full life cycle analyses are
48   needed to assess the energy efficiency of such approaches, and the viability of scaling them up to
49   climatically-important levels, taking account of associated environmental and socio-economic implications.
50
51   A different mitigation option using seaweeds relates to their use as a dietary supplement for ruminants to
52   suppress methane production. *In vitro* studies have given promising results (Dubois et al., 2013; Machado et
53   al., 2016; Machado et al., 2018). However, because the potential scale of real-world benefits have yet to be
54   quantified, there is *low confidence* in this approach as a mitigation option.
55

5.5.1.2.4 Land-sea integrated eco-engineering

Land-based nutrient management could, in theory, be used to enhance carbon storage in coastal seas and deeper waters, by increasing the amount of refractory dissolved organic carbon (Jiao et al., 2011; Jiao et al., 2014b; Jiao et al., 2018b). This idea is supported by a statistical analysis of the relationship between organic carbon and nitrate in various natural environments (Taylor and Townsend, 2010) as well as by experimental results in estuarine and offshore waters (Yuan et al., 2010; Jiao et al., 2011; Jiao et al., 2014b). Delivery of nutrients from agricultural fertilizers and sewage discharge to coastal waters may currently promote the microbial breakdown of river-derived terrestrial dissolved organic carbon, reducing carbon storage (Liu et al., 2014). Thus reducing nutrient inputs in the future may expand carbon storage by favouring the microbial carbon pump, in addition to the multiple co-benefits of reduced nutrient loads related to harmful algal blooms, oxygenation and ocean acidification (Miranda et al., 2013; Jiao et al., 2018a; Zhang et al., 2018). Although there is some evidence for the impact of dissolved organic carbon variations on global scale climate (Rothman et al., 2003) the benefits of this approach have yet to be determined quantitatively and uncertainties remain regarding the longevity of removal and associated carbon accounting (measurement, reporting and verification). Until such issues are better resolved, there is *low confidence* that stimulation of refractory dissolved organic carbon production could provide an operational long-term mitigation measure.

5.5.1.2.5 Control of sediment disturbance, enhanced weathering and other geochemical approaches

Anthropogenic sediment disturbance, through fishing, dredging and the installation of offshore structures, affects the security of carbon storage in shelf sea sediments (Hale et al., 2017). Management of such activities might therefore increase carbon retention, over relatively large areas of shelf seas (Avelar et al., 2017; Luisetti et al., 2019). However, there is a lack of data and understanding of the complex processes that affect carbon storage in the potentially-mobile fraction of marine sediments (van de Velde et al., 2018); exceptions are provided by Hu et al. (2016) and Diesing et al. (2017). Due to these uncertainties, there is currently *low confidence* that control of sediment disturbance can be used for climate mitigation.

There is theoretically greater potential for carbon removal by 'enhanced weathering' using mineral additions to coastal waters (and the open ocean) (Rau, 2011; Renforth and Henderson, 2017). These approaches are based on increasing the naturally-occurring uptake of CO_2 by carbonates (e.g., calcite and dolomite) or silicate minerals (such as olivine). Such rock-weathering currently sequesters ~0.25 GtC yr^{-1}, on land and at sea (Taylor et al., 2015) and provides the longterm control of atmospheric CO_2 concentrations. It could be enhanced by adding ground minerals to beaches (Montserrat et al., 2017) or the sea surface. Other geochemical approaches for adding alkalinity that are less directly based on natural processes (Rau et al., 2012; GESAMP, 2019) are not considered here.

Enhanced weathering methods might be used to reduce local impacts, e.g., for coral reefs (Albright et al., 2016b; Feng et al., 2016), as well as contributing to wider mitigation of climate change. However, their climatic benefits would be difficult to quantify, with other constraints on their development and deployment relating to the governance, cost, and uncertain environmental impacts of large-scale application (Gattuso et al., 2018). The combination of these factors results in *low confidence* that enhanced weathering can provide a viable and acceptable climate mitigation approach.

5.5.1.3   Climate Mitigation in the Open Ocean

Recent reviews of the scope for using natural processes in the open ocean for climate mitigation are provided by Keller (2019a) and GESAMP (2019). The summary assessment given here is limited to direct and indirect biologically-based approaches, consistent with the scoping of this report and the major governance constraints on the large-scale application of open ocean interventions.

Current net primary production by marine phytoplankton is estimated to be 58 ± 7 GtC yr^{-1} (Legendre et al., 2015), similar to terrestrial primary production and around 6 times greater than anthropogenic emissions (Le

1   Quere et al. (2016). However, over 99% of the biologically-fixed carbon returns to the atmosphere over a
2   range of timescales (Cartapanis et al., 2018).
3
4   The direct method of increasing marine productivity involves adding land-derived nutrients that may
5   currently limit primary production, particularly iron. This approach has been investigated experimentally, by
6   modelling and by observations of natural system behaviour (Keller et al., 2014a; Bowie et al., 2015;
7   Tagliabue et al., 2017). The thirteen experimental studies to date (seven in the Southern Ocean, five in the
8   Pacific, and one in the sub-tropical Atlantic) have shown that primary production can be, but is not always,
9   enhanced by the addition of iron (Boyd et al., 2007; Yoon et al., 2016; GESAMP, 2019).
10
11   The difficulties arise in demonstrating the time-scale of additional carbon removal, and in obtaining
12   information on the consequences of the fertilization for other marine ecosystem components, including
13   ocean acidification and other potential side-effects (Williamson and Turley, 2012). Modelling studies
14   (Aumont and Bopp, 2006) indicate that the climatic benefits could be relatively short-lived. Furthermore,
15   public and political acceptability for ocean fertilization is low (Williamson et al., 2012; Boyd and Bressac,
16   2016; Williamson and Bodle, 2016; Fuentes-George, 2017; McGee et al., 2018). Ocean iron fertilization is
17   regulated by the London Protocol, with amendments prohibiting such action unless constituting legitimate
18   scientific research authorized under permit (see Section 5.5.4.1). There are additional governance constraints
19   for the Southern Ocean where ocean iron fertilization is theoretically considered to be most effective
20   (Robinson et al., 2014).
21
22   Open ocean fertilization by macro-nutrients (e.g., nitrate) has also been proposed, with modelled potential
23   for gigaton-scale carbon removal (Harrison, 2017). Similar technical and governance considerations apply
24   with regard to the quantification of mitigation benefits, the monitoring of potential adverse impacts, and the
25   political acceptability of large-scale deployment. This approach would also involve higher costs, because of
26   the much greater quantities of nutrients required (Williamson and Turley, 2012).
27
28   The indirect method of enhancing marine productivity uses physical devices to increase upwelling, thereby
29   increasing the supply of a wide range of naturally-occurring nutrients from deeper water. This technique
30   risks releasing additional CO_2 to the atmosphere, reducing its potential for climate mitigation (Bauman et al.,
31   2014). There may also be other undesirable climatic consequences, including disruption of regional weather
32   patterns and long-term warming rather than cooling, if enhanced upwelling is deployed at large scale
33   (Kwiatkowski et al., 2015).
34
35   Because of the many technical, environmental and governance issues relating to marine productivity
36   enhancement, by either direct fertilization or upwelling, there is *low confidence* that such open ocean
37   manipulations provide a viable mitigation measure.
38
39   **5.5.2   *Ocean-based Adaptation***
40
41   The IPCC Fifth Assessment Report (AR5) concluded, with *high agreement* but *limited evidence*, that climate
42   change impacts on coastal human settlements and communities could be reduced through coastal adaptation
43   activities (Wong et al., 2014a). The limited evidence of the context-specific application of adaptation
44   principles to support the assessment was highlighted as a knowledge gap for future research. This assessment
45   reports progress made with developing such evidence and assesses human adaptation response to climate
46   change in ecosystems, coastal communities and marine environments.
47
48   Components of human adaptation responses include risk assessment, risk reduction, and pathways towards
49   resilience (Cross-Chapter Box 2; Chapter 1.6). Residual risk remains where hazard, vulnerability and
50   exposure intersect, subsequent to an adaptation pathway response. Here we focus on adaptation responses
51   within ecosystems and in human systems, as framed in Chapter 1, and defined by:
52
53   • *Nature-based* or *ecosystem-based adaptation* (5.5.2.1). The use of biodiversity and ecosystem services as
54   part of an overall adaptation strategy to help people to adapt to the adverse effects of climate change.
55   Ecosystem-based adaptation uses the range of opportunities for the sustainable management,
56   conservation, and restoration of ecosystems to provide services that enable people to adapt to the impacts
57   of climate change (Narayan et al., 2016; Moosavi, 2017).

- *Human systems - Built environment adaptation* (5.5.2.3.1) Adaptation solutions pertaining to coastal built infrastructure and the systems that support such infrastructure (Mutombo and Ölçer, 2016; Forzieri et al., 2018).

- *Human systems - Socio-institutional adaptation* (5.5.2.3) Adaptation responses within human social, governance and economic systems and sectors (Oswald Beiler et al., 2016; Thorne et al., 2017). This includes, but is not limited to *community-based adaptation* by coastal communities (5.5.2.3.2) based on empowering and promoting the adaptive capacity of communities, through appropriate use of context, culture, knowledge, agency, and community preferences (Archer et al., 2014; Shaffiril et al., 2017)

To avoid duplication, detailed consideration of adaptation responses to sea level rise and extreme events (including heat waves, and compound and cascading events) are avoided here, as they are covered by Chapter 4 and Chapter 6, respectively. Tables 5.7 and 5.8 provide a summary assessment of climate change impacts, human adaptation response and benefits in ecosystems and human systems respectively. Details of the assessed literature are in Supplementary Material Table 5.7. Climate drivers and impacts reported in the adaptation literature are consistent with those reported in Sections 5.2 and 5.3. Physical impacts include the disruption of physical coastal processes, like sediment dynamics, leading to, for example, erosion, flooding and coastal infrastructure damage (see Tables 5.7 and 5.8). Ecological impacts include the loss of ecosystems and biodiversity (Sections 5.2.3, 5.2.4, 5.3), which affected provision of ecosystem services, like coastal protection or food provision. The most commonly reported non-climate human drivers are growing human coastal populations (Elliff and Silva, 2017; van Oppen et al., 2017a; Gattuso et al., 2018) with poorly planned or managed urban development (Barbier, 2015; Wigand et al., 2017), land use change (Robins et al., 2016a), loss of ecosystems (Runting et al., 2017), socio-economic vulnerability (Broto et al., 2015; Bennett et al., 2016) of many coastal communities, ineffective governance and knowledge gaps for implementation.

5.5.2.1   Ecosystem-based Adaptation

This section assesses adaptation response in coastal ecosystems, beginning with biological adaptation in species, and followed by a summary assessment of ecosystem-based adaptation as a response to climate change.

5.5.2.1.1   Biological adaptation
There are many studies on biological climate change adaptation responses (Crozier and Hutchings, 2014; Miller et al., 2017; Diamond, 2018). Sections 5.2.3 and 5.3.3 discuss three main types of biological adaptation, broadly defined: evolutionary (genetic) adaptation through natural selection; phenotypic plasticity (acclimatization), within an organism's lifetime; and individual or population mobility towards more favourable conditions. There are, however, expected to be limits to such natural adaptation, and large variations between species and populations (Gienapp and Merilä, 2018).

An accurate understanding of climate change impacts upon species, their sensitivity and adaptive capacity and consequent ecological effects (considering both indirect as well as direct impacts) is used to estimate extinction risk, so that an appropriate management response can be developed (Butt et al., 2016). Ecosystem based adaptation takes these complex interactions into account (Hobday et al., 2015), including the disruptive impacts of alien invasive species (Ondiviela et al., 2014; Wigand et al., 2017) . Effective adaptation action, therefore, contains a broader consideration than historical conservation practices (*medium evidence, high agreement*), including the development of international collaborations and databases to improve ocean-scale understanding of climate change impacts (Okey et al., 2014; Young et al., 2015). A key knowledge gap relates to the critical thresholds for irreversible change for species (Powell et al., 2017).

Table 5.7: Summary of reported Adaptation responses (A), the Impacts (I) they aimed to address, and the expected Benefits (B) in coastal ecosystems within Physical, Ecological, Social, Governance, Economic and Knowledge categories. For further details of impacts on ecosystems see Section 5.3. Legend: a + sign indicates *robust evidence*, a triangle indicates *medium evidence* and an underline indicates *limited evidence*. Dark blue cells indicate *high agreement*, blue indicates *medium agreement* and light blue indicates either *low agreement* (denoted by presence of a sign) if sufficient papers were reviewed for an assessment or no assessment (if less than three papers were assessed per cell). The papers used for this assessment can be found in Supplementary Material SM5.5.

	Impacts (I)	Adaptation responses (A)	Coral Reefs I	A	B	Mangroves I	A	B	Salt marshes/ Wetlands I	A	B	Estuaries I	A	B	Sandy Beaches/ Dunes I	A	B	Multiple ecosystems I	A	B	Benefits (B)
Physical	Coastal physical processes disrupted	Supporting physical processes	Δ			Δ	Δ	Δ	+	+	_	Δ	+	_	Δ	+		+			Physical processes supported
	Catchment physical processes disrupted	Hard engineering responses											_								Coastal infrastructure resilience increased
	Coastal infrastructure damage	Soft engineering responses and buffers									_				Δ						Improved infrastructure functionality
	Disruption of urban systems	Integrated hard and soft engineering													Δ						Increased structural heterogeneity
	Land subsidence	Managed retreat and coastal realignment																			
Ecological	Ecosystem degradation and loss	Ecosystem restoration and protection	+	Δ	+	+	+	+	Δ			_	_	Δ	Δ	_					Ecosystem/ ecological resilience supported
	Biodiversity and genetic diversity loss	Bioengineering									Δ	_									Physical processes supported
	Habitat range shifts	Assisted evolution and relocation		Δ								Δ									Coastal infrastructure resilience increased
	Sub-lethal species impacts	Nature based solutions					Δ														Increased biodiversity
	Invasive alien species				Δ																Habitat range shifts accommodated
																					Improved organismal fitness
																					Genetic heterogeneity supported
																					Strengthened socio-ecological system
Social	Decreased access to ecosystem services	Improving access to/ storage of natural resources				Δ				+		Δ									Access to sustainable ecosystem services
	Local decline in agriculture and fisheries	Improving agricultural or fisheries practices																			Improved access to community services
	Increasing living costs	Supporting natured-based industries																			Increasing resilience in human systems
	Livelihoods impacts	Sustainable resource use													_						Improved socio-economic services
	Increased food insecurity	Maintaining or switching livelihoods								_											Improved employment and livelihoods
	Public health risks increased	Community participatory programmes							_												Improved health
	Cultural and traditional knowledge impacts	Developing adaptive networks																			Improved community participation
	Gender-related impacts	Sustainable household management																			Better informed communities
	Increased social vulnerability	Improving access to community services																			Improved integration of knowledge systems
	Decreased access to local government services	Empowering communities and addressing inequality																			Empowering women and children
	Socio-economic entrapment and decline	Building socio-ecological resilience																			Increased adaptive capacity
	Global declines in food stocks																				Improved disaster preparedness
	Public areas access restrictions																				Empowered communities
	Decline in perceived value of human systems																				Improved community cohesion
	Conflict and migration																				Reduced inequality

	Impacts (I)	Adaptation responses (A)	Coral Reefs (I A B)	Mangroves (I A B)	Salt marshes/ Wetlands (I A B)	Estuaries (I A B)	Sandy Beaches/ Dunes (I A B)	Multiple ecosystems (I A B)	Benefits (B)
Governance	Capacity challenges	Adopting/ mainstreaming sustainability policies	▪	▪			▪	▪	Political and institutional capacity developed
	Increased geopolitical tensions	Improving disaster response programmes		▪			▪		Strengthened participatory governance
	Growing inequalities	Improving implementation and coordination of policies	▪	▪					Better planning processes supported
		Development controls				▪			Improved coordination and decision making
		Evidence-based implementation					▪		Improved implementation and policies
		Improving ICM/ MPAs	▪					▪	Better communication
		Horizontal/ vertical integration of governance	▪						Improved transparency and trust
		Developing partnerships and building capacity		▪			▪	▪	Climate justice advanced
		Improving access to community services							Reduced conflict
		Pursuing climate justice							Improved security
							▪		Improved adaptive management
									Development supported
Economic	Increased business and living costs	Improving financial resources availability		▪	▪	▪			Increased revenue/ income
	Business disruptions and losses	Improving access to insurance products							Increased financial resources available
	Decreased value of assets/ products	Economic diversification						▪	Reduced operational and capital costs
		Improving access to international funding programmes							Investment strengthened
Knowledge	Uncertainty for decision makers	Better monitoring and modelling		▪		▪		▪	Informed decision making tools
		Improving planning processes		▪	▪		▪		Improved co-production of knowledge
		Improving forecasting and early warning systems							Improved relevance of products
		Improving decision support frameworks		▪			▪ ▪		Improved education and outreach
		Improving participatory processes							Improved awareness
		Coordinating top down and bottom up approaches							
		Integrating knowledge systems						Δ	
		Improving location and context specific knowledge							
		Improving scientific communication		▪					
		Stakeholder identification, outreach and education		▪		▪			

1 *5.5.2.1.2   Adaptation in coral reefs*
2 Coral reefs are currently threatened by the continuous global degradation of warm-water coral reef
3 ecosystems and the failure of traditional conservation actions to revive most of the degrading reefs
4 (Rinkevich, 2008; Miller and Russ, 2014). Interventions to rehabilitate degraded coral reef ecosystems can
5 be categorized as preventive ('passive' restoration) or adaptive ('active' restoration) (Miller and Russ, 2014;
6 Linden and Rinkevich, 2017) (see Box 5.5).

8 Inspired by silviculture (forestation) approaches to terrestrial ecosystem restoration, studies (Rinkevich,
9 1995; Rinkevich, 2005; Rinkevich, 2006; Rinkevich, 2008; Bongiorni et al., 2011) have proposed a two-step
10 restoration strategy for warm-water coral reefs termed gardening of denuded coral reefs. In the first step, a
11 large pool of coral colonies (derived from coral nubbins and fragments, and from sexually derived spat) are
12 farmed in underwater nurseries, preferably on mid-water floating devices installed in sheltered zones, in
13 which coral material can be cultured for up to several years. In the second step, nursery-grown coral
14 colonies, together with recruited associated biota, are transplanted to degraded reef sites (Shafir and
15 Rinkevich, 2008; Mbije et al., 2010; Shaish et al., 2010b; Shaish et al., 2010a; Bongiorni et al., 2011;
16 Horoszowski-Fridman et al., 2011; Linden and Rinkevich, 2011; Mbije et al., 2013; Cruz et al., 2014;
17 Chavanich et al., 2015; Horoszowski-Fridman et al., 2015; Lirman and Schopmeyer, 2016; Montoya Maya
18 et al., 2016; Ng Chin Soon et al., 2016; Lohr and Patterson, 2017; Rachmilovitz and Rinkevich, 2017).
19 Active restoration of coral reefs, while still in its infancy and facing a variety of challenges (Rinkevich,
20 2015b; Hein et al., 2017), has been suggested to potentially improve the ecological status of degraded coral
21 reefs and the socio-economic benefits that the reefs provide (Rinkevich, 2014; Rinkevich, 2015b; Linden and
22 Rinkevich, 2017).

24 Ecological engineering approaches may promote coral reef adaptation (Rinkevich, 2014; Forsman et al.,
25 2015; Coelho et al., 2017; Horoszowski-Fridman and Rinkevich, 2017; Linden and Rinkevich, 2017;
26 Rachmilovitz and Rinkevich, 2017). They also include: augmenting functional diversity, including that of
27 the microbiome (Casey et al., 2015; Horoszowski-Fridman and Rinkevich, 2017; Shaver and Silliman,
28 2017); transplantating whole habitats (Shaish et al., 2010b; Gómez et al., 2014); and enhancing genetic
29 diversity (Iwao et al., 2014; Drury et al., 2016; Horoszowski-Fridman and Rinkevich, 2017). Active
30 restoration can contribute to reef rehabilitation in all major reef regions (Rinkevich, 2014; Rinkevich,
31 2015b). However, there is limited evidence on how resistant these manipulated corals are to global change
32 drivers (Shaish et al., 2010b; Shaish et al., 2010a) or how the nursery time affects biological traits like
33 reproduction in coral transplants (Horoszowski-Fridman et al., 2011). Coral epigenetics may also be used as
34 an adaptive management tool for reef rehabilitation (*low confidence*), as suggested by studies on coral
35 adaptation (Brown et al., 2002; Horoszowski-Fridman et al., 2011; Palumbi et al., 2014; Putnam and Gates,
36 2015; Putnam et al., 2016).

38 Research on active coral reef restoration (Box 5.5) suggests the potential to help rehabilitate degraded coral
39 reefs, provided that the underlying drivers of the impacts are mitigated (*high confidence*). Ongoing and new
40 research in active coral reef restoration may further improve active reef restoration outcomes (Box 5.5) (*low
41 confidence*). However, these coral reef restoration options may be ineffectual if global warming exceeds
42 1.5°C relative to pre-industrial levels (Hoegh-Guldberg et al., 2018; IPCC, 2018).

45 [START BOX 5.5 HERE]

47 **Box 5.5: Coral Reef Restoration as Ocean-based Adaptation**

49 Anthropogenic global change is impacting all warm-water corals and the reef structures (Section 5.2.2.3.3;
50 IPCC AR5 WGII). These impacts are rapidly increasing in scale and intensity, exposing coral reefs to
51 enhanced degradation rates and diminishing capacities to maintain ecological resilience, to absorb
52 disturbances, and to adapt to the changes (Box 5.1) (Graham et al., 2014; Rinkevich, 2015a; Harborne et al.,
53 2017). With the growing awareness that traditional reef conservation measures are insufficient to address
54 climate change impacts on coral reefs (Section 5.2.2.1), adaptation interventions to enhance the resilience of
55 coral reefs are being called for (Rinkevich, 1995; Rinkevich, 2000; Barton et al., 2017). Intervention
56 strategies that are still at the 'proof-of-concept' stage, include: 'assisted colonization' - actively moving
57 species that are confined to disappearing habitats (Hoegh-Guldberg et al., 2008; Chauvenet et al., 2013);

1   'assisted evolution' - developing corals resistant to climate change via accelerated natural evolution
2   processes (van Oppen et al., 2015); assisted coral chimerism (Rinkevich, 2019); novel coral–symbiont
3   associations (McIlroy and Coffroth, 2017); and coral microbiome manipulation (Bourne et al., 2016; Sweet
4   and Bulling, 2017; van Oppen et al., 2017b). In contrast, the 'coral gardening' approach - coral farmed in
5   nurseries and transplanted using a range of tactics to increase survivability, growth rates and reproduction
6   (Rinkevich, 2006; Rinkevich, 2014) - is already in use. Other interventions that have already been
7   implemented in some coral reefs, such as the use of artificial reefs (Ng Chin Soon et al., 2017) are limited in
8   impacts, and all are also revealing considerable challenges (Riegl et al., 2011; Coles and Riegl, 2013;
9   Ferrario et al., 2014).
10
11 Many of the alternative interventions that aim to increase the climate-resilience of coral reefs involve
12 culturing, selectively breeding and transplanting corals to enhance the adaptability of reef organisms to
13 climate change, for example, by supporting the natural poleward range expansion of corals (West et al.,
14 2017; Vergés et al., 2019). Advances in reef restoration techniques have been made in the last two decades
15 (Rinkevich, 2014; Lirman and Schopmeyer, 2016), but assessments of the effectiveness of these techniques
16 have mostly focused on the short-term feasibility of the technique (Frias-Torres and van de Geer, 2015;
17 Lirman and Schopmeyer, 2016; Montoya Maya et al., 2016; Jacob et al., 2017; Rachmilovitz and Rinkevich,
18 2017), while longer-term evaluation in the context of all the pillars of sustainable development (Section
19 5.4.2) is limited (Rinkevich, 2015b; Barton et al., 2017; Flores et al., 2017; Hein et al., 2017). These
20 alternative interventions, primarily the coral gardening approach, face two challenges. The first is scaling up;
21 currently, these interventions have been tested at scales of hundreds of meters, while application at larger
22 scale is lacking (Rinkevich, 2014). The second challenge (Box 5.5, Figure 1) is the effectiveness of active
23 reef restoration to mitigate or rehabilitate global change impacts (Shaish et al., 2010a; Schopmeyer et al.,
24 2012; Coles and Riegl, 2013; Hernández-Delgado et al., 2014; Rinkevich, 2015a; Wilson and Forsyth, 2018)
25 and whether it can keep up with rising sea levels (Perry et al., 2018), especially in low lying ocean states.
26
27 Altogether, coral reefs of the future will not resemble those of today because of the projected decline and
28 changes in the composition of corals and associated species in the remaining reefs (Section 5.3.4, Box 5.5
29 Figure 1) (Rinkevich, 2008; Ban et al., 2014) *(high confidence)*. The very high vulnerability of coral reefs to
30 warming, ocean acidification, increasing storm intensity and sea level rise under climate change (AR5
31 WG2), including enhanced bioerosion (Schönberg et al., 2017) *(high confidence)* point to the importance of
32 considering both mitigation (Section 5.5.1) and adaptation (section 5.3.3.6) for coral reefs. Extensive
33 research has explored adaptation measures involving the cultivation and transplantation of corals; however,
34 the literature contains *limited evidence* on the comprehensive analysis of the relative costs and benefits of
35 these interventions across the economic, ecological, social and cultural dimensions (Bayraktarov et al., 2016;
36 Flores et al., 2017; Linden and Rinkevich, 2017).
37
38
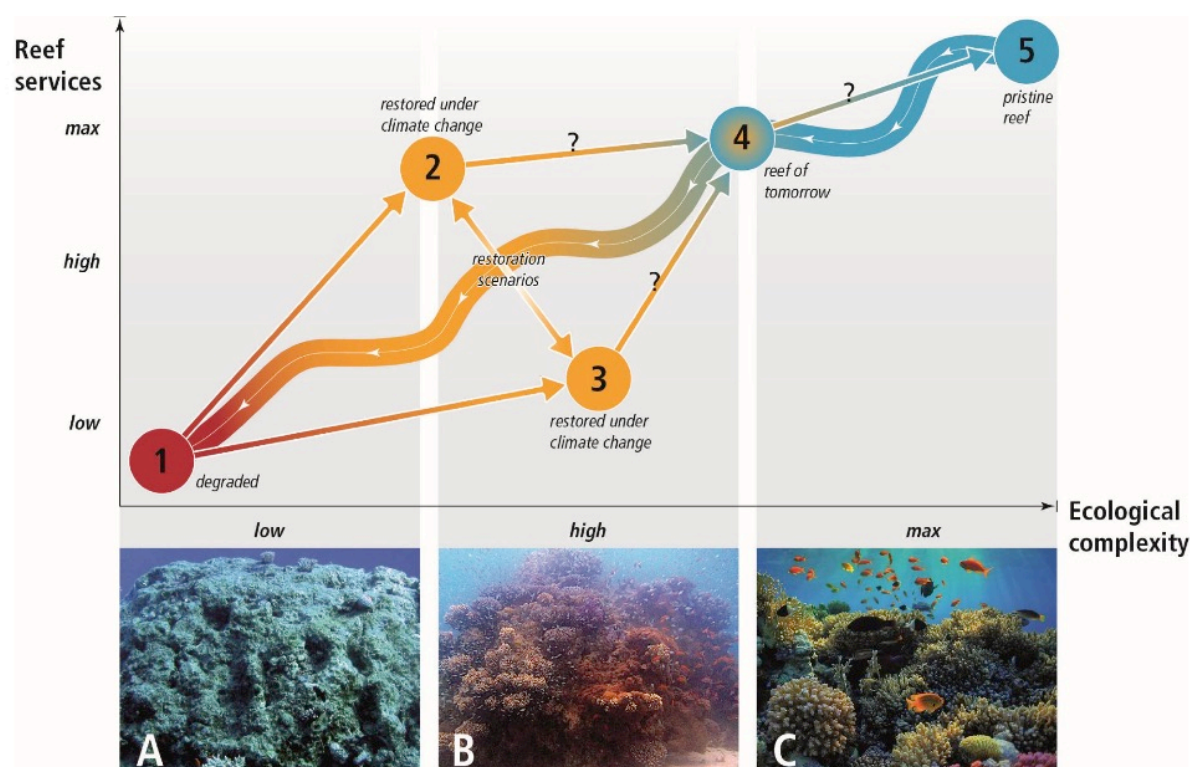
Box 5.5, Figure 1: Coral reef restoration as an ocean-based adaptation tool to climate change. The squiggly line represents non-linear ecological statuses along a trajectory and 5 reef states (circles 1-5; in varying ecological complexity [x-axis] and service levels [y-axis]) including two extreme statuses (a pristine versus a highly degraded state, circles 5 and 1, respectively). Two 'restored reef-state' scenarios (circles 2, 3), lead to the state of the restored 'reef of tomorrow' (circle 4). The route from the state of the 'reef of tomorrow' (circle 4) to a pristine state (circle 5) is doubtful (the question mark) and is still at a theoretical level. The routes from the two 'restored reef-state' scenarios to the 'reef of tomorrow' are under investigations (the question marks). Based on Rinkevich (2014)(Figure 1). A-C represent different reef statuses. A = a denuded knoll at the Dekel Beach, Eilat, Israel before reef transplantation (November 2005; Photo: Y. Horoszowski-Fridman); B = the same knoll, restored (June 2016; photo by Shai Shafir). More than 300 nursery-grown colonies of 7 coral species were transplanted during three successive transplantations (years 2005, 2007, 2009). In 2016 the knoll was surrounded by reef inhabiting schools of fish. C = a pristine reef, not existing under current and anticipated reef conditions. Restoration scenarios are developed along paths from a degraded reef (low ecological complexity, minimal reef services) toward a healthy 'reef of tomorrow', passing through two restored reef states that are impacted by climate change (Shaish et al., 2010a; Schopmeyer et al., 2012; Hernández-Delgado et al., 2014; Rinkevich, 2015a). The employment of ecological engineering approaches may help in moving the ecological states from either restored reef to the 'reef of tomorrow' status (*medium confidence*).

[END BOX 5.5 HERE]

5.5.2.1.3   Adaptation in mangroves and other coastal ecosystems

Mangroves provide significant ecosystem services, including localized coastal protection from extreme storm events (Section 5.4.1), supporting services through increased sedimentation rates (Hayden and Granek, 2015) and provisioning services for local communities, e.g. habitats for nurseries to support fisheries. Mangroves provide limited carbon mitigation, in terms of global emissions reduction, and substantial job creation (Table 5.7) co-benefits (for example through Reducing Emissions from Deforestation and Forest Degradation programs) when managed properly (5.4.1, 5.5.1.1), and there is evidence of their value in supporting aquaculture and fishery initiatives (Huxham et al., 2015; Ahmed and Glaser, 2016a).

Mangrove ecosystem-based adaptation responses most commonly reported included ecosystem restoration (Sierra-Correa and Cantera Kintz, 2015; Romañach et al., 2018) and management such as mangroves re-planting through community participation programmes (Nanlohy et al., 2015; Nguyen et al., 2017; Triyanti et al., 2017). Mangrove ecosystem-based adaptation has been reported to provide multiple co-benefits in terms of improvement in support for coastal physical processes, including: shoreline stabilisation (Hayden and Granek, 2015; Nanlohy et al., 2015); ecological functioning (Sierra-Correa and Cantera Kintz, 2015; Miller et al., 2017) with improved ecosystem services (Alongi, 2015; Nanlohy et al., 2015; Palacios and

1 Cantera, 2017); carbon mitigation (5.5.1.1); supporting livelihoods (Nanlohy et al., 2015; Nguyen et al.,
2 2017); and reductions in coastal infrastructure damage and community vulnerability to climate change
3 impacts. Managed retreat to counter coastal squeeze (Section 5.3) through improved governance, creation of
4 finance and land use planning can allow mangroves to move up the shoreline contour or down the latitudinal
5 gradient (Sierra-Correa and Cantera Kintz, 2015; Ward et al., 2016; Romañach et al., 2018). Therefore,
6 mangrove ecosystem-based adaptation responses can strengthen coastal ecosystem services through
7 shoreline stabilization and provide multiple co-benefits for coastal communities, like job creation and
8 improved access to ecosystem services (*high confidence*).
9
10 There are, however, examples where community mangrove restoration projects have resulted in maladaptive
11 outcomes, in which the resulting ecosystem degradation could not provide the ecosystem services required
12 (Nguyen et al., 2017; Romañach et al., 2018). Such maladaptation can be a result of poor governance
13 processes or a lack of community compliance with restoration plans. These examples emphasize the value of
14 designing effective governance to implement adaptation responses with broad community participation to
15 improve the climate risk reduction outcomes and co-benefits (Sierra-Correa and Cantera Kintz, 2015;
16 Nguyen et al., 2017) (*medium evidence, high agreement*).
17
18 Mangrove and other coastal ecosystems restoration and management can be applied through reducing non-
19 climatic hazards (Gilman et al., 2008; Ataur Rahman and Rahman, 2015; Sierra-Correa and Cantera Kintz,
20 2015; Ahmed and Glaser, 2016a; Nguyen et al., 2017; Romañach et al., 2018). Coastal and catchment
21 development, including wetland transformation and degradation (Miloshis and Fairfield, 2015; Schaeffer-
22 Novelli et al., 2016; Watson et al., 2017a; Schuerch et al., 2018a), the disruption of physical processes
23 impacting sedimentation rates (Watson et al., 2017a) and coastal squeeze compound coastal climate change
24 impacts like erosion, flooding and saltwater intrusion (Ondiviela et al., 2014; Miloshis and Fairfield, 2015;
25 Schaeffer-Novelli et al., 2016; Wigand et al., 2017) (Section 5.3). This reduces the ability of these
26 ecosystems to provide protection from wave and storm impacts, whilst positive feedbacks may occur that
27 cause a net release of carbon into the atmosphere, e.g. in saltmarshes (Wong et al., 2014a) (Section 5.4.1). In
28 some cases, effective inventions requires management at a broad spatial scale that includes a variety of
29 ecosystems, for example, including ecosystems like mussel beds on the seaward side of seagrass beds to
30 reduce wave energy and erosion (Ondiviela et al., 2014). Where sediment accretion matches the sea level
31 rise rate, wetlands and salt marshes provide effective coastal protection and other important ecosystem
32 services (*high confidence*).
33
34 Coastal dune systems are widely transformed globally. Human disturbance and the limited stabilizing ability
35 of dune vegetation are key causes of degradation (Onaka et al., 2015; Ranasinghe, 2016; MacDonald et al.,
36 2017; Pranzini, 2017; Salgado and Martinez, 2017; Vikolainen et al., 2017; Gracia et al., 2018), while
37 restoration efforts can be supported by both hard (Sutton-Grier et al., 2015; Pranzini, 2017) and soft (Sutton-
38 Grier et al., 2015; Vikolainen et al., 2017) engineering responses. Reduced coastal erosion (Sánchez-Arcilla
39 et al., 2016; Goreau and Prong, 2017; Vikolainen et al., 2017; Carro, 2018; Gracia et al., 2018) and flood
40 risk (Onaka et al., 2015; MacDonald et al., 2017; Nehren et al., 2017) through maintaining dunes as natural
41 buffers against wave energy (Nehren et al., 2017) can increase resilience to climate change impacts (Sutton-
42 Grier et al., 2015; Magnan and Duvat, 2018). Engineered responses and sand replenishment are considered
43 complementary approaches (Onaka et al., 2015; Martínez et al., 2017). Section 4.4.4.1 provides an overview
44 of sediment-based adaptation response measures, including cost estimates for beach nourishment and dune
45 maintenance, a discussion of co-benefits and drawbacks of combining hard and soft infrastructure measures,
46 and challenges with sourcing sediment for beach replenishment. In some cases dune restoration and sand
47 replenishment projects have not been successful, due to fire damage (Shumack and Hesse, 2017) or the rapid
48 loss of sand within replenishment schemes due to coastal processes and stakeholder rejection of adaptation
49 activities (Pranzini, 2017). Coastal dune restoration and beach replenishment are effective responses against
50 coastal erosion and flooding, where sufficient materials and space to implement are available (*medium
51 confidence*).
52
53 *5.5.2.1.4   Ecosystem-based adaptation*
54 There is a growing body of literature regarding the effectiveness and economics of ecosystem-based
55 adaptation (EBA). In addition to building resilience to climate change, EBA is expected to bring a wide
56 range of co-benefits that include increasing ecological complexity, with multiple ecosystem services, and
57 other economic co-benefits (Perkins et al., 2015; Perry, 2015; Moosavi, 2017; Scarano, 2017). The cost-

1   effectiveness of EBA approaches varies between marine ecosystem types; for example, coral reefs (Perkins
2   et al., 2015; Beetham et al., 2017; Elliff and Silva, 2017; Beck et al., 2018; Comte and Pendleton, 2018) and
3   salt-marshes (Ondiviela et al., 2014; Miloshis and Fairfield, 2015; Schaeffer-Novelli et al., 2016; Wigand et
4   al., 2017) performed best at reducing wave heights, whilst salt-marshes and mangroves were two to five
5   times cheaper than submerged breakwaters for wave heights of less than half a meter. Although low regrets,
6   win-win approaches like EBA are supported in the literature (Watkiss et al., 2014; FAO, 2018b), syntheses
7   of experience from context-specific practical implementation of EBA and assessment of their cost-
8   effectiveness are limited (Narayan et al., 2016). Therefore, EBA can be a cost-effective approach for
9   securing climate change-related ecosystem services with multiple co-benefits (*medium evidence, high*
10   *agreement*).
11
12   The application of EBA approaches can be more effective when incorporating local knowledge and
13   Indigenous knowledge and cultural practices into adaptation responses (Ataur Rahman and Rahman, 2015;
14   Perkins et al., 2015; Sutton-Grier et al., 2015; Sánchez-Arcilla et al., 2016; van der Nat et al., 2016). The
15   application of synergistic combinations of adaptation responses in multiple ecosystems can provide a range
16   of co-benefits, and this approach is strengthened when combined with socio-institutional approaches
17   (Kochnower et al., 2015; MacDonald et al., 2017). Research to improve and refine EBA approaches and
18   increase their specificity to local context is important for their effectiveness in reducing climate risks and
19   generating co-benefits (Sutton-Grier et al., 2015). Conversely, a lack of inclusion of local communities and
20   economic undervaluation of specific coastal and marine ecosystems, compounded by gaps in scientific data,
21   can undermine the potential effectiveness of EBA approaches (Perkins et al., 2015; Hernández-González et
22   al.; Narayan et al., 2016; Roberts et al., 2017).
23
24   Despite the abundance of EBA examples in the literature, knowledge gaps pertaining to their implementation
25   and limitations remain. Developing this literature could help with understanding context specific application
26   of EBA and improve their effectiveness (*medium confidence*).
27
28   *5.5.2.2   Human Systems:*
29
30   Many of the world's great cities lie within the coastal region, and climate change impacts put these cities,
31   their inhabitants and their economic activities at risk. Section 5.5.2.2 assesses the impacts of climate change,
32   adaptation response and benefits upon human systems, including coastal communities, built infrastructure,
33   fisheries and aquaculture, coastal tourism, government and health systems. Table 5.8 provides a summary of
34   the assessment, with citations provided in the Supplementary Material Table 5.7.
35
36   Poorly planned (Ataur Rahman and Rahman, 2015), located (Abedin et al., 2014; Betzold and Mohamed,
37   2017; Linkon, 2018) and managed urban settlements or human systems, driven by growing human coastal
38   populations (Perkins et al., 2015; Moosavi, 2017; Carter, 2018) and compounded by the disruption of coastal
39   and catchment physical processes (Nagy et al., 2014; Broto et al., 2015; Marfai et al., 2015; Kabisch et al.,
40   2017) and pollution (Zikra et al., 2015; Peng et al., 2017) are major human drivers of change compounding
41   the impacts of climate change.
42
43   Coastal Communities, Built Infrastructure and Fisheries and Aquaculture (Table 5.8) are likely to be
44   significantly affected through the disruption of coastal physical processes (DasGupta and Shaw, 2015;
45   Betzold and Mohamed, 2017; Hagedoorn et al., 2019) leading to coastal erosion, flooding, salt-water
46   intrusion and built infrastructure damage (Dhar and Khirfan, 2016; Hobday et al., 2016a; Jurjonas and
47   Seekamp, 2018) (*robust evidence, high agreement*). Ecosystem degradation and biodiversity loss will further
48   compound impacts in Coastal Communities and Fisheries and Aquaculture (Ataur Rahman and Rahman,
49   2015; Petzold and Ratter, 2015; Dhar and Khirfan, 2016), with sub-lethal species impacts like changes in the
50   productivity and distribution of fisheries target species reported for the latter (Gourlie et al., 2018; Nursey-
51   Bray et al., 2018; Pinsky et al., 2018) (*high confidence*). This is likely to result in decreased access to
52   ecosystem services (Asch et al., 2018; Cheung et al., 2018b; Finkbeiner et al., 2018) (*medium evidence, high*
53   *agreement*), local declines in agriculture and fisheries (Cvitanovic et al., 2016; Faraco et al., 2016) (*high*
54   *confidence*) and livelihood impacts (Harkes et al., 2015; Busch et al., 2016; Valmonte-Santos et al., 2016)
55   (*high confidence*) in Coastal Communities and Fisheries and Aquaculture, particularly increased food
56   insecurity and health risk in the latter (*high confidence*). These livelihood impacts are likely to increase

1   social vulnerability (*high confidence*). Businesses within Coastal Communities are likely to experience
2   disruptions and losses (*robust evidence, high agreement*).
3
4   *5.5.2.2.1   Coastal Communities*
5   This section describes a range of adaptation responses reported at the level of the individual or community.
6   Hard engineering responses included small scale hard infrastructure coastal defenses (Betzold and
7   Mohamed, 2017; Jamero et al., 2018), design responses at the household-level (Ataur Rahman and Rahman,
8   2015; Linkon, 2018) and retreat (Marfai et al., 2015). Ecosystem restoration and protection, particularly in
9   mangroves (Ataur Rahman and Rahman, 2015; Bennett et al., 2016; Jamero et al., 2018; Hagedoorn et al.,
10 2019) through community participation programmes (Barbier, 2015; Petzold and Ratter, 2015; Bennett et al.,
11 2016; Dhar and Khirfan, 2016; Jamero et al., 2018) was strongly supported in the literature as a means to
12 improve access to or storage of natural resources (*medium evidence, high agreement*).
13
14

Table 5.8: Summary of reported Adaptation responses (A), the Impacts (I) they aimed to address, and the expected Benefits (B) in human systems within Physical, Ecological, Social, Governance, Economic and Knowledge categories. Legend: a + sign indicates *robust evidence*, a triangle indicates *medium evidence* and an underline indicates *limited evidence*. Dark blue cells indicate *high agreement*, blue indicates *medium agreement* and light blue indicates either *low agreement* (denoted by presence of a sign) if sufficient papers were reviewed for an assessment or no assessment (if less than three papers were assessed per cell). Papers used for this assessment can be found in supplementary material SM5.6.

	Impacts (I)	Adaptation responses (A)	Coastal Communities I	A	B	Built Infrastructure I	A	B	Fisheries and Aquaculture I	A	B	Coastal Tourism I	A	B	Government I	A	B	Health I	A	B	Benefits (B)
Physical	Coastal physical processes disrupted	Supporting physical processes	+			+			+												Physical processes supported
	Catchment physical processes disrupted	Hard engineering responses		+		+	+														Coastal infrastructure resilience increased
	Coastal infrastructure damage	Soft engineering responses and buffers	+	_		+				_											Improved infrastructure functionality
	Disruption of urban systems	Integrated hard and soft engineering	Δ			_	+														Increased structural heterogeneity
	Land subsidence	Managed retreat and coastal realignment							_												
Ecological	Ecosystem degradation and loss	Ecosystem restoration and protection	+	+					+	+	_	_				_					Ecosystem/ ecological resilience supported
	Biodiversity and genetic diversity loss	Bioengineering	Δ						Δ	+											Physical processes supported
	Habitat range shifts	Assisted evolution and relocation																			Coastal infrastructure resilience increased
	Sub-lethal species impacts	Nature based solutions	_						Δ	_	+	_									Increased biodiversity
	Invasive alien species																				Habitat range shifts accommodated
																					Improved organismal fitness
																					Genetic heterogeneity supported
																					Strengthened socio-ecological system
Social	Decreased access to ecosystem services	Improving access to/ storage of natural resources	Δ	Δ					Δ	+											Access to sustainable ecosystem services
	Local decline in agriculture and fisheries	Improving agricultural or fisheries practices	+	_					+	+											Improved access to community services
	Increasing living costs	Supporting natured-based industries													_						Increasing resilience in human systems
	Livelihoods impacts	Sustainable resource use	+	_					+			Δ									Improved socio-economic services
	Increased food insecurity	Maintaining or switching livelihoods	_			_			+	_	+										Improved employment and livelihoods
	Public health risks increased	Community participatory programmes	_	+					+	+								_			Improved health
	Cultural and traditional knowledge impacts	Developing adaptive networks																			Improved community participation
	Gender-related impacts	Sustainable household management		_	_																Better informed communities
	Increased social vulnerability	Improving access to community services	+						+												Improved integration of knowledge systems
	Decreased access to local government services	Empowering communities and addressing inequality								Δ											Empowering women and children
	Socio-economic entrapment and decline	Building socio-ecological resilience								Δ											Increased adaptive capacity
	Global declines in food stocks																				Improved disaster preparedness
	Public areas access restrictions																				Empowered communities
	Decline in perceived value of human systems									_											Improved community cohesion
	Conflict and migration																				Reduced inequality

	Impacts (I)	Adaptation responses (A)	Coastal Communities I	A	B	Built Infrastructure I	A	B	Fisheries and Aquaculture I	A	B	Coastal Tourism I	A	B	Government I	A	B	Health I	A	B	Benefits (B)
Governance	Capacity challenges	Adopting/ mainstreaming sustainability policies					Δ			−						−					Political and institutional capacity developed
	Increased geopolitical tensions	Improving disaster response programmes								−											Strengthened participatory governance
	Growing inequalities	Improving implementation and coordination of policies								+	−					+	−				Better planning processes supported
		Development controls																			Improved coordination and decision making
		Evidence-based implementation															−				Improved implementation and policies
		Improving ICM/ MPAs								+						+					Better communication
		Horizontal/ vertical integration of governance																			Improved transparency and trust
		Developing partnerships and building capacity	−							−							Δ				Climate justice advanced
		Improving access to community services								−											Reduced conflict
		Pursuing climate justice																			Improved security
										+											Improved adaptive management
																					Development supported
Economic	Increased business and living costs	Improving financial resources availability								−											Increased revenue/ income
	Business disruptions and losses	Improving access to insurance products	+							+											Increased financial resources available
	Decreased value of assets/ products	Economic diversification																			Reduced operational and capital costs
		Improving access to international funding programmes																			Investment strengthened
Knowledge	Uncertainty for decision makers	Better monitoring and modelling					+	+		+	Δ			−		Δ					Informed decision making tools
		Improving planning processes				Δ	Δ														Improved co-production of knowledge
		Improving forecasting and early warning systems								Δ											Improved relevance of products
		Improving decision support frameworks				Δ				+			−			−					Improved education and outreach
		Improving participatory processes				Δ				+											Improved awareness
		Coordinating top down and bottom up approaches																			
		Integrating knowledge systems	+							Δ											
		Improving location and context specific knowledge	Δ																		
		Improving scientific communication								−											
		Stakeholder identification, outreach and education	−							Δ						−					

1   Social responses include increasing climate change awareness, improving participatory decision-making
2   through bottom-up approaches, community organization for action and engagements with local management
3   authorities (Dutra et al., 2015; Tapsuwan and Rongrongmuang, 2015; Galappaththi et al., 2017; Ray et al.,
4   2017; Cinner et al., 2018; Hagedoorn et al., 2019). In coastal communities, and indeed in most other sectors,
5   despite consensus on the importance of cooperation in tackling climate change (Elrick-Barr et al., 2016),
6   adaptation progress may be hampered by competing economic interests and worldviews (Hamilton and
7   Safford, 2015), which can be compounded by limited climate change knowledge (Nanlohy et al., 2015).
8   Factors like home ownership and a general future planning ability support resilience (Elrick-Barr et al.,
9   2016). Climate change adaptation capacity is shaped by historical path dependencies, local context and
10 international linkages, while action is shaped by science, research partnerships and citizen participation
11 (Hernández-Delgado, 2015; Sheller and León, 2016). Locally-context-specific data to guide appropriate
12 adaptation response remains a knowledge gap (Abedin and Shaw, 2015; Hobday et al., 2015; Lirman and
13 Schopmeyer, 2016; Williams et al., 2016)
14
15 Coastal and oceanic adaptation responses are greatly complicated by the presence of competing interests
16 (either between user-groups, communities or nations), where considerations other than climate change need
17 to be incorporated into cooperation agreements and policy (Wong et al., 2014a). The deployment of either
18 built or natural protection systems, or adopting a 'wait and see' approach, is subject to the social acceptance
19 of these approaches in communities (Poumadère et al., 2015; Sherren et al., 2016; Torabi et al., 2018).
20 Similarly, the willingness to move away from climate change-impacted zones is dependent upon a range of
21 other socio-economic factors like age, access to resources and crime (Bukvic et al.; Rulleau and Rey-Valette,
22 2017). Adaptation to climate change includes a range of non-climatic and social variables that complicate
23 implementation of adaptation plans (*robust evidence, high agreement*).
24
25 Improving community participation and integrating knowledge systems (local, traditional and scientific)
26 supports coastal community adaptation responses (*high confidence*), providing improved co-production of
27 knowledge (*medium evidence, high agreement*), improved community awareness (*medium evidence, medium*
28 *agreement*) and better-informed, more cohesive coastal communities (*limited evidence, medium agreement*).
29
30 *5.5.2.2.2   Built Infrastructure*
31 Built infrastructure impacts are most frequently addressed through hard engineering approaches including:
32 construction of groins, seawalls, revetments, gabions and breakwaters (Friedrich and Kretzinger, 2012;
33 Vikolainen et al., 2017); improving drainage and raising the height of roadways and other fixed-location
34 infrastructure (Perkins et al., 2015; Becker et al., 2016; Colin et al., 2016; Asadabadi and Miller-Hooks,
35 2017; Brown et al., 2018a); erosion control systems (Jeong et al., 2014); and the relocation of infrastructure
36 (Friedrich and Kretzinger, 2012; Colin et al., 2016). Nature-based responses are increasingly being reported
37 as complementary and supporting tools (van der Nat et al., 2016; Kabisch et al., 2017; Gracia et al., 2018)
38 using ecological engineering (Perkins et al., 2015; van der Nat et al., 2016; Moosavi, 2017) combined with
39 innovative construction strategies (Moosavi, 2017).
40
41 When implemented together, hard and soft engineering responses provide social (Gracia et al., 2018)
42 (Martínez et al., 2018; Woodruff, 2018) and ecological (Perkins et al., 2015; van der Nat et al., 2016; Gracia
43 et al., 2018) co-benefits with reduced damage costs (Jeong et al., 2014). Constraints on implementation
44 include the space and extra cost required by ecological infrastructure, sub-optimal performance when
45 impacted by natural physical processes that are disrupted (Gracia et al., 2018) or restrictions associated with
46 governance (Vikolainen et al., 2017). Adaptation planning including local communities can improve
47 implementation and help fill knowledge gaps (Kaja and Mellic, 2017; Moosavi, 2017; Martínez et al., 2018;
48 Mikellidou et al., 2018). Benefits include increased resilience in coastal infrastructure and better informed
49 decision-making tools (*medium confidence*),
50
51 *5.5.2.2.3   Adaptation in fisheries and aquaculture*
52 Sixty percent of assessed species are projected to be at high risk from both overfishing and climate change
53 by 2050 (RCP8.5), particularly tropical and sub-tropical species (Cheung et al., 2018b). Overfishing is one
54 of the most important non-climatic drivers affecting the sustainability of fisheries (Islam et al., 2013; Heenan
55 et al., 2015b; Faraco et al., 2016; Dasgupta et al., 2017; Cheung et al., 2018b; Harvey et al., 2018). Pursuing
56 sustainable fisheries practices under a low emissions scenario would decrease risk by 63%. This highlights

1   the importance of effective fisheries management (Gaines et al., 2018). Eliminating overfishing would,
2   however, require reducing current levels of fishing effort, with a potential short-term reduction in catches
3   impacting livelihoods and the food security of coastal communities (Hobday et al., 2015; Dey et al., 2016;
4   Rosegrant et al., 2016; Campbell, 2017; Finkbeiner et al., 2018). Despite consensus on the effectiveness of
5   eliminating overfishing in supporting climate change adaptation in fisheries (*robust evidence, high
6   agreement)*, successful adaptation outcomes remain aspirational.
7
8   Range shifts under ocean warming (Section 5.2.3) will alter the distribution of fish stocks across political
9   boundaries, thus demand for transboundary fisheries management will increase. Redistribution of
10 transboundary fish stocks between countries (Ho et al., 2016; Gourlie et al., 2017; Asch et al., 2018) could
11 destabilize existing international fisheries agreements and increase the risk of international conflicts (Section
12 5.4.2). Adaptation to reduce risks in international fisheries management could involve improving planning
13 for cooperative management between countries informed by reliable predictions (Payne et al., 2017) and
14 projections (Pinsky et al., 2018) of species shifts and associated uncertainties. Cooperative international
15 fisheries arrangements, such as flexible fishing effort allocation and adaptive frameworks (Colburn et al.,
16 2016; Cvitanovic et al., 2016; Faraco et al., 2016) may also improve the robustness of fisheries management
17 (Miller et al., 2013). Thus, although range shifts pose significant challenges to transboundary fisheries
18 management, proactive planning and adjustment of fisheries management arrangements, informed by
19 scientific projections, could help improve adaptive capacity (*medium confidence*). The effectiveness of
20 incorporating MPAs as an adaptation strategy to climate change can be improved by considering climate
21 impacts in the design of MPAs (*medium, high agreement*).
22
23 Improving integrated coastal management and better planning for marine protected areas by incorporating
24 projected shifting biological communities, abundance and life history changes (Álvarez-Romero et al., 2018)
25 due to climate change could contribute towards improved fisheries adaptive management by, for example,
26 increasing resilience of habitats, providing refugia for species with shifting distributions and by conserving
27 biodiversity (Faraco et al., 2016; Valmonte-Santos et al., 2016; Dasgupta et al., 2017; Le Cornu et al., 2017;
28 Roberts et al., 2017; Asch et al., 2018; Cheung et al., 2018b; Harvey et al., 2018; Jones et al., 2018; O'Leary
29 and Roberts, 2018) (Sections 5.2.3, 5.3, 5.4.1), but MPAs may also reduce access to subsistence fishers,
30 increasing their vulnerability to food insecurity (Bennett et al., 2016; Faraco et al., 2016). The global area of
31 MPAs is rapidly increasing towards the United Nations' target of 10% of the global ocean. While this is
32 encouraging, it is estimated that only 2% of the ocean is well enough managed, as described in (Edgar et al.,
33 2014), to meet conservation goals (Sala et al., 2018). Improving the implementation and coordination of
34 policies, and improving integrated coastal management and MPAs have emerged in the literature as
35 important adaptation governance responses (*robust evidence, medium agreement*).
36
37 Governance responses to support adaptation in fisheries communities include conducting vulnerability
38 assessments, improving monitoring of ecosystem indicators and evaluating management strategies (Himes-
39 Cornell and Kasperski, 2015b; Busch et al., 2016). Socio-economic factors like access to alternative income,
40 mobility, gender and religion collectively shape a community's adaptation response (Arroyo Mina et al.,
41 2016). In West Africa, the industrial fishery response to climate change-induced reductions in landings was
42 the expansion of fishing grounds, which increased operational costs (Belhabib et al., 2016). This response is
43 not available to artisanal and local fishing communities, who are considered highly vulnerable (Kais and
44 Islam, 2017). Access to finance to support these communities or their governments could help them reach
45 novel fishing grounds, and, therefore, potentially reduce their vulnerability. Food security linked to fisheries
46 depends on stock recovery, but also on access to and distribution of the harvest, as well as gender
47 considerations (Béné et al., 2015). Hence, granting preferential access to dependent coastal communities
48 should be considered in examining policy options. Other adaptation responses include improved fishing gear
49 and technology, use of Fish Aggregating Devices and uptake of insurance products (Zougmoré et al., 2016).
50 See (FAO, 2018b) for a summary of possible adaptation responses. Community response as a part of climate
51 change adaptation for local fisheries is an important element in assessing adaptive capacit**y** (*medium
52 evidence, good agreement*),
53
54 Fisheries management strategies depend heavily upon data collection and monitoring systems. These include
55 the accuracy of data collected in respect of predicting environmental conditions, over time scales from
56 months to decades (Dunstan et al., 2017), effective monitoring and evaluative mechanisms (Le Cornu et al.,

1 2017; Gourlie et al., 2018), controlling for aspects of fish population dynamics like recruitment success and
2 fish movement (Mace, 2001). Seasonal to decadal climate prediction systems allow for skillful predictions of
3 climate variables relevant to fisheries management strategies (Hobday et al., 2016b; Payne et al., 2017).
4 Effective fisheries adaptation responses will require knowledge development including better monitoring,
5 modelling and improving decision support frameworks (*medium evidence, high agreement*) and improving
6 forecasting and early warning systems (*medium evidence, medium agreement*).

8 In considering a participatory decision-making approach for fisheries management that responds to climate
9 change, Heenan et al. (2015b) provided a number of key elements that contribute towards a successful
10 outcome. These include expert knowledge of climate change threats to fish habitats, stocks and landings, the
11 necessity of transdisciplinary collaboration and stakeholder participation, broadening the range and scope of
12 fisheries systems and increased commitment of resources and capacity. This was considered in the context of
13 the ability of developing countries to sustainably exploit fisheries resources and related ecosystems. More
14 research is required on socio-ecological responses to climate change-impacts on fishery communities,
15 including such aspect as like risk reduction, adaptive capacity through knowledge attainment and social
16 networks, developing alternative skills and participatory approaches to decision-making (Dubey et al., 2017;
17 Shaffril et al., 2017; Finkbeiner et al., 2018). Important fisheries adaptation responses in relation to
18 knowledge management include improving participatory processes (*robust evidence, high agreement*),
19 integrating knowledge systems (*medium evidence, high agreement*), and stakeholder identification, outreach
20 and education (*medium evidence, medium agreement*). Ecosystem based adaptation, community
21 participatory programs, and improving agricultural and fisheries practices are very strongly supported in the
22 literature (*high confidence*).

24 Less still is known about how climate change will affect the deep oceans and its fisheries (Section 5.2.3 and
25 5.2.4), the vulnerability of its habitats to fishing disturbance and future effects on resources not currently
26 harvested (FAO, 2019). Johnson et al. (2019) concluded that in a 20- to 50-year timeframe, the effectiveness
27 of virtually all North Atlantic deep-water and open ocean area-based management tools can be expected to
28 be affected. They concluded that more precise and detailed oceanographic data are needed to determine
29 possible refugia, and more research on adaptation and resilience in the deep sea is needed to predict
30 ecosystem response times.

32 As with fisheries, community- and ecosystem-based adaptation responses, an integrated coastal management
33 framework is considered useful for planning for anticipated challenges for aquaculture (Ahmed and Diana,
34 2015b; FAO, 2018b). Where in-situ adaptation is not possible, translocation and polyculture (Ahmed and
35 Diana, 2015a; Bunting et al., 2017) have been suggested as appropriate responses, but this would suit
36 commercial rather than subsistence interests. Policy, economic, knowledge and other types of support are
37 required to build socio-ecological resilience of vulnerable coastal aquaculture communities (Harkes et al.,
38 2015; Bunting et al., 2017; Rodríguez-Rodríguez and Bande Ramudo, 2017), which requires a deep
39 understanding of the nature of stressors and a commitment for collective action (Galappaththi et al., 2017).
40 Climate resilient pathway development (see Cross-Chapter Box 2) is considered a useful framework for Sri
41 Lankan shrimp aquaculture (Harkes et al., 2015). Another example of successful aquaculture adaptation is
42 the employment of near-real time monitoring technology to track the carbonate chemistry in water to reduce
43 bio-erosion in shellfish from acidification (Barton et al., 2015; Cooley et al., 2016). Numerous adaptation
44 responses are available for aquaculture, but some options, like translocation and technological responses
45 may not be available to subsistence-based communities (*medium evidence*).

47 An example of eco-engineering-based adaptation option in seaweed aquaculture under climate change is
48 artificial upwelling, as shown by experiments and observations. Artificial upwelling powered by green
49 energy (solar, wind, wave, or tidal energy) to seaweeds (Jiao et al., 2014b; Zhang et al., 2015; Pan and
50 Schimel, 2016) can moderate the amount of deep water upwelled to the euphotic zone to just meet the
51 demands of nutrients and DIC by the seaweed for photosynthesis, while avoiding the acidification and
52 hypoxia that often occur in natural upwelling systems (Jiao et al., 2018a; Jiao et al., 2018b) (*high
53 confidence*). Such artificial upwelling based eco-engineering may also gradually release the 'bomb' of rich
54 nutrients and hypoxia in the bottom water, which could otherwise breakout following storms (Daneri et al.,
55 2012) (*high confidence*).

5.5.2.3.4   Coastal Tourism

The coastal tourism economic sector is highly sensitive to climate change. Tourism response, in terms of mitigating carbon emissions and adapting to climate change impacts, are assessed here. Coastal tourism is likely to be impacted by ecosystem degradation and loss (*limited evidence, medium agreement*), which underscores the importance of nature-based tourism. An example of coastal erosion in Latin America illustrates this, whereby sea level rise interacting with non-climate change impacts including sand mining, inappropriate development and habitat destruction (e.g. mangroves), resulted in declines in tourism (Rangel-Buitrago et al., 2015). The management recommendation was appropriate legislation with a marine spatial planning emphasis, enforcement, sustainable funding mechanisms and support networks for decision-making.

Climate change impacts upon tourism are nuanced and not restricted to just physical impacts on tourism establishments (Biggs et al., 2015). Understanding the drivers of tourist choices could help support adaptation in the industry through marine spatial planning processes (Papageorgiou, 2016). For example, in an survey ranking mitigation and adaptation responses in Greece, tourists prioritised rational energy use, energy efficiency and water saving measures (Michailidou et al., 2016b). Location specific information of tourist choices could help shape local industries. In one example from the Thailand dive industry, climate change adaptation responses of participants were reported to be based on misconceptions about climate change and personal observations (Tapsuwan and Rongrongmuang, 2015). To improve community-based adaptation, efforts aimed at broadening the level of awareness about climate change could improve decision-making processes (Tapsuwan and Rongrongmuang, 2015). Tourist behaviour is shaped by changing ocean physical processes and degrading ecosystems at tourist destinations, which drive destination changes, economic flows and market share adjustments. (Bujosa et al., 2015; De Urioste-Stone et al., 2016).

It is very likely that climate change will have direct and nuanced impacts upon coastal tourism. Improving decision support frameworks (*low evidence, medium agreement*) for better-informed decision making tools could contribute towards increasing resilience in coastal tourism (*low evidence, limited agreement*).

5.5.2.2.5   Government responses

Government responses included adopting and mainstreaming sustainability policies, including investments and policies for climate change (Aylett, 2015; Buurman and Babovic, 2016) and applying the precautionary principle in the absence of precise scientific guidance (Johnson et al., 2018). Developing adequate governance and management systems (Johnson et al., 2018), strengthening capacity (Gallo et al., 2017; Paterson et al., 2017), increasing cooperation (Nunn et al., 2014; Gormley et al., 2015) and aligning policies of local authorities (Porter et al., 2015; Gallo et al., 2017; Rosendo et al., 2018) could help to improve implementation (Sano et al., 2015; Elsharouny, 2016). This includes planning for marine protected areas and improving integrated coastal management (Abelshausen et al., 2015; Roberts et al., 2017; Rosendo et al., 2018) by incorporating climate science (Hopkins et al., 2016; Johnson et al., 2018) to optimize priority marine habitats (Gormley et al., 2015; Jones et al., 2018). An advantage of integrated coastal management is that it helps manage the interactions between multiple climate and non-climatic drivers of coastal ecosystems and sectors. Incorporating stakeholder participation with Local Knowledge and Indigenous Knowledge could help to reduce the risk of maladaptation, and increase buy-in for implementation (Serrao-Neumann et al., 2013). Improving participatory processes strengthens governance decision making and flexible risk management processes (Gerkensmeier and Ratter, 2018; Rosendo et al., 2018), while stimulating bi-directional knowledge flow and improving social learning (Abelshausen et al., 2015).

Technology for environmental monitoring, for example using drones (Clark, 2017), web-based coastal information systems (Mayerle et al., 2016; Newell and Canessa, 2017), the Internet of Things and Machine Learning solutions promise to improve the local scale knowledge base, which should improve climate adaptation planning and resilience effort and environmental management decisions (Conde et al., 2015). Where such knowledge gaps persist, the implementation of climate change adaptation measures could proceed on the basis of a set of general principals of best practice (Sheaves et al., 2016; Thorne et al., 2017).

Benefits of effective government adaptation response includes the promotion of sustainable use, development and protection of coastal ecosystems (Rosendo et al., 2018) and the protection of biodiversity through setting appropriate conservation priorities (Gormley et al., 2015). Improved governance includes

1   consideration of social processes in risk management (Gerkensmeier and Ratter, 2018; Rosendo et al., 2018)
2   and improved systematic conservation planning (Johnson et al., 2018). At a local level, this translates into
3   sustained service delivery (Aylett, 2015), improved rationality and effective policy-making (Serrao-
4   Neumann et al., 2013; Rosendo et al., 2018).
5
6   Improving the implementation and coordination of policies and improving integrated coastal management
7   are both considered important climate change adaptation governance responses (*robust evidence, high*
8   *agreement*), as are developing partnerships and building capacity (*medium evidence, high agreement*) and
9   adopting or mainstreaming sustainability policies (*limited evidence, medium agreement*). Benefits include
10   improved ecosystem resilience, better planning processes, implementation and policies (all *limited evidence,*
11   *medium agreement*).
12
13   *5.5.2.3   Ocean-based Climate Change Adaptation Frameworks*
14
15   Adaptation action in pursuit of a climate resilient development pathway is likely to have a deeper
16   transformative outcome than stepwise or *ad hoc* responses (Cross-Chapter Box 2 in Chapter 1). Recent
17   literature highlighting the effectiveness of components of adaptation planning includes quantitative
18   assessments of vulnerability in ecosystems (Kuhfuss et al., 2016), species (Cheung et al., 2015; Cushing et
19   al., 2018), and communities (Islam et al., 2013; Himes-Cornell and Kasperski, 2015b), and integrated
20   assessments of all of the above (Peirson et al., 2015; Kaplan-Hallam et al., 2017; McNeeley et al., 2017;
21   Ramm et al., 2017; Mavromatidi et al., 2018). Seasonal and decadal forecasting tools have improved rapidly
22   since AR5, especially in supporting management of living marine resources (Payne et al., 2017) and
23   modelling to support decision-making processes (Čerkasova et al., 2016; Chapman and Darby, 2016; Jiang
24   et al., 2016; Justic et al., 2016; Joyce et al., 2017; Mitchell et al., 2017). Decision-making processes are
25   supported by economic evaluations (Bujosa et al., 2015; Jones et al., 2015), evaluations of ecosystem
26   services (MacDonald et al., 2017; Micallef et al., 2018), participatory processes (Byrne et al., 2015) and
27   social learning outcomes, the development of adaptation pathways, frameworks and decision making
28   (Buurman and Babovic, 2016; Dittrich et al., 2016; Michailidou et al., 2016a; Osorio-Cano et al., 2017;
29   Cumiskey et al., 2018), and indicators to support evaluation of adaptation actions (Carapuço et al., 2016;
30   Nguyen et al., 2016) through monitoring frameworks (Huxham et al., 2015). Climate change adaptation
31   responses are more effective when developed within institutional frameworks that include effective planning
32   and across-sector integration.
33
34   Evidence-based decision-making for climate adaptation is strongly supported in the literature (Endo et al.,
35   2017; Thorne et al., 2017) through better understanding of coastal ecosystems and human adaptation
36   responses (Dutra et al., 2015; Cvitanovic et al., 2016), as well as consideration of non-climate change-related
37   factors. Relevant research includes the topics of: multiple-stakeholder participatory planning (Archer et al.,
38   2014; Abedin and Shaw, 2015); trans-boundary ocean management (Gormley et al., 2015; Williams et al.,
39   2016); ecosystem-based adaptation (Hobday et al., 2015; Dalyander et al., 2016; McNeeley et al., 2017;
40   Osorio-Cano et al., 2017); and community-based adaptation with socio-economic outcomes (Merkens et al.,
41   2016). Research on applying 'big data' and high end computational capabilities could also help develop a
42   comprehensive understanding of climate and non-climate variables in planning for coastal adaptation
43   (Rumson et al., 2017). New knowledge from these research areas could substantially improve planning,
44   implementation and monitoring of climate adaptation responses for marine systems, if research processes are
45   participatory and inclusive (*medium confidence*).
46
47   Despite such interest, evaluations of the planning, implementation and monitoring of adaptation actions
48   remain scarce (Miller et al., 2017). In a global analysis of 401 local governments, only 15% reported on
49   adaptation actions (mostly large cities in high income countries), and 18% reported on planning towards
50   adaptation policy (Araos et al., 2016). Thus, integrated adaptation planning with non-climate change related
51   impacts remains an under-achieved ambition, especially in developing countries (Finkbeiner et al., 2018).
52   Challenges reported for adaptation planning include uncoordinated, top-down approaches, a lack of political
53   will, insufficient resources (Elias and Omojola, 2015; Porter et al., 2015), and access to information (Thorne
54   et al., 2017).
55

1   Characteristics of successful adaptation frameworks include: a robust but flexible approach, accounting for
2   deep uncertainty through well-coordinated participatory processes (Dutra et al., 2015; Jiao et al., 2015;
3   Buurman and Babovic, 2016; Dittrich et al., 2016); well-developed monitoring systems (Barrett et al., 2015;
4   Bell et al., 2018b); and taking a whole systems approach (Sheaves et al., 2016), with the identification of co-
5   benefits for human development and the environment (Wise et al., 2016). The coastal adaptation framework
6   literature is dominated by Australian, North American and European cities, with fewer studies from African
7   and Caribbean sites, least developed countries and small island developing states (Kuruppu and Willie, 2015;
8   Torresan et al., 2016).
9
10   In contrast with the many examples of proposed frameworks for climate-resilient coastal adaptation, few
11   studies have assessed their success, possibly due to the time-lag between implementation, monitoring,
12   evaluation and reporting. Nevertheless, there is substantial support for 'no regrets' approaches addressing
13   both proximate and systematic underlying drivers of vulnerability (Sánchez-Arcilla et al., 2016; Pentz and
14   Klenk, 2017; Zandvoort et al., 2017) with leadership, adaptive management, capacity and the monitoring and
15   evaluation of actions considered useful in governance responses (Dutra et al., 2015; Doherty et al., 2016).
16   More extensive learning processes could help build decision-makers' capacity to tackle systemic drivers,
17   guide pursuance climate change appropriate policies (FAO, 2018b) and to scrutinize potentially maladaptive
18   infrastructural investments (Wise et al., 2016). More effective coordination across a range of stakeholders,
19   within and between organizations, especially in developing countries, would strengthen the global coastal
20   adaptation response (*medium confidence*).
21
22   *5.5.2.4   The Role of Education and Local Knowledge in Adapting to Climate Change.*
23   Education can help improve understanding of issues related to climate change and increase adaptive capacity
24   (Fauville et al., 2011; Marshall et al., 2013; von Heland et al., 2014; Pescaroli and Magni, 2015; Tapsuwan
25   and Rongrongmuang, 2015; Wynveen and Sutton, 2015). Participatory processes can facilitate the
26   development of networks between coastal communities and environmental managers for the purposes of
27   developing and implementing adaptation strategies (Wynveen and Sutton, 2015). Education, combined with
28   other forms of institutional support empowers fisheries and aquaculture communities (Table 5.8) to make
29   informed adaptation decisions and take action (*medium evidence, medium agreement*).
30
31   Local knowledge and Indigenous knowledge systems can complement scientific knowledge by, for example,
32   improving community ability to understand their local environment (Andrachuk and Armitage, 2015),
33   forecast extreme events (Audefroy and Sánchez, 2017) and help to increase community resilience (Leon et
34   al., 2015; Sakakibara, 2017; Cinner et al., 2018; Panikkar et al., 2018). Committing resources could
35   strengthen local level adaptation planning (Alam et al., 2016; Novak Colwell et al., 2017) through the
36   inclusion of cultural practices (Audefroy and Sánchez, 2017; Fatorić and Seekamp, 2017) and Indigenous
37   Knowledge systems (Kuruppu and Willie, 2015; von Storch et al., 2015). Local Knowledge can, however,
38   act as a barrier to adaptation where there is a strong dependency upon such knowledge for immediate
39   survival, to the detriment of long-term adaptation planning (Marshall et al., 2013; Metcalf et al., 2015).
40   There is evidence, however, to suggest that vulnerability in fisheries communities and coastal tourism
41   operators with high levels of Local Knowledge is reduced where they have a correspondingly high level of
42   adaptive capacity (Marshall et al., 2013). Resource users with high levels of Local Knowledge may also be
43   able to identify signals of change within their environment, and recognize the need to adapt. In these
44   instances, fishers with higher Local Knowledge are expected to demonstrate a higher adaptive capacity than
45   fishers with lower Local Knowledge, and can be expected to progress towards developing new strategies to
46   combat the impacts of climate change (Kittinger et al., 2012). In these instances, Local Knowledge acts to
47   promote adaptation (*medium confidence*).
48
49   Localized, individual-scale behaviors can aggregate rapidly and contribute to the global adaptation response.
50   This can be supported by clear messaging that clarifies the role of individuals, households and local
51   businesses in addressing climate change. Coastal communities can improve the co-production of climate
52   change knowledge (*medium evidence, good agreement*) through the integration of knowledge systems (Table
53   5.8). In fisheries and aquaculture, better-informed decision-making tools (*medium evidence, medium
54   agreement*) are supported by improved participatory processes (*high confidence*), integrating knowledge
55   systems (*medium evidence, good agreement*) and improving decision support frameworks (*medium evidence,
56   medium agreement*).

5.5.2.5   Costs and Limits for Coastal Climate Change Adaptation

Challenges persist in conducting economic assessments for built infrastructure adaptation due to complicated uncertainties such as the accuracy of climate projections and limited information regarding paths for future economic growth and adaptation technologies. Annual investment and maintenance costs of protecting coasts were projected to be USD 12 billion to USD 71 billion (Hinkel et al., 2014), which was considered significantly less than damage costs in the absence of such action. In an analysis of twelve Pacific island countries, 57% of assessed built infrastructure was located within 500m of coastlines, requiring a replacement value of USD 21.9 billion. Substantial coastal adaptation costs (and international financing) are likely to be required in these countries (*medium confidence*).

In West African fisheries, loss of coastal ecosystems and productivity are estimated to require 5–10% of countries' Gross Domestic Product in adaptation costs (Zougmoré et al., 2016). Similarly, for Pacific Islands and Coastal Territories, fisheries adaptation will require significant investment from local governments and the private sector (Rosegrant et al., 2016), with adaptation costs considered beyond the means of most of these countries (Campbell, 2017). In Small Island Developing States, tourism could provide the funding for climate change adaptation, but concerns with creating investment barriers, assumptions around cost-effectiveness and consumer driven demand remain barriers (Hess and Kelman, 2017). Marine Protected Areas, with multiple co-benefits, are considered a cost-effective strategy (Byrne et al., 2015). In 2004, the annual cost of managing 20 – 30% of global seas as MPAs was estimated at between USD 5 – 19 billion, with the creation of approximately one million jobs (Balmford et al., 2004).

Estimating adaptation costs is challenging because of wide ranging regional responses and uncertainty (Dittrich et al., 2016). Despite these challenges, the protection from flooding and frequent storms that coral reefs provide has been quantified by (Beck et al., 2018), who estimated that without reefs, damage from flooding and costs from frequent storms would double and triple respectively, while countries from Southeast Asia, East Asia and Central America could each save in excess of USD 400 million through good reef management. Although quantifying global adaptation costs remains challenging because of a wide range of regional responses and contexts, it is likely that managing ecosystems will contribute towards reducing costs associated with climate change associated coastal storms (*medium confidence*). Further research evaluating natural infrastructure is required (Roberts et al., 2017) to better understand costs and benefits of EBA.

There is a broad range of reported barriers and limits to climate change adaptation for both ecosystems and human systems. Coastal ecosystem-based adaptation can be physically constrained by space requirements and coastal squeeze (Sutton-Grier et al., 2015; Robins et al., 2016a; Sánchez-Arcilla et al., 2016; Ahmed et al., 2017; Peña-Alonso et al., 2017; Salgado and Martinez, 2017; Triyanti et al., 2017; Schuerch et al., 2018a), while the pace of climate change may exceed the adaptive capacity of ecosystems e.g. sea level rise may outpace the vertical reef accretion rate (Beetham et al., 2017; Elliff and Silva, 2017; Joyce et al., 2017). One technical limit for coral reef adaptation is that tools have not yet been developed for large-scale implementation (van Oppen et al., 2017a). Ecosystems may also have physiological and ecological constraints which are exceeded by climate change impacts (Miller et al., 2017; Wigand et al., 2017), and the recovery periods of natural systems (Gracia et al., 2018) and for ecological succession (Salgado and Martinez, 2017) may be outpaced by climate change impacts. The performance of ecosystems in EBA projects may be inhibited by the poor condition of the ecosystem (Nehren et al., 2017), highlighting the importance of effective implementation (Salgado and Martinez, 2017).

Social and cultural norms with conflicting and competing values (Miller et al., 2017), public lack of knowledge on climate change and distrust of information sources (Wynveen and Sutton, 2015), as well as populations increasingly distanced from, and unconcerned about nature (Romañach et al., 2018), may constrain ecosystem-based adaptation response. Examples of governance adaptation constraints include: inadequate policy, governance and institutional structures (Sánchez-Arcilla et al., 2016; Miller et al., 2017; Wigand et al., 2017); limited capacity (Sutton-Grier et al., 2015; Thorne et al., 2017); ineffective implementation (Nguyen et al., 2017; Comte and Pendleton, 2018); and poor enforcement (Nguyen et al., 2017). Governance constraints are compounded by lack of finances (Miller et al., 2017), financial costs of

1   design and implementation (Gallagher et al., 2015) and the high cost of coastal land (Gracia et al., 2018),
2   although ecosystem-based adaptation is considered cheaper than man-made structures (Nehren et al., 2017;
3   Salgado and Martinez, 2017; Vikolainen et al., 2017; Gracia et al., 2018).
4
5   Knowledge limitations can include a lack of data (Sutton-Grier et al., 2015; Wigand et al., 2017; Romañach
6   et al., 2018), for example, when an absence of baseline data may undermine coastline management (Perkins
7   et al., 2015). Scale-relevant information may be required for local decision-making (Robins et al., 2016a;
8   Thorne et al., 2017) and to comply with localized design requirements (Vikolainen et al., 2017). Other
9   knowledge barriers include inherent uncertainties in models (Schaeffer-Novelli et al., 2016) and complexity
10 of coastal systems (Wigand et al., 2017). A more nuanced knowledge barrier is the disconnect between
11 scientific, community and decision-making processes (Romañach et al., 2018).
12
13 Substantial knowledge gaps are reported for ecosystem-based adaptation, including: restoration of coral reef
14 systems as an adaptation tool (Comte and Pendleton, 2018); managing mangrove and human response to
15 climate change (Ward et al., 2016); advancing coastal EBA science by quantifying ecosystem services
16 (Hernández-González et al.); and evaluating natural infrastructure (Roberts et al., 2017). Few syntheses of
17 the context-specific application and cost-effectiveness of EBA approaches are to be found in the literature
18 (Narayan et al., 2016).
19
20 Human systems have similar limitations. Improved understanding of limitations in built-infrastructure, beach
21 nourishment and nature-based adaptation responses, especially with respect to cost effectiveness and
22 resilience, would substantially aid shoreline stabilisation attempts (Mackey and Ware, 2018). For artisanal
23 fisheries, a range of physical and socio-institutional limits and barriers to adaptation have been reported,
24 including increasing occurrence and severity of storms limiting fishing time, technologically poor boats and
25 fishing equipment and lack of access to credit and markets, among others (Islam et al., 2013). Conflicting
26 interests and values of stakeholders (Evans et al., 2016), the path-dependent nature of organisations and
27 resistance to change (Evans et al., 2016) and inadequate collaboration and public awareness (Oulahen et al.,
28 2018) have been reported as socio-institutional barriers. A knowledge gap persists in understanding how
29 such limits and barriers interact to suppress adaptation response.
30
31 In some communities, climate change may not be prioritised in the face of chronic, daily challenges to
32 secure livelihoods (Esteban et al., 2017; Fischer, 2018) or risk severity may be underestimated due to a high
33 frequency of exposure in the recent past (Esteban et al., 2017). In a world with competing risks and urgent
34 priorities, some local inhabitants appear to be unable to avoid, or are willing to carry, the risk associated with
35 a climate impact in order to meet other, more pressing needs. This example reflects the reality of many poor,
36 informal settlement dwellers in coastal areas around the world (*medium confidence*). Other human system
37 barriers to effective adaptation action include insufficient climate change knowledge, inappropriate coping
38 strategies, high dependency upon natural resources, level of exposure to hazards and weak community
39 networks (Islam et al., 2013; Nanlohy et al., 2015; Lohmann, 2016; Koya et al., 2017; Senapati and Gupta,
40 2017; Cumiskey et al., 2018).
41
42 In summary, we conclude that the broad range of reported barriers and limits to climate change adaptation
43 for ecosystem and human system adaptation responses (*high confidence*). Limitations include the space that
44 ecosystems require, non-climatic drivers and human impacts that need to be addressed as part of the
45 adaptation response, the lowering of adaptive capacity of ecosystems because of climate change, and slower
46 ecosystem recovery rates relative to the recurrence of climate impacts, availability of technology, knowledge
47 and financial support, and existing governance structures (*medium confidence*). (5.5.2.5)
48
49 *5.5.2.6   Summary*
50
51 There has been a substantial amount of literature focused on coastal and oceanic adaptation since AR5.
52 Socio-institutional adaptation responses are the more numerous of the three types of adaptation responses
53 assessed in this chapter. There is broad agreement that hard engineering responses are optimally supported
54 by ecosystem-based adaptation approaches, and both approaches should be augmented by socio-institutional
55 approaches for adaptation (*high confidence*) (Nicholls et al., 2015; Peirson et al., 2015; Sánchez-Arcilla et
56 al., 2016; van der Nat et al., 2016; Francesch-Huidobro et al., 2017; Khamis et al., 2017). In planning

adaptation responses, awareness-raising and stakeholder engagement processes are important for buy-in and ownership of responses (*robust evidence, high agreement*) as is institutional capacity within local government organizations, whose importance in coastal adaptation initiatives has been emphasized in the recent literature (*robust evidence, high agreement*). With all three types of adaptation, basic good governance and effective implementation of service delivery processes are prerequisites for successful adaptation planning and response.

5.5.3   *Governance Across All Scales*

There are many global, regional, national and local governance structures with interests in climate-driven ocean warming, acidification, de-oxygenation and sea-level rise, and their impacts on marine ecosystems and dependent communities (Galland et al., 2012; Stephens, 2015; Fennel and VanderZwaag, 2016; Diamond, 2018). The legal, policy and institutional response is therefore shared by many institutions developed for a number of distinct but inter-related fields, including governance regimes for ocean systems, climate change, marine environment, fisheries and the environment generally. A changing ocean poses several scale-related challenges for these governance institutions and processes, arising from:

- The global and transboundary scales of the major changes to ocean properties (temperature, circulation, oxygen loss, acidification, etc.), with variability in their local expression;
- The regional scales of changes in ecosystem services following from the changes in ocean properties (including services provided to humans living far from the coasts);
- The global scales of land-based drivers of those changes (both greenhouse gas emissions and changes in ecosystems services), which often motivate policy responses (primarily at the national level) and behavioural responses (primarily at the community level);
- The scale-dependent need for coordinated responses by the different governance structures, to ensure their overall effectiveness (see also Chapter 1)

For all of these challenges, the scales of the climate-related issues may be poorly matched to the scales of most governance institutions and processes, making effective responses or proactive initiatives difficult. Sections 5.2 to 5.4 provide evidence, through case histories and thematic overviews, that illustrates these four types of challenges. In some cases, more than one type of challenge is illustrated in a single example, such as when a change in an amount or availability of an ecosystem service is discussed in the context of factors influencing the vulnerability of socio-ecological systems to climate change (Sections 5.2. 5.3 and 5.4).

Existing ocean governance structures for the ocean already face multi-dimensional challenges because of climate change, and this trend of increasing complexity will continue (Galaz et al., 2012). Current international governance regimes and structures for fisheries and the ocean environment do not yet adequately address the issues of ocean warming, acidification and deoxygenation (Oral, 2018); Box 5.6). At the time of the initial development and adoption of these legal and governance regimes, minimal attention was given to climate change and the effects of carbon dioxide emissions on the ocean, with associated impacts on the interacting physical, chemical, biological properties of the ecosystems, and the resulting risks and vulnerabilities of dependent communities and economic sectors. In particular, the governance of ocean areas beyond national jurisdiction (ABNJ) is a major challenge (Levin and Le Bris, 2015); the collaborative structures and mechanisms for environmental assessment in ABNJ need further development (Warner, 2018) (*high confidence*). Negotiations are currently ongoing regarding a new international agreement for marine biodiversity of ABNJ (UNEP, 2016).

Table 5.9: Ocean Governance and Climate Change: Major Issues

Area of Governance	Major Legal Instruments	Major Issues and Actions
Marine Environment Generally	UNCLOS, CBD, CITES, WHC,	UNCLOS imposes obligations on state parties to take action to combat the main sources of ocean pollution. Tools and techniques in UNCLOS may need adjustment in response to the emerging challenges created by ocean climate

	MARPOL and other IMO legal instruments, regional seas conventions and other legal instruments	change (Redgwell, 2012). However, success of the umbrella regulatory framework of UNCLOS depends heavily on the further development, modification and implementation of detailed regulations by relevant international, regional and national institutions (Karim, 2015). The London Protocol to the London Convention was amended in 2006 to address the issue of carbon dioxide storage processes for sequestration. Two subsequent amendments concern sharing transboundary sub-seabed geological formations for sequestration projects, and ocean fertilization and other marine geoengineering. One of these new amendments prohibits ocean fertilization except for research purposes (Dixon et al., 2014). The issue of ocean acidification has been considered within the framework of the OSPAR Convention, the CCAMLR Convention (Herr et al., 2014), and the CBD (Hennige et al., 2014) ; this issue is discussed further in Box 5.6. The CBD has also considered regulatory issues relating to ocean fertilization and other (marine) geoengineering (Williamson and Bodle, 2016). In 2018, the CBD adopted *Voluntary Guidelines for the Design and Effective Implementation of Ecosystem-Based Approaches to Climate Change Adaptation and Disaster Risk Reduction*. However, even if Parties to the Convention choose to adopt the voluntary guidelines, there is no mechanism to implement them beyond their exclusive economic zones in the water column and their extended continental shelves (if recognized) in the seabed Most of the 29 world heritage listed coral reefs are facing severe heat stress (Heron, 2017) and the WHC may play a role for coral reef protection.
Climate Change	UNFCCC, Paris Agreement, MARPOL Convention and other legal instruments	Existing international legal instruments do not adequately address climate change challenges for the open ocean and coastal seas (Galland et al., 2012; Redgwell, 2012; Herr et al., 2014; Magnan et al., 2016; Gallo et al., 2017; Heron, 2017). Nevertheless, ocean and coastal areas will benefit from the overall UNFCCC goal for preventing dangerous interference with the climate system. A study of the 161 national pledges for climate change mitigation and adaptation (NDCs) identified 'gaps between scientific [understanding] and government attention, including on ocean deoxygenation, which is barely mentioned' (Gallo et al., 2017). In 2011, the MARPOL convention was amended to include technical and operational measures for the reduction of greenhouse gas emissions from ships. However, the effectiveness of these provisions depends on the national implementation by flag, port and coastal states, with no international enforcement authority (Karim, 2015).
Fisheries	UNCLOS, UN Fish Stocks Agreement, FAO Compliance Agreement, FAO PSMA, Regional Fisheries Agreements and other legal instruments	The impact of climate change on marine fisheries is expected to be very significant (Sections 5.3, 5.4) (FAO, 2018a; FAO, 2019), with adverse impacts on food security, livelihood and national development in many coastal countries; least developed countries seem particularly vulnerable (Blasiak et al., 2017). Regional fisheries management systems need to address these emerging challenges (Brooks et al., 2013). The ecological and socio-ecological criteria and standards for performance can be set at regional levels where Regional Fisheries Management Organizations have been established, but their effectiveness is variable depending on the characteristics of regulatory instruments and other factors (Ojea et al., 2017). The current international regulatory framework for fisheries management has a responsiveness gap, since it does not fully incorporate issues related to the fluctuating and changing distribution of fisheries (Pentz and Klenk, 2017; Pinsky et al., 2018). However, some regional fisheries management organisations (RFMOs) have initiated processes to improve the equity of sharing fishery resources affected by climate change (Aqorau et al., 2018). A climate-informed ecosystem-based fisheries governance approach has been suggested for enhancing climate change

		resilience of marine fisheries in the developing world (Heenan et al., 2015a), but robust and effective management, policy, legislation and planning based on flexibility and scientific understanding will be required for coastal fisheries (Gourlie et al., 2017). The existing failing condition of many stocks, coupled with maladaptive responses to climate change, may create serious challenges for the sustainability of global fisheries; improved fisheries governance can offset some of these challenges (Gaines et al., 2018). The fisheries agreements and the provisions in UNCLOS have helped RFMOs to increase the sustainability of fisheries on stocks in or migrating through international waters, and equity of access to them. Because the distribution of many stocks changes with changes in physical oceanic conditions (particularly temperature and current regimes), many of the measures and access arrangements negotiated and adopted by the RFMOs have reduced effectiveness in a changing climate. New arrangements have been difficult to negotiate, in part because of concerns that the distributions and productivities will continue to change as climate change continues to drive changes on ocean conditions (Blasiak et al., 2017; Ojea et al., 2017; Pentz and Klenk, 2017; Aqorau et al., 2018; Pinsky et al., 2018).

Acronyms and organizations: CBD, Convention on Biological Diversity; CCAMLR, Convention on the Conservation of Antarctic Marine Living Resources; CITES, Convention on International Trade in Endangered Species of Wild Fauna and Flora; IMO: International Maritime Organization; London Convention: Convention on the Prevention of Marine Pollution by Dumping of Wastes and Other Matter; London Protocol: 1996 Protocol to the Convention on the Prevention of Marine Pollution by Dumping of Wastes and Other Matter; MARPOL Convention: International Convention for the Prevention of Pollution from Ships; NDCs: Nationally Determined Contributions; OSPAR Convention: Convention for the Protection of the Marine Environment of the North-East Atlantic; UN Fish Stocks Agreement: The Agreement for the Implementation of the Provisions of the United Nations Convention on the Law of the Sea of 10 December 1982 relating to the Conservation and Management of Straddling Fish Stocks and Highly Migratory Fish Stocks; UNCLOS: United Nations Convention on the Law of the Sea; UNFCCC: United Nations Framework Convention on Climate Change; WHC, World Heritage Convention: Convention Concerning the Protection of the World Cultural and Natural Heritage; FAO Compliance Agreement: The Agreement to Promote Compliance with International Conservation and Management Measures by Fishing Vessels on the High Seas; FAO PSMA: The Agreement on Port State Measures

The following changes in governance may improve the ability of governance institutions and processes to address the challenges identified above:

- Cooperation on regional and global scales through various types of agreements of varying degrees of formality for States and other participants in governance
- Increasing the voice and role in decision-making for non-governmental participants such as Indigenous Peoples, social and labour organizations
- Increasing the horizontal integration of decision-making across industry and societal sectors, under processes such as "integrated management" and "marine spatial planning"
- Increasing resource mobilisation at the community scale to enable communities to experiment and innovate to address the challenges, and then to share their experiences with other communities and build cooperative approaches to promote strategies with successful outcomes

These governance innovation strategies have the potential to increase the ability of the governance institutions and processes to successfully respond to all four types of scale-related challenges listed earlier. However, any of them also have the potential to fail to address their intended concerns effectively if implemented inappropriately, or to create new challenges as the initial priorities are addressed. In some countries, lack of capacity of the existing governance institutions, lack of access to basic facilities, insufficient income diversification and illiteracy are major hindrance for ocean governance in a changing climate (Bennett et al., 2014; Salik et al., 2015; Weng et al., 2015; Karim and Uddin, 2019; Sarkodie and Strezov, 2019) (*high confidence*)

1   Additional considerations identified by recent studies of ocean-related mitigation and adaptation include the
2   need for: early warning and precautionary management; multi-level and multi-sectorial governance
3   responses; holistic, integrated and flexible management systems; integration of scientific and local
4   knowledge as well as natural, social and economic investigation; identification and incorporation of a set of
5   social indicators and checklists; adaptive governance; and incorporation of climate change effects in marine
6   spatial planning (Hiwasaki et al., 2014; Kettle et al., 2014; Hernández-Delgado, 2015; Himes-Cornell and
7   Kasperski, 2015a; Pittman et al., 2015; Colburn et al., 2016; Creighton et al., 2016; Hobday et al., 2016a;
8   Audefroy and Sánchez, 2017; Gissi et al., 2019; Tuda et al., 2019). Diverse adaptations of governance are
9   being tried, and some are producing promising results (Sections 5.2, 5.3 and 5.4). However, rigorous further
10 evaluation is needed regarding the effectiveness of these adaptations in achieving their goals in addressing
11 specific governance challenges. Robust conclusions on the effectiveness of specific types of governance
12 adaptations in various socio-ecological contexts would require a targeted assessment of ocean (and
13 terrestrial) governance in a changing climate, possible as a key part of IPCC AR6.
14
15
16 [START BOX 5.6 HERE]
17
18 **Box 5.6: Policy Responses to Ocean Acidification: Is there an international Governance Gap?**
19
20 Ocean acidification is not specifically mentioned in the Paris Agreement on climate change (UNFCCC,
21 2015) and has only been given limited attention to date in other UNFCCC discussions. Nevertheless, ocean
22 acidification is widely considered to be part of the climate system: it is one of seven state-of-the-climate
23 indicators used by the World Meteorological Organization (WMO, 2019); it featured strongly in IPCC AR5,
24 being covered by both WGI and WGII; its impacts are assessed in many sections of this Chapter; and
25 concerns regarding ocean acidification have been raised through many international governance structures,
26 including the United Nations Convention on the Law of the Sea (UNCLOS), the Convention on Biological
27 Diversity (CBD), the United Nations Environment Programme (UNEP), and the Intergovernmental
28 Oceanographic Commission of the United Nations Educational, Scientific and Cultural Organization (IOC-
29 UNESCO).
30
31 Although many bodies have interests in ocean acidification, no unifying treaty or single instrument has been
32 developed (Herr et al., 2014; Harrould-Kolieb and Hoegh-Guldberg, 2019) and there has been only limited
33 governance action that is specific to the problem (Fennel and VanderZwaag, 2016; Jagers et al., 2018).
34 Exceptions to this generalization are the development of coordinated monitoring through the Global Ocean
35 Acidification Observing Network (Newton et al., 2015), with associated scientific support through the
36 International Atomic Energy Agency (IAEA) (Osborn et al., 2017; Watson-Wright and Valdés, 2018); and
37 UN Sustainable Development Goal 14.3, with its non-binding, and relatively general, commitment to
38 'minimise and address the impacts of ocean acidification, including through enhanced scientific cooperation
39 at all levels'.
40
41 One possible response to the fragmented responsibilities for ocean acidification governance would be the
42 development of a new UN mechanism specifically to address ocean acidification (Kim, 2012). This option
43 would take time and political will, and has not been widely supported (Harrould-Kolieb and Herr, 2012).
44 One pragmatic approach could be enhancing the involvement of UNFCCC with acidification governance
45 (Herr et al., 2014) together with increased use of multilateral environment agreements (Harrould-Kolieb and
46 Herr, 2012) (*medium confidence*).
47
48 UNFCCC action to stabilize the climate by reducing CO_2 emissions also necessarily addresses the problem
49 of ocean acidification, which is primarily caused by anthropogenic CO_2 dissolving in seawater and lowering
50 pH. Nevertheless, there are also distinct ocean acidification mitigation and adaptation issues, including:
51
52 • Climate mitigation measures that might be focused on greenhouse gases other than CO_2
53 • pH-associated thresholds or tipping-points (Hughes et al., 2013; Good et al., 2018) that have implications
54   for scenario-modelling of emission reductions (Steinacher et al., 2013)

1   • The large-scale use of bioenergy with carbon capture and storage (BECCS) as a mitigation option, if this
2   involved sub-seafloor CO_2 storage, with risk of leakage and hence ocean acidification impacts (Blackford
3   et al., 2014)
4   • The use of other CO_2 removal techniques (negative emissions) such as ocean fertilization (Section
5   5.5.1.3), or solar radiation management, without CO_2 emission reductions; both approaches would
6   worsen ocean acidification (Williamson and Turley, 2012; Keller et al., 2014a).

8   Adaptation to climate change could also include a more integrated approach to reduce ocean acidification
9   impacts (Section 5.5.2). Proposed adaptation actions for ocean acidification (Kelly et al., 2011; Billé et al.,
10   2013; Strong et al., 2014; Albright et al., 2016a) include reduction of pollution and other stressors (thereby
11   strengthening resilience); water treatment (e.g., for high-value aquaculture); and the use of seaweed
12   cultivation and seagrass restoration to slow longterm pH changes (although short-term variability may be
13   increased) (Sabine, 2018). These measures are generally applicable to relatively limited spatial scales; whilst
14   they may succeed in 'buying time', their future effectiveness will decrease unless underlying global drivers
15   are also addressed (*high confidence*).

17   [END BOX 5.6 HERE]

20   **5.6   Synthesis**

22   This chapter has documented an extraordinary array of observed changes in the open ocean, deep-sea and
23   coasts. It draws on evidence from thousands of references from the literature, millions of observations and
24   hundreds of simulations of the past and future scenarios. The ocean climate and its state, ecosystems and
25   human systems have changed (Section 4.2.2.6, 5.2, 5.3, 5.4, Figure 5.24) and are projected to change further.
26   The ocean is a highly connected environment allowing water and living organisms to move freely. Change
27   is observed across physical conditions that pose hazards to ecosystems in all regions from the surface to the
28   deepest parts (Figure 5.24). All types of human and managed systems that have been covered in this chapter
29   have evidence of mostly negative impacts but also some positive, some very significantly, some less so
30   (Figure 5.24). Overall the multiple lines of evidence from the literature and the assessment in this chapter's
31   Executive Summary point to profound and pervasive changes on regional and global scales (Figure 5.24).

33   The level of knowledge and confidence of the changes in the marine environment that are particularly
34   relevant to ecosystems and human systems ranges from *virtually certain* to *low confidence* (see Figure 5.24).
35   Many of the observed changes in some variables can be directly attributed to human influence from rising
36   greenhouse gases and other anthropogenic forcings (Section 5.2.2 and 5.2.3). For other variables and in some
37   systems the evidence is less direct, but the cascading of risks from changing ocean, marine ecosystems and
38   dependent communities remains robust when considered as a whole. The observed and projected changes in
39   the ocean systems that are covered in this chapter are consistent with our understanding of ocean chemistry
40   and circulation, and our knowledge of the ecosystems responses. In many cases, the assessments of risk level
41   of ecosystems for the recent past and long term future are based on multiple lines of evidence, combining
42   ecological and physiological knowledge (from experiments, direct observations and model projections) with
43   the major climate drivers (e.g., Sections 5.2.5 and 5.3.4). Globally, all the marine ecosystems assessed here
44   have elevated risk for biodiversity, ecosystem function, structure and services with increasing greenhouse
45   gas emissions (Figure 5.16) (*high confidence*). These risks result from ocean warming, stratification,
46   acidification, deoxygenation, sea level rise and associated changes as well as interactions with non-climatic
47   human drivers. Most importantly, all the coastal ecosystems that were assessed, where linkages between
48   natural systems and human communities are the strongest, had increased risk, and none saw a risk reduction
49   from a warming climate (*high confidence*).
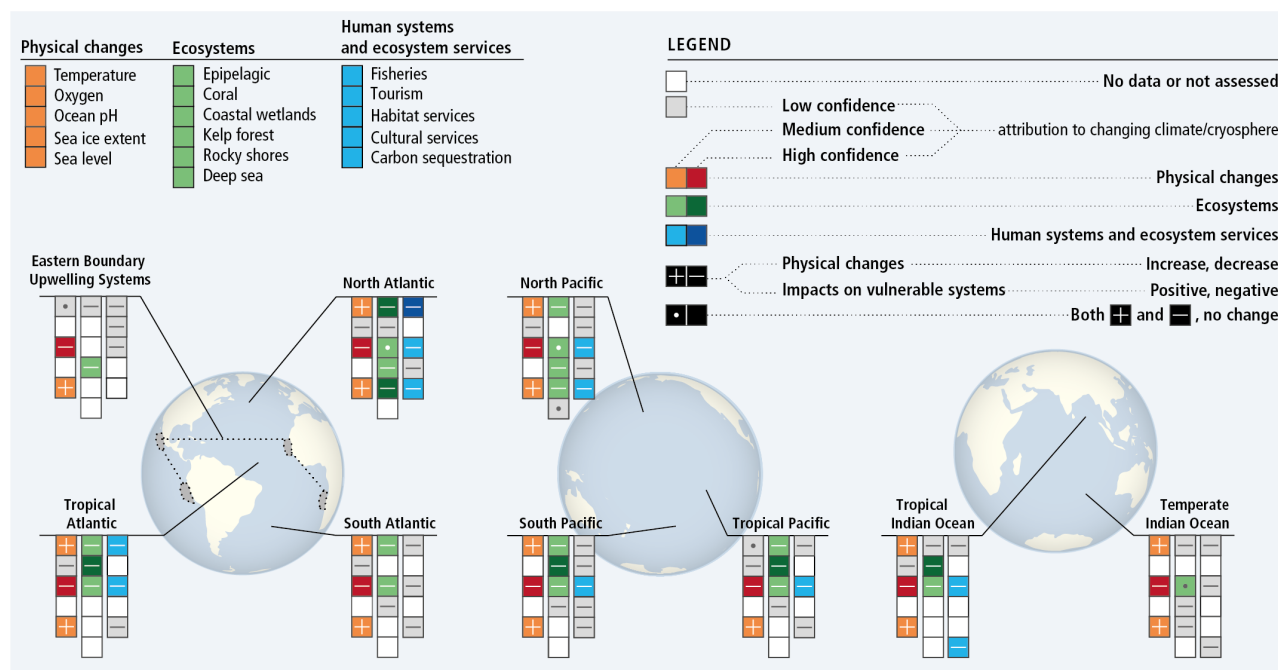
Figure 5.24: Synthesis of regional consequences and impacts in ocean regions assessed in Chapter 5. For each region, physical changes (red boxes), impacts on key ecosystems (green boxes), and impacts on human sectors and ecosystem services (blue boxes) are shown. Physical changes shown are attributable to rising greenhouse gas concentrations at either global or regional scales with the indicated confidence; attribution is less certain at regional scales due to higher internal variability. Physical changes refer to averages for each of the named regions. For physical changes, + or – refers to an increase or decrease in amount or frequency in the measured parameter. For impacts on ecosystems, human sectors and ecosystems services, + or – depicts a positive or negative impact, respectively. A dot represents both positive and negative impacts are observed. The underlying information is given for ocean regions in SM 5.11. {4.2.2.6, 5.2.2, 5.2.3, 5.3.3, 5.4,5.6}

The observed and projected changes in the open ocean and coastal seas have consequences on human communities and affect all aspects of wellbeing and have social, economic and environmental costs (Section 5.4, *high confidence*). The range and diversity of impacts is striking, with varying consequences for the wider community when analysed across the key marine ecosystems services. These consequences clearly affect the capacity for human society to achieve the Sustainable Development Goals (e.g., Figure 5.22). The evidence of climate change in the ocean is a pervasive thread through all types of coupled human-natural systems and projections amplify these observed impacts with the least impact from lower emission scenarios.

Risk-reduction responses and their governance through adaptation at the local scale are the most common responses to climate change from ocean systems (Section 5.5.2). It is clear that there are many choices for reducing risk of climate change. Many of the actions have benefits and relatively few dis-benefits, while others have large dis-benefits and marginal effectiveness (Section 5.5.1, Figure 5.23, Table 5.7 and Table 5.8). Many of the risk reduction approaches are limited in their capacity to reduce the risks of climate change, or are at best temporary solutions, which is a significant challenge to adapting to climate change (*high confidence*). In particular, the effectiveness of the assessed risk reduction measures are minimal under high greenhouse gas emission scenarios, highlighting the critical importance of mitigation. The assessment points to the increased effectiveness and importance of a portfolio of different types of mitigation and adaptation options. Governance is also a critical element in the portfolio of options and occurs at local, national and international scales. Such responses can be more effective with the support of scientific information, Local knowledge and Indigenous knowledge, and the consideration of local context and the inclusion of stakeholders.

5.7   Key Uncertainties and Gaps

1   This chapter was designed around three guiding questions (Section 5.1). These guiding questions mean that
2   the report covers both regional and global scales of the ocean and many aspects of human systems, including
3   governance and institutions, and adaptation pathways for dependent communities. This assessment is new
4   linking together a broad and complex set of ocean disciplines and therefore also provides a unique
5   perspective on key uncertainties and gaps in these systems. These gaps limit the extent of the assessments
6   that were possible in this report. Notable outstanding uncertainties and gaps from this assessment include the
7   following:
8
9   **Physical and biogeochemical processes:** While the Earth system is better monitored and the relevant data
10 are more accessible than the other areas of assessment there is considerable room to improve these
11 capabilities. For example, gaps remain in predictive modelling of climate change in coastal areas, deep ocean
12 temperature and salinity measurements for sea-level and closure of the energy budget, and oxygen and
13 carbon measurements dense enough to measure de-oxygenation of the world ocean and track the
14 mechanisms driving the ocean carbon cycle. Our capacity to understand and model net primary productivity
15 and the rates of carbon burial in coastal sediments are also significant weaknesses. Projections of future
16 changes in the Earth system depend on the use of Earth System Models (ESM), in which there are
17 uncertainties arising from physical or ecological processes that are either omitted or incompletely
18 understood. Most ESMs still rely on relatively simple representations of ocean biogeochemical cycling and
19 the linkages to ocean ecosystem structure and function (Section 5.2.3). Other examples of under-assessed
20 biogeochemical process in the ocean that may have implications for the Earth system under climate change
21 include the fate of methane in the deep ocean (Section 5.2.4). Open ocean primary productivity and its
22 projections requires critical corroborating measurements and improved understanding of its drivers to project
23 changes in ocean productivity with higher confidence (Sections 5.2.2 and 5.2.3).
24 **Biological processes and monitoring:** There are a number of marine environments (e.g., on the deep sea
25 floor) and ecosystem components (e.g., viruses and protists) where insufficient scientific understanding
26 limits the assessments of risks to *low confidence or no assessment.* Examples of gaps include the narrow
27 range of climate and non-climatic hazards and their interactions in simulation models, the linkages between
28 single organisms to communities of organisms, knowledge of climate feedbacks in biological systems
29 (Section 5.3.4, Section 5.2.4), and the capacity and limits of biological adaptation for many ecosystems
30 (Section 5.2.2, 5.2.3, 5.2.4, 5.3). Increasing observational capacity can help provide the data to improve
31 understanding and modelling of these important biophysical responses to climate change.
32
33 **Variance in human systems and effectiveness of responses:** The wide range of contributing factors
34 (physical, social and economic) that interact with localised climate projections make projecting site-specific
35 costs of impacts and benefits of adaptation difficult. There were few examples in the literature evaluating
36 implemented adaptation actions, and there was *low confidence* in their reliability and provenance, thus
37 largely precluding any assessments of their cost effectiveness. This lack of evidence on costs and benefits
38 particularly affected assessments in Section 5.4 and 5.5. Adaptation responses to climate change have been
39 undertaken by communities, industry and governments. However, their effectiveness for mitigating the risks
40 of climate change (e.g., different types of adaptation response on the coasts, Section 5.5.2) is largely
41 unassessed here, and consequently precludes a global understanding of the capacity in the world to address
42 the risks of climate change in coastal seas, open ocean and the deep sea. A partial solution would be
43 establishing an appropriate ocean and coasts database, including costs-benefits, for these types of studies.
44
45
46
47

1 [START FAQ5.1 HERE]
2
3 **FAQ5.1: How is life in the sea affected by climate change?**
4

5 Climate change poses a serious threat to life in our seas, including coral reefs and fisheries, with impacts on
6 marine ecosystems, economies and societies, especially those most dependent upon natural resources. The
7 risk posed by climate change can be reduced by limiting global warming to no more than 1.5°C.

8
9 Life in most of the global ocean, from pole to pole and from sea surface to the abyssal depths, is already
10 experiencing higher temperatures due to human-driven climate change. In many places, that increase may be
11 barely measurable. In others, particularly in near-surface waters, warming has already had dramatic impacts
12 on marine animals, plants and microbes. Due to closely-linked changes in seawater chemistry, less oxygen
13 remains available (in a process called ocean deoxygenation). Seawater contains more dissolved carbon
14 dioxide, causing ocean acidification. Non-climatic effects of human activities are also ubiquitous, including
15 over-fishing and pollution. Whilst these stressors and their combined effects are *likely* to be harmful to
16 almost all marine organisms, food-webs and ecosystems, some are at greater risk (FAQ5.1, Figure 1). The
17 consequences for human society can be serious unless sufficient action is taken to constrain future climate
18 change.

19
20 Warm-water coral reefs host a wide variety of marine life and are very important for tropical fisheries and
21 other marine and human systems. They are particularly vulnerable, since they can suffer high mortalities
22 when water temperatures persist above a threshold of between 1 – 2°C above the normal range. Such
23 conditions occurred in many tropical seas between 2015 and 2017 and resulted in extensive coral bleaching,
24 when the coral animal hosts ejected the algal partners upon which they depend. After mass coral mortalities
25 due to bleaching, reef recovery typically takes at least 10–15 years. Other impacts of climate change include
26 sea level rise, acidification and reef erosion. Whilst some coral species are more resilient than others, and
27 impacts vary between regions, further reef degradation due to future climate change now seems inevitable,
28 with serious consequences for other marine and coastal ecosystems, like loss of coastal protection for many
29 islands and low-lying areas and loss of the high biodiversity these reefs host. Coral habitats can also occur in
30 deeper waters and cooler seas, and more research is needed to understand impacts in these reefs. Although
31 these cold water corals are not at risk from bleaching, due to their cooler environment, they may weaken or
32 dissolve under ocean acidification, and other ocean changes.

33
34 Mobile species, such as fish, may respond to climate change by moving to more favorable regions, with
35 populations shifting poleward or to deeper water, to find their preferred range of water temperatures or
36 oxygen levels. As a result, projections of total future fishery yields under different climate change scenarios
37 only show a moderate decrease of around 4% (~3.4 million tons) per degree Celsius warming. However,
38 there are dramatic regional variations. With high levels of climate change, fisheries in tropical regions could
39 lose up to half of their current catch levels by the end of this century. Polar catch levels may increase
40 slightly, although the extent of such gains is uncertain, because fish populations that are currently depleted
41 by over-fishing and subject to other stressors may not be capable of migrating to polar regions, as assumed
42 in models.

43
44 In polar seas, species adapted to life on or under sea-ice are directly threatened by habitat loss due to climate
45 change. The Arctic and Southern Oceans are home to a rich diversity of life, from tiny plankton to fish, krill
46 and seafloor invertebrates to whales, seals, polar bears or penguins. Their complex interactions may be
47 altered if new warmer-water species extend their ranges as sea temperatures rise. The effects of acidification
48 on shelled organisms, as well as increased human activities (e.g., shipping) in ice-free waters, can amplify
49 these disruptions.

50
51 Whilst some climate change impacts (like possible increased catch levels in polar regions) may benefit
52 humans, most will be disruptive for ecosystems, economies and societies, especially those that are highly
53 dependent upon natural resources. However, the impacts of climate change can be much reduced if the world
54 as a whole, through inter-governmental interventions, manages to limit global warming to no more than
55 1.5°C.
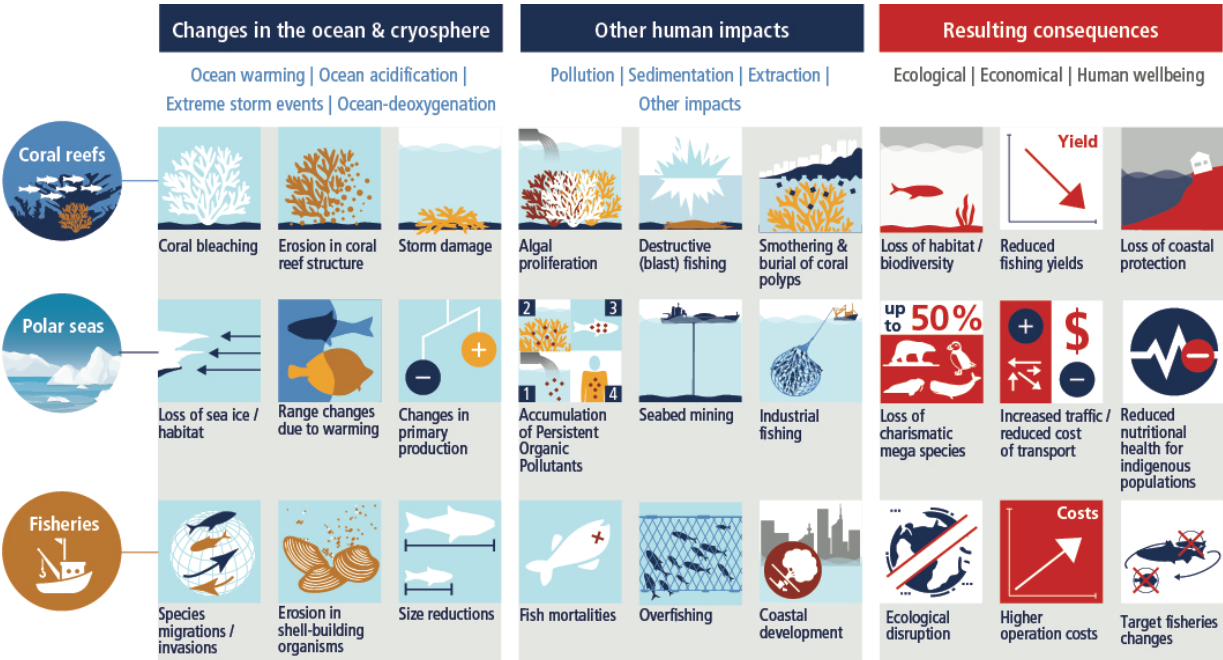
FAQ5.1, Figure 1: Summary schematic of the impacts and resulting consequences of climate change (warming, acidification, storminess and deoxygenation) and other human impacts, on coral reefs, polar seas and fisheries, discussed in this FAQ.

[END FAQ5.1 HERE]

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